World Geomorphological Landscapes

Colin K. Ballantyne
John E. Gordon *Editors*

Landscapes and Landforms of Scotland

Springer

World Geomorphological Landscapes

Series Editor

Piotr Migoń, Institute of Geography and Regional Development, University of Wrocław, Wrocław, Poland

Geomorphology – 'the Science of Scenery' – is a part of Earth Sciences that focuses on the scientific study of landforms, their assemblages, and surface and subsurface processes that moulded them in the past and that change them today. Shapes of landforms and regularities of their spatial distribution, their origin, evolution, and ages are the subject of geomorphology. Geomorphology is also a science of considerable practical importance since many geomorphic processes occur so suddenly and unexpectedly, and with such a force, that they pose significant hazards to human populations. Landforms and landscapes vary enormously across the Earth, from high mountains to endless plains. At a smaller scale, Nature often surprises us creating shapes which look improbable. Many geomorphological landscapes are so immensely beautiful that they received the highest possible recognition – they hold the status of World Heritage properties. Apart from often being immensely scenic, landscapes tell stories which not uncommonly can be traced back in time for millions of years and include unique events. This international book series will be a scientific library of monographs that present and explain physical landscapes across the globe, focusing on both representative and uniquely spectacular examples. Each book contains details on geomorphology of a particular country (i.e. The Geomorphological Landscapes of France, The Geomorphological Landscapes of Italy, The Geomorphological Landscapes of India) or a geographically coherent region. The content is divided into two parts. Part one contains the necessary background about geology and tectonic framework, past and present climate, geographical regions, and long-term geomorphological history. The core of each book is however succinct presentation of key geomorphological localities (landscapes) and it is envisaged that the number of such studies will generally vary from 20 to 30. There is additional scope for discussing issues of geomorphological heritage and suggesting itineraries to visit the most important sites. The series provides a unique reference source not only for geomorphologists, but all Earth scientists, geographers, and conservationists. It complements the existing reference books in geomorphology which focus on specific themes rather than regions or localities and fills a growing gap between poorly accessible regional studies, often in national languages, and papers in international journals which put major emphasis on understanding processes rather than particular landscapes. The World Geomorphological Landscapes series is a peer-reviewed series which contains single and multi-authored books as well as edited volumes.

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Series Editor Preface

Landforms and landscapes vary enormously across the Earth, from high mountains to endless plains. At a smaller scale, Nature often surprises us by creating shapes which look improbable. Many physical landscapes are so immensely beautiful that they have received the highest possible recognition—they hold the status of World Heritage properties. Apart from often being immensely scenic, landscapes tell stories which not uncommonly can be traced back in time for tens of millions of years and include unique events. In addition, many landscapes owe their appearance and harmony not solely to natural forces. For centuries, or even millennia, they have been shaped by humans who modified hillslopes, river courses, and coastlines, and erected structures which often blend with the natural landforms to form inseparable entities.

These landscapes are studied by Geomorphology—'the Science of Scenery'—a part of Earth Sciences that focuses on landforms, their assemblages, the surface and subsurface processes that moulded them in the past and that change them today. The shapes of landforms and the regularities of their spatial distribution, their origin, evolution, and ages are the subject of research. Geomorphology is also a science of considerable practical importance since many geomorphic processes occur so suddenly and unexpectedly, and with such a force, that they pose significant hazards to human populations and not uncommonly result in considerable damage or even casualties.

To show the importance of geomorphology in understanding the landscape, and to present the beauty and diversity of the geomorphological sceneries across the world, we have launched a new book series *World Geomorphological Landscapes*. It aims to be a scientific library of monographs that present and explain physical landscapes, focusing on both representative and uniquely spectacular examples. Each book will contain details on geomorphology of a particular country or a geographically coherent region. This volume is dedicated to Scotland, part of the world where modern geoscience was born in the late eighteenth century thanks to the insightful thinking of James Hutton; some of key sites first interpreted by Hutton are presented in this book, although in a more geomorphological context. As with most books in the series, this one also starts with chapters providing the geological and temporal context, setting the stage for systematic presentation of the geomorphology of specific regions of Scotland. The regional part is as comprehensive as possible, so that the reader can virtually visit the northern archipelagos of Shetland and Orkney, jump across the islands of the Hebrides, travel through the glacially moulded Scottish Highlands to the largely inherited landscape of Buchan, enjoy the glacial topography of the Midland Valley, be introduced to the less-known geomorphology of the Southern Uplands, and get acquainted with the diverse Scottish coasts. Even underwater landscapes around the Scottish landmass are covered. The effects of successive glaciations are the main theme in many regions, but we also learn about the pre-Quaternary roots of Scotland's topography, going back to really distant times of the Palaeozoic, as well as the evidence of the multitude of landscape changes that postdate the demise of the last ice sheet and glaciers.

The World Geomorphological Landscapes series is produced under the scientific patronage of the International Association of Geomorphologists—a society that brings together geomorphologists from all around the world. The IAG was established in 1989 and is an independent scientific association affiliated with the International Geographical Union and the International Union of Geological Sciences. Among its main aims are to promote geomorphology and to foster dissemination of geomorphological knowledge. I believe that this lavishly illustrated series, which sticks to the scientific rigour, is the most appropriate means to fulfil these aims and to serve the geoscientific community. To this end, my great thanks go to Colin Ballantyne and John Gordon, who agreed to coordinate the book and contributed many chapters themselves, based on their enormous experience in geomorphological research across the lowlands, highlands, and islands of Scotland. Their editorial work was absolutely first class, and they ensured that all geographical gaps were expertly filled. Reading many chapters brought back my own distant memories of seeing a fraction of Scotland's superb geomorphology, under expert guidance of Philip Ringrose and Michael Thomas, whose hospitality is also acknowledged here. I am also grateful to all the other contributors to the book, who agreed to add the task of writing chapters to their busy agendas and delivered high-quality final products. Collectively, they have shown us that Scottish geomorphological landscapes are some of the finest in the world and are, besides single malts, the key attractions of this part of the globe.

Wrocław, Poland

Piotr Migoń
Series Editor

Foreword

Scotland is renowned for its landscapes of mountains, lowlands, islands, and its intricate coast. Within an hour or so, a visitor can travel from the rolling uplands and coastlands of the east to the steep and rocky mountains of the west, or from the gentle central lowlands to the open landscapes of the northwest with their isolated mountains rising above a rugged plain, or even from the sweeping beaches of the east to the fretted and dramatic coast of the fjords in the west. By bridge or ferry, the visitor can move to the world of islands, each with its own distinctive landscape. And if the visitor wonders at this remarkable variety within such a small area, then he or she enters the world of geomorphology, namely the study of landforms, their origins, and the processes that have shaped them in the past and are shaping them today.

Perhaps it is no surprise that Scotland's landscapes have stimulated scientists over the centuries and that several world-leading figures cut their teeth in Scotland. Indeed, in the late eighteenth century and nineteenth century, Scotland was a veritable hotbed of activity that transformed geomorphology. James Hutton, whose treatise *Theory of the Earth* was published in Edinburgh in 1788, is seen internationally as the founder of both geomorphology and geology. His ideas, developed by John Playfair, underpinned Charles Lyell's hugely influential three-volume *Principles of Geology* published in the 1830s. In 1864, James Croll was the first to develop the astronomical theory of the ice ages; no mean glaciologist and fluvial geomorphologist himself, Croll was stimulated by the wish to understand Scotland's glacial landforms and interglacial sediments. The Geikie brothers were notable for their demonstration of how field studies could link with theory to aid geomorphological understanding. In 1865, Archibald Geikie published what was probably the first-ever regional geomorphology, *The Scenery of Scotland*. A few years later in 1874, James Geikie produced his book on *The Great Ice Age*, a path-breaking study of Scotland's glacial deposits that reflected the vicissitudes of the last ice age. And even Charles Darwin was involved, although it must be admitted that his marine explanation of the Parallel Roads of Glen Roy has since been rejected!

This book, entitled the *Landscapes and Landforms of Scotland*, consists of 29 chapters, each written by authors with expertise and empathy for particular topics or regions and reflecting the latest research. The book begins with early chapters on the geological evolution and the evolving physical environment of Scotland. Here we learn that Scotland first became a geological entity ~420 million years ago. There are rocks telling of the glacial, desert, and humid tropical climates experienced as Scotland migrated from near the South Pole to its present mid-latitude location in the Northern Hemisphere. The main outlines of the present topography and drainage have evolved in the last 65 million years, first under fluvial conditions and then in the last 2.6 million years, through successive glaciations. Currently, we learn how Scotland is being moulded by rivers, frost, mass movements, wind, and the sea.

Many chapters in the book are devoted to particular regions of iconic interest. These will be a boon to geomorphologists, teachers, students, visitors, and all those with an interest in the landscape of Scotland. Thus in the west and north, there are regional chapters covering the geomorphology of Shetland, Orkney and Caithness, the Outer Hebrides, and the islands of Skye, Mull, Rùm, Arran, Islay, Jura, Colonsay, Tiree, and Coll; for the western mainland, there are chapters on the far northwest and Wester Ross. Other regional chapters cover the Grampian Highlands (east, central and west), the uplands of the Southern Uplands, and the Solway coast and lowlands. There are chapters devoted to the Midland Valley, a glacially streamlined landscape punctuated by upstanding volcanic plugs such as that crowned by Edinburgh Castle and another that is home to the gannets on Bass Rock. Visitors to iconic mainland sites will be pleased to find a chapter covering Loch Ness in the Great Glen, one on the Parallel Roads of Glen Roy, one on Loch Lomond, and one on the Cairngorm Mountains in the heart of the national park of the same name. In the last location, the juxtaposition of tor-studded ancient uplands with dramatic cliffs bounding glacially scoured corries and troughs is a wonderful example of selective erosion by ice sheets and a dramatic contrast to the glacially dissected landscape of western Scotland.

Other chapters cover specific geomorphological topics. Thus, we can read of landslides in the Northwest Highlands, fluvial landforms in Glen Feshie, the origin of the ancient palaeosurface of Northeast Scotland, glacifluvial landforms in the Midland Valley, and beaches and dunes of Eastern Scotland, especially along the Moray coast and from Aberdeenshire to East Lothian. Of special interest is one chapter that opens up a new world; it describes and interprets the geomorphology of the extensive continental shelf surrounding Scotland. Here, the surveys associated with the petroleum and fishing industries have revolutionised our understanding of Scotland's geomorphological evolution. For example, we can now see that the last ice sheet extended to the Atlantic margin of the continental shelf, with a huge outlet glacier flowing in a trough from The Minch in Northwest Scotland and depositing a large fan of debris at the shelf edge. The final chapter in the book focuses on the geomorphological heritage in Scotland and its conservation. Overall, the book is a powerful statement of the importance of geomorphology in underpinning so much of what makes Scotland's landscapes so special.

I would like to end with a note on behalf of future readers to thank all those authors who have contributed to chapters and also to the editors, Colin K. Ballantyne and John E. Gordon, who have organised and contributed to such a significant volume. I hope readers and visitors will enjoy learning about the geomorphology of Scotland, especially since it comes with the privilege of learning from experts who have devoted so much of their energy to its study. The book is significant not only for the subject of geomorphology, but also as a wider inspiration to those who treasure the landscape and culture of Scotland.

David Sugden
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We wish to thank the editor of the World Geomorphological Landscapes series, Piotr Migoń, for his excellent guidance and advice during the compilation of this book. Piotr reviewed every chapter and made perceptive comments on each one; we particularly valued his continuous support and remarkably rapid turnaround of all submitted material. We also owe a great debt to Graeme Sandeman, formerly cartographer at the University of St Andrews, for producing many of the maps and diagrams in the book; without his skill and forbearance, the final production of several chapters would have been greatly delayed. Finally, we thank Emeritus Professor David Sugden of the University of Edinburgh for agreeing to write the Foreword to this volume. David is the doyen of geomorphology and glaciology in Scotland, and we are honoured to have his endorsement.

An edited volume such as this would not be possible without the commitment of our 21 contributing authors, who have persevered in producing their chapters despite the difficult circumstances and pressures arising from the 2020 coronavirus pandemic. These authors have themselves added hugely to our understanding of the landscapes and landforms of Scotland, and we are privileged to have received their contributions. We particularly thank them for their patience as each chapter evolved through successive reviews and edits and for responding positively to our numerous queries and suggestions.

The chapters in this book were externally peer reviewed by a small but distinguished group of individuals, and we are delighted to acknowledge the contributions of Riccardo Arosio, Doug Benn, Clare Boston, John Brown, Ben Chandler, Rodger Connell, Simon Cook, Pete Coxon, Roger Crofts, Rob Ferguson, Andrew Finlayson, Callum Firth, Con Gillen, Martin Kirkbride, Stephen Livingstone, Jon Merritt, Julian Orford, Brice Rea, Jim Rose, and David Smith to the review process, which resulted in substantial improvements to many of the draft chapters.

We also thank the staff at Springer Nature, in particular Manjula Saravanan and Banu Dhayalan, and project manager Madanagopal Deenadayalan and the team at Scientific Publishing Services, Chennai, India, for overseeing and expediting the production of the book on behalf of Springer Nature.

The editors would like to express their unanimity and solidarity in co-editing this book, which has benefitted greatly from equal sharing of editorial tasks and responsibilities. Writing, editing, and compilation of this volume took over 18 months, and it was not always straightforward. We are both fortunate in having understanding wives, Rebecca Trengove and Janet Gordon, who are inured by experience to having husbands who spend long hours closeted in their studies. Rebecca and Janet provided vital support throughout and deserve our warmest thanks of all.

Colin K. Ballantyne
John E. Gordon

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Editors and Contributors

About the Editors

Colin K. Ballantyne is Emeritus Professor of Physical Geography at the University of St Andrews and Fellow of the Royal Society of Edinburgh, the Royal Scottish Geographical Society, the Geological Society, and the British Society for Geomorphology. He has published over 200 papers on various aspects of geomorphology and Quaternary science, many of which concern Scotland. His recent books include *The Quaternary of Skye* (2016), *Periglacial Geomorphology* (2018), and *Scotland's Mountain Landscapes: A Geomorphological Perspective* (2019). He is Honorary Life Member of the Quaternary Research Association and Recipient of the Clough Medal of the Edinburgh Geological Society, the Saltire Earth Science Medal, the Gordon Warwick Medal and Wiley Award of the British Society for Geomorphology, the President's Medal, Coppock Research Medal and Newbigin Prize of the Royal Scottish Geographical Society, and the Lyell Medal of the Geological Society.

John E. Gordon is Honorary Professor in the School of Geography and Sustainable Development at the University of St Andrews, Scotland. His research interests include geoconservation, the Quaternary of Scotland, and mountain geomorphology and glaciation in North Norway and South Georgia. He has published many academic papers and popular articles in these fields and is co-author/co-editor of books including *Quaternary of Scotland* (1993), *Antarctic Environments and Resources* (1998), *Earth Science and the Natural Heritage* (2001), *Land of Mountain and Flood: the Geology and Landforms of Scotland* (2007) and *Advances in Scottish Quaternary Studies*, a special issue of *Earth and Environmental Science Transactions of the Royal Society of Edinburgh* (2019). He is a Fellow of the Royal Scottish Geographical Society, an Honorary Member of the Quaternary Research Association, and a Deputy Chair of the Geoheritage Specialist Group of the IUCN World Commission on Protected Areas.

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1 Landscapes and Landforms of Scotland: A Geomorphological Odyssey

John E. Gordon and Colin K. Ballantyne

Abstract

The landscapes and landforms of Scotland are renowned for their outstanding geodiversity, the outcome of a long and complex geological evolution, Cenozoic uplift and etchplanation, and modification by glacial and interglacial processes during the Quaternary. The Scottish landscape has provided the stimulus for over two centuries of groundbreaking research in geology and geomorphology, beginning with the seminal work of James Hutton (1726–1797) and the subsequent development of uniformitarianism as a geological paradigm by Charles Lyell (1797–1875). From 1840, Scottish researchers played a major part in the recognition of the role of Quaternary glaciation in fashioning the landscape, and other nineteenth-century Scottish pioneers developed such concepts as glacio-isostasy, multiple Pleistocene glaciations and the astronomical theory of climate change. We trace the subsequent history of key geomorphological developments in Scotland before outlining the rationale for the chapters in this book: (i) four systematic chapters that set the context and chronology for those that follow; (ii) 17 regionally focused chapters that encompass particular landscapes; and (iii) six thematic-based chapters that highlight particular aspects of Scotland's geomorphology. The final chapter addresses geoconservation, and the approach and measures adopted to protect Scotland's exceptional geoheritage.

Keywords

Geomorphology • Geology • Geodiversity • Geoheritage • Scotland • Landscape • History of geomorphology

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1.1 Introduction: The Landscapes of Scotland

The story of the origin of our scenery ... leads us back into the past farther than imagination can well follow ... Archibald Geikie (1865), p. 351.

Scotland is renowned for the great variety of its scenery. In the far northwest, monolithic sandstone inselbergs surmount a platform of Archaean gneiss; on the Isle of Skye, serrated battlements of Palaeogene gabbro rise above deeply gouged glacial troughs, and huge basalt landslides descend to the sea; on the granite of the Cairngorms, glacial troughs carve through an ancient tor-studded plateau scalloped by cirques; and even the lowlands of central Scotland are interrupted by steep lava scarps, sills and volcanic plugs; whilst at the coast, sandy beaches and bays backed by dunes and raised (emerged) beaches contrast starkly with rocky promontories and spectacular cliffs (Fig. 1.1). This remarkable landscape diversity occurs within a small country only 80,240 km^2 in extent; few (if any) parts of the Earth's surface encompass such variety within an equivalent area.

Such landscape heterogeneity is the outcome of not only a long and complex geological evolution that spans the Archaean to the Cenozoic, but also the varied processes that have operated over the past 400 Ma to mould the land surface to its present form. Most of the large-scale elements of Scotland's relief had evolved by the end of the Neogene Period (2.59 Ma), but the detailed form of the landscape was radically modified during the multiple glacial-interglacial cycles of the Quaternary (2.59 Ma to the present). Successive episodes of icefield or ice-sheet glaciation reshaped much of the pre-glacial landscape, carving alpine landforms across the western Highlands and Hebrides, selectively incising older palaeosurfaces in the Eastern Grampian Highlands and Southern Uplands, scouring the basement gneisses of the far northwest and depositing a mantle of glacigenic deposits over the sedimentary rocks of the Midland Valley. During ice-free intervals, the landscaping

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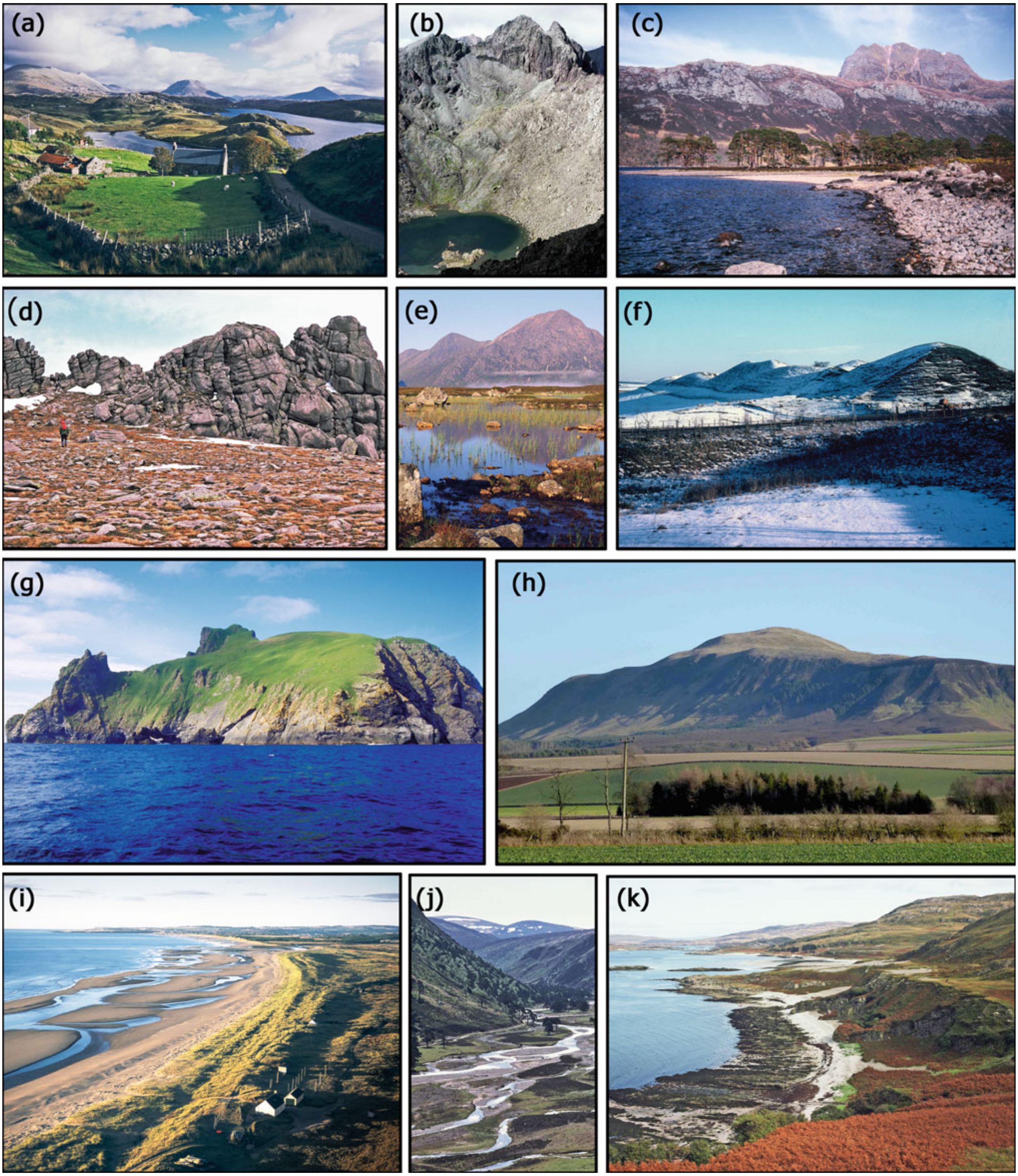

Fig. 1.1 Landscapes and landforms of Scotland, a land of outstanding geodiversity. **a** Ice-scoured landscape on Archaean Lewisian gneiss, with Cambrian quartzite inselbergs in the background, NW Sutherland. **b** Landscape of mountain glacier erosion with cirques and arêtes in Palaeogene volcanic rocks, Cuillin Hills, Isle of Skye. **c** Inselberg in Proterozoic Torridonian sandstone resting on an ancient erosion surface on Archaean Lewisian gneiss, Wester Ross. **d** Granite tor, Cairngorm Mountains. **e** Ice-scoured montane basin of Rannoch Moor, Western Grampian Highlands. **f** Carstairs Kames esker system, Midland Valley. **g** Coastal cliffs in Palaeogene volcanic rocks, St Kilda. **h** Scarp of the Carboniferous Midland Valley Sill, Lomond Hills, Fife. **i** Beach and dune coastal landscape, St Cyrus, Montrose Bay, eastern Scotland. **j** Wandering gravel-bed river, River Feshie, Strath Spey. **k** Raised and intertidal shore platforms, Isle of Jura. (Images: **a**–**d**, **f**, **i**–**k** John Gordon; **e** Lorne Gill/NatureScot; **g** David Donnan/NatureScot; **h** Colin Ballantyne)

effects of glaciation were modulated by periglacial activity, renewed rock weathering, rivers, paraglacial landslides and coastal processes. These changes occurred against a backdrop of fluctuations in sea level and changes in vegetation cover, from tundra and cold desert to boreal forest and, during interglacials, temperate deciduous forest. Moreover, the impact of successive glaciations was uneven: in the west, fast-flowing, erosive glaciers fed by abundant snowfall extensively dissected the pre-glacial terrain, whereas in the east some areas exhibit only limited and selective modification of the pre-glacial landscape. Nowhere is this better illustrated than on the Scottish coastline: the west coast is fretted by glacially deepened fjords and sounds and fringed by hundreds of islands and islets, but such features are lacking along Scotland's eastern seaboard.

1.2 Exploring Scotland's Geodiversity

1.2.1 The Scottish Pioneers

Given its remarkable geological and geomorphological diversity, it is not surprising that the Scottish landscape has stimulated a vast body of research for more than two centuries and continues to yield findings of global significance for geoscience. Perhaps the most fundamental of these appeared in 1788 in volume one of the *Transactions of the Royal Society of Edinburgh* under the title *Theory of the Earth; or an investigation of the laws observable in the composition, dissolution and restoration of land upon the globe.* The author of this ambitious exposition was James Hutton (1726–1797; Fig. 1.2a), now regarded by many as the father of modern geology. Hutton's claim to geological paternity rests not merely on his formidable skills in interpreting the rock record in various parts of Scotland, but in his realization that the rocks and landforms of Scotland represent not some past cataclysm, but the slow operation through aeons of time of the processes presently acting on and within the Earth: the paradigm, later termed *uniformitarianism,* that underpins all subsequent geological and geomorphological research. This insight required rejection of the then-accepted biblical timescale of creation some 6000 years ago and enabled Hutton to stare into the depths of deep time; his treatise concludes with the stirring words: 'the result, therefore of our present enquiry is, that we find no vestige of a beginning, no prospect of an end'.

Hutton was a prolix writer, but his ideas were promoted after his death with the publication of the splendid *Illustrations of the Huttonian Theory of the Earth* (1802) by his disciple, John Playfair (1748–1819), and formed much of the basis of the hugely influential three-volume *Principles of Geology*, published by another Scot, Charles Lyell (1797–1875; Fig. 1.2b), between 1830 and 1833. This monumental treatise formed the bedrock of geological enquiry across the world for half a century (the final edition appeared in 1875) and exhibits a keen appreciation of the role of surface processes in the fashioning of landforms and landscapes. With its demonstration of the great age of the Earth, Lyell's *Principles* played a fundamental role in influencing Charles Darwin's ideas on the evolution of species.

Recognition that Scotland had been glaciated was first anticipated by Robert Jameson (1774–1854; Fig. 1.2c), Professor of Natural History at the University of Edinburgh. In lectures Jameson presented in 1827 that were recorded in the notes of his student, James Forbes (1809–1868; Fig. 1.2d), Jameson referred to the existence of moraines as evidence of former Scottish glaciers (Cunningham 1990). Vindication of his views came in 1840 when Swiss glaciologist, Louis Agassiz, travelled by stagecoach from Glasgow to Fort William in the western Highlands in the company of William Buckland, president of the Geological Society. As their coach rumbled slowly northwards, Agassiz noted countless examples of the landforms that he had observed on the forelands of Swiss glaciers: moraines, erratics, roches moutonnées and striae. Their journey culminated in Glen Roy, where three shorelines or 'parallel roads' cut across both valley sides. Dismissing Darwin's interpretation of these features as the uplifted products of marine erosion, Agassiz concluded that they represented shorelines formed at the margin of an ancient glacier-dammed lake, an interpretation that implies the former existence of a glacier hundreds of metres thick in neighbouring Glen Spean. Agassiz dashed off a brief account that was published in *The Scotsman* newspaper, noting that he had seen near Ben Nevis '…the most distinct morains and polished rocky surfaces' and concluding that '…the existence of glaciers in Scotland at early periods can no longer be doubted' (Agassiz 1840, p. 3). Whilst Agassiz travelled onwards to Ireland, Buckland visited Lyell at the latter's estate in Angus. Lyell was quickly converted to what became known as the Glacial Theory and together with Buckland conducted an energetic search for further evidence of glaciation in Scotland.

Agassiz, Buckland and Lyell presented papers on their momentous findings at meetings of the Geological Society in November and December 1840 but met with a critical response, so much so that Lyell recanted. But the seed was sown: only six years later the pioneering Scottish glaciologist James Forbes published an account of glacial phenomena on the Isle of Skye (Forbes 1846), and by the 1870s, few geologists were in any doubt that glaciers had not only occupied Scotland, but had formerly extended over extensive tracts of Europe and North America.

Fig. 1.2 Key figures in the history of geomorphology in Scotland. **a** James Hutton. **b** Charles Lyell. **c** Robert Jameson. **d** James Forbes. **e** Archibald Geikie. **f** James Geikie. **g** John Horne (left) and Ben Peach (right). **h** Andrew Ramsay. **i** James Croll. **j** Thomas Jamieson. **k** Brian Sissons. (Images: **a** Sir Henry Raeburn, *James Hutton, 1726–1797. Geologist*, National Galleries of Scotland. Purchased with the aid of the Art Fund and the National Heritage Memorial Fund 1986. Image rights: Copyright National Galleries of Scotland, photography by AIC photography services. **b** Photograph by John Watkins. Wellcome Collection. Attribution 4.0 International (CC BY 4.0). **c** Stipple engraving by J. Jenkins, 1847, after K. Macleay. Wellcome Collection. Attribution 4.0 International (CC BY 4.0). **d** Unattributed engraving from Shairp et al. (1873). **e** Unattributed photo. Wellcome Collection. Attribution 4.0 International (CC BY 4.0). **f** Photo by Gutebrunst from Newbiggin and Flett (1917); **g** Courtesy of the British Geological Survey, UKRI sourced: https://geoscenic.bgs.ac.uk/asset-bank/action/quickSearch?CSRF=D5QFYdsKdJtoS2OixvJW&newSearch=true&quickSearch=true&includeImplicitCategoryMembers=true&keywords=P008682. **h** Photograph by D. Hains from Geikie (1895). **i**: Unattributed photo from Irons (1896); **j** Courtesy of the University of Aberdeen from 'Wilson's Aberdeen Portraits, George Washington Wilson and Co', L fAa Y24 Wil; **k** Murray Gray)

1.2.2 1865–1914: Archibald Geikie, James Geikie and the Geological Survey of Scotland

The year 1865 marked the publication of a remarkable book entitled *The Scenery of Scotland, Viewed in Connexion with its Physical Geology.* This book has a claim to be the first to focus on the geomorphology of any part of the Earth. It contains not only an analysis of the role played by geology in determining the shape of the land surface, but also over 100 pages devoted to glacial landforms and deposits, together with detailed descriptions of the contribution of rock weathering, frost, soil erosion, rivers and coastal processes in the evolution of the Scottish landscape. The final edition, published in 1901, brilliantly encompasses the sum of nineteenth-century understanding of the geomorphology of Scotland.

The author of this book was Archibald Geikie (1835–1924; Fig. 1.2e). Like his equally eminent brother, James Geikie (1839–1915; Fig. 1.2f), Archibald Geikie was not only a geologist of the first rank, but also an innovative geomorphologist, whose understanding of glacigenic deposits and their representation in the stratigraphic record provided a foundation for much later work. James Geikie authored several seminal books on geological topics, but his crowning achievement was *The Great Ice Age* (1874; third edition 1894), which drew heavily on his research in Scotland and demonstrated that the stratigraphic complexity of glacial deposits implied that the Quaternary Period incorporated multiple glacial stages separated by intervening temperate intervals. Together, the Geikie brothers not only employed the diversity of Scottish landscapes to illustrate and enrich their research, but also established a tradition of geomorphological enquiry that persists to the present. They were also pioneers in learning from modern glacier environments, conducting a Geological Survey expedition to Norway in 1865 (Worsley 2006). Another Scot, Andrew Crombie Ramsay (1814–1891; Fig. 1.2h), was a strong influence on Archibald Geikie. Ramsay, who worked for many years with the Geological Survey in Wales, was a key figure in recognizing and promoting the role of glacial erosion in shaping the landscape, and particularly in forming glacial troughs, rock basins, cirques, fjords and the lochans in knock-and-lochan topography, including many classic examples in Scotland (Ramsay 1862).

The establishment of the Geological Survey of Scotland in 1867, with Archibald Geikie as its first director, lent impetus to the mapping and understanding of Scotland's geology and landscape. Early work focused on the Midland Valley and adjacent areas, but in 1875 attention turned to the Highlands. The complexities of Highland geology were unravelled by a remarkable team of survey geologists, including James Geikie, Charles Clough (1852–1916) and two individuals who collaborated extensively, Ben Peach (1842–1926) and John Horne (1848–1928; Fig. 1.2g). Peach and Horne not only resolved the 'Highland controversy' by demonstrating thrust-faulting of Moine rocks across the stable foreland of the Hebridean terrane, but also made major contributions to Quaternary geology, notably on Orkney, Shetland and the Northwest Highlands, paralleled by pioneering research on the Outer Hebrides by Geikie and the landscape of Argyll by Clough. The manager of records at the Survey was James Croll (1852–1916; Fig. 1.2i), the father of the orbital theory of glacial-interglacial oscillations, subsequently enshrined as the Croll-Milanković theory and now widely accepted as representing the pacemaker of climate change during the Quaternary. Also briefly associated with the Scottish Survey was the Irishman, William Bourke Wright (1876–1939), who contributed innovative work on the coastal platforms and igneous rocks of the Hebrides, and whose book *The Quaternary Ice Age* (1914; second edition 1936) became the standard reference text on the Quaternary for nearly half a century and makes frequent reference to the glacial landforms and deposits of Scotland, and to the pattern of isostatic recovery of the land surface.

Another pioneering Scottish figure in the nineteenth century was Thomas Francis Jamieson (1829–1913; Fig. 1.2j), factor of Ellon Castle Estate north of Aberdeen and Fordyce Lecturer in Agriculture at the University of Aberdeen. Jamieson is credited with establishing the principle of glacio-isostasy from his work on the carse deposits in the Forth valley and was an early advocate of glacial erosion in shaping the landscape. He also made significant contributions to understanding the glaciation of NE Scotland and was the first to work out in detail the sequence of formation of the ice-dammed lakes and Parallel Roads of Glen Roy.

1.2.3 The Twentieth Century (1914–2000)

The interwar years (1918–1939) saw the publication of further maps and memoirs by the Geological Survey of Scotland, notably the prescient work of Edward Bailey (1881–1965) on the Tertiary Igneous Province in the Inner Hebrides and the igneous rocks of the Glen Coe and Ben Nevis region. During this period, only a handful of papers addressed Scotland's geomorphology, notably a 1923 study of the Moray Firth coast by the Scottish geographer, Alan G. Ogilvie (1887–1954). A focus on Scottish geomorphology was reignited after the Second World War by David Linton (1906–1971); although Linton is best remembered for his work on long-term landscape evolution (then termed 'denudation chronology'), he published a dozen seminal papers on river capture, glacial erosion and tors in Scotland, some of which have continued resonance today. Interest in the later Quaternary was reawakened by a survey of the Late-glacial history of Scotland (1955) by John Kaye Charlesworth (1889–1972), whose magisterial two-volume book *The Quaternary Era* (1957) also refers to copious Scottish examples. A particularly outstanding (and still much-cited) contribution to the post-war study of geomorphology in Scotland was made by the French geomorphologist, Alain Godard, whose eclectic doctoral thesis, *Recherches de Géomorphologie en Écosse du Nord-Ouest* (1965), encompassed the influence of lithology on landscape evolution, identification of palaeosurfaces, recognition of the influence of pre-glacial weathering, neotectonics, the effects of successive Pleistocene glaciations and postglacial landscape modification by periglacial activity and landslides. Notable contributions to the documentation of Scotland's coastal geomorphology during the second half of the century included publication of *The Coastline of Scotland* (1963) by J. Alfred Steers and the *Beaches of Scotland* surveys commissioned by the Countryside Commission for Scotland between 1969 and 1984 (Ritchie and Mather 1984).

The most influential figure in the development of geomorphology in Scotland in the second half of the twentieth century was Brian Sissons (1926–2018; Fig. 1.2k). Between 1958 and 1982, he published some 80 papers on aspects of Scottish geomorphology, together with two influential books: *The Evolution of Scotland's Scenery* (1967) and a briefer update, *The Geomorphology of the British Isles: Scotland* (1976). His principal achievements included a re-evaluation of the significance of meltwater channels and associated glacifluvial landforms, deciphering the complex sequence of raised and buried shorelines in eastern Scotland and the significance of rock platforms along the western seaboard, and detailed reconstruction of the extent and palaeoclimatic implications of the glaciers that formed in Scotland during the Loch Lomond (≈Younger Dryas) Stade of ~12.9–11.7 ka, as well as reassessment of the sequence of Lateglacial events associated with the ice-dammed lake sequence in Glen Roy (Ballantyne and Gray 1984). Equally importantly, he nurtured a school of over 30 Ph.D. students, many of whom continued to contribute research on aspects of the geomorphology and Quaternary history of Scotland, and some of whom remain active at present.

This period also saw a significant expansion in studies of modern glacier environments by geomorphologists based in Scotland. Among these, Chalmers Clapperton, Robert Price and David Sugden applied knowledge gained from their work on glacier processes in Iceland, Greenland, Alaska, Svalbard and the Antarctic to provide new insights into the formation of Scotland's glacial landforms and landscapes, both in their research and teaching (Price 1973, 1983; Sugden and John 1976).

The status of research at the end of the twentieth century on aspects of the geomorphology and Quaternary evolution of Scotland was summarized in the Quaternary, coastal, fluvial, mass-movement, and karst and caves volumes of the Geological Conservation Review (Gordon and Sutherland 1993; Waltham et al. 1997; Werritty and McEwen 1997; May and Hansom 2003; Cooper 2007). These encapsulated contemporary understanding of glaciation, Quaternary environmental changes, postglacial landscape evolution and modern coastal and river processes, with detailed accounts of key sites.

1.2.4 Recent Research

The geomorphological themes explored by Sissons and others have continued to engage geomorphologists working in Scotland over the past 40 years, with increasingly sophisticated research on raised shorelines and sea-level change, and the landforms, dimensions and dynamics of Loch Lomond Stadial glaciers, but new or revigorated research has also addressed many other aspects of Scotland's landscape evolution during this period, often reflecting technological advances in geochronology and remotely-sensed data. There have, *inter alia*, been groundbreaking discoveries regarding Cenozoic landscape evolution and the geomorphology of the continental shelf; a revolution in our understanding of the dimensions and dynamics of the last Scottish Ice Sheet; rigorous re-examination and formal classification of Quaternary stratigraphy; a growing awareness of the importance of paraglacial landscape modification; enhanced understanding of river processes and channel changes, and coastal sediment budgets; and heightened appreciation of the importance of geoheritage and geoconservation. Many of these topics were reviewed in a recent journal issue edited by Gordon and Werritty (2019) and permeate the chapters of this book. Several recent books also offer introductions to the geological framework and geomorphology of Scotland, notably Trewin (2002), McKirdy et al. (2007), Gillen (2013), Upton (2015) and Ballantyne (2019). The pace of recent research on the landscapes and landforms of Scotland is relentless; though with James Hutton we can identify the beginning, there is certainly no immediate prospect of an end.

1.3 Revealing Scotland's Geodiversity: The Structure of This Book

As with other volumes in this series, this account of Scotland's landscapes and landforms contains three categories of chapters, each with a different focus: (i) systematic chapters that set the context and chronology for those that follow;

(ii) regionally focused chapters that encompass particular landscapes; and (iii) thematic-based chapters that focus on particular aspects of Scotland's geomorphological heritage (Fig. 1.3).

The systematic chapters begin with an account of Scotland's diverse geology and geological evolution, which extends from the Archaean to the Cenozoic (Chap. 2). Scotland contains some of the oldest rocks in Europe, dating back to ~3.2 Ga, and evidence of several subsequent orogenic events; the present landmass of Scotland did not exist as a single geological entity until ~420 Ma, near the end of the Caledonian Orogeny. Rocks or sediments of all of the Phanerozoic geological periods are represented, from Cambrian to Quaternary, either onshore or in offshore basins. The rock record developed under a range of climates from glacial to hot desert and humid tropical as those parts of the crust that now form Scotland migrated from near the South Pole in Neoproterozoic times, across the Equator to Scotland's present latitude of 55–61°N. Although some of the major relief elements in Scotland may have developed as early as 400 Ma, the present configuration of the Scottish landscape owes much to magmatic activity, crustal uplift, deep weathering and erosion operating under humid subtropical to temperate conditions during the Palaeogene and Neogene Periods (65.5–2.59 Ma). These processes produced the present broad patterns of relief and drainage and are locally manifest in the survival of palaeosurfaces and evidence for etchplanation (Chap. 3). Radical transformation of the end-Neogene landscape occurred during the dramatic climatic shifts of the Quaternary Period, when Scotland experienced the alternation of recurrent glacial and interglacial stages. Cold periods were dominated by the effects of glaciation and periglaciation, and each return to temperate conditions was characterized by paraglacial landscape response, renewed weathering and the operation of slope, fluvial and coastal processes. As Chap. 4 demonstrates, our understanding of events prior to the last ice-sheet glaciation is limited, but the last ice-sheet glaciation (~35–14 ka) and more limited glaciation during the ensuing Loch Lomond Stade (~12.9–11.7 ka) provide analogues of glacial events earlier in the Quaternary. Similarly, geomorphological changes during the Holocene Epoch (present interglacial) provide insights into the nature of landscape evolution during earlier interglacials. Finally, Chap. 5 briefly considers the geomorphological processes that are affecting the Scottish landscape at present, and how these are responding to climate change.

Selection of regions for inclusion in the chapters that follow (Fig. 1.3) was guided primarily by their diversity: all differ in their geomorphological attributes. In part, these differences reflect the influence of geology and geological evolution: Scotland encompasses five major geological terranes, each of which is geologically and geomorphologically distinct, and representation of these terranes partly guided selection of the regional chapters. A second consideration was geographical: Scotland's major archipelagos, for example, exhibit great diversity of geology, relief and landscape, and deserve individual chapters. A final criterion that guided selection of key regions was the literature base: whereas some parts of Scotland are generously endowed with recent publications covering a wide range of landforms, others are relative *terra incognita* in terms of the publication record.

At less than a third of the size, the present land area of Scotland is dwarfed by the area of adjacent offshore shelves (~286,500 km^2). Drawing on a wealth of recent high-resolution marine geophysical and bathymetric data, Chap. 6 explores the geomorphology of this submarine realm, including moraines, grounding-zone wedges, tunnel valleys, streamlined bedforms, iceberg ploughmarks, pockmarks and submarine landslides, and it highlights the importance of submarine landforms in unravelling the glacial history of the last and earlier Scottish ice sheets. Chapters 7–11 are devoted to the iconic geomorphology of Scotland's main islands and archipelagos, beginning with the northern isles of Shetland and Orkney, the visible manifestation of the much larger (but largely submerged) Orkney-Shetland Platform. Shetland (Chap. 7) differs from Orkney (Chap. 8) not only in terms of its greater geological complexity, but also because during the last glaciation it formed an independent centre of ice dispersal. Orkney shares its geology and glacial history with neighbouring Caithness (Fig. 1.3), so both have been included in Chap. 8. Both Orkney and Shetland are exposed to extreme storm and wave conditions on the Atlantic margin, and host spectacular coastal landforms.

The Outer Hebrides (Chap. 9) also represent the exposed parts of a major submarine platform but are almost entirely composed of the resistant Archaean basement rocks (Lewisian gneiss) of the Hebridean terrane, which form wide expanses of low-lying, glacially scoured knock-and-lochan terrain, surmounted by rounded ice-moulded hills and inselbergs, and with a western fringe of calcareous sandy beaches and coastal plains. By contrast, those islands that form part of the Palaeogene Hebridean Igneous Province (Skye, Rùm, Mull and Arran; Chap. 10) are unique in Scotland in illustrating the effects of successive Pleistocene glaciations on terrain underlain by complex igneous geology and the aftermath of glaciation in terms of postglacial landscape evolution on such terrain. Our geomorphological journey to the Hebrides concludes with visits to the islands of Islay, Jura, Colonsay, Tiree and Coll, which collectively host internationally renowned raised shorelines that include

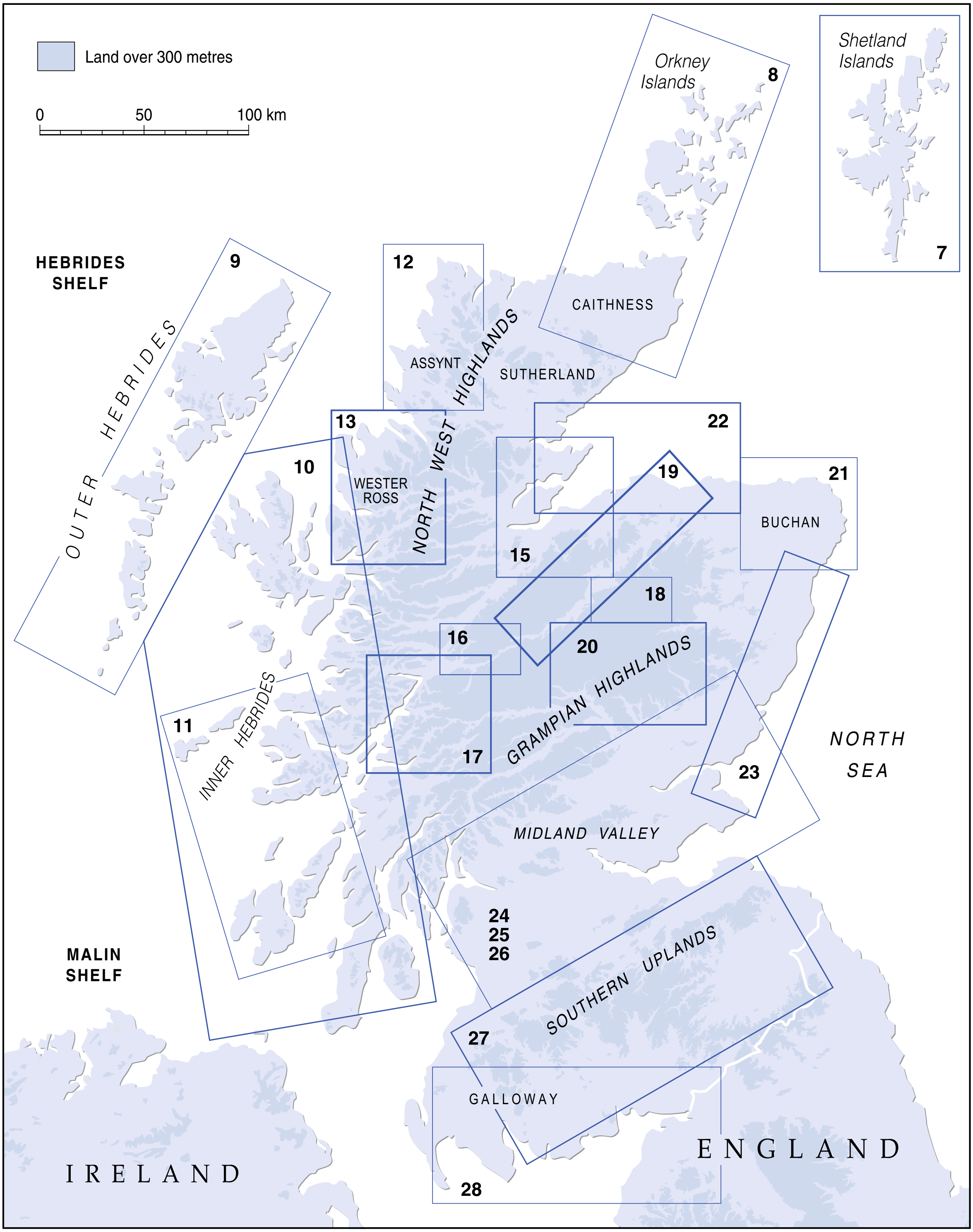

Fig. 1.3 Approximate areas covered by the regional and thematic chapters in this book (numbered boxes). Chapter 6 (Offshore geomorphology) covers the offshore shelf and Chap. 14 (not shown) covers the entire North West Highlands

both raised rock platforms and sequences of raised gravel beaches (Chap. 11).

Chapters 12 (Assynt and Sutherland) and 13 (Wester Ross) cover the spectacular mountainous terrain that straddles the Moine Thrust Zone in the NW Highlands. West of the thrust zone, mountains and inselbergs of Neoproterozoic sandstones and Cambrian quartzite rise above an undulating basement of Archaean gneiss; east of the thrust zone are the glaciated mountains of Moine metasedimentary rocks. All have been transformed by successive glaciations into a landscape of intersecting troughs and glacial breaches that extend westward into some of the finest fjords in Scotland. Chapters 17 (Western Grampian Highlands) and 20 (Central and Eastern Grampian Highlands) illustrate the contrast between the landscapes of intensive glacial dissection that characterize the western Highlands and the landscapes of selective linear glacial erosion that typify more easterly locations, where pre-glacial palaeosurfaces are widely preserved, mainly on Dalradian metasedimentary rocks. The preservation of granite palaeosurfaces and tors is spectacularly illustrated by the Cairngorm Mountains (Chap. 18), a landscape of striking contrasts where cirques and deep glacial troughs are incised into a plateau of essentially pre-glacial origin.

The Midland Valley terrane of central Scotland represents an archetypal landscape of differential erosion. Here, drift-covered undulating lowlands on sedimentary rocks are interrupted by high ground underlain by stacked lava flows, sills and volcanic plugs that have proved resistant to the passage of successive Quaternary ice sheets. The geomorphological diversity of this region is explored in Chaps. 24, 25 and 26 that cover, respectively, the glacial geomorphology of the Highland boundary zone, the legacy of glacial meltwater rivers on the landscape and the effects of ice-moulding in creating a glacial landscape of drumlins, ribbed moraine and crag-and-tail landforms. The Southern Uplands terrane to the south is mainly composed of a Caledonian forearc accretionary complex and forms a rolling tableland, dissected by fluvial valleys and glacial troughs. The upland landscapes of this distinctive region are explored in Chap. 27, and the glacial and coastal characteristics of the adjacent Solway coast and lowlands are described in Chap. 28.

The thematic chapters of this book have been chosen to represent particular facets of Scotland's geomorphology. The legacy of pre-glacial landscape evolution is explored in Chap. 21, which describes the palaeolandscape of the Buchan region of NE Scotland, where the effects of glaciation are muted and evidence for pre-glacial landscape evolution, such as basins and saprolite covers, is exceptionally well-preserved. Chapter 15 focuses on the rich legacy of glacial and glacifluvial geomorphology of the Inverness and Moray Firth area. Chapter 16 is set in the Lochaber area of the Western Grampians, where shorelines and lake deposits in Glen Roy and neighbouring valleys record the history of glacier-dammed lakes during the Loch Lomond Stade; this is one of the most-visited and most intensively researched geomorphological locations in Scotland. Finally, five thematic chapters explore aspects of the postglacial evolution of the Scottish landscape. Chapter 14 records over 400 rock-slope failures in the NW Highlands and considers their causes, timing and influence on long-term landscape evolution, whilst Chap. 19 is devoted to the fluvial geomorphology of the catchment of the River Spey and its much-studied tributary, the River Feshie. The origins, development and present status of the spectacular beaches, spits and barrier islands of the Moray Firth coast and eastern Scotland are explored in Chaps. 22 and 23. The volume concludes with Chap. 29 on geoconservation, which outlines the approach and measures adopted to protect Scotland's exceptional geoheritage and key geomorphological sites.

1.4 Conclusion

The outstanding diversity of Scotland's landscapes and landforms has captured the imagination of geologists and geomorphologists for over two centuries, leading to groundbreaking insights of international significance, such as the concepts of uniformitarianism, the Glacial Theory, thrust-faulting, glacio-isostatic uplift, the significance of meltwater channels and the formation of raised shorelines. The Late Pleistocene glacial history of Scotland between ~35 and 11 ka has been established in greater detail than probably for any other part of the globe. Yet despite the contribution of geologists and geomorphologists working in Scotland to our understanding of geomorphological processes and landscape evolution, there remain gaps that became evident during the compilation of this book. The nature of pre-glacial landscape evolution during the Cenozoic Era remains to be unravelled for large parts of the country, and only a limited number of studies provide insights into Quaternary landscape changes prior to the last ice-sheet glaciation. Within Scotland, much recent research has been focused on particular topics (such as the last glaciation and glacial-interglacial transition, and sea-level change) and particular locations, such as the Isle of Skye, the Cairngorms and Glen Roy. Some parts of the Scottish landscape, such as parts of the NW Highlands and the Southern Uplands, have received little attention during the present century. There remains much to be discovered about (and from) the landscapes and landforms of Scotland that, as James Hutton presciently suggested, 'may afford the human mind both information and entertainment' (Hutton 1785, p. 30). This applies not only from a research perspective, but through enhancement of wider public awareness and appreciation of the origins of Scotland's scenery and its significance both for geoscience and as an asset for its aesthetic and cultural values, enjoyment and inspiration.

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Part I

Physical Environment

2 Scotland: Geological Foundations and Landscape Evolution

John E. Gordon and Philip Stone

Abstract

Scotland displays remarkable geological diversity, comprising igneous, sedimentary and metamorphic rocks that date back to the Archaean and span all the major periods of the geological timescale. This geodiversity reflects global plate tectonics, the position of Scotland near convergent and divergent plate margins at different times during its history, the drift of the Scottish landmass as a small part of larger continental entities across the surface of the globe through different morphoclimatic zones and depositional environments, and long-term changes in global climate and sea level. The Scottish landmass was essentially assembled ~425 Ma, from the amalgamation of the Laurentia and Baltica–Avalonia plates during the Caledonian Orogeny as the Iapetus Ocean closed. This brought together different crustal fragments comprising Archaean igneous and metamorphic rocks, and Proterozoic and Early Palaeozoic sedimentary and metasedimentary rocks. Orogenesis was accompanied by the emplacement of igneous intrusions, and uplift and erosion of the Caledonian mountain chain. Deposition of thick accumulations of terrestrial sediments during the Devonian was followed later by Palaeozoic and Mesozoic marine, deltaic and continental sedimentation and Palaeozoic igneous activity. During the Palaeogene, central igneous complexes and extensive lava fields formed in the Hebridean Igneous Province due to the presence of an underlying mantle plume that also initiated the opening of the North Atlantic Ocean to the west. The influence of this geodiversity is widely expressed in the geomorphology of the present landscape as a consequence of differential weathering and erosion, modulated by variable tectonic uplift, tilting and downwarping, mainly during post-Caledonian times and particularly during the Cenozoic.

Keywords

Geological history • Plate tectonics • Caledonian Orogeny • Palaeogene volcanism • Differential weathering and erosion • Geology and landscape

2.1 Introduction

This chapter outlines the development of the geological framework that provides the foundation for the geomorphology of Scotland, and the processes and events that have shaped the present landscape (Fig. 2.1). In recent decades, new dating methods, geophysical models, and studies of Mesozoic and Cenozoic deposits in offshore sedimentary basins have provided fresh insights into the links between geology, tectonics, surface processes, climate and long-term landscape evolution, highlighting the antiquity of many features in the present landscape (Hall 1991; Hall and Bishop 2002; Fame et al. 2018; Chap. 3).

Although located today on the passive continental margin of NW Europe, the geological development of Scotland has given rise to remarkable geodiversity for a country of its size, with the presence onshore and in offshore sedimentary basins of rocks spanning all the major geological time periods since the Archaean (Trewin 2002; McKirdy et al. 2007; Stone 2008; Gillen 2013). Several factors account for this geodiversity, which reflects the combined influence of geological and geomorphological processes operating over a wide range of timescales.

First, the broad geological and topographic framework of Scotland is the result of major global tectonic events. These

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Fig. 2.1 Scotland: topography and main locations mentioned in the text. (Map base by Eric Gaba, NordNordWest, Uwe Dedering from Wikimedia Commons CC BY-SA 3.0)

include the repeated formation and break-up of supercontinents, the closure of the Early Palaeozoic Iapetus Ocean and the associated crustal shortening and uplift during the Caledonian Orogeny, Mesozoic rifting and subsidence in the North Sea Basin and the Minches, and finally the opening of the North Atlantic Ocean in the Palaeogene (Table 2.1). The effects of these global-scale events have also imparted more local geological influences on the landscape and landforms through the formation of faults, shatter zones and structural grain, and the juxtaposition of resistant and weaker rocks, all subsequently exploited by surface processes of differential weathering and erosion, particularly during the Cenozoic.

Second, the position of Scotland near an active plate margin at various times during its history has resulted in episodes of mountain building, volcanism, large-scale crustal deformation, uplift and erosion, all of which have left a legacy both in the rock record and in the present landscape.

Third, as a result of plate tectonics, the Scottish landmass has drifted across the surface of the globe and through different climatic zones and morphogenetic and sedimentary

Table 2.1 Summary of the geological history of Scotland (Main sources: Trewin 2002; Woodcock and Strachan 2012; Gillen 2013)

Era	Period	Age (Ma)	Latitude of Scotland[a]	Main events and palaeoenvironments
Cenozoic	Quaternary	2.6–0	57°N	Repeated growth and decay of mountain glaciers and ice sheets; glacial stripping of saprolites and cumulative development of landforms of glacial erosion; formation of deglacial and postglacial landforms
	Neogene	23–2.6	50–57°N	Phases of uplift and erosion; weathering continued under warm-temperate, humid climate; climate cooling intensified after ~3 Ma
	Palaeogene	66–23	45–50°N	Opening of the northern North Atlantic Ocean (~56 Ma); extensive intrusive and extrusive igneous activity in the Hebridean Igneous Province (~61–55 Ma); differential uplift, tilting and erosion of the Scottish landmass; deep chemical weathering initially under humid, subtropical climate; stripping of saprolites and formation of etchplains
Mesozoic	Cretaceous	145–66	40–45°N	Warm, shallow seas covered much of Scotland, apart from the Highlands and Southern Uplands
	Jurassic	201–145	35–40°N	The Highlands formed an upland area of reduced relief, with rivers draining to deltas in the North Sea Basin; deposition of sediments in shallow seas around the margins of the uplands
	Triassic	252–201	25–35°N	Hot, semi-arid environment with deserts present around the margins of the Highlands; deposition of the New Red Sandstone
Palaeozoic	Permian	299–252	15–25°N	Hot, arid environment with deserts present around the margins of the Highlands; deposition of the New Red Sandstone; crustal extension and rifting initiated the North Sea Basin
	Carboniferous	359–299	5°S–15°N	Scotland drifted north into equatorial latitudes; tropical forests, intermittent shallow shelf seas and deltas were present in the subsiding Midland Valley; formation of coal measures in low-lying coastal swamps; eruption of central volcanoes and lavas in the Midland Valley; erosion of the Caledonian mountains continued
	Devonian	419–359	10–5°S	Erosion of the Caledonian mountains accompanied by the formation of large alluvial fans in hot, semi-arid conditions; large inland lake basins formed in NE Scotland and Orkney (Orcadian Basin); volcanic activity in the Midland Valley and Lochaber; Acadian orogenic event

(continued)

Table 2.1 (continued)

Era	Period	Age (Ma)	Latitude of Scotland[a]	Main events and palaeoenvironments
	Silurian	444–419	15–10°S	Closure of the Iapetus Ocean as Baltica, then Avalonia, collided with Laurentia, joining the crustal units of Scotland with those of England and Wales along the Iapetus Suture; Scandian orogenic event; uplift and erosion of the Caledonian mountains; Moine sediments folded and metamorphosed; major crustal dislocations along the Moine Thrust Zone; turbidite sediments deposited in the Iapetus Ocean with ensuing accretionary deformation (Southern Uplands terrane); volcanic activity and granite intrusions in the Highlands and volcanic activity in the Midland Valley
	Ordovician	485–444	30–15°S	The Iapetus Ocean was closing rapidly; limestones formed on the shelf of a warm shallow sea in NW Scotland; turbidite sediments deposited in the Iapetus Ocean with ensuing accretionary deformation (Southern Uplands terrane); Grampian orogenic event as the Midland Valley Arc collided with Laurentia; uplift of the Caledonian mountains accompanied by metamorphism and folding of Dalradian sediments and intrusion of gabbros and granites in the Grampian Highlands
	Cambrian	541–485	30°S	Sandstones, then limestones formed in warm, shallow shelf seas in NW Scotland; the Iapetus Ocean reached its maximum extent
Neoproterozoic		1000–541	Close to South Pole	~800–500 Ma: Dalradian sediments deposited in marine basins, mainly on the continental margin; glaciations occurred on a global scale; the Iapetus Ocean opened
			South of Equator	~1000–875 Ma: Moine sediments deposited mainly in shallow seas
Mesoproterozoic		1600–1000	10–30°N to 15–20°S	~1200–950 Ma: Torridonian sandstones mainly formed from sediments eroded from the mountains of the Grenville Orogeny (~1100–1000 Ma) on the Laurentian plate and deposited in alluvial, fluvial and lacustrine environments on its continental margin ~1200–1000 Ma: Lewisian rocks exposed at the surface following uplift and erosion
Palaeoproterozoic Neoarchaean Mesoarchaean		3200–1600	Unknown	~2300–1700 Ma: Laxfordian event ~2400–2000 Ma: formation of Scourie Dykes ~2900–2300 Ma: Badcallian event ~3200–2800 Ma: formation of the Lewisian Gneiss Complex by metamorphism of older igneous and subsidiary sedimentary rocks

[a]All latitudes are generalisations; there is a range within individual time periods and not all schemes agree. The overall pattern is one of northward drift of Scotland during the last ~1000 Ma

environments, ranging from near-polar to tropical, further enhancing the geodiversity. Moreover, long-term global climate change (Williams et al. 2007) has influenced surface processes, particularly the overall global cooling from the Early Eocene climate optimum to the Quaternary glaciations.

Fourth, long-term trends in eustatic sea level (Haq and Schutter 2008; Wright et al. 2020), arising from tectonic processes and the growth and decay of continental ice sheets, have resulted in significant changes in the position of the coastline.

The combination of these factors has resulted in Scotland now comprising rock types that range from among the oldest in Europe to the most recent, and with a wide variety of igneous, metamorphic and sedimentary origins (Fig. 2.2). As a result of weathering and erosion over long periods of geological time, the characteristics of the different rocks and their geological histories are reflected today at a variety of scales in the diversity of landscapes, landforms and coastal configurations.

2.2 Palaeogeography and Plate Tectonics

The global operation of plate tectonics has caused the relative configuration of the Earth's continental masses and oceans to change continuously over geological time. Oceans have opened and closed, and continents have fragmented and merged, such rifting and collisions being accompanied by igneous activity and the uplift and formation of ancient and modern mountain ranges. The global perspective has been recently reviewed and illustrated by Torsvik and Cocks (2017), and there is an extensive literature pertaining to the tectonic development of Scotland (Trewin 2002; Jones and Blake 2003; Hunter and Easterbrook 2004; Macdonald and Fettes 2007; Woodcock and Strachan 2012; Chew and Strachan 2014). Scotland's basic geological framework owes much to these processes and to the geographical location of Scotland near continental plate margins at times of continental rupture and collision, and at other times locked within continental interiors.

For most of the Earth's geological history, Scotland did not exist as a single entity. Around 425 Ma ago, towards the end of the Caledonian Orogeny, the disparate crustal fragments that now comprise the present landmass came together, all as parts of larger geological units (Oliver 2002; Holdsworth et al. 2012). Events during the Caledonian Orogeny were fundamental in shaping the present geology, and hence landscape of Scotland; their legacy is clearly evident today in the structural grain of much of the landscape and consequently in the character and orientation of major landforms.

The fundamental geological framework of Scotland consists of five main units or terranes (Fig. 2.3):

- The Hebridean terrane forms the north-west foreland to the Caledonian orogenic belt and is bounded on its eastern side by the Moine Thrust Zone. It comprises the Archaean to Palaeoproterozoic Lewisian Complex and its unconformable cover of Mesoproterozoic to Cambro-Ordovician sedimentary rocks.
- The Northern Highlands terrane to the north-west of the Great Glen Fault is composed mainly of metasedimentary rocks of the Neoproterozoic Moine Supergroup.
- The Grampian Highlands terrane between the Great Glen Fault and the Highland Boundary Fault is dominated by Neoproterozoic to Cambrian metasedimentary and volcanic rocks of the Dalradian Supergroup, locally intruded by large felsic and mafic plutons during the Caledonian Orogeny.
- The Midland Valley terrane is a structural graben between the Highland Boundary Fault and the Southern Upland Fault, and contains a range of Palaeozoic sedimentary and igneous rocks.
- The Southern Uplands terrane, to the south of the Southern Upland Fault, is an accretionary complex of Ordovician and Silurian oceanic sediments that were stripped from the underlying oceanic crust as it entered a subduction zone.

As a result of differential erosion, the boundaries between the five terranes, the Moine Thrust Zone, the Great Glen Fault, the Highland Boundary Fault and the Southern Upland Fault, now form major topographic features (Fig. 2.4). The assembly of the different terranes was driven by global plate tectonics and involved processes of ocean closure, continental collision, and the lateral movement of geographically separate terranes along strike-slip faults. This was accompanied, and followed, by the continental drift of northern Scotland from a position close to the South Pole at ~600 Ma, northwards across the Equator and on to its present location (Fig. 2.5; Table 2.1).

Around 1000 Ma, the ancient northern part of Scotland lay within the supercontinent of Rodinia (Li et al. 2008). Debris eroded from the mountains of Rodinia was transported by rivers and deposited in basins on the continental margins and adjacent ocean floor to form the Torridonian and Moine rocks, respectively (Table 2.1). Rodinia started to break up ~750 Ma, with northern Scotland incorporated into the Laurentian plate, but the main continental fragments later re-assembled to form another supercontinent, Pannotia. Northern Scotland lay near the South Pole on the eastern edge of Laurentia, which included North America and Greenland (Fig. 2.5a). As Pannotia broke up and Laurentia drifted northwards, the Iapetus Ocean developed between Laurentia and a southern continent, Gondwana (Fig. 2.5b). Northward drift of Scotland continued for over 500 million

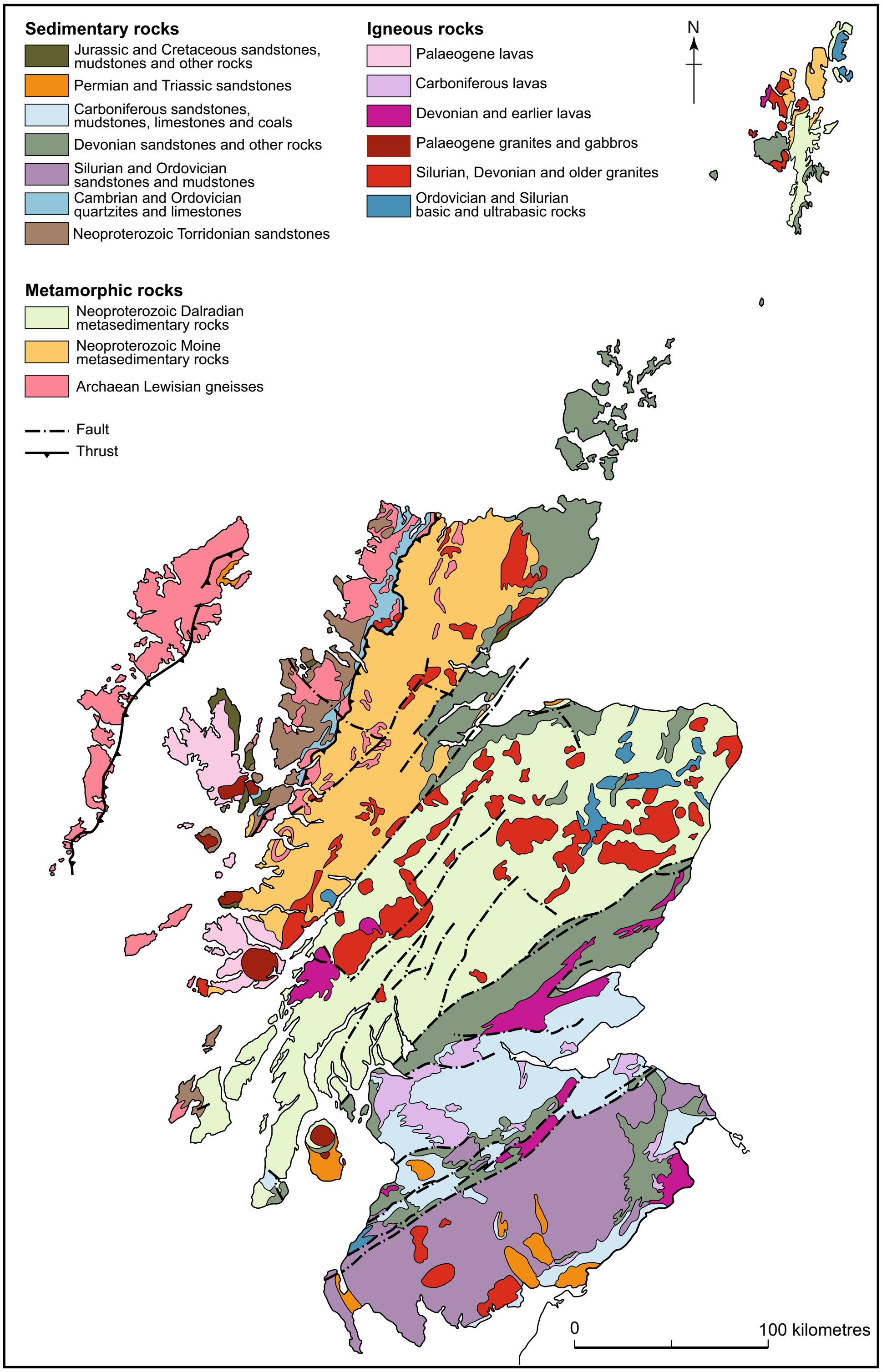

Fig. 2.2 Simplified version of the BGS 1:625 k Bedrock Geology map of Scotland. (Permit Number CP20/008 Slightly Modified from BGS © UKRI. All Rights Reserved)

years until ~55 Ma. Since then, and as the westernmost part of Eurasia, Scotland has moved broadly eastwards as the North Atlantic Ocean has progressively widened.

The Neoproterozoic and Early Palaeozoic (1000–425 Ma) geology of Scotland developed as the Iapetus Ocean opened and closed. At the ocean's northern margin, deep- and shallow-water sediments were deposited in two broadly contemporaneous sub-parallel zones adjacent to Laurentia forming, respectively, what were to become the Dalradian rocks of the Grampian Highlands and the Cambro-Ordovician succession of the NW Highlands. The Iapetus Ocean separated Scotland from England and Wales

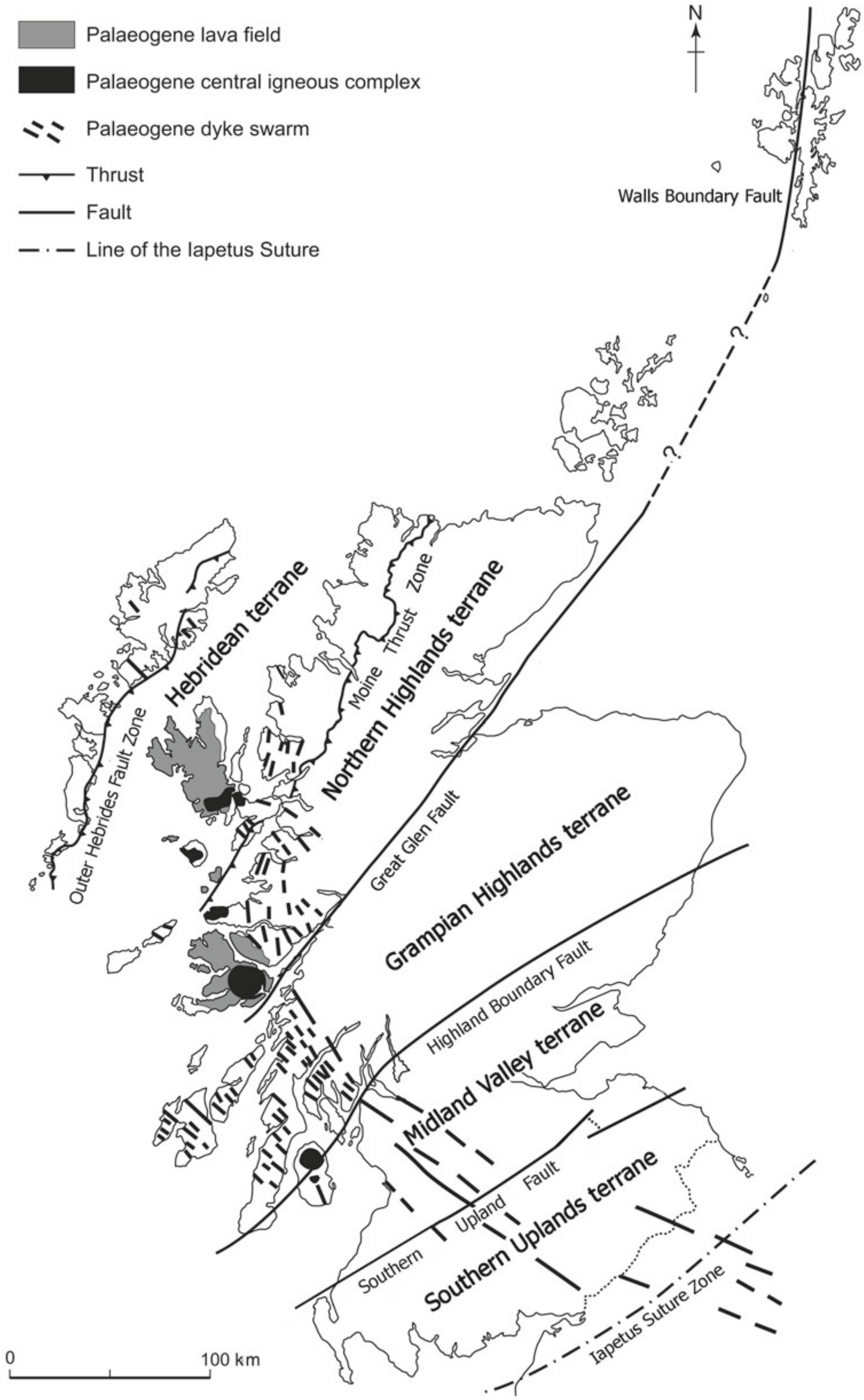

Fig. 2.3 The major geological features and terranes of Scotland

which were at that time part of Avalonia, a micro-continent to the south that had rifted from Gondwana and was detached from it by the opening Rheic Ocean. By ~425 Ma, however, the Iapetus Ocean had closed, accompanied by tectonic events that had a fundamental effect on the shaping of Scotland (Fig. 2.5c–e).

As the opposing continental plates of Laurentia, Avalonia and Baltica converged and the Iapetus Ocean closed, the

Fig. 2.4 Aerial view of Loch Lomond from the south. The Highland Boundary Fault (white line) separates the Devonian sedimentary rocks of the Midland Valley terrane (foreground) from the more resistant Dalradian metamorphic rocks of the Grampian Highlands terrane to the north. Duncryne Hill (D) is a resistant Early Carboniferous volcanic vent. (Permit Number CP20/008, Photograph P001246 digitized with grant-in-aid from SCRAN the Scottish Cultural Resources Access Network, Reproduced with permission of BGS © UKRI. All Rights Reserved—Sourced: https://geoscenic.bgs.ac.uk/asset-bank/action/viewAsset?id=2543&index=0&total=1&view=viewSearchItem)

crustal fragments that now comprise Scotland were brought together. The closure of the Iapetus Ocean also joined Scotland with England, along the major structure now known as the Iapetus Suture Zone (Fig. 2.3). Despite its geological significance, the Iapetus Suture Zone has no obvious surface expression and is buried beneath the younger rocks of northern England. However, many of the more recent extensional faults that cut NE–SW across northern England have been shown by geophysical modeling to be rooted into the suture, demonstrating a post-Caledonian history of reactivation.

The crustal shortening, thickening and deformation that accompanied ocean closure and continental collision were focused into a series of mountain building events between ~490 and ~390 Ma that constitute the Caledonian Orogeny. The initial collision between a volcanic island arc, the Midland Valley Arc (which now forms the basement to the Midland Valley terrane) and the Grampian Highlands terrane (part of Laurentia) occurred at ~480–465 Ma and produced the Grampian orogenic event (Fig. 2.5d). During this event, the Dalradian sediments were buried deep in the crust and metamorphosed at temperatures of over 600 °C and pressures of at least 800 MPa (≡ ~25 km depth). The compressive forces produced major recumbent folds (-nappes) in the Grampian Highlands (Stephenson and Gould 1995). Consequently, in some areas, the original stratification dips steeply or is overturned, as in the Tay Nappe at Ben Lawers.

The later Scandian orogenic event of ~435–420 Ma resulted from the 'hard' continent–continent collision of Baltica and that part of Laurentia containing the Northern Highlands (Fig. 2.5e). During this event the Moine

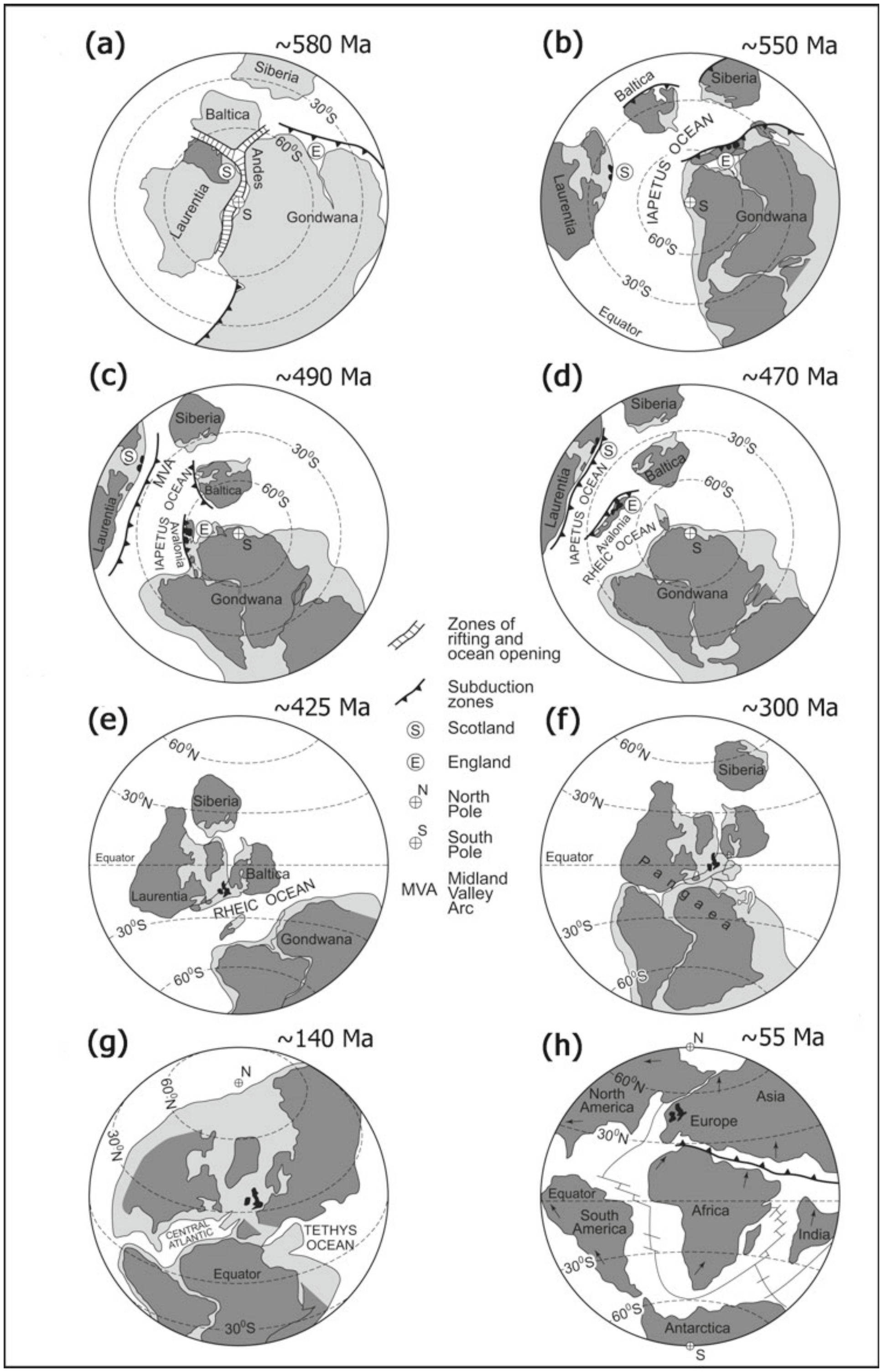

Fig. 2.5 Global palaeocontinental reconstructions showing changes in the position of Scotland at selected times and relative to past continental landmasses. (**a**–**g**: after Oliver (2002) Chronology and terrane assembly, new and old controversies. In: Trewin NH (ed) The geology of Scotland, 4th edn. The Geological Society, London, pp 201–211; Holdsworth et al. (2012) Geological framework of Britain and Ireland. In: Woodcock N, Strachan R (eds) Geological history of Britain and Ireland. Wiley-Blackwell, Chichester. Copyright (2012) with permission from John Wiley and Sons conveyed through Clearance Center, Inc; and Torsvik et al. (1996) Continental break-up and collision in the Neoproterozoic and Palaeozoic—a tale of Baltica and Laurentia. Earth-Sci Rev 40:229–258. Copyright (1996), with permission from Elsevier); **h**: after Dietz and Holden (1970) Reconstruction of Pangaea: Breakup and dispersion of continents, Permian to Present. J Geophys Res 75:4939–4956. Copyright (1970) by the American Geophysical Union, Copyright (2012) John Wiley and Sons, with permission from John Wiley and Sons conveyed through Copyright Clearance Center, Inc)

sedimentary succession of the Northern Highlands terrane was buried and metamorphosed. At its western limit, a series of low-angle thrusts and related folds were generated by the westward movement of older rocks over the top of younger ones by at least 70 km and perhaps as much as 100 km, such that Moine rocks (and some of their Lewisianoid basement) now rest on Cambro-Ordovician rocks. The Moine Thrust (Fig. 2.3) is the largest and best known of these crustal dislocations and is clearly displayed at Knockan Crag in Assynt, within the NW Highlands UNESCO Global Geopark, where there is on-site interpretation of its formation. The structures now seen at the surface were originally formed at a depth of several kilometres in the crust, but their presence has been revealed by subsequent erosion of the overlying rocks. The Caledonian 'docking' of the continents also involved significant lateral movements of hundreds of kilometres, generally along an array of northeast-trending strike-slip faults (Fig. 2.6). These movements united formerly geographically separate terranes, such as the Northern Highlands and Grampian Highlands terranes. Farther south, Avalonia finally docked in a 'soft' oblique collision with Laurentia around at ~425 Ma (Fig. 2.5e), with the leading edge of Laurentia (the Southern Uplands terrane) over-riding and depressing the margin of Avalonia.

The uplift that accompanied and followed the Grampian and Scandian orogenic events formed the Caledonian Mountain Belt, the uplifted and eroded roots of which now occur in the Scottish Highlands, the Appalachians, Newfoundland, Greenland, Scandinavia and Svalbard. Farther south, as the Iapetus Ocean closed, the sediments that had accumulated on its floor were incorporated into an extensive thrust belt that formed above the subduction zone carrying the ocean crust beneath Laurentia; this accumulation of tectonically imbricated deep-marine strata formed the Southern Uplands accretionary complex. Igneous activity occurred synchronously with the subduction of the Iapetus Ocean floor beneath the Laurentian continental margin and masses of magma rose up into the overlying crust to crystallize as the numerous granite intrusions of the Grampian Highlands and Northern Highlands (Fig. 2.2). The final stages of the collision between Laurentia and Avalonia (~410–390 Ma) also involved intrusion of granite plutons and dykes of variable composition into the rocks of the Southern Uplands terrane.

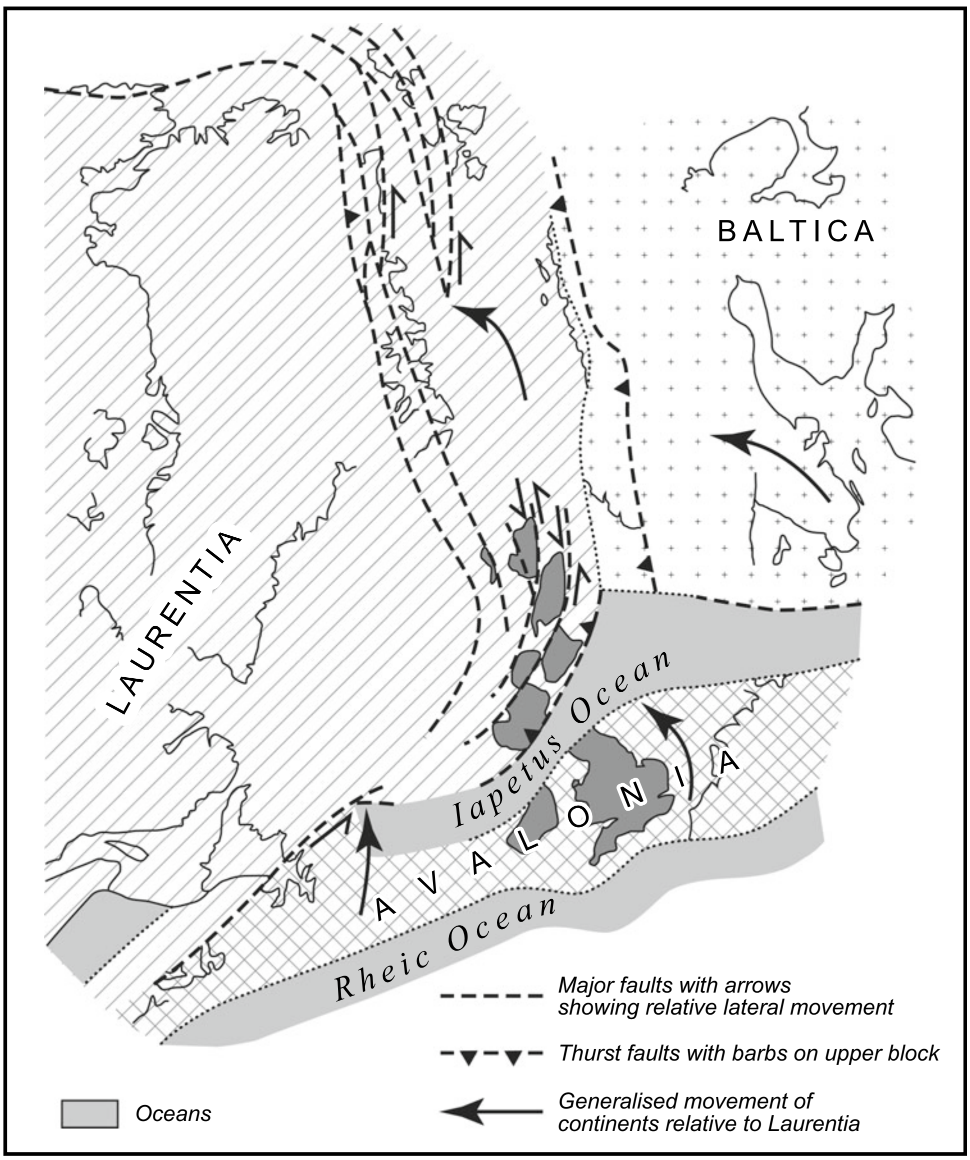

Fig. 2.6 Schematic reconstruction of the closure of the Iapetus Ocean and the assembly of the main geological terranes along major strike-slip faults. (Adapted from Stephenson et al. (1999) Caledonian igneous rocks of Great Britain. Geological Conservation Review Series 17. Joint Nature Conservation Committee, Peterborough, with permission from JNCC, and Soper et al. (1992) Sinistral transpression and the Silurian closure of Iapetus. J Geol Soc Lond 149:871–880 © Geological Society of London 1992)

By the beginning of the Devonian, following the closure of the Iapetus Ocean, the geological framework of Scotland was essentially in place, although granite intrusion and eruption of lavas continued over a wide area until ~390 Ma. Far-field stresses from the mid-Devonian Acadian orogenic event produced deformation along the major terrane-bounding faults and the formation of the NE–SW trending Sidlaw Anticline and Strathmore Syncline at 399–393 Ma (Mendum 2012), fold structures now reflected in the topographic form of the northern part of the Midland Valley. Thereafter, erosion further lowered the Caledonian mountains and exposed the granite intrusions, such as those of the Cruachan-Etive mountains and the Cairngorms. As a component of the Laurentia–Avalonia continent, Scotland thereafter continued its northward drift towards equatorial regions, later becoming part of a new supercontinent, Pangaea, following closure of the Rheic Ocean to the south during the Carboniferous (Fig. 2.5f); further far-field stresses from the deformation linked to closure of the Rheic Ocean (the Variscan Orogeny) produced folding in the Carboniferous rocks of the Midland Valley.

During the Late Palaeozoic and Mesozoic, extension and rifting of the crust led to the formation of new sedimentary basins and episodes of resurgent igneous activity. Sediments were deposited in the subsiding basins under a range of climatic conditions and depositional environments. The resulting strata include the terrestrial and lacustrine Devonian sandstones and conglomerates of the Old Red Sandstone Supergroup in the Midland Valley, Caithness and Orkney, the shallow marine and deltaic Carboniferous sequences of the Midland Valley, the Permo-Triassic desert sandstones (the New Red Sandstone) of Moray, the Midland Valley and the Southern Uplands outliers, and the Jurassic and Cretaceous shallow marine sedimentary rocks of Skye and eastern Sutherland (Fig. 2.2). The deposited sediment probably originated partly from erosion of the Caledonian mountains, but considerable sediment recycling must have occurred since there is a lack of evidence of direct derivation of material from the Scottish Highlands in the now-adjacent Devonian and Carboniferous successions. During post-Caledonian times, the emplacement of igneous intrusions and widespread eruptions of lavas and tuffs reflected crustal tension associated with distant plate movements. Notable representatives are the Devonian lavas of the Sidlaw and Ochil hills and the Carboniferous sills and lavas of the Midland Valley. From about the Late Permian onwards, crustal extension and rifting also initiated the North Sea sedimentary basins.

Early in the Jurassic, Pangaea began to break up with the opening of a southern Atlantic Ocean between America and Africa after ~180 Ma (Fig. 2.5g). Crustal extension eventually spread northwards into the western Scotland zone that had experienced extensional tectonics since the Triassic, and by the Middle Jurassic a shallow marine basin had developed along the west coast of Scotland. By ~65 Ma the opening of the northern part of the Atlantic was well under way, leading eventually to the detachment of Scotland from North America and Greenland at ~56 Ma as the crust thinned and separated above an underlying mantle plume (Fig. 2.5 h). Between ~61 and ~55 Ma, during the Early Palaeogene, the mantle plume gave rise to extensive magmatic activity in the Hebrides and on the northwest continental shelf. This resulted in the formation of central igneous complexes that represent the plutonic roots of large volcanoes, and the spread of extensive lava fields from fissure eruptions (Emeleus and Bell 2005; Figs. 2.2 and 2.3). Episodic tectonic and thermal uplift associated with the continental break-up also elevated parts of the Scottish landmass during the Palaeogene and Neogene. Landscape modification proceeded through differential weathering and erosion under relatively warm and humid conditions during these times (Chap. 3), and subsequently through glacial and periglacial processes during the course of repeated Quaternary glaciations (Chap. 4). Nevertheless, many pre-glacial features still occur in the present landscape.

2.3 Landscape Evolution

Multiple stages of uplift, weathering and erosion during post-Caledonian times, and particularly during the Cenozoic, shaped many of the elements of the present landscape, but there is evidence that depths of exhumation of basement rocks in the Highlands did not exceed 1–2 km (Hall 1991; Hall and Bishop 2002; Holford et al. 2010; Fame et al. 2018; Hall et al. 2019; Chap. 3). Following extensive erosion during the Devonian, the Highlands probably formed an area of moderately low relief during the later Palaeozoic and Mesozoic. Differential erosion over long timescales exploited faults and weaker rocks and emphasized the underlying structural grain of the landscape. It is unclear how much this ancient landscape pattern, formed ~400 Ma ago, has influenced the current topography, but some features are undoubtedly inherited from that time. Examples include the Great Glen, the main Grampian watershed and the margins of various basins in the Highlands and the Moray Firth (Hall 1991), while the Midland Valley probably existed by ~300 Ma. During the Late Cretaceous, a time of high eustatic sea level, much of Scotland, apart probably from high ground in the Highlands and Southern Uplands, was submerged beneath a shallow tropical shelf-sea (Harker 2002). Remnants of the associated sedimentary rocks (chalk, sandstones and residual flint beds) now occur in Morvern and on Mull, Eigg, Skye and Raasay (Emeleus and Bell 2005).

Apart from the foreland of NW Scotland, the overall NE–SW topographic grain of much of the Scottish landscape is inherited from the Caledonian Orogeny, and in particular from the alignment of major Caledonian faults. Valleys and drainage patterns partly reflect this geological inheritance and are partly superimposed across it (Hall 1991). Uplift of the landmass of Scotland by up to a kilometre accompanied Palaeogene magmatic activity along the western seaboard and was followed by renewed uplift during the Neogene (Holford et al. 2010; Stoker et al. 2010). While uplift was generally greatest in the west and in the Cairngorms, it varied spatially across Scotland between different crustal blocks, as did accompanying exhumation and overall eastward tilting of the Highlands (Ringrose and Migoń 1997; Fame et al. 2018); some areas, such as Buchan (Chap. 21), apparently remained stable. The uplift in the west and eastward tilting of the landmass led to the location of the main pre-glacial north–south watershed in the western Highlands, with longer east-flowing rivers and short, steep west-flowing ones (Chap. 3). Extensive dissection of much of the upland landscape by rivers accompanied the uplift, particularly in the west, initiating contrasts in landscape between the west and east Highlands that were later emphasized by glacial erosion (Chap. 4). Intense erosion was focused on the Palaeogene central volcanoes on Skye, Rùm, Mull and Arran, exposing their root intrusions that originally lay about 2 km below the surface. In contrast, the North Sea to the east has been an area of overall subsidence and sediment accumulation since the Late Permian.

During the Palaeogene and Neogene, the climate in Scotland was warmer and more humid than today. Under these conditions, the surface layers of the bedrock were deeply chemically weathered, and pockets of saprolite have survived subsequent erosion and glaciation, particularly in tectonically stable areas of NE Scotland (Chap. 21). Generally, however, as the landmass of Scotland underwent phases of differential uplift during the Palaeogene and Neogene, the saprolite covers were stripped by subaerial erosion and slope retreat, and later by successive Quaternary ice sheets, forming a series of glacially modified and tilted etch plains and basins (Hall 1991; Ringrose and Migoń 1997; Hall and Bishop 2002). These palaeosurfaces are most extensive in the eastern Grampian Highlands, NE Scotland and the Southern Uplands (Chaps. 3, 18, 20 and 27), although surface fragments and summit accordances are also evident in parts of the glacially dissected western and northern Highlands (Godard 1965; see Fig. 2.8b below). Because uplift and precipitation were generally greater in the west, the pre-glacial landscape there was probably more extensively dissected by rivers; also, glacial erosion was more 'selective' in the east than in the west (Chap. 4). Differences in rock resistance to weathering and erosion also produced topographic basins, which are a distinctive feature of NE Scotland and parts of the Highlands (e.g. Howe of Cromar in Aberdeenshire, Rannoch Moor in Lochaber and the Naver basin in Sutherland; Hall 1991). Over the last 50 Ma, following a global climate 'optimum' in the Early Eocene, a long-term trend of stepwise cooling intensified at ~3 Ma (Zachos et al. 2001), leading to the onset of Quaternary glaciations in Scotland.

2.4 The Geological Framework and Its Influence on the Landscape

Geologically, Scotland may be divided into six geographical regions that correspond broadly with the five major terranes and the Hebridean Igneous Province (Fig. 2.3). Each experienced a distinctive geological evolution that is reflected in the rocks present and the influence of these rocks on the landscape as moderated through Cenozoic tectonic activity and the processes of differential weathering and erosion (Sissons 1976; Gillen 2013; Ballantyne 2019).

2.4.1 The Hebridean Terrane

The Hebridean terrane includes the Outer Hebrides and the NW foreland to the Caledonian orogenic belt (Fig. 2.3). This region contains the oldest rocks in Scotland and some of the oldest in Europe, the Lewisian gneisses, which originated as igneous intrusions emplaced deep in the Earth's crust at ~3,200–2,800 Ma (Fig. 2.7a, b). These dominantly sheeted intrusions of tonalite, granodiorite and gabbro have been intensely deformed and metamorphosed during a long and complex history involving the amalgamation of earlier terranes, and represent the eroded remnants of the early Laurentian continental mass encompassing what is now Greenland and Canada. The region, extending from Cape Wrath to Knoydart, and including the Outer Hebrides and the islands of Coll, Tiree and Iona, and parts of SE Skye and SW Islay, was largely unaffected by the Caledonian Orogeny and, in parts, the structural grain in the rocks is orientated WNW–ESE, an ancient trend which is reflected in the alignment of some of the glens, lochs and fjords along the northwest seaboard. Some of these are aligned along faults (e.g. Loch Maree and Loch Shin), whereas others follow major shear zones (e.g. Loch Laxford). Two major metamorphic, tectonic and intrusive events during the Archaean and Palaeoproterozoic reflect ancient plate movements: the Badcallian (~2900–2300 Ma) and the Laxfordian (~2300–1700 Ma) events; these were separated by an intervening period of crustal tension during which tholeiitic dolerite dykes (the Scourie Dykes) were emplaced at ~2400–2000 Ma (Fig. 2.7a). Subsequent prolonged uplift and erosion exposed the Lewisian gneiss at the surface at ~1200–1000 Ma, forming a low hilly landscape prior to deposition of the overlying Torridonian sandstones and

Cambro-Ordovician rocks. Very substantial erosion, perhaps some tens of kilometres, is implied by exposure at the surface of these high-grade metamorphic rocks, which were originally produced deep in the continental crust. Today, the Lewisian landscape is characterized by rocky, ice-scoured hills developed on an ancient erosion surface at least in part exhumed from beneath Torridonian and Cambrian rocks in NW Scotland. This landscape appears to have developed as an etch surface with varying depths of chemical weathering exploiting weaknesses in the rock. Removal of the saprolite cover by glacial erosion has formed the characteristic knock-and-lochan topography on low ground (Krabbendam and Bradwell 2014; Fig. 2.7b, c).

Unconformably overlying the Lewisian gneiss complex are the Torridonian sandstones (~1200–900 Ma), which were mainly derived from erosion of Grenvillian (~1100–

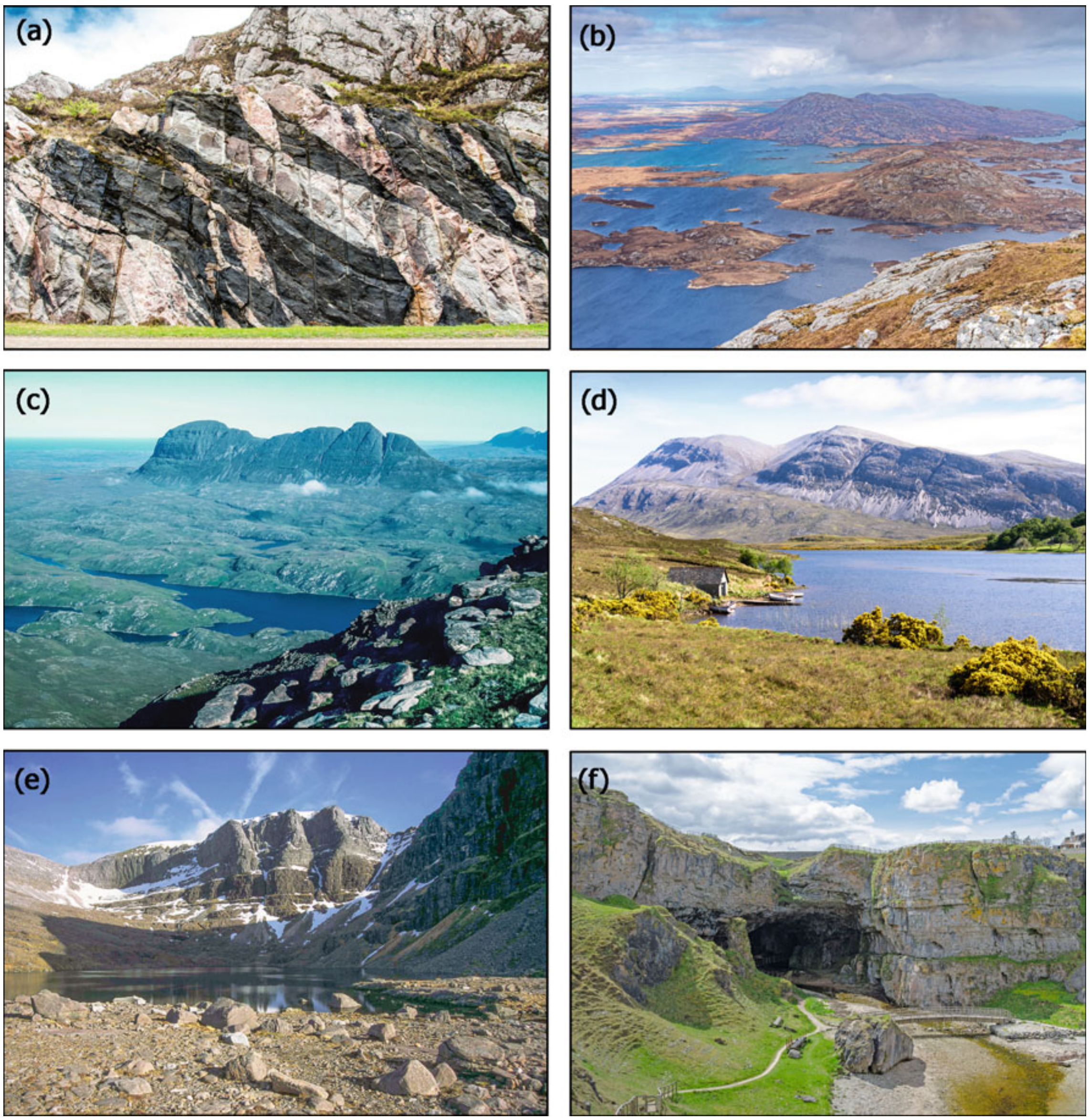

Fig. 2.7 Geology and landscape of the Hebridean terrane. **a** Lewisian gneiss exposed south of Rhiconich, NW Scotland. Dark basic dykes cut the pale grey gneiss and are themselves cut by veins of granite and pegmatite. **b** Ice-scoured landscape on Lewisian gneiss, North Uist, Outer Hebrides. **c** The Torridonian sandstone inselberg of Suilven (731 m) in Assynt rises abruptly above an ancient platform of glacially scoured Lewisian gneiss. **d** The Cambrian quartzite of Arkle (787 m), NW Sutherland, unconformably overlies an ancient landscape of Lewisian gneiss. **e** The 350 m high headwall of Coire Mhic Fhearchair, Beinn Eighe, Wester Ross, comprises Torridonian sandstone overlain by paler Cambrian quartzite. **f** Smoo Cave near Durness is a large coastal cave in Cambro-Ordovician limestone of the Durness Group, partly formed by karstic solutional processes and partly by marine erosion. (Images: **a**–**e** John Gordon; **f** Florian Fuchs, Wikimedia Commons CC BY 3.0)

1000 Ma) orogenic mountain belts on the Laurentian part of Rodinia, and have an aggregate thickness of up to 10 km. They originated in alluvial-fan, fluvial and lacustrine environments in rift valleys or in more extensive alluvial braidplains (Stewart 2002; Kinnaird et al. 2007). The Torridonian rocks rest unconformably on a pre-existing landscape fashioned out of Lewisian gneisses that formed a low-relief residual erosion surface (Fig. 2.7c). In places, however, valleys in the ancient Lewisian landscape can be seen to be infilled by Torridonian sandstone (e.g. at Slioch on the northeast shore of Loch Maree). Today, the eroded and resistant remnants of these ancient sandstones form the spectacular inselberg mountains that rise abruptly above the ice-scoured Lewisian basement between Wester Ross and Assynt (Fig. 2.7c). They, and isolated Lewisian mountains farther north, formed through uplift, retreat and dissection of a west-facing scarp, possibly of Early Cenozoic or much earlier origins.

The Torridonian succession comprises three groups, mostly composed of sandstone but with locally significant inclusions of mudstone, siltstone and breccia-conglomerate. The oldest strata form the Mesoproterozoic Stoer Group (~1200 Ma) of breccia and conglomerate, red arkosic sandstone, and siltstone, with sediment partly derived from Lewisian rocks. The basal breccias infill depressions in the underlying Lewisian gneiss, whilst sedimentary features throughout the group indicate deposition in an arid environment of seasonal braided rivers, shallow ephemeral lakes, and dune fields. The younger, Neoproterozoic part of the Torridonian succession was derived more widely from the Grenvillian mountains. It comprises the Sleat Group of coarse-grained, grey, fluvial sandstone and lacustrine mudstone, and the Torridon Group, of mainly coarse, red, arkosic sandstone originally deposited on broad alluvial plains. Like the Stoer Group, the Sleat Group rests unconformably on the Lewisian rocks, whilst the Torridon Group conformably overlies the Sleat Group but extends elsewhere to unconformably overlie the Stoer Group and the Lewisian.

During Early Cambrian times, quartz-rich sands were deposited in a shallow shelf-sea bordering Laurentia at the northwest margin of the Iapetus Ocean. Quartzite from the resulting succession now forms distinctive pale grey caps on Torridonian sandstone mountains such as Beinn Eighe and An Teallach in Wester Ross, and forms the upper slopes of more northern hills such as Arkle and Foinaven in NW Sutherland (Fig. 2.7d, e). Elsewhere, the Cambro-Ordovician succession includes the Durness Group limestones which locally provide important base-rich soils. Cave systems occur in the limestone near Durness (Fig. 2.7f) and in Assynt, where they contain ice-age faunal remains dating between ~50 and ~26 ka, Holocene faunal remains, and signs of human occupation from ~4.5 ka (Chap. 12).

The Torridonian and Cambro-Ordovician sedimentary successions are tilted but have only been substantially deformed in the Moine Thrust Zone, a major detachment that extends from Loch Eriboll, east of Durness, to Skye (Fig. 2.3). The thrust zone is a complex array of interconnected, low-angle thrust faults that carried deformed Moine rocks and their Lewisianoid foundation westward onto the Lewisian basement and its sedimentary cover (Fig. 2.8a). Most of the thrust movement took place during the Scandian orogenic event.

The generally subdued topography of the Outer Hebrides reflects the ancient erosion surface of the Lewisian gneisses, but the hills of Harris include more resistant granite sheet intrusions emplaced at ~1700 Ma, one of the last events in the long history of the Lewisian Complex. The hills along the eastern margins of the Uists (e.g. Beinn Mhòr and Hecla on South Uist) form part of a thrust zone, termed the Outer Hebrides Fault Zone, that formed initially during the Grenvillian event but was reactivated during the Scandian event. However, their upstanding dissected scarp topography is mainly due to their uplift as a tilted fault block to the west of the Minch Fault, a Mesozoic fault whose trace defines the eastern margin of the Outer Hebrides for much of its length (Fig. 2.3). The Mesozoic basins that lie to the east of the Minch Fault below The Minches and Sea of the Hebrides were formed by crustal stretching over some 200 million years, extension that culminated in the opening of the North Atlantic farther west. Triassic, Jurassic and some Cretaceous sediments were deposited in these basins and are exposed on Skye and Raasay (Fig. 2.2).

2.4.2 The Northern Highlands Terrane

The Northern Highlands terrane extends from the Moine Thrust Zone to the Great Glen Fault (Fig. 2.3). The rocks of this terrane belong mainly to the Moine Supergroup. Like the Torridonian sequence farther west, these rocks originated through erosion of a mountainous continental block of Grenvillian age and accumulated mostly in shallow-marine basins around 1000–900 Ma. The sands, silts and muds formed sequences up to 12 km thick in subsiding basins along the eastern margin of the Laurentian continent. In contrast to the Torridonian and Cambro-Ordovician rocks, which lay beyond the main effects of the Caledonian Orogeny, the Moine rocks show evidence of a long and complex history of deformation and metamorphism that has converted the original sandstones, siltstones and mudstones into deformed and metamorphosed psammites, schistose semipelites and pelites, respectively. The duration of the tectonic activity ranged from ~800 Ma to the Scandian phase of the Caledonian Orogeny at ~425 Ma. The latter involved the substantial westward thrust movements that culminated in

Fig. 2.8 Geology and landscape of the Northern Highlands terrane. **a** The Glencoul Thrust (white line), one of the classic structures of the Moine Thrust Zone. Cambrian rocks (C) lie between the Lewisian basement (LB) and the overlying Lewisian rocks (L) of the Glencoul Thrust Sheet. **b** Ice-scoured mountain ridges and valleys in Knoydart are underlain by rocks of the Moine Supergroup; view NE towards Loch Hourn. Note the general accordance of summits ~900–1000 m above sea level. **c** Gently dipping Devonian sandstones are exposed in the coastal cliffs and sea stacks at Duncansby, Caithness, with part of the undulating Caithness plain in the background. **d** Morven (706 m; right) and Maiden Pap (484 m; left) in SE Caithness form prominent hills of relatively resistant Devonian conglomerates. **e** Resistant Devonian volcanic rocks form the cliffed coast of Eshaness, western Shetland. **f** Aerial view looking northeast along the glacially excavated Great Glen, which follows the line of the Great Glen Fault Zone. (Images: **a**, **c**, **e** © Lorne Gill/NatureScot; **b** John Gordon; **d** © Andrew Tryon from Wikimedia Commons CC BY-SA 2.0—https://www.geograph.org.uk/p/4457053; **f** Permit Number CP20/008, Photograph P000742 digitized with grant-in-aid from SCRAN the Scottish Cultural Resources Access Network, Reproduced with permission of BGS © UKRI. All Rights Reserved. Sourced: https://geoscenic.bgs.ac.uk/asset-bank/action/viewAsset?id=2041&index=0&total=1&view=viewSearchItem)

the development of the Moine Thrust Zone. Although the Moine Thrust itself is the best-known feature, other thrusts can be clearly seen, for example, north of Loch Glencoul (Fig. 2.8a). The topography of this region is characterized by long, broadly east–west orientated mountain ridges and intervening glens from Knoydart north through Kintail, Affric, Strathfarrar and the Fannich Mountains (Fig. 2.8b). Farther north, a major strath (broad, shallow valley) follows the NW–SE orientated Loch Shin fault.

In the northeast part of the Northern Highlands terrane, the lowlands of Caithness and Orkney are underlain by generally flat-lying Early to Middle Devonian strata of the Old Red Sandstone Supergroup (Fig. 2.8c). These were deposited as fluvial and lacustrine sediments in the large inland Orcadian Basin at a time when Scotland lay south of the Equator and experienced hot, semi-arid conditions. Alluvial fan breccias and conglomerates built up on the irregular basal unconformity cut across Proterozoic rocks, and were then covered by aeolian sands or transgressive lacustrine sediments. Lake transgression and regression occurred repeatedly, giving rise to a thinly interbedded succession of limestone, mudstone and sandstone with annual cyclicity recorded in some parts.

Some individual mountains of more resistant rocks rise above the Orcadian Basin strata that underlie the low-lying topography of Caithness: Morven and Maiden Pap (Devonian conglomerates; Fig. 2.8d), Scaraben (Moine quartzite) and Ben Loyal (Silurian syenite). Elsewhere, Jurassic sediments of the offshore Moray Firth Basin succession extend onto parts of the east coasts of Caithness and Sutherland, and include the coal deposits, sandstones, siltstones, shales and thin limestones that are exposed along the east Sutherland coast between Golspie and Helmsdale.

In the eastern half of Shetland, the grain of the topography closely follows the mainly north–south Caledonian structural trends in the underlying Moine and Dalradian rocks and dictates the pattern of valleys and inlets (voes). West of the Walls Boundary Fault, which forms a major north–south feature in Shetland, the relief is more varied, reflecting the range of rock types: Archaean gneisses, Dalradian and Moine metasedimentary rocks, Silurian–Devonian granites and diorites, and Devonian sandstones and lavas (Fig. 2.8e). These lithologies were either involved in, or are products of, the Caledonian Orogeny. Fragments of the ocean floor and underlying crust of the Iapetus Ocean, probably of Cambrian age, are preserved on Unst and Fetlar as an ophiolite complex obducted westwards onto the edge of the Laurentian continent (Flinn 2014).

The Walls Boundary Fault of Shetland is probably the structural extension of the Great Glen Fault, which on the mainland of Scotland has given rise to one of the most distinctive lineaments in the Scottish landscape (Fig. 2.3). It has a complex history of movement, involving several phases of strike-slip displacement. The fault may follow the line of a basement structure initiated in Precambrian times but much of the displacement was achieved incrementally at ~430–400 Ma towards the end of the Caledonian Orogeny. However, lateral movements continued periodically as late as the Jurassic, linked to opening of the North Sea Basin, and during the Cenozoic (Le Breton et al. 2013). As now exposed, the Great Glen Fault is a zone of intense fracturing and shearing ~1.0–1.5 km wide; its cumulative sinistral displacement is most likely of the order of 200–300 km. This zone of weak rocks was selectively excavated by pre-glacial rivers and later by successive Quaternary ice sheets to form a major topographic feature and the basins now occupied by Lochs Ness, Oich, Lochy and Linnhe (Fig. 2.7f).

2.4.3 The Grampian Highlands Terrane

The Grampian Highlands terrane extends between the Great Glen Fault and the Highland Boundary Fault (Fig. 2.3), and comprises a range of rock types that form the Dalradian Supergroup, now mainly represented as metasedimentary and metavolcanic rocks. The sedimentary rocks were originally deposited as sands, silts and lime-rich muds in marine sedimentary basins on the southeast margin of the Laurentian continent at ~800–510 Ma. Contemporaneous igneous activity produced a variety of mafic volcanic rocks and volcanigenic sediments. The presence of glacimarine tillites in the succession on Islay and the Garvellach Islands indicates episodes of ice sheet glaciation that occurred when Scotland lay in southern polar latitudes (Table 2.1). The Dalradian sequence was strongly deformed during the Grampian orogenic event, and the structural trend imposed by orogenesis is reflected in the prominent NE–SW structural grain of the present topography (Fig. 2.9a). This alignment is particularly evident along major faults that have been exploited by erosion: Loch Laidon, Loch Ericht, Glen Truim and upper Strathspey, for example, all follow the line of the Ericht-Laidon Fault (Fig. 2.9b).

Various suites of igneous rock were intruded into the Grampian Highlands terrane during and following the two main orogenic events of the Caledonian Orogeny. Those intruded during the Grampian event include layered basic-ultrabasic intrusions and granites in Aberdeenshire. Most of the granite masses in the Grampian Highlands, such as those of the Cairngorms and Rannoch Moor, were emplaced later, at ~425–410 Ma, after the Scandian event. These voluminous intrusions reflect the subduction and partial melting of ocean crust beneath the continental margin of Laurentia (Oliver et al. 2008). Associated volcanic activity produced the Devonian volcanic complexes of Ben Nevis and Glen Coe (Chap. 17), and the extensive Lorn Plateau lavas. Ben Nevis and Glen Coe in Lochaber were originally interpreted as classic examples of caldera subsidence, created when volcanoes collapsed in on themselves as the underlying magma chambers were emptied by explosive eruptions. More recently, however, the summit area of Ben Nevis has been reinterpreted as a downfaulted remnant of a thick sequence of volcanigenic debris flows that once covered a much wider area of the SW Highlands (Muir and Vaughan 2018). The volcanic rocks of both Ben Nevis and Glen Coe now form the highest ground in these areas (Fig. 2.9c).

The Caledonian intrusions coincided with late-orogenic uplift and erosion that led to rapid unroofing of granite masses, typically in only a few million years. The Cairngorm granite was probably exposed at the surface by ~390 Ma and the overall form of the massif (Fig. 2.9d) and the

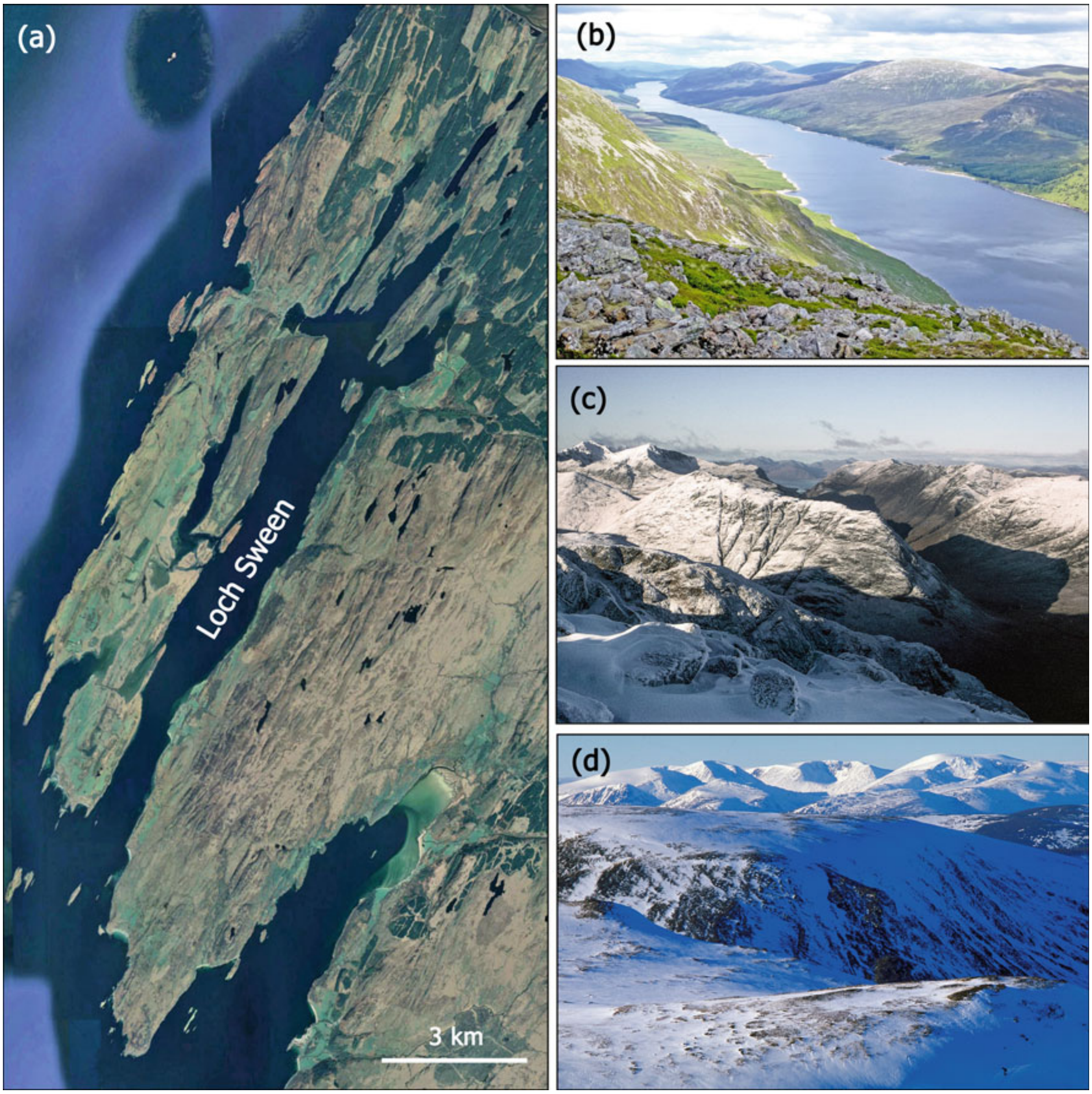

Fig. 2.9 Geology and landscape of the Grampian Highlands terrane. **a** The NE–SW alignment of the topography and coastline in the Loch Sween area, SW Argyll, illustrates the structural grain of the Dalradian metasedimentary rocks; north is at the top of the image. **b** Loch Ericht occupies a glacial trough excavated along the line of the Ericht-Laidon fault. **c** The mountains of Glen Coe are formed of resistant volcanic rocks, deeply dissected by glacial erosion. **d** The predominantly granite massif of the Cairngorms forms a high topographic area that includes several summits above 1200 m in altitude. Cirques are incised into the edges of the summit palaeosurface. (Images: **a** Google Earth™; **b** Colin Ballantyne; **c** John Gordon; **d** © Lorne Gill/NatureScot)

precursors of the present glens probably date from that time (Thomas et al. 2004). Weathering and erosion exploited lines of weakness associated with localized hydrothermal alteration of the granite, forming the major glens that were much later deepened by glacial erosion (Chap. 18). Most of the granites form areas of high ground including the Cairngorms, Lochnagar, Bennachie in Aberdeenshire and the mountains around upper Loch Etive, although deep weathering and erosion of biotite-rich granite has formed the topographic basin occupied by Rannoch Moor. Towards the east, the large-scale form of the landscape is dominated by extensive palaeosurfaces at different altitudes, broad straths and topographic basins largely formed during the Palaeogene and Neogene, although with inherited elements dating back to the Devonian (Chap. 3). Although underlain by metamorphic and igneous rocks similar to those that form the Grampian mountains, Buchan has been downwarped and forms relatively lower ground (Chap. 21).

In post-Caledonian times, the Moray Firth and North Sea areas formed subsiding basins that accumulated sediments. Devonian, Permian, Triassic and Jurassic sedimentary rocks now crop out on the basin margins along the Moray Firth coastlands both east and north of Inverness. By Permian and Triassic times, Scotland had drifted across the Equator into the northern arid belt, and dune-bedded sandstones accumulated in terrestrial basins in the Elgin area, and more extensively farther south around, for example, Mauchline and Dumfries in the Midland Valley and Southern Uplands terranes, respectively (see below).

The southern boundary of the Grampian Highlands terrane is defined by the Highland Boundary Fault, which forms a distinctive geological as well as topographic boundary, separating the Highlands from the Midland Valley (Fig. 2.4). The line of the fault is marked by a prominent scarp that stretches from Stonehaven to Helensburgh, whence the line of the fault can be traced to Arran. Within the fault zone are about ten structurally isolated slivers containing, variously, ultramafic rock, amphibolite, mafic lavas, and associated sediments. These broadly ophiolitic assemblages, known as the Highland Border Complex, range in age from Neoproterozoic to Ordovician. They are relics of the ocean crust that was destroyed as the Midland Valley Arc, founded on a fragment of continental crust, collided with the margin of Laurentia, initiating the Grampian phase of the Caledonian Orogeny.

2.4.4 The Midland Valley Terrane

The Midland Valley is a rift valley between the Highland Boundary Fault and the Southern Upland Fault (Fig. 2.3). A striking feature of the landscape is its varied relief, comprising hills, upland plateaux and areas of low ground, and the general correspondence of higher ground with igneous rocks and lower ground with sedimentary rocks (Fig. 2.10). For example, lavas, sills and volcanic plugs generally form conspicuous hills, such as the Ochil, Gargunnock and Lomond hills, that overlook the weaker surrounding sedimentary rocks and add greatly to the diversity of the landscape character (Fig. 2.11a).

The Midland Valley terrane is thought to have a deep basement formed from volcanic-arc rocks and their underlying foundation of continental crust. This assemblage is not seen at outcrop, where Upper Palaeozoic rocks predominate with inliers of Ordovician and Silurian rocks adjacent to the terrane's southern boundary. The inlier at Ballantrae comprises ophiolitic ultramafic rocks, gabbros and mafic lavas, ranging in age from Late Cambrian to Early Ordovician (the Ballantrae Complex) that were emplaced at the margin of the Midland Valley by ~465 Ma (Stone 2014a; Stone and Rushton 2018). This obduction of oceanic crust onto the margin of a continental block was part of the initiation of the closure of the Iapetus Ocean, which proceeded thereafter by subduction of the oceanic crust beneath the Midland Valley block, which by that time was effectively part of the Laurentian continental margin. Continuing subduction resulted in the accumulation of the forearc accretionary complex of the Southern Uplands terrane (see below), which by the end of the Silurian had developed into an uplifted massif bordering the southern margin of the Midland Valley.

The Ballantrae Complex is overlain by an Ordovician to Silurian sedimentary succession of conglomerate, sandstone and limestone, whilst several other inliers along the southern margin of the terrane contain Silurian successions of deep-water, turbidite sandstone passing up into shallow marine and ultimately lacustrine sandstone and mudstone. Apart from these local exceptions, the rock at outcrop in the Midland Valley is predominantly Devonian and Carboniferous in age and of sedimentary and volcanic origin, laid down in a series of large, fault-bounded basins. Crustal tension associated with distant plate movements not only resulted in subsidence as sediment accumulated during these periods, but also provided zones of weakness that allowed basic magmas to rise intermittently towards the surface, creating volcanoes, fissure eruptions and intrusions of igneous rocks.

By the middle of the Silurian, the Iapetus Ocean had effectively closed, but supra-subduction zone volcanism continued to affect the Midland Valley, with the eruption of volcanic arc lavas continuing into the Devonian. The eruptive centres were surrounded by a shallow sea and the lavas interdigitate with marine sediments. In the south of the Midland Valley this relationship is represented by the lavas and Devonian strata of the Pentland Hills inlier. Farther north, Devonian volcanic activity formed the basaltic to andesitic lavas and volcanic breccias of the Ochil and Sidlaw hills (Fig. 2.10), whilst in SE Scotland extensive lavas were erupted in the Cheviot area.

For at least the early part of the Devonian, the central part of the Midland Valley terrane was probably still an exposed massif. At that time Scotland lay in a semi-arid zone south of the Equator and thick accumulations of terrestrial sediments were deposited in alluvial fans and by braided rivers: to the north in basins adjacent to the rapidly eroding Caledonian Mountains and to the south adjacent to the newly uplifted Southern Uplands. These sedimentary basins were partly initiated by sinistral strike-slip tectonics and in them accumulated alluvial sediments that form the sandstones, siltstones and conglomerates of the Old Red Sandstone Supergroup. These rocks underlie Strathmore and the lower ground between Dundee and Stonehaven in the northern part of the terrane, and crop out along parts of the Ayrshire coast, in the Lanark basin, and across parts of the eastern Southern Uplands (Figs. 2.2 and 2.11b).

By Carboniferous times, Scotland had drifted north to lie in equatorial latitudes (Table 2.1), and the Midland Valley formed part of an extensive coastal plain covered by

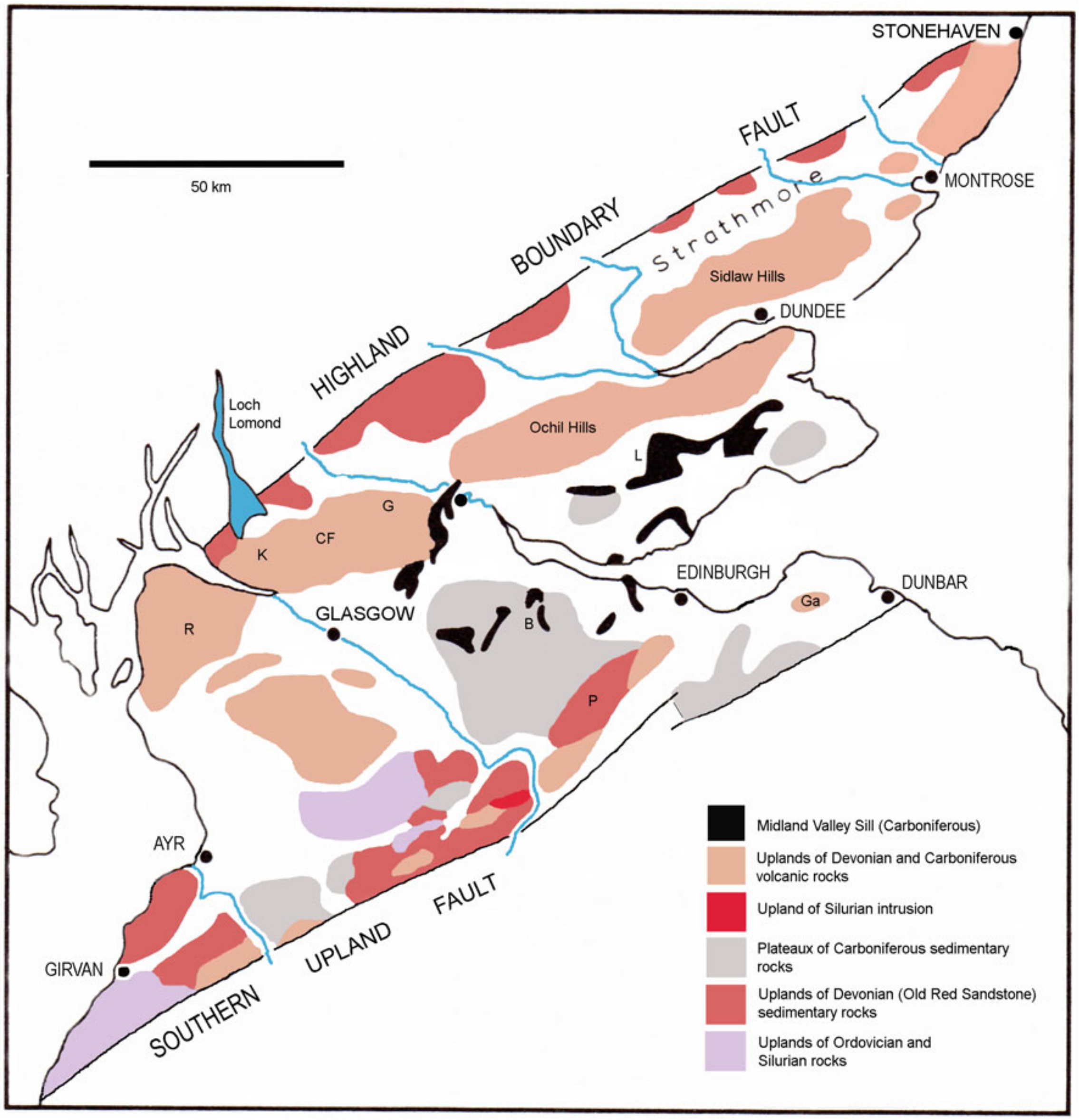

Fig. 2.10 Geology and relief in the Midland Valley. B: Bathgate Hills; CF: Campsie Fells; G: Gargunnock Hills; Ga: Garleton Hills; K: Kilpatrick Hills; L: Lomond Hills; P: Pentland Hills; R: Renfrewshire Hills. (Adapted from Sissons 1976)

primitive rainforest and swamps. Changes in relative sea level, in part driven by the waxing and waning of ice sheets across the southern polar portion of Gondwana, saw the Midland Valley repeatedly inundated by shallow tropical seas. As sea level fluctuated, cyclic accumulations of sediments were laid down in warm, shallow seas, coastal lagoonal swamps and river deltas, subsequently forming sequences of limestone, mudstone, sandstone and coal measures. The pattern of sequential changes in sea level, in concert with basin subsidence, was repeated many times. The swamp vegetation was buried and compressed beneath overlying sediments and progressively transformed into the seams of coal that have been exploited in the coalfields of Ayrshire, Central Scotland, Midlothian and Fife. The sedimentary rocks were later tilted and folded as exemplified along the coast of Fife and near Dunbar in East Lothian (Fig. 2.11c). This deformation occurred in several phases around 300 Ma and represents the far-field effects of the Variscan Orogeny. At this time Scotland lay in the centre of the huge landmass of Laurussia (Laurentia, Northern Europe, Russia and Siberia) and by Permo-Triassic times, aeolian desert sands were deposited, and are now preserved in parts of Ayrshire and on Arran (Fig. 2.11d).

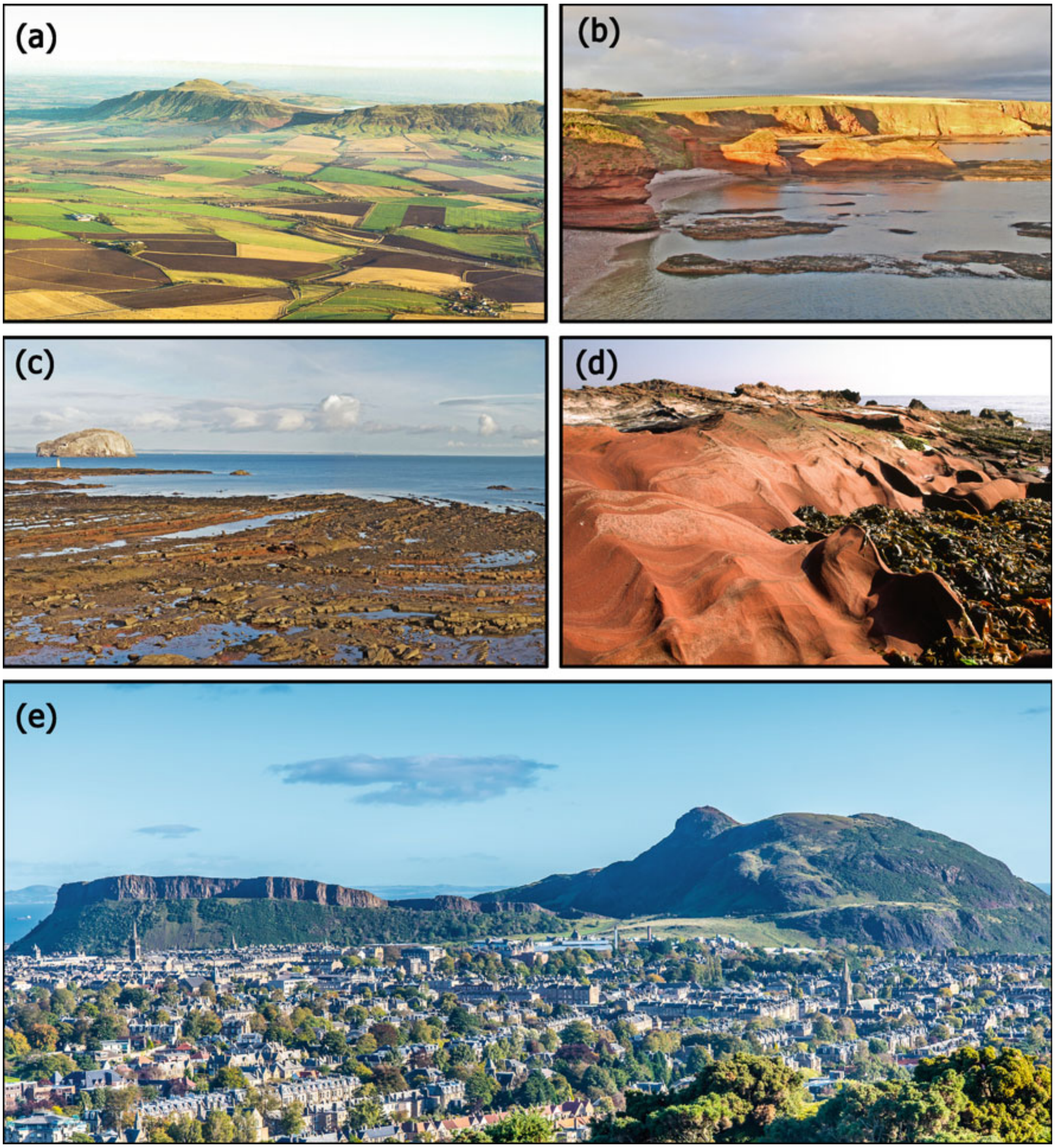

Fig. 2.11 Geology and landscape of the Midland Valley terrane. **a** The Lomond Hills in Fife form a prominent escarpment defined by part of the Midland Valley Sill, intruded into sandstone and limestone during the Carboniferous. The cone-shaped summits of the hills are volcanic plugs. **b** Devonian sandstones and conglomerates form coastal cliffs near Arbroath. **c** Shore platform cut across folded Carboniferous sedimentary rocks, Seacliff, East Lothian. The Bass Rock in the background is a Carboniferous phonolite volcanic plug. **d** Permian desert sandstone, Arran. **e** The remnants of a Carboniferous volcano (Arthur's Seat, right) and sill (Salisbury Crags, left) rise above the surrounding less-resistant sedimentary rocks and provide a spectacular backdrop to the city of Edinburgh. (Images: **a** Permit Number CP20/008, Photograph P002927 digitized with grant-in-aid from SCRAN the Scottish Cultural Resources Access Network, Reproduced with permission of BGS © UKRI. Sourced: https://geoscenic.bgs.ac.uk/asset-bank/action/viewAsset?id=4218&index=0&total=1&view=viewSearchItem; **b**, **c**, **e** John Gordon; **d** © Lorne Gill/NatureScot)

Throughout Carboniferous times, intrusive igneous rocks were emplaced and volcanic lavas and tuffs were erupted extensively in the Midland Valley, in response to the onset of broadly extensional tectonics, accompanied by a progressively increasing component of dextral strike-slip. These mostly mafic igneous rocks now form many of the hill masses and individual hills, or 'laws', which characterize the region. Lavas and subsidiary tuffs form the Garleton Hills in East Lothian, the Bathgate Hills in West Lothian, the Renfrewshire Hills, the Kilpatrick Hills, the Campsie Fells and the Gargunnock Hills (Fig. 2.10). The volcanic vents now represented by such imposing features as Arthur's Seat in Edinburgh (Fig. 2.11e) and North Berwick Law in East Lothian also formed at this time, as did other intrusions, such as Salisbury Crags in Edinburgh, Traprain Law and the Bass Rock in East Lothian (Fig. 2.11c), the Midland Valley Sill and numerous vents and plugs in Fife. Over time, the igneous rocks have been tilted by earth movements and modified by erosion, which has removed the volcanic cones, leaving the more resistant feeder necks as upstanding hills, and lowered the adjacent weaker sedimentary rocks to a greater extent than the stronger lavas and sills. As a consequence, conspicuous scarp slopes of the more resistant volcanic rocks often rise abruptly above the less resistant surrounding sedimentary rocks. A particularly good example is the quartz-dolerite Midland Valley Sill that underlies much of the Lomond Hills (Fig. 2.11a) and Forth basin, and forms a resistant cap over weaker sedimentary rocks. At Stirling, eroded remnants of this gently dipping dolerite sill provide spectacular platforms for Stirling Castle and the Wallace Monument.

The southern margin of the Midland Valley terrane is marked by a prominent break of slope coincident with the faults that juxtapose it with the high ground of the Southern Uplands. Running southwest from near Dunbar, on the North Sea coast, the Lammermuir Fault forms the steep northern edge of the Lammermuir Hills and the Moorfoot Hills. Farther southwest, the topographical break steps north to the Southern Upland Fault (sensu stricto) and thence continues southwest to Ballantrae on the Firth of Clyde, where it forms the southern boundary of the eponymous ophiolite complex. There has probably been significant sinistral strike-slip movement on the Southern Upland Fault in addition to the evident downthrow to the north. Elsewhere, fault scarps are prominent landscape features, notably along the southern edges of the Ochil Hills and the Campsie Fells.

2.4.5 The Southern Uplands Terrane

The Southern Uplands terrane extends from the Southern Upland Fault to the Iapetus Suture Zone, which lies several kilometres south of the boundary between Scotland and England (Fig. 2.3). Most of the uplands in this region comprise dissected undulating plateaux underlain by Ordovician and Silurian greywacke sandstones, mudstones and shales that were originally deposited as turbiditic sands, silts and muds on the floor of the Iapetus Ocean during the Late Ordovician and Early Silurian (Fig. 2.12a). Throughout this interval, the Iapetus Ocean was being subducted northwards beneath the Midland Valley terrane, by then effectively the margin of Laurentia. As the oceanic crust was subducted, its cover of sediment was sheared off and incorporated into a series of thrust sheets, younger thrust slices being successively added beneath the previously incorporated slices to build up an imbricate accretionary complex (Stone 2014b and references therein). Between twenty and thirty individual thrust slices have been identified, each representing sedimentation over no more than a million years. As the complex developed, the thrust slices were rotated towards the vertical so that strata dip steeply across much of the region (Fig. 2.12b). This has created the 'Southern Uplands paradox', whereby the strata in each individual thrust slice become younger towards the north, but overall the youngest, Silurian beds are in the south and the oldest, Ordovician beds are in the north. Both the strike of individual beds and the major faults marking the boundaries of the imbricate thrust slices are generally aligned NE–SW.

When incorporated into the accretionary complex, the component strata were deformed under low-grade metamorphic conditions (up to 300 °C and 400 MPa) with, at some periods, an element of sinistral strike-slip added to the subduction process. The sequential nature of the principal deformation means that it was diachronous across the terrane, older in the north than in the south, and very little deformation can be unequivocally ascribed to the final closure of the Iapetus Ocean. This is usually described as a 'soft collision' which saw the leading edge of the Southern Uplands accretionary complex override the margin of Avalonia, the original southern margin of the ocean. Nevertheless, significant regional uplift seems likely to have occurred.

The strong NE–SW structural grain developed within the accretionary complex is now a controlling influence on the orientation of many of the region's valleys. Fault zones, often associated with mudstone units when strike-aligned, afford opportunities for differential erosion within the otherwise relatively uniform successions of greywacke sandstone. Major faults guide valleys such as Moffat Dale (Chap. 27) and Glen App, south of Ballantrae, and the same trend is apparent in the surface topography along the margins of the Tweed valley (e.g. north of Hawick); these features have been generally emphasized by glacial erosion. Away from the major faults, the relatively uniform nature of the dominant greywacke sandstones is reflected in the rolling topographic form of the hills (Fig. 2.12a), but this is interrupted by granite massifs in the southwest of the terrane. After closure of the Iapetus Ocean, igneous activity accompanied the final stages of the Caledonian Orogeny and several granite plutons, including those underlying Criffel, Carsphairn and Cairnsmore of Fleet, were intruded during the Early Devonian at ~410–390 Ma. These now, together with the surrounding, thermally metamorphosed sandstones, form prominent hills (Fig. 2.12c). By contrast, the Loch Doon granite now

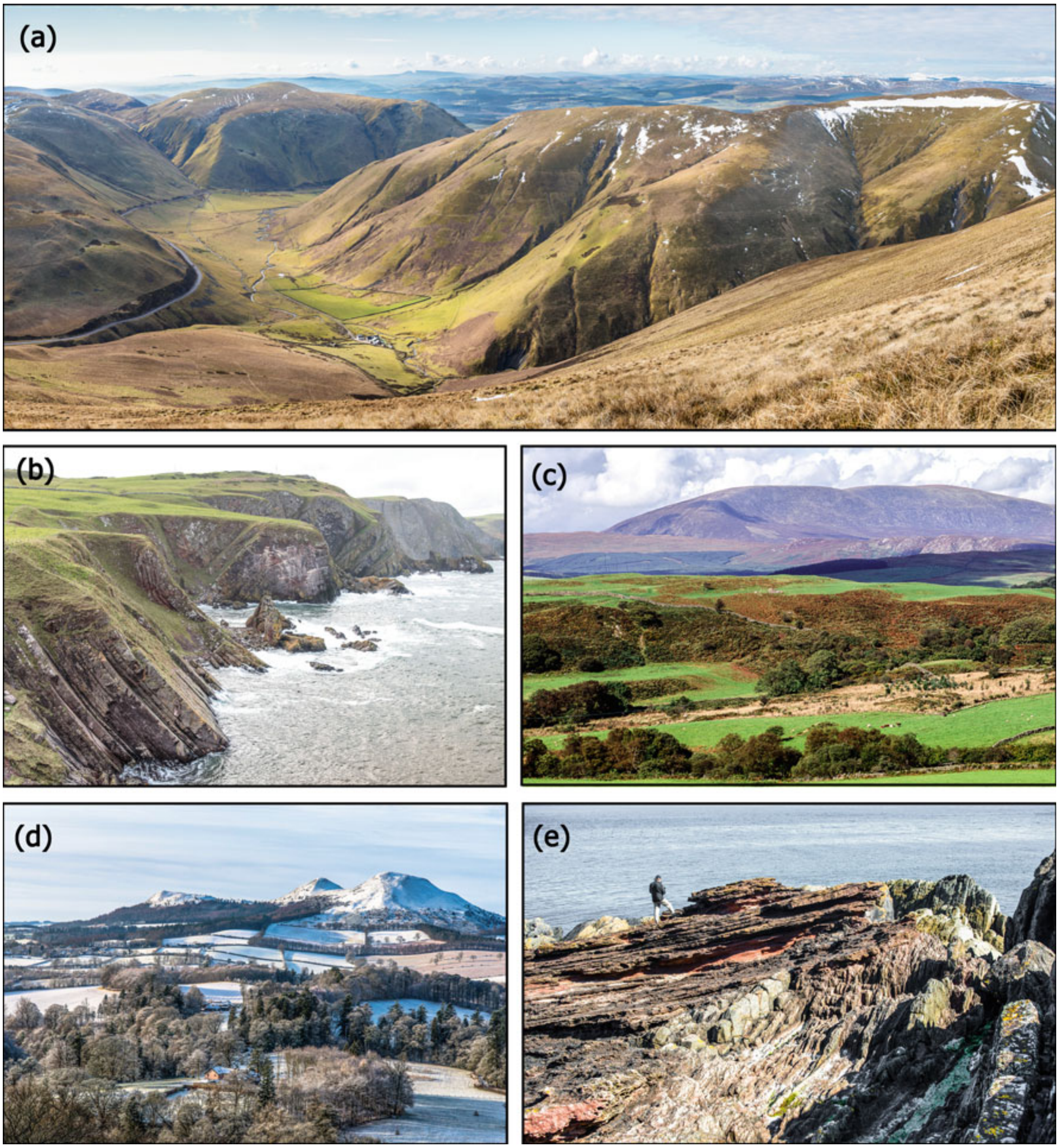

Fig. 2.12 Geology and landscape of the Southern Uplands terrane. **a** Dalveen Pass, a glacial trough carved through the rolling hills of the Southern Uplands, follows a north–south aligned fault which cuts across the NE–SW trend of the Southern Uplands strike faults. **b** Steeply dipping and folded Silurian greywacke sandstones and siltstones form coastal cliffs near St Abb's Head. **c** The granite pluton of Cairnsmore of Fleet (711 m). **d** The Eildon Hills are the eroded remnants of a laccolith intruded into Silurian greywackes. **e** Hutton's classic section at Siccar Point, Berwickshire, showing the unconformity between steeply dipping Silurian greywacke sandstones and overlying Devonian sandstones. (Images: **a**, **b**, **d**, **e**: John Gordon; **c** © Lorne Gill/NatureScot)

forms a basin as a product of deep weathering and erosion during the Cenozoic, surrounded by a metamorphic aureole of resistant, thermally altered greywacke sandstones that culminate in Merrick to the west and Corserine to the east (Chap. 27). Later intrusions that now have a topographic expression include the Eildon Hills, near Melrose, the remnants of a trachyte laccolith intruded into Upper Old Red Sandstone rocks during the Carboniferous (Fig. 2.12d).

The Annan, Nith and Ken valleys, orientated NNW–SSE, are superimposed across the strike of the Ordovician and

Silurian rocks of the Southern Uplands, following the trend of later faulting that controlled the development of Devonian and Permian extensional basins. The sedimentary rocks deposited in these basins are amongst the younger strata that now unconformably overlie the Lower Palaeozoic accretionary complex (Fig. 2.12e). Carboniferous limestones deposited in the east of the area reflect the northern limits of the Northumberland Trough. Permo-Triassic desert sandstones were probably once widespread, certainly over the western parts of the Southern Uplands (and the adjacent Midland Valley) but are now preserved only in the several fault-controlled basins, notably around Dumfries, Thornhill (in Nithsdale), Lochmaben (in Annandale) and Stranraer. The Rivers Annan and Nith occupy valleys excavated along a line of basins infilled with Permian sediments, while the low ground of The Merse (lower Tweed valley) is underlain by Carboniferous sedimentary rocks.

The Iapetus Suture Zone that marks the geological southern boundary of the Southern Uplands terrane represents the tectonic base of the accretionary prism, now resting on Avalonian continental crust. The suture now lies buried beneath younger rocks in northern England but has been imaged geophysically at depth. The topographic southern boundary of the Southern Uplands is marked by the outcrop of unconformable younger rocks, mostly of Carboniferous age, that span the Anglo-Scottish border and form much of the northern coastline of the inner Solway Firth.

2.4.6 The Hebridean Igneous Province

A major Palaeogene magmatic event strongly influenced the geology of Scotland's western seaboard. Crustal fracturing and geothermal uplift were associated with the development of the Iceland mantle plume and the opening of the North Atlantic Ocean. The proximity of western Scotland to a major magmatic province saw extensive volcanic activity in the Inner Hebrides and parts of the western seaboard during the Early Palaeogene (~61–55 Ma), with a peak at ~60–58 Ma. Fissure eruptions, similar to those in Iceland today, produced large thicknesses (4 km) of basalt lavas in three major lava fields: on Skye and Canna, on Eigg and Muck, and on Mull (Fig. 2.3). These now form the distinctive trap (stepped) landscapes of these islands (Fig. 2.13a). Where the lavas and related sub-volcanic sills are underlain by weaker Mesozoic sedimentary rocks, as in northern Skye, extensive landslides have occurred (Chap. 10).

A number of intrusive igneous complexes represent the roots of volcanic centres that developed at different times during the period of Palaeogene magmatism. These extend from Arran and Ailsa Craig in the south to Skye in the north (Fig. 2.3) and St Kilda offshore to the north-west (Fig. 2.13b); other volcanic centres lie concealed offshore. Large stratovolcanoes probably formed above some of the centres, but these superstructures have been completely removed by erosion, and the former presence of the volcanoes is now represented by their cores of plutonic rock, the unroofed granites, gabbros and peridotites that form the mountains of Arran, Rùm Skye and Mull (Chap. 10). On Skye, the layered gabbros, eucrites and peridotites, steeply dipping intrusive basaltic cone sheets and dykes of the Cuillin Hills have weathered and been eroded into jagged mountain ridges, contrasting with the more rounded forms of the granitic Red Hills (Fig. 2.13c). Similar contrasts are evident on Rùm between the layered mafic and ultramafic rocks of the Rùm Cuillin and the felsic rocks of the Western Hills. On the Ardnamurchan peninsula, three successive igneous centres involving the intrusion of ring dykes and cone sheets form a magnificent topographic feature when viewed from the air (Fig. 2.13d). Palaeogene magmatic activity also resulted in the intrusion of numerous dyke swarms, mainly sourced from the Skye and Mull centres, that trend southeast across Argyll and the western part of the Midland Valley (Fig. 2.3). Individual dykes, commonly several metres across, are often more resistant to erosion than the surrounding country rocks and so form conspicuous features in the landscape and are especially well-developed in southern Arran (Fig. 2.13e; Chap. 10).

2.5 Conclusion

The exceptional diversity of landscapes and landforms in Scotland today reflects the interactions between the geological foundations, plate tectonics, long-term changes in global climate and the varied processes of weathering and erosion over many millions of years. The broad outlines of the landscape, as well as many of its finer details, are the product of geological and tectonic controls. In particular, the Caledonian Orogeny and its aftermath of igneous activity and erosional denudation established the geological framework and landscape template for much of the mainland; later Palaeozoic and Palaeogene volcanism strongly influenced the landscapes of the Midland Valley and Inner Hebrides, respectively; and Cenozoic uplift, crustal warping, erosion and etchplanation under warmer, more humid climates than today established many of the main features of the topography. Over millions of years, differential weathering and erosion have exploited lines of weakness and variations in rock resistance and have emphasiszd the inherent structural

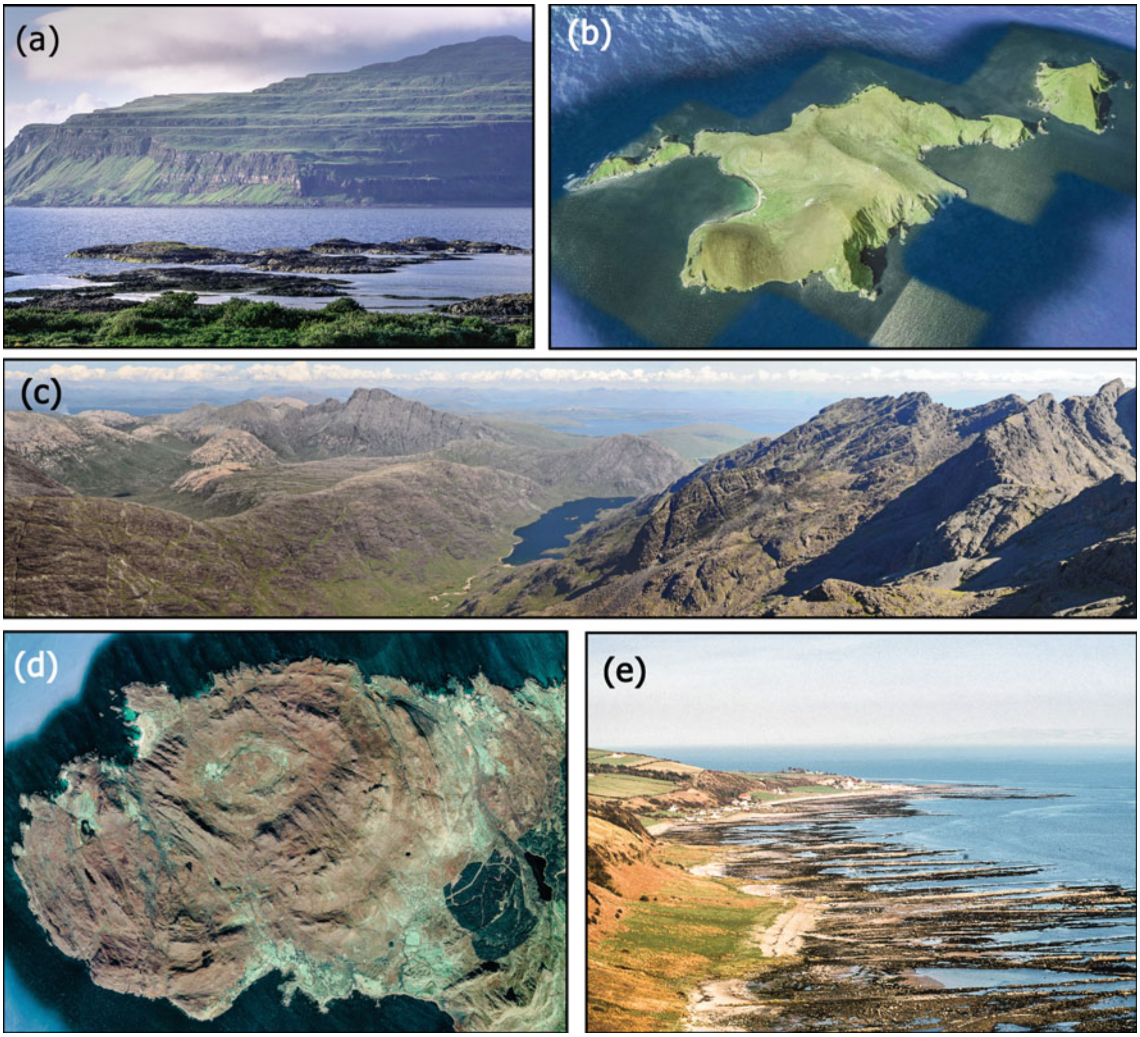

Fig. 2.13 Geology and landscape of the Hebridean Igneous Province. **a** Trap landscape on Mull formed of stacked Palaeogene lavas. **b** The islands of St Kilda comprise the partly submerged remnants of a Palaeogene volcanic centre. **c** The gabbros of the Cuillin central igneous complex on Skye, cut by numerous basaltic dykes and cone sheets that have selectively weathered and eroded, form rugged mountain ridges (right) that contrast with the more rounded Red Hills granite complex (left). **d** The successive intrusions and ring dykes of the Ardnamurchan Palaeogene volcanic complex are clearly reflected in the form of the present topography. **e** Dyke swarm cutting Triassic sandstones, Kildonan, Arran. Note the cross-cutting of individual dykes that form the ribs running out to sea. (Images: **a** © Lorne Gill/NatureScot; **b**, **d** Google Earth™; **c** John Gordon; **e** Permit Number CP20/008 BGS © UKRI. All Rights Reserved. Source: https://earthwise.bgs.ac.uk/index.php?title=File:P580470.jpg&filetimestamp=20150722203927&)

grain that is now apparent in the topography in many areas. Many older elements persist in the present landscape and are also expressed in the links between rocks and relief highlighted by differential weathering and erosion during the Palaeogene and Neogene, and subsequently by glacial erosion and interglacial processes during the Quaternary. The present Scottish landscape is essentially a palimpsest of pre-existing landscapes strongly influenced by their geological evolution and modified to varying degrees by Pleistocene and Holocene geomorphological processes.

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3 Long-Term Denudation and Geomorphology in Scotland

Adrian M. Hall

Abstract

Long-term geomorphology has received little recent attention in Scotland. Palaeosurfaces and major landforms, including valleys and basins, can be linked to the sub-Caledonian and sub-Permian basement unconformities. The present topography was sculpted from the sub-Palaeocene unconformity after kilometre-scale uplift in response to Early Palaeogene magmatism. Recent results from thermochronology indicate an eastward decrease in denudation across Highland Scotland that is supported by landscape persistence in eastern areas since the Devonian. A sequence of planation surfaces developed from the Late Eocene onwards in intervals of slow uplift and high sea levels. The highest extensive surface, the Eastern Grampian Surface, has been uplifted to an elevation of 500–840 m. The surfaces formed by etch processes operating in response to changing base levels under warm-to-cool, humid environments, with extensive forests and widespread deep weathering. After uplift, each planation surface was modified through valley incision, backwearing of scarps and downwearing. Pliocene sea-level fall likely led to formation of coastal platforms that included precursors of the present Hebridean strandflat.

Keywords

Tectonics • Unconformity • Weathering • Denudation • Planation surfaces • Inselbergs • Topographic basins • Drainage patterns • Long-term relief development

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3.1 Introduction

The focus of this chapter is on the shaping of Scotland's scenery before the Pleistocene. The macro-scale topography has a long evolution, its inherited features remain prominent in today's scenery and its gross form has guided the build-up of glaciers and the pattern and intensity of glacial erosion through the Pleistocene. The origins of the present relief can be traced in erosional unconformities that emerge from beneath sedimentary and volcanic cover of Precambrian to Palaeocene age. The present topography was shaped during and after Palaeogene uplift in response to tectonic, climatic and sea-level forcing. The major landforms developed mainly in the Neogene through the interplay of base-level changes, differential weathering and fluvial erosion.

3.2 Deep Time: Unconformities, Erosion and Inherited Relief

Scotland has an unusually diverse geology for a small part of Earth's surface. Plate fragments, or terranes, with rocks that extend back in time to the Archaean, were assembled into their present configuration during the Palaeozoic (Chap. 2). The terranes are separated by major vertical structures in the crust, such as the Great Glen Fault (Fig. 3.1a), that have guided lateral and vertical displacement through the Phanerozoic. The geological terranes approximate to, but do not match, the main morpho-structural blocks that have conditioned denudation and relief development since the Cretaceous: the Orkney-Shetland Platform (OSP) , the Outer Hebrides Platform (OHP), the Northern Highlands, the Grampian Highlands, the Midland Valley and the Southern Uplands. The Palaeogene Hebridean Igneous Province (HIP), scattered across western Scotland, has its own distinctive topographic expression. The morpho-structural blocks stand above the upper Palaeozoic to Mesozoic rift basins of the Inner Hebrides and Moray Firth. Sediments in

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World Geomorphological Landscapes, https://doi.org/10.1007/978-3-030-71246-4_3

these basins, in the North Sea and on the North Atlantic shelf, were largely sourced from land areas centred on present-day Scotland during and after the Devonian (Trewin 2002).

The morpho-structural blocks carry extensive unconformities that mark the terminations of long periods of erosion and the onset of burial of basement by terrestrial and marine sediment. Collectively, the unconformities record the depth and pattern of erosion and burial over time (Fig. 3.1a). The oldest stack of unconformities is on the Lewisian foreland of the NW Highlands, with unconformities that lie beneath the Stoer Group ($\sim$1.18 Ga) and Torridon Group ($\sim$1.04 Ga) conglomerates and sandstones, Cambrian ($\sim$540 Ma) quartzites, Permo-Triassic ($\sim$265 Ma) sandstones and Palaeocene ($\sim$63 Ma) basalt. The presence of five tiered unconformities in less than 1 km height range indicates that denudation of the Lewisian basement since $\sim$ 1.0 Ga has been limited. Despite kilometre-scale burial and loading by Precambrian sedimentary cover, the Moine Thrust nappes and younger sedimentary rocks, the basement has returned to approximately the same level after each period of erosion and unloading.

Across Scotland north of the Highland Boundary Fault, the Caledonian Orogeny is the starting point for relief development (Chap. 2). Profound erosion in the Moine and Dalradian gneisses and schists and rapid exhumation of plutonic intrusions in the Caledonian mountain belt culminated in the formation of the post-orogenic, sub-Caledonian (Devonian) unconformity. This unconformity is exposed extensively in eastern parts of the Northern and Grampian Highlands, with only a small core area lacking Devonian outliers (Fig. 3.1a); since $\sim$380 Ma, basement erosion has been limited (Hall 1991; Macdonald et al. 2007). Permo-Triassic sediments were deposited in rift basins

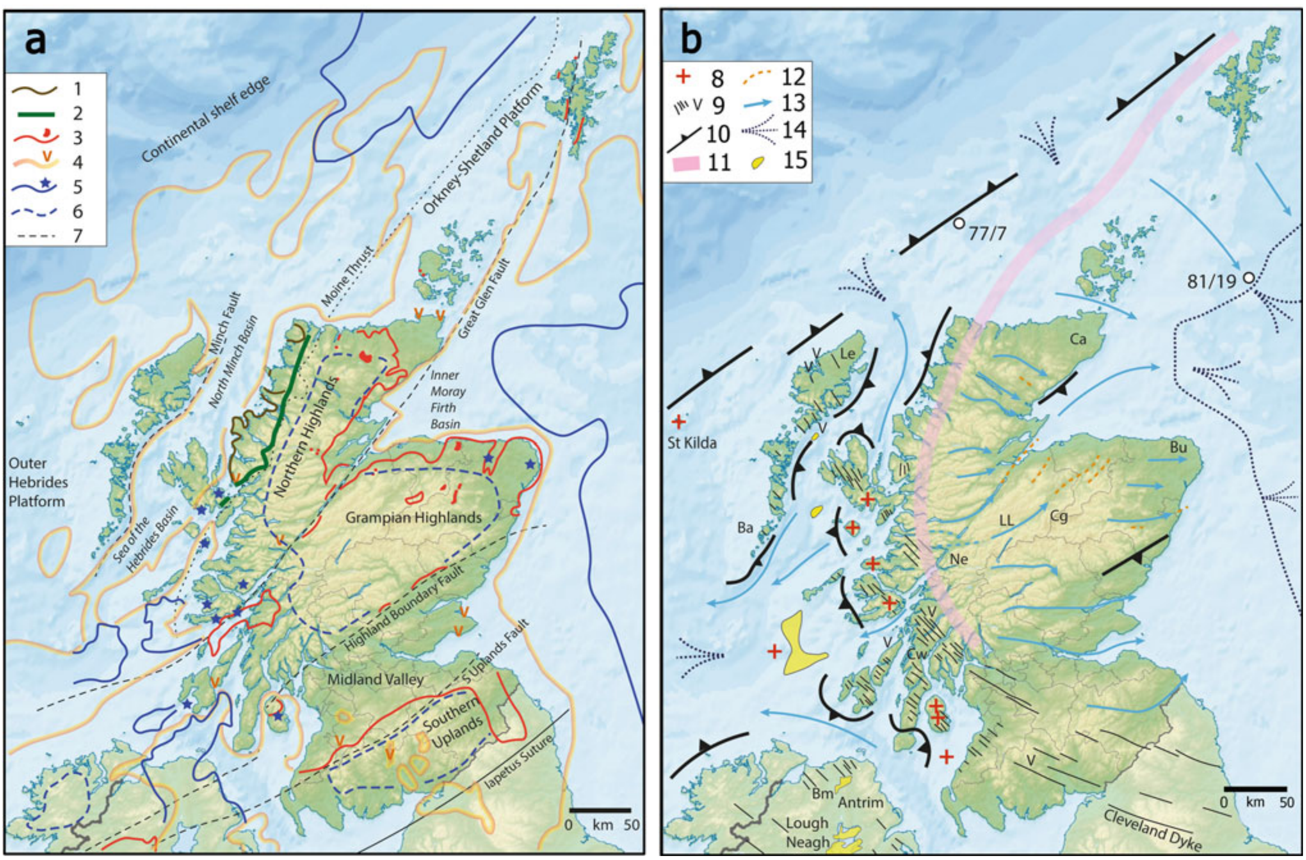

Fig. 3.1 Morpho-structural setting of Scotland. **a** Present limits of unconformities. 1: Sub-Torridonian unconformity. 2: Sub-Cambrian unconformity. 3: Sub-Caledonian (Devonian) unconformity and outliers. 4: Sub-Permian unconformity; V: Permian vent. 5: Sub-Palaeocene unconformity. 6: Approximate maximum limit of the Late Cretaceous transgression; Late Cretaceous outliers marked by stars. 7: Major fault. **b** Palaeogene uplift and drainage. Dotted line shows the position of the Early Eocene shoreline in the western North Sea, after Knox (2002). 8: Major intrusive igneous centre. 9: Palaeocene dykes; V: vesicular dykes. 10: Major scarp. 11: Main watershed. 12: Exhumed Devonian valley. 13: Major drainage routeway. 14: Sediment routeway. 15: Late Oligocene basin. Also shown are BGS boreholes (77/7 and 81/19) referred to in the text. Ba: Barra; Bm: Ballymoney; Bu: Buchan; Cg: Cairngorms; Le: Lewis; LL: Loch Laggan; Ne: Ben Nevis. (Base digital elevation map on a Creative Commons licence at CC-BY-SA-3.0, GFDL)

developed during crustal extension that heralded the opening of the proto-Atlantic. The edge of the Permian cover conforms largely to the present outline of the Scottish land area (Glennie 2002). Early Permian volcanic vents in Lochaber and Ayrshire (Fig. 3.1a) indicate that surrounding land surfaces remain close to Early Mesozoic erosion levels (Hall 1991). The offlap from the Devonian unconformity suggests that Devonian and Carboniferous sedimentary cover was thinned, locally removed and recycled by erosion in the Late Palaeozoic. Similarly, the farther offlap of the Late Cretaceous Chalk edge indicates removal of Permo-Triassic cover during and since the Mesozoic. The youngest unconformity is represented by the present land surface.

The sub-Torridon Group unconformity has low relief near Cape Wrath, whereas farther south the ancient land surface becomes hilly and, locally, mountainous (Stewart 1972). In contrast, the sub-Cambrian unconformity surface is remarkably planar for ~150 km on the Lewisian foreland; it represents a fragment of the former Great Unconformity on Laurentia. The sub-Caledonian unconformity is generally hilly across the Northern and Grampian Highlands, but lower-relief pediment surfaces exist in Caithness (Chap. 8). The sub-Permian unconformity was strongly influenced by fault-block tectonics. For example, on the eastern flank of the Moffat basin in the Southern Uplands, the unconformity is buried beneath breccias originally deposited as scree at the base of steep fault scarps (Brookfield 1980). Hence, the buried unconformity surfaces differ widely in morphology.

Where unconformities are of high relief, inheritance of exhumed topography in the present landscape is generally confined to valley segments at the edges of outliers. Persistence of landforms is more widespread where major features have maintained their gross form and position during later denudation. Examples include sub-Torridonian gneiss hills and valleys around Loch Maree (Stewart 1972), sub-Devonian valleys in fracture zones around the Cairngorms (Hall and Gillespie 2016) and the sub-Cretaceous quartzite inselberg of Mormond Hill in Buchan (Chap. 21). The persistence of exhumed hills, valleys and basins in the Cenozoic topography is a consequence of the reactivation of the same rock controls on differential weathering and erosion that originally brought these features into existence.

3.3 Morphogenetic Setting for Palaeogene and Neogene Relief Development

Deposition of chalk continued in western Scotland, Northern Ireland and the North Sea Basin until ~70 Ma (Hopson 2005). During the Late Cretaceous sea-level highstand, the Scottish land area was small and of low relief, contributing little clastic sediment to flanking basins (Fig. 3.1a). Similar conditions existed in SW Norway, where contemporary relief was <500 m (Sømme et al. 2013). The low relief of the sub-Palaeocene unconformity constrains the patterns and magnitude of uplift and denudation later in the Cenozoic. The timings of the main events that have shaped the scenery of Scotland through this time are summarised in Fig. 3.2.

3.3.1 Magmatism and Tectonics

Cenozoic tectonics involved a main phase of uplift in the Palaeogene driven initially by magmatism in the HIP and later by the passage of the Icelandic plume, with magmatic underplating of topography (Mudge 2014). Early magmatism in the HIP is recorded by the eruption of basalts at 61–58 Ma. Later volcanism was associated with the emplacement of major igneous centres in western Britain at 60–55 Ma (Bell and Williamson 2002; Chap. 10). Minor magmatic activity continued on the OHP until ~45 Ma (Faithfull et al. 2012). In the HIP, the sub-Palaeocene surface was broken and warped by fault movements and uplift, with kilometre-scale displacement around igneous centres (Hall 1991). Combined uplift and lava accumulation raised Palaeocene basalt plateaux to elevations of 2 km or more on Skye (Jolley 1997) and Mull (Bell and Williamson 2002).

Regional uplift produced extensive emergence of the Scottish land area. Erosion stripped Permo-Triassic cover from the OSP (Morton et al. 2004) and large volumes of sand accumulated in the Faroe-Shetland and North Sea Basins. Eastward tilting in the inner Moray Firth Basin may have led to removal by erosion of ~1.3 km of mainly Mesozoic section (Mackay et al. 2005), although the sub-Caledonian unconformity around the Moray Firth does not show tilting of this magnitude (Fig. 3.1a). Major uplift of the Southern Uplands and northern England also occurred at this time, with removal of ≥ 1.2 km of Mesozoic section across the English Lake District (Łuszczak et al. 2018). Present elevations across Scotland indicate that the mainland rose as a block, with minor tilt towards the east (Fig. 3.1b). The broad pattern of uplift across the northern Highlands is seen in its summit envelope surface (Fig. 3.3). Deformation has occurred towards block margins in NE Scotland, the eastern Midland Valley and towards the North Channel between NE Ireland and SW Scotland. The start of seafloor spreading in the NE Atlantic at ~55 Ma led to basin subsidence along the Atlantic shelf and in the North Sea (Mackay et al. 2005). Rates of sediment input declined sharply in the North Sea (Liu and Galloway 1997), consistent with slowing of uplift and relief reduction on the OSP and in eastern Scotland. The OHP remained emergent (Evans et al. 1997).

Episodes of plate reorganisation in the North Atlantic led to further uplift during the Neogene (Stoker et al. 2005). Continuous flexural uplift and denudational isostasy likely also occurred, as on other Atlantic passive margins (Rouby

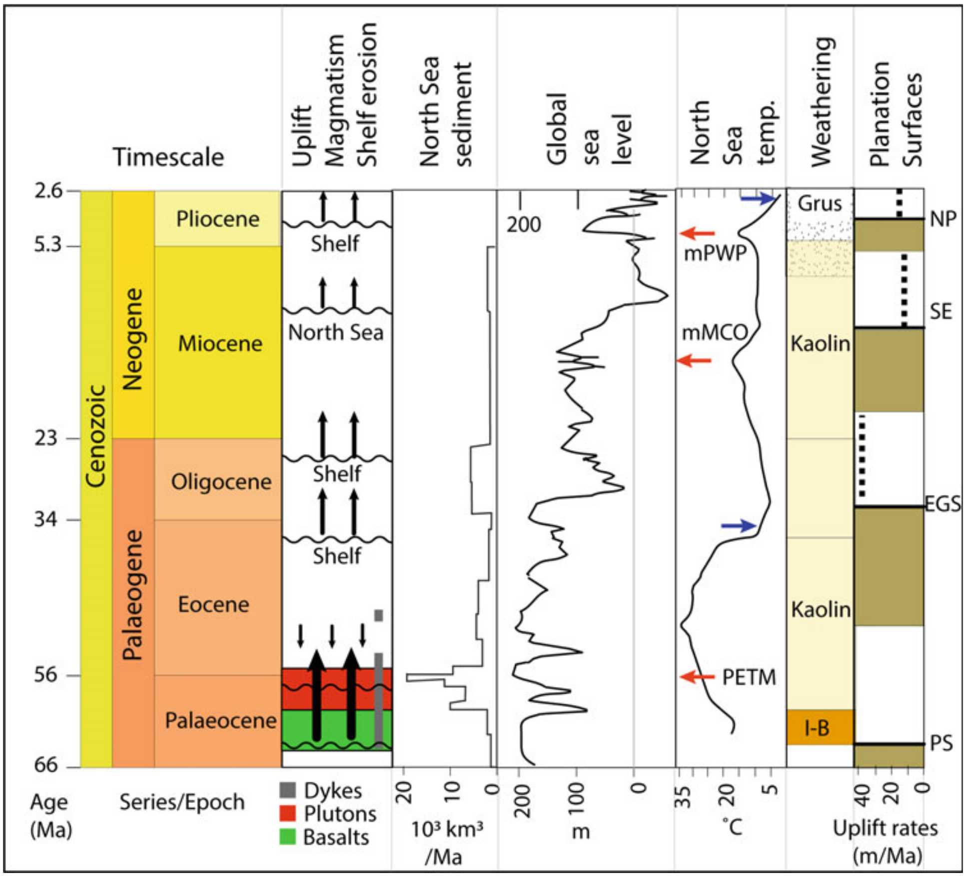

Fig. 3.2 Chronological framework for Cenozoic (Palaeogene and Neogene) environmental change. Main unconformities on the Atlantic shelf (wavy lines) from Stoker et al. (2010). See text for timing of magmatism in the Hebridean Igneous Province. North Sea sediment volumes from Liu and Galloway (1997). Global sea-level curve from Haq et al. (1987). North Sea temperatures from Buchardt (1978). PETM: Palaeocene–Eocene Thermal Maximum; mMCO: mid-Miocene Climatic Optimum; mPWP: mid-Pliocene Warm Period. Weathering types and planation surfaces modified after Hall (1991). NP: Niveau Pliocène; SE: Surface Écossaise; EGS: Eastern Grampian Surface; PS: sub-Palaeocene unconformity; I-B: Inter-basaltic lateritic horizons. Post-Eocene uplift rates (vertical dashed lines) are derived from the relative elevations of the outer edges of each surface since its uplift from base level

et al. 2013), induced by earlier and continuing erosion centred on western Scotland and the OSP and by sediment transfer to, and loading on, North Atlantic and North Sea shelves. In the Late Oligocene, uplift and erosion affected western Britain, with terrestrial sedimentation in small extensional basins (Evans et al. 1991). The edge of the North Atlantic shelf was transgressed in the Early Miocene (Evans et al. 1997). A younger, erosional unconformity developed on the shelf in the Early Pliocene ($\sim$4 Ma), with subsidence along the shelf edge and seaward tilting of the OHP (Stoker 2002). Large volumes of sediment were transferred to the North Atlantic shelf edge in the Miocene and Early Pliocene (Stoker et al. 2010), but in the North Sea, Scandinavian sources dominated sediment supply following uplift of the Scandic mountains in the mid-Miocene. Erosion of the OSP continued to supply material to adjacent shelves through the Pliocene (Ottesen et al. 2018); drowning of the OSP and the OHP occurred mainly during the Pleistocene.

3.3.2 Sea Level

Global sea levels were up to 200 m above present through much of the Palaeogene, with the North Sea reaching its largest extent during the Eocene (Anell et al. 2012), but fell sharply in the Oligocene and dropped again after the mid-Miocene and mid-Pliocene temperature highs (Fig. 3.2). Whilst the amplitude of these fluctuations was much less than the kilometre-scale uplift experienced on the land area, the effect of falling Neogene sea levels was likely profound on the shelves surrounding Scotland. First, shoreline positions shifted across gently inclined shelf surfaces (Mudge 2014). Second, large volumes of sediment released by Palaeogene erosion were likely stored on inner shelves under high Eocene sea levels and later transferred to the North Atlantic shelf edge during later emergence. Third, global sea-level fall in the Pliocene likely led to the development of extensive, emerged shoreline features, similar to the wide coastal ramps developed on Hercynian platforms that stand today at elevations up to 200 m above sea level in western France (Pedoja et al. 2018) and 115 m in SW England (Gunnell 2020). Such features in Scotland included precursors of the strandflat identified in the Inner and Outer Hebrides (Chaps. 9 and 11), later modified, lowered and locally erased by Pleistocene glacial and marine erosion (Dawson et al. 2013).

3.3.3 Climate, Vegetation and Weathering

The location of Scotland on the eastern North Atlantic margin maintained high precipitation throughout the Cenozoic. Mean annual temperatures (MATs) remained subtropical through the Palaeogene (Liu et al. 2018). Towards the Eocene–Oligocene boundary ($\sim$34 Ma), there was a step change towards the modern ice-house climate, with build-up of the Antarctic ice sheet and a sharp fall in sea-surface temperatures in the North Sea (Fig. 3.2). Temperatures recovered to reach 15.5–20 °C in Denmark (Larsson et al. 2011) for a period of $\sim$12 Ma during the Late Oligocene and Miocene. By the Early Pliocene (4–5 Ma), MATs had fallen to −1 °C in the high Arctic (Csank et al. 2011). During the mid-Pliocene Warm Period at $\sim$3 Ma, sea-surface temperatures in the southern North Sea rose briefly to levels similar to, or warmer than the present (Williams et al. 2009), before continued global cooling and the onset of pan-Arctic glaciation at $\sim$3.6 Ma (Knies et al. 2014).

The Cenozoic vegetation record for Scotland is fragmentary. During the Palaeocene, vertical zonation of forest developed over a $\sim$2 km elevation range in the HIP (Jolley 1997), and swamps containing broadleaved and evergreen taxa as well as *Taxodium* (swamp-cypress) covered the inner Moray Firth area (Kender et al. 2012). During the Late Oligocene, cool temperate swamps formed in lowland Antrim, in NE Ireland, with upland coniferous forest (Mitchell 2004), and a similar biome existed in the Inner Hebrides (Evans et al. 1991). Lignite layers of likely Early Pliocene age in BGS borehole 81/19 in the outer Moray Firth contain pollen of *Carya* (hickory) (Andrews et al. 1990). Late Pliocene deciduous forests in western Ireland were of high species diversity, with hickory accompanied by other species now exotic to the British Isles, such as swamp-cypress, *Liriodendron* (tulip tree) and *Nyssa* (sour-gum) (Coxon 2005). Dynamic forest biomes likely dominated throughout the Palaeogene and Neogene, modulating earth-surface processes.

Chemical weathering developed widely below forest floors under warm-to-temperate, humid climates but the character of weathering profiles changed as climates cooled through the Cenozoic. In the Palaeocene, laterite and iron crusts developed under humid subtropical monsoon climates in Antrim (Hill et al. 2000) and in the HIP (Bain et al. 1980). Basalts were weathered to clay to a depth of 80 m during the Eocene and Early Oligocene at Ballymoney in Antrim (Mitchell 2004). In borehole 77/7 near North Rona (Fig. 3.1b), kaolinite-rich, highly weathered basement occurs beneath Late Oligocene sediments (Evans et al. 1997). In Buchan, stable isotope ratios for kaolin clays in >25 m deep weathering profiles indicate groundwater temperatures of 23±5 °C, consistent with wider evidence for intense and deep Palaeogene weathering (Hall et al. 2015; Chap. 21). Clay mineral stratigraphic studies of North Sea sediments show abundant smectite in the Palaeocene and Eocene derived from alteration of volcanic ash sourced from the HIP and other eruptive centres (Huggett and Knox 2006). Climatic cooling from the Late Miocene onwards is associated with an increase in illite and chlorite in North Sea sediments (Nielsen et al. 2015). This change has been linked to the onset of formation of geochemically immature, grus-type weathering profiles in NE Scotland, characterised by low clay contents, retention of little-altered primary minerals such as feldspar and biotite, and varied clay mineral assemblages that remain strongly influenced by parent rock type (Hall et al. 1989). Saprolites and associated supergene minerals in Scotland currently lack absolute dating, unlike in other parts of Europe (Dill et al. 2010a, b), limiting our understanding of rates of long-term denudation and landform development.

3.4 Cenozoic Denudation

Patterns and rates of Cenozoic rock removal across Scotland may be reconstructed using evidence from thermochronology, sediment volumes and the landform record.

Early models from apatite fission track and AHe thermochronometry have provided three contrasting scenarios for the cooling history of the Scottish Highlands: (i) cooling following Caledonian exhumation without significant burial (e.g. Persano et al. 2007); (ii) burial in the Late Palaeozoic to Early Mesozoic followed by uninterrupted cooling (e.g. Thomson et al. 1999); or (iii) multiple cycles of post-Caledonian burial and exhumation from the Late Palaeozoic through the Cenozoic (e.g. Holford et al. 2010). Recent work on low-temperature AHe and (U-Th-Sm)/He thermochronology in the west-central Highlands supports the last scenario, providing evidence for three main cooling phases: at the end of the Caledonian Orogeny, during Permo-Triassic extension and in the Early Palaeogene (Fame 2017; Amin 2020); these correspond with phases of uplift and rapid erosion identified in the geological record. Two main findings stand out for the Cenozoic cooling history: (i) widespread Early Palaeogene cooling is recorded in the HIP (Dobson et al. 2010), along the western edge of the

Highlands (Amin 2020) and in the Southern Uplands (Łuszczak et al. 2018); and (ii) Neogene cooling was limited but remains poorly constrained in magnitude, space and time (Fame 2017; Łuszczak et al. 2018; Amin 2020).

Estimates of the maximum depth of Cenozoic denudation around Ben Nevis in the western Highlands fall within the range 0.7–1.7 km (Fame 2017; Amin 2020). Estimated denudation depths decrease eastward towards the Cairngorms, where sub-Devonian landforms are preserved (Hall and Gillespie 2016). Cenozoic denudation was near zero in Buchan, where Cretaceous Chalk flints survive (Chap. 21). However, estimates based on different thermochronological techniques differ widely, with 0.3 km of Cenozoic denudation estimated at Loch Laggan in the Western Grampians (Fame 2017), but 0.9–1.5 km in the Cairngorms (Amin 2020). Moreover, apparent cooling ages also vary significantly over short (10–20 km) distances. For example, on the Outer Hebrides, estimated maximum Cenozoic denudation varies from 3 km on Barra to 1.5 km on Lewis (Amin 2020). Whilst differential denudation may be real and attributable in part to block movements, further research is needed.

Sediment volumes in the North Sea Basin indicate the loss of an average ~1.0 km of rock from the Scottish Highlands through the Cenozoic (Hall 1991). Similar estimates are derived from modelled subsidence histories for sedimentary basins around Scotland (Jones et al. 2002). Where the sub-Caledonian unconformity stands close to the present erosion level, losses were mainly through the removal of sedimentary cover. Revised calculations that include the Rockall and Faroe-Shetland basins have provided significantly higher estimates of 2.0–2.4 km for Highland denudation (Wilkinson 2017). However, taking account of the potential contributions to these basins from erosion in east Greenland, around Palaeogene igneous centres on the Atlantic shelf and after inversion of Mesozoic basins on the shelf, these estimates appear large. For example, removal of ~1.85 km thick Mesozoic cover is estimated for the West Orkney Basin through the Cenozoic (Evans 1997). Adoption of the higher estimates requires removal of great thicknesses of former Mesozoic cover from the Highlands for which independent evidence is lacking (Hudson 2011).

Palaeocene dyke swarms are exposed at elevations of up to 900 m in the western Highlands, 400 m in the Midland Valley and 600 m in the Southern Uplands (Fig. 3.1b). Where dykes lack evidence of near-surface emplacement, deep erosion below the sub-Palaeocene surface is likely (George 1966). However, where dykes show amygdales and segregation vesicles or are associated with small vents, cooling occurred within a few hundred metres of the present surface. Examples occur on Lewis (Chap. 9), Cowal in the SW Highlands (Hall 1991) and along the Cleveland Dyke in the Southern Uplands (Dagley et al. 2008); these areas experienced more limited uplift in the Palaeocene and less denudation thereafter.

On the OHP, uplift of mountains on Harris led to deep Palaeocene erosion but block movements across major faults maintained north Lewis at low elevations (Chap. 9). In NE Ireland, the preservation of Cretaceous Chalk below Palaeocene lavas indicates that later erosion in Antrim has been confined to the volcanic cover, amounting to the loss of 300–700 m of rock (Holford et al. 2009); depths of Cenozoic erosion across the SW Hebrides were likely comparable. The surface of the OSP has been entirely reshaped, with levelling of Palaeogene uplands after kilometre-scale erosion (Chap. 7). Across much of Scotland, the depth of Palaeocene-Early Eocene denudation requires that major epigene landforms are younger features.

3.5 Landforms

The scenery of Scotland includes four main generations of relief: the ancient exhumed elements (Sect. 3.2), the major pre-Pleistocene Cenozoic landforms, the Pleistocene glacial and marine features, and the landforms and sediments that post-date the Last Glacial Maximum (Chap. 4). The Cenozoic (Palaeogene and Neogene) landforms considered here were variously modified and dissected during Pleistocene cold stages. The form of the older features is most evident in eastern areas where the imprint of glacial erosion was lighter, but the gross form of the Neogene relief is evident everywhere.

3.5.1 Drainage Patterns and Major Valleys

The broad pattern of Palaeogene drainage across Scotland is indicated by sediment routeways to adjacent shelves and by broad straths set deep within the Highlands and Southern Uplands (Fig. 3.1b). During Palaeocene uplift, it is possible that the headwaters of eastward-draining rivers in the NW Highlands locally extended west of the Moine Thrust but firm evidence of a major sediment input from this source area is lacking from the heavy mineralogy of contemporaneous North Sea sandstones (Morton et al. 2004). The positions of the main watersheds are long established, with sediment transport routes indicating origins in the Devonian and Permian.

Early models of Cenozoic drainage evolution identified discordance between drainage routes and regional structural trends. This discordance was attributed to: (i) superimposition of drainage from a formerly extensive cover of Cretaceous rocks, following uplift (Bremner 1942; Linton 1951); and (ii) drainage development consequent to the emergence of tilted marine erosion surfaces in the Eocene (George

1966). Current understanding of the history of uplift and denudation indicates that the present summit surface of the Highlands stands below the sub-Palaeocene surface and that Scotland remained emergent throughout the Cenozoic. The emphasis on discordance was likely also overstated as major faults and shears in Scotland have multiple orientations and the centripetal drainage around the inner Moray Firth includes partly exhumed sub-Devonian valley systems aligned along SW–NE Caledonian structural trends (Hall 1991; Fig. 3.1b). Drainage towards the North Sea and inner Moray Firth developed mainly in response to patterns of Palaeogene uplift. In the Inner Hebrides, drainage was away from igneous centres and across down-faulted and down-warped basalts now found on the seabed (Fyfe et al. 1993). In the Northern Highlands, precursors of the present Shin, Brora and Helmsdale rivers became incised into the uplifted shoulder facing the Moray Firth (Hall 1991). The existence of proto-Dee and proto-Don river systems is indicated by the distinctive Dalradian garnet assemblage of Late Palaeocene sands offshore (Morton et al. 2004). The proto-Tay-Forth drainage fed sediment to deltas in the west-central North Sea (Gatliff et al. 1994).

3.5.2 Planation Surfaces

Stepped sequences of planation surfaces are ubiquitous features on passive margins and provide important insights into long-term relief development (Picart et al. 2019). The outstanding early analysis of Godard (1965) established that a staircase of surfaces exists in the Scottish Highlands, separated locally by scarps, but his landscape models have received little detailed attention by geologists (Jarman 2007). Also, unlike in Scandinavia (Ebert et al. 2011), no detailed GIS analyses exist of the regional topography in Scotland. Here we focus on three well-preserved surfaces in Scotland, north of the Highland Boundary Fault (Fig. 3.3).

- The Niveau Pliocène (90–180 m above sea level) is a peripheral surface that extends inland along valleys. The surface incorporates inherited Palaeogene facets in Lewis and Buchan, but widespread grus-type weathering indicates a mainly Pliocene reshaping. The Niveau Pliocène has been raised along the Minch, Helmsdale and Banff Faults and disrupted by later block faulting in Lewis (Chap. 9), the Sea of the Hebrides (Le Coeur 1999) and Orkney (Chap. 8).
- The Surface Écossaise (180–300 m) is more extensive, penetrating deep into the Northern Highlands and extending towards the main watershed. Similar headward extension along straths and basins occurs in the Central and Eastern Grampians. The Surface Écossaise has been raised along the Minch and Helmsdale Faults, but regional tilting is limited.
- The Eastern Grampian Surface (500–840 m) includes extensive high plateaux around the Cairngorms (Hall 1991; Fig. 3.4a; Chap. 20). On the Monadhliath, this surface was dislocated by later faulting (Ringrose and Migoń 1997). On the Mounth (SE Grampians), this surface has been gently tilted towards the North Sea.

Remnants of the Niveau Pliocène and the Surface Écossaise indicate initiation at base level, followed by extension inland along river valleys, and subsequent uplift and dissection. As the highest fragments of planation surfaces cross-cut Palaeocene dykes in western Scotland (Godard 1965), all surfaces postdate Palaeogene magmatism. Dating controls on surface initiation remain limited. Correlation with offshore unconformities that mark the terminations of phases of erosion after uplift suggests initial formation of the Eastern Grampian Surface in the ~15 Ma long period of relative tectonic stability in the late Eocene to mid-Oligocene, the Surface Écossaise in the mid-Miocene and the Niveau Pliocène by ~ 4 Ma (Fig. 3.2).

3.5.3 Hills

Isolated hills or inselbergs occur widely in the Highlands, either singly or as groups. Some inselbergs reflect localised rock resistance, whereas others appear to be products of backwearing of surrounding slopes. A few exhumed inselbergs also occur, notably the sub-Devonian inselberg of Scaraben in Caithness (Chap. 8). Inselbergs of resistance are most common on quartzites and include the Paps of Jura (Chap. 11) and Schiehallion and numerous hills in the lowlands of NE Scotland (Hall 1986). Inselbergs of this type are also found, however, on a range of other rocks, including volcanic necks in the Midland Valley, metamorphosed grits of the Highland Border Complex (e.g. Ben Ledi), slates (e.g. Hills of Foudland in Buchan) and Devonian conglomerates (e.g. Ben Griam Mòr in Sutherland). The considerable heights of many of these hills indicate prolonged differential denudation. The quartzite inselberg of Mormond Hill in Buchan is, in outline, a Mesozoic relic (Merritt et al. 2003). The most spectacular examples of inselbergs of position occur in the NW Highlands, where a chain of isolated hills in Torridonian sandstone and, towards the north, Cambrian quartzite, stretches from Stac Pollaidh to Cranstackie, rising steeply above the Niveau Pliocène on the surrounding Lewisian lowlands (Fig. 3.4b; Chap. 12). Most inselbergs of position, however, are developed wholly in crystalline rocks. Notable examples include Morven and Mount Keen, rising above the plateaux flanking the Dee valley, and the larger hill masses of Ben Hope and Ben Hee in the Northern

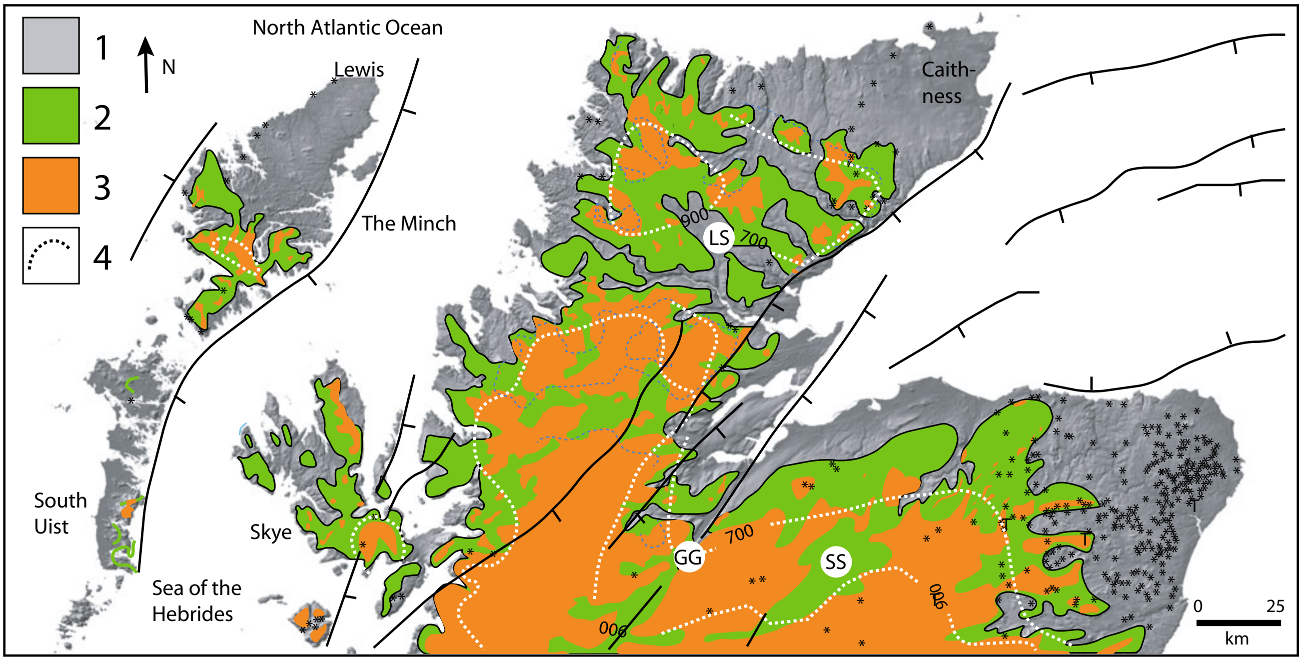

Fig. 3.3 Denudation and planation in the northern Highlands. 1: Niveau Pliocène. 2: Surface Écossaise. 3: Relief standing above the Surface Écossaise. 4: Summit envelope surface contours. *: Relict saprolite. GG: Great Glen; LS: Loch Shin; SS: Strathspey

Highlands. These inselbergs often occur within areas of rocks of relatively uniform resistance where their elevation cannot be accounted for by Neogene fault movements, and isolation through the encroachment of surrounding planation surfaces is likely (Godard 1965).

3.5.4 Basins

Topographic basins typically have broad floors of low relief enclosed by inward-facing slopes broken only by narrow drainage exits. Basins are a major element of the Highland scenery and are found at elevations from close to sea level to over 900 m in the Cairngorms. Basins also occur in the Howe of Fife in the Midland Valley and in upper Nithsdale in the Southern Uplands. The large basins of the Northern Highlands (Jarman 2007; Fig. 3.4c) and the Central Grampians (Hall 1991), including Rannoch Moor (Chap. 17), contain or are flanked by stepped planation surfaces and so have a development that spans at least the Neogene period. In NE Scotland, the basins are smaller, with floor areas of 10–60 km^2. Many basin floors are dissected and where the floors lie well above present valley floors, as in the Atholl basin, dissection began before the Pleistocene. Basins are rare west of the main watershed, where steep valley gradients encouraged pre-glacial valley incision and intense glacial dissection of relief (Haynes 1983; Chap. 4).

Many basins are long-established features, set deeply within the surrounding hills. The Cabrach basin is an exhumed Devonian form. Elsewhere in NE Scotland, basins up to 300 m deep are developed largely in biotite-bearing basic and granitic rocks that are susceptible to chemical weathering (Fig. 3.4d). Basin margins commonly coincide with litho-structural boundaries (Hall 1991). Grus is common beneath basin floors and linear weathering zones up to 50 m deep are preserved at the base of bounding scarps (Hall 1986). However, the large topographic basins around Lochs Naver and Shin do not show simple litho-structural controls on their positions and forms (Godard 1965). Whilst these basins were perhaps initiated by deep weathering, scarps at basin margins appear to have retreated across more resistant lithologies during Neogene basin-floor lowering.

3.6 Models of Long-Term Relief Development

The Scottish and Norwegian sections of the North Atlantic passive margin have broadly similar Phanerozoic tectonic histories, but with a dominance of Palaeogene uplift in Scotland versus Mio-Pliocene uplift in Norway. Debate over the long-term development of the Norwegian margin is currently polarised around two end-member models: (i) persistence of mountains since the Caledonian Orogeny with long-term denudation of dynamic topography modulated through variations in eustasy and climate (e.g. Huuse 2002; Nielsen et al. 2009; Pedersen et al. 2018); and (ii) episodic tectonism and uplift of the continental margin after

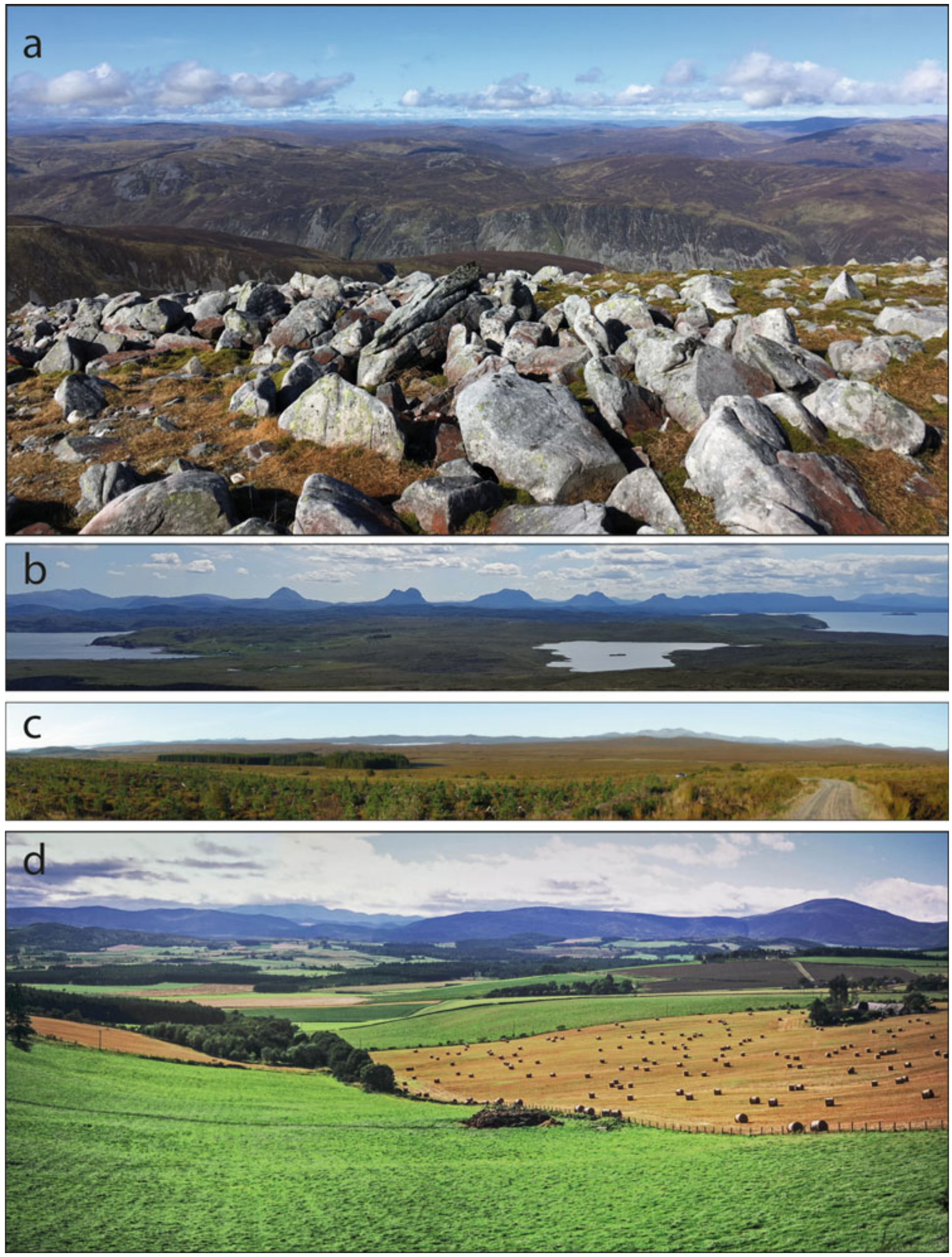

Fig. 3.4 Erosion surfaces and landforms. **a** View northwest from the highest point on Beinn a' Ghlo across the Eastern Grampian Surface on the Gaick plateau. **b** View southeast from Point of Stoer towards the chain of inselbergs of position that rise above the glacially eroded surface of the Niveau Pliocène. **c** Panoramic view west from near Crask across the Surface Écossaise on the floor of the Loch Shin basin and towards the main watershed. **d** View southeast across the Tarland basin towards the inselberg, Morven (right), that rises above the Eastern Grampian Surface. (Images: **a** Colin Ballantyne; **b**, **c** Adrian Hall; **d** John Gordon)

Mesozoic rifting, with formation of peneplains at base level, later uplifted but not significantly eroded (e.g. Lidmar-Bergström and Näslund 2002; Japsen et al. 2018). How does Scotland compare?

The post-Caledonian history of Scotland is not one of simple persistence of topography. Major episodes of uplift, mountain building and cooling associated with deep erosion occurred in the Early Devonian and Late Permian. By the Late Devonian and Late Triassic, the relief had been much reduced, allowing marine transgression of basin margins in the Early Carboniferous and Early Jurassic (Hall 1991). Persistence is seen, however, through the Mesozoic in palaeogeographic reconstructions of Scotland, when the Outer Hebrides, the Northern and Grampian Highlands and the Southern Uplands remain buoyant and shed sediment to surrounding basins (Hudson and Trewin 2002). However, by the culmination of the Late Cretaceous marine transgression, the residual land area of Scotland was small and low lying (Fig. 3.1b); no mountain topography remained from the Palaeozoic.

After kilometre-scale uplift in the Early Palaeocene, later uplift was episodic. The focus of deep erosion in western Scotland and its eastward decline across the Highlands, as indicated by thermochronology, is consistent, however, with additional, but subordinate flexural uplift of the western shoulder of Scotland driven by denudational isostasy (Fame et al. 2018). The elevation of the Eastern Grampian Surface at its eastern edge indicates minimum uplift of ~500 m since the mid-Eocene in parts of the Eastern Grampians. The younger surfaces indicate Neogene uplift rates of 13–22 m Ma^{-1} (Fig. 3.2), comparable to those in England and Ireland (Walsh et al. 1999). Generally, however, Cenozoic uplift and erosion depths in Scotland remain poorly constrained, falling between the temporal and spatial resolutions of low-temperature thermochronology (>10 Ma and >0.5 km depth) and cosmogenic isotope inventories (<1 Ma and <10 m depth).

Planation surfaces were initiated at base level during periods of relative tectonic stability and high eustatic sea level (Fig. 3.2). Surfaces were extended inland by headward erosion along short (100–250 km) river systems in southern Scotland (Sissons 1960), the Grampian Highlands (Hall and Bishop 2002) and the Northern Highlands (Godard 1965). Remnants of deep weathering covers and kaolinitic terrestrial sediments in basins around Scotland indicate that etch processes were fundamental in development of low-relief surfaces (Thomas 1995). Topography often became closely adjusted to litho-structural controls, principally mineralogy and fracturing, which influenced patterns and rates of chemical weathering (Godard 1962; Hall 1986). Planation surfaces, however, were subjected to continued weathering after uplift, as shown by the presence of deep grusses at elevations up to 800 m in the Cairngorms (Hall and Mellor 1988), and to significant erosion, with minimum rates of 10 m Ma^{-1} in western Scotland (Le Coeur 1999). Internal relief of hills and basins developed on uplifted surfaces during downwearing. Hence, the present upland surfaces are not preserved uplifted base-level peneplains, but instead represent dynamic etchsurfaces, modified after uplift during 2–20 Ma long periods of downwearing and headward extension, and largely stripped of regolith by Pleistocene glacial erosion (Chap. 4).

3.7 Conclusions

Problems of denudation and relief development over the last 400 Ma in Scotland have received little attention. Tiered unconformities indicate limited erosion of crystalline basement since the Devonian. By the end of the Cretaceous, relief was low. Major uplift along the North Atlantic margin, driven mainly by Palaeogene magmatism, produced contemporaneous kilometre-scale erosion in the western Highlands, but depths of Cenozoic erosion thinned eastwards. A staircase of younger etchplains developed close to base level in the late Eocene, mid-Miocene and Pliocene. Each surface was uplifted after formation, dissected, lowered and modified by weathering and erosion. Coastal platforms developed under falling sea levels in the Pliocene, including precursors of the Hebridean strandflat.

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The Quaternary in Scotland

4

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Abstract

The Quaternary Period in Scotland was characterized by major climatic shifts and the alternation of glacial and temperate conditions over a wide range of timescales. The extent of multiple glaciations prior to the Mid-Pleistocene Transition (1.25–0.70 Ma) is uncertain, but thereafter up to ten major episodes of ice-sheet expansion occurred. Glacial erosion by successive glaciers and ice sheets created a range of terrain types: glaciated mountains, zones of areal scouring, landscapes of selective linear erosion, drift-mantled terrain of differential erosion and areas of limited glacial modification. The last ice sheet (∼35–14 ka) extended to the shelf edge in the west and was confluent with the Fennoscandian Ice Sheet in the North Sea Basin; during its existence, it experienced marked changes in configuration, in part driven by the development of major ice streams. Subsequent glaciation during the Loch Lomond Stade (∼12.9–11.7 ka) was restricted to a major icefield in the western Highlands and smaller glaciers in peripheral mountain areas. Contrasting glacial landsystems occupy terrain inside and outside the limits of the Loch Lomond Stadial glaciers. Postglacial landscape changes have been characterized by Lateglacial periglaciation and paraglacial landscape modification, mainly in the form of rock-slope failure and the accumulation (then later erosion) of paraglacial sediment stores and incision and terracing of glacigenic valley fills. Shore platforms of various ages formed around rock coastlines during the Quaternary, glacio-isostatic uplift has resulted in the formation of Lateglacial and Holocene raised beaches, and reworking of glacigenic deposits has provided sediments for present-day beach and dune systems.

Keywords

Pleistocene • Holocene • Late Devensian • Loch Lomond Stade • Glaciation • Ice streams • Periglaciation • Glacial erosion • Glacial landsystems • Permafrost • Paraglacial landscape modification • Rock-slope failure • Fluvial landforms • Sea-level change • Rock platforms • Raised beaches • Beach and dune systems

4.1 Introduction

The Quaternary Period comprises the Pleistocene Epoch (2.59 Ma to 11.7 ka) and Holocene Epoch (11.7 ka to the present). The Pleistocene is conventionally subdivided into the Early (2.59–0.78 Ma), Middle (0.78–0.13 Ma) and Late (130–11.7 ka) Pleistocene.

The Quaternary was characterized by dramatic and rapid climatic shifts. These resulted in alternation of glacial stages (stadial episodes), when Scotland experienced a climate of arctic severity, with interglacial stages (and interstadial episodes), when more temperate conditions prevailed. Cold periods were dominated by the effects of glaciation and periglaciation, whereas each return to temperate conditions was characterized by paraglacial landscape response and the operation of weathering, slope, fluvial, coastal and aeolian processes. These changes occurred against a backdrop of fluctuations in local relative sea level (from about −130 to +40 m compared to present sea level) and changes in biome, from tundra and cold desert to boreal forest and,

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during interglacials, temperate deciduous forest. Although the major relief elements of Scotland were established before the Pleistocene (Chap. 3), the effects of successive glacial cycles transformed much of the Scottish landscape.

Our understanding of the chronology and nature of landscape changes in Scotland during much of the Quaternary is limited, because successive ice sheets covered all of the present land area, removing most of the terrestrial stratigraphic record of earlier events and environmental conditions. There is consequently a considerably greater body of published literature concerning the last ice sheet and its aftermath (the last ~35 ka) than all of the earlier Quaternary. Conversely, however, our growing understanding of the extent and dynamics of the last ice sheet provides insights into earlier periods of ice-sheet glaciation. Similarly, the brief return to cold conditions after the retreat of the last ice sheet during the Loch Lomond (≈Younger Dryas) Stade of ~12.9–11.7 ka provides an analogue for periods of more restricted mountain glaciation and the Holocene (present interglacial) yields insights into the nature of landscape evolution during earlier interglacial periods.

Throughout the Quaternary, erosion by glacier ice during successive cold periods operated in synergy with nonglacial processes during intervening interglacial and interstadial periods. Glacial erosion steepened slopes and deepened valleys, glacial deposition mantled low ground with sediment and loading by glacier ice depressed the lithosphere and stressed the near-surface crust. Every subsequent deglaciation was accompanied and followed by isostatic rebound and paraglacial landscape adjustment in the form of rockfall activity, rock-slope failures and reworking of glacigenic sediment by hillslope, fluvial and coastal processes. During interglacials, much of the resulting paraglacial sediment was stored on the land surface or in nearshore locations. Such sediment and the products of interglacial weathering were then reworked by later glaciers or ice sheets, often being transported to offshore depocentres on the adjacent shelf or shelf edge. The Quaternary landscapes of Scotland are therefore not simply the outcome of successive episodes of glacial and glacifluvial erosion and deposition, but the product of landscape evolution during multiple glacial-interglacial cycles. Moreover, cumulative mass losses from eroded areas and corresponding mass gains in peripheral sediment sinks progressively influenced regional patterns of uplift and subsidence by redistributing overburden loading on the lithosphere.

This chapter focuses first on the nature of landscape evolution during the Early and Middle Pleistocene, before addressing: (i) the dynamics and chronology of the last ice sheet; (ii) glaciation during the Loch Lomond Stade; then (iii) the characteristics of Late Devensian glacial landsystems. The final parts of the chapter consider the trajectories of landscape evolution during both the Lateglacial period (between final local deglaciation and the onset of the Holocene) and the Holocene itself. Key locations are depicted in Fig. 4.1.

4.2 Early and Middle Pleistocene

4.2.1 Introduction

Although the main topographic features of Scotland were established before the Quaternary (Chap. 3), the end-Neogene landscapes of Scotland differed fundamentally from those of the present. The North Sea coastline lay seaward from its present location (Huuse 2002) and many indentations of the Scottish coastline (firths, fjords, sounds and inner seas) did not exist in their present form; some islands, such as Skye and Mull, were still part of the Scottish mainland. The end-Neogene landscape was dominated by dendritic river systems flowing towards extensive coastal lowlands and mantled by thick covers of soil, sediment and saprolite. Much of the modification of this landscape was accomplished by successive Pleistocene glaciers and ice sheets, but weathering, slope, fluvial and coastal processes continued to operate over long periods, particularly during the Early Pleistocene. The rapid tempo of environmental changes over comparatively brief timescales, coupled with changing vegetation cover and shifts in sea level, is likely to have enhanced rates of erosion by nonglacial processes, contributing to landscape transformation. The extraordinary dynamism of Pleistocene environments in Scotland typified that of glaciated areas along the North Atlantic passive margin.

Neotectonic movements of the crust operated throughout the Pleistocene in Scotland, driven by the far-field effects of intra-plate stresses generated by mid-Atlantic ridge push and the Alpine orogeny, the transfer of large volumes of sediment onto adjacent offshore shelves and basins, and glacio-isostatic loading and unloading of the crust by the growth and shrinkage of successive ice sheets (Stewart et al. 2000). The long-term effects of such movements are illustrated by evidence for vertical movement of fault blocks in Orkney, Moray and the Hebrides, and by uplift of a low-level Pliocene planation surface that formed around the margins of the Highlands to elevations of up to 150 m in eastern Lewis, Caithness and Buchan (Chap. 3). Similarly, high-level (10–40 m OD) shore platforms, often buried under till, are preserved on coastlines peripheral to the centres of ice accumulation (Smith et al. 2019; Sect. 4.9.2). Fault scarps provide evidence of minor faulting since the retreat of the last ice sheet (Firth and Stewart 2000; Smith et al. 2009) and significant seismic activity ($M_L \geq 4$) continues to affect the western Highlands, Inner Hebrides and Midland Valley at present (Musson 2007).

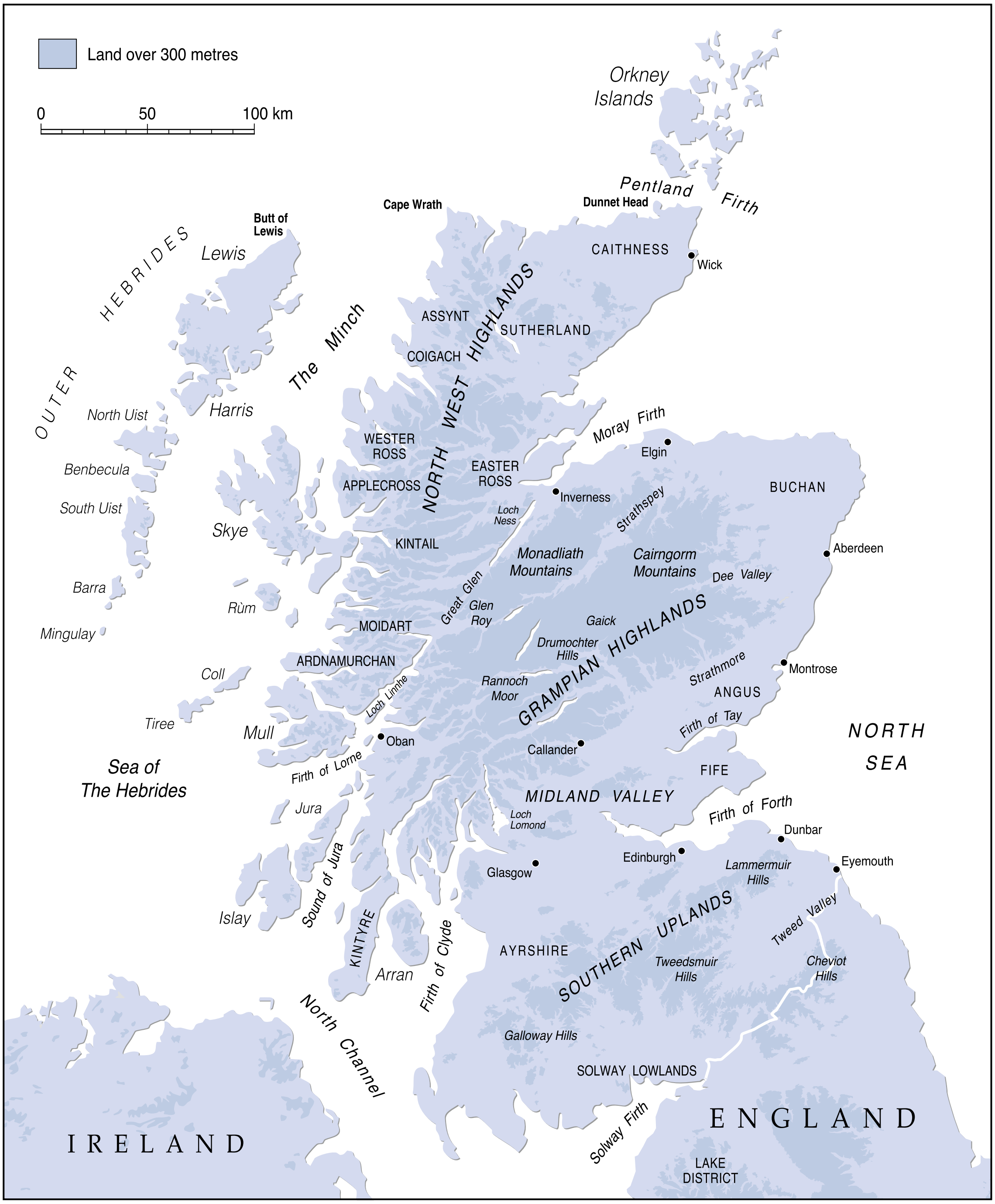

Fig. 4.1 Key locations mentioned in the text. The Shetland Islands (not shown) are located ~170 km north of the Scottish mainland

The $\delta^{18}0$ record for North Atlantic marine sediments constrains the first-order timing and intensity of Pleistocene environmental change in Scotland (Fig. 4.2). In the Early Pleistocene, global climate forcing was driven by 41 ka orbital cycles, but during the Mid-Pleistocene Transition (MPT) at ~1.25–0.70 Ma there was a shift towards forcing driven by 100 ka cyclicity (Head and Gibbard 2005; Willeit et al. 2019) that initiated the growth of extensive ice sheets. In the Early Pleistocene, global sea level reached a few metres higher than present during brief interglacial periods and declined no lower than −60 m during glacial stages. After the MPT, sea-level lowstand reached −130 m during glacial maxima due to the huge volumes of water sequestered in ice sheets (Spratt and Lisiecki 2016).

The Early Pleistocene was dominated by two climate types (Hall et al. 2019; Fig. 4.2): an interglacial or

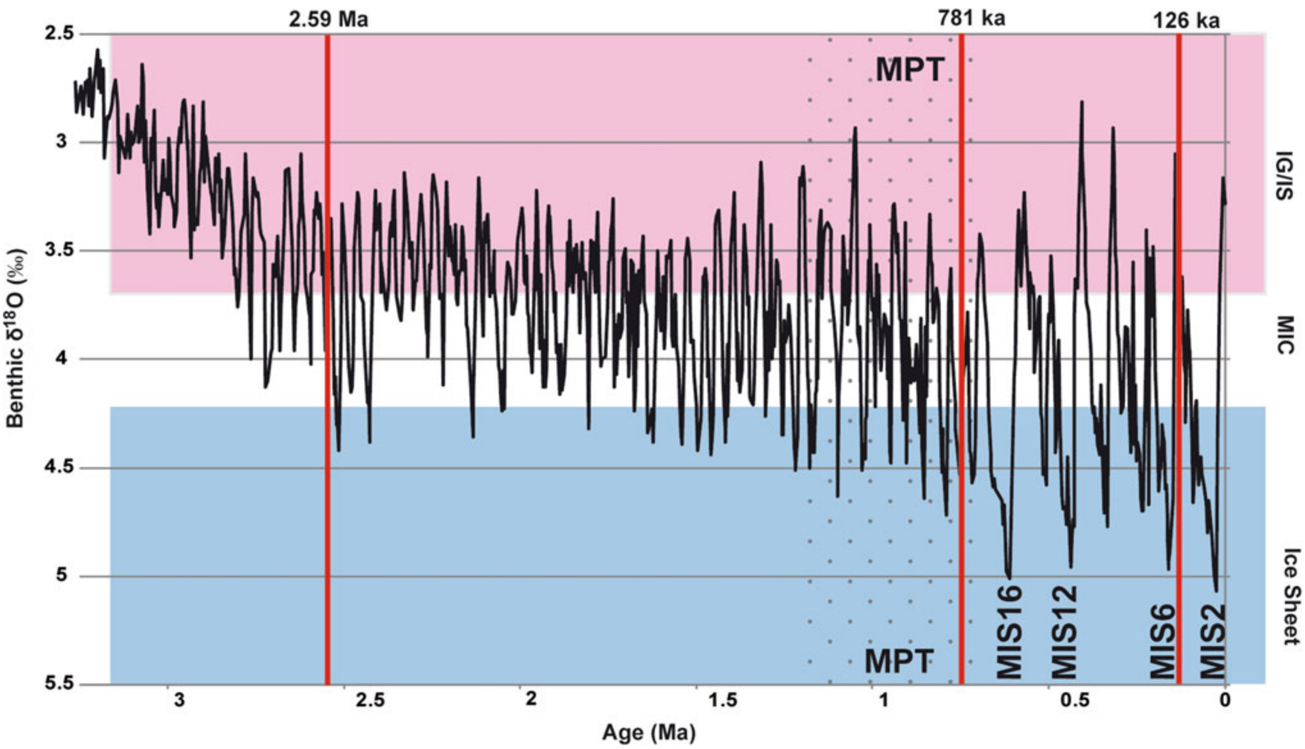

Fig. 4.2 The δ^{18}O record for benthic foraminifera from marine core DSDP 607, interpreted as a proxy for glacier extent in Scotland. Values of <3.7$^0/_{00}$ δ^{18}O are interpreted as indicating interglacial (IG) or interstadial (IS) conditions in Scotland; 3.7–4.2$^0/_{00}$ δ^{18}O as indicating periods of mountain ice caps (MIC) and limited glacier cover; >4.2$^0/_{00}$ δ^{18}O as indicating development of extensive ice sheets. MPT: Mid-Pleistocene Transition. MIS: Marine Isotope Stage. (Modified from Hall et al. 2019 Earth Env Sci Trans R Soc Edinb 110:537 © The Royal Society of Edinburgh)

interstadial regime, when temperate conditions prevailed and glaciers were absent in Scotland; and a stadial regime, when glaciers developed in mountain areas and periglacial conditions affected lower, peripheral locations (Murton and Ballantyne 2017). Owing to their position on the leeward margin of the North Atlantic Ocean, Scotland's mountains provided near-optimal conditions for the rapid development of icefields, ice caps and valley glaciers in response to moderate (~3–6 °C) falls in summer temperature, as evidenced by the formation of such glaciers during the Loch Lomond Stade (Golledge et al. 2008; Sect. 4.5). The maximum extent of glacier ice during the Early Pleistocene is uncertain. Evidence provided by iceberg scour marks in stacked marine sediments suggests repeated extension of grounded ice margins into the western North Sea Basin after 2.53 Ma and the central North Sea Basin after 1.87 Ma, implying that ice sheets developed over Scotland from the earliest Pleistocene (Rea et al. 2018). These may, however, have been of limited extent and duration, as Early Pleistocene sediments on the edge of the East Shetland Platform appear to lack a glacial component until 1.1–1.0 Ma (Buckley 2017). In the west-central North Sea Basin, the Early Pleistocene Aberdeen Ground Formation is truncated by an extensive erosional unconformity, the Upper Regional Unconformity (URU), which formed at ~1.2 Ma and probably represents the earliest expansion of the Fennoscandian Ice Sheet into the central North Sea area (Reinardy et al. 2017).

The increased intensity and duration of cooling events after the beginning of the Middle Pleistocene implies that up to ten subsequent episodes of ice-sheet expansion occurred in Scotland (Fig. 4.2). By analogy with the extent and thickness of the last ice sheet (Sect. 4.4), it is likely that the ice sheets that developed in marine isotope stages (MIS) 16, 12 and 8–6 had maximum surface elevations of 1.5–2.0 km or more and extended far across the adjacent shelves to, or close to, the Atlantic shelf break in the west (Stoker et al. 1993). Eastwards extension of successive Middle Pleistocene ice sheets across the North Sea Basin is indicated by increasing influx of sediment of Scottish origin over the URU, seven generations of tunnel valleys within these sediments and several sets of mega-scale glacial lineations within the Middle to Late Pleistocene sequence (Stewart et al. 2013). During successive Middle Pleistocene glacial maxima, the Scottish and Fennoscandian Ice Sheets were probably confluent across a shifting, broad zone in the North Sea Basin (Bendixen et al. 2017; Sect. 4.4.1). Extension of Scottish ice as far south as East Anglia during the Middle

Pleistocene is demonstrated by the presence of indicator heavy minerals and erratics of Scottish provenance in coeval till deposits (Lee et al. 2015).

It is likely that the patterns of ice flow during periods of Middle Pleistocene ice-sheet glaciation were similar to those of the last ice sheet, with primary centres of ice dispersal located over the Scottish Highlands and Southern Uplands and persistent local ice divides over the Outer Hebrides, the Orkney-Shetland Platform, Skye, Mull, the Cairngorm Mountains and the Galloway Hills (Sect. 4.4.1). Similarly, the locations of the main ice divides and flow paths are likely to have migrated during the lifetimes of Middle Pleistocene ice sheets (Hughes et al. 2014). As with the last ice sheet, successive Middle Pleistocene ice sheets probably exhibited 'binge and purge' cycles, with build-up of cold-based ice over upland areas alternating with drawdown of ice evacuated by persistent or transient ice streams (Hubbard et al. 2009).

4.2.2 Early and Middle Pleistocene Terrestrial Stratigraphy

In Scotland, no Early Pleistocene sediments have been confirmed onshore. In a few locations in Buchan and Caithness, however, sediments or fossils of probable Early Pleistocene age have been reworked into Late Pleistocene glacigenic sequences (Merritt et al. 2003).

The Middle Pleistocene stratigraphic record in Scotland includes glacigenic deposits and periglacial features that represent cold stages back at least to MIS 8, together with terrestrial sediments with a floral, faunal or pedological record of interstadial and interglacial conditions. The most complete stratigraphic record of Middle Devensian events is preserved in the Kirkhill area of Buchan, where at least three stadial-interstadial-interglacial cycles are represented (Hall et al. 2019). The lowermost cycle probably pre-dates MIS 7 and includes glacial and glacifluvial material deposited at different times by ice moving either towards or away from the North Sea coast; these sediments are overlain in turn by a periglacial gelifluctate (solifluction deposit) then a podzolic palaeosol that formed under humid-temperate conditions. The overlying sediments of the second cycle, assigned to MIS 6–5e, record an initial return to permafrost conditions (represented by ice-wedge pseudomorphs and gelifluctate), followed by glaciation (represented by a till) and development of a palaeosol of inferred last interglacial age. The final cycle, of Devensian (MIS 5d–2) age, is represented by ice-wedge pseudomorphs indicating the return of permafrost, tills deposited by the last ice sheet, then Lateglacial gelifluctate.

At a few other sites, glacigenic deposits of Middle Devensian age underlie organic deposits of probable last interglacial (MIS 5e) age. On the NW coast of Shetland, for example, a till indicating SE–NW ice movement underlies a peat layer of probable last interglacial age, suggesting that during MIS 6 ice flowed radially outwards from the Orkney-Shetland Platform (Hall et al. 2002; Chap. 7). Across Moray and Buchan, and into lower Strathspey, tills of probable MIS 6 age that locally underlie last interglacial palaeosols and organic deposits contain erratics and exhibit fabrics that suggest directions of ice flow similar to those of the last ice sheet (Merritt et al. 2003). In Caithness, lower till units of uncertain age record both former ice flow from the mountains to the west and northeastward flow of ice from the Moray Firth (Hall and Riding 2016). Collectively, these and other sites suggest that the pattern of flow of the MIS 6 ice sheet was broadly similar to that associated with the last ice sheet.

4.2.3 Long-Term Glacial Erosion

Given the frequency of glacial episodes in Scotland after ~2.6 Ma (Fig. 4.2) and the growth and demise of successive ice sheets after ~0.78 Ma, it is likely that all large-scale glacial erosional landforms in Scotland had formed before the last interglacial and that Late Pleistocene (~130–11.7 ka) glacial erosion accomplished only limited landscape modification. This is evident in the preservation of older bedrock landforms, such as shore platforms, sea cliffs, meltwater gorges and the scars of major rock-slope failures. Moreover, the geomorphological legacy of successive episodes of Pleistocene glaciation was markedly non-uniform across the country. Topographic asymmetry and exposure to dominantly westerly airmasses produced strong longitudinal climatic gradients across Scotland. Snowfall and air temperature were highest on the Atlantic margins (Barr et al. 2017) and consequently the glaciers in western Scotland had high ice discharges and were predominantly warm-based (Hubbard et al. 2009). The overall duration of Pleistocene ice cover was also greater near western centres of ice dispersal, particularly during periods of limited glaciation similar to that of the Loch Lomond Stade, when marine-terminating icefields developed in the western Highlands, but only small cirque and valley glaciers occupied the comparatively snow-starved eastern Highlands (Sect. 4.5). Glacial dissection of mountain areas is consequently most strongly developed in the Hebrides, western Highlands and SW Southern Uplands (Chaps. 10, 13, 17 and 27).

Farther east, as across the Eastern Grampians, Cairngorm Mountains, the lowlands of Buchan and much of the Southern Uplands, successive ice sheets were predominantly slow-moving and cold-based, and consequently accomplished very limited modification of palaeosurfaces (Hall and Sugden 1987; Chaps. 18, 20, 21 and 27). In such areas, effective glacial erosion was confined to corridors of fast-flowing ice within glacial troughs and landscapes of selective linear glacial erosion evolved (Rea 1998).

The effect of glacial erosion across lowlands was conditioned by the underlying lithology. Across the resistant gneisses of NW Scotland and the Outer Hebrides, for example, glacial abrasion and plucking predominated, forming knock-and-lochan landscapes of low, ice-scoured bedrock hillocks and intervening depressions, typically with patchy drift cover (Krabbendam and Bradwell 2014; Chaps. 9 and 12). In the Midland Valley, differential glacial erosion created a landscape of drift-covered lowlands underlain by sedimentary rocks, interspersed with higher ground formed of resistant lavas, sills and volcanic plugs that have commonly been streamlined in the direction of ice flow and shaped into crag-and-tail forms (Chap. 26).

The western Highlands and mountainous parts of Skye, Rùm, Mull and Arran are characterized by a glacial landscape of deep troughs, cirques, rock basins, fjords, arêtes and areas of glacially roughened bedrock (Fig. 4.3a–c). In this region, successive episodes of glacial erosion operating on a fluvially dissected landscape removed all but small fragments of pre-Quaternary palaeosurfaces. A repeatedly established, former mountain ice-cap zone in the western Highlands is marked by lake-filled rock basins in a 15–30 km wide zone on both sides of the former main Highland ice shed. In the NW Highlands, eastward migration of the former ice divide (Hughes et al. 2014) resulted in the formation of numerous glacial breaches that cut through the present north–south watershed. Equally conspicuous are the glacial breaches, troughs and rock basins that radiate from the Rannoch Moor basin in the Grampian Highlands, suggesting that this area formed a major centre of ice dispersal during successive glaciations (Chap. 17). Despite the pervasive effects of glacial erosion, however, some summits and plateaux in the western Highlands retain a cover of little-modified blockfields that pre-dates at least the last ice sheet (Fabel et al. 2012; Hopkinson and Ballantyne 2014). The preservation of such blockfields implies that the last ice sheet was polythermal, with cold-based ice occupying the highest ground whilst warm-based, erosive ice flowed through adjacent troughs (Fame et al. 2018) and it is likely that this was also true of Early and Middle Pleistocene ice caps and ice sheets. Along the western seaboard from Knoydart to Argyll, however, summit blockfields are absent and glacially moulded bedrock often extends to mountain summits.

Across much of the eastern Highlands the polythermal nature of the last and earlier ice sheets is manifest in the striking contrast between deep glacial troughs and well-preserved, extensive plateau palaeosurfaces such as those of the Cairngorms, Monadhliath, Gaick and Eastern Grampians (Chaps. 3, 18 and 20; Fig. 4.4a–c). Similar undulating tablelands dissected by glacial troughs also form high ground in the Southern Uplands (Chap. 27), though here the effects of glacial erosion are more muted (Fig. 4.4 d, e). Such topographic contrasts imply that during successive glaciations cold-based ice that was frozen to the underlying substrate persisted on high ground, feeding fast-moving, warm-based glaciers that occupied (and progressively eroded) adjacent and intervening troughs (Sugden 1968; Glasser 1995; Rea 1998). Some plateau margins in the eastern Highlands, such as the Cairngorms, are extensively scalloped by cirques. Others, such as the Gaick plateau, are almost devoid of cirques, but flanked by steep trough heads.

Blockfields and other forms of frost-weathered regolith (gelifractate) mantle most plateaux in the eastern Highlands, and on the Cairngorms plateau numerous granite tors have survived burial by the last and earlier ice sheets, though some have been modified by glacial erosion (Hall and Phillips 2006; Phillips et al. 2006). Across the Highlands as a whole, the lower limit of blockfields declines away from the centres of ice dispersal, indicating the fundamental importance of englacial thermal boundaries in dictating upper limits to effective glacial erosion (Hall and Glasser 2003). There is, for example, a consistent decline in the lower limit of blockfields away from former ice-shed locations in the northern Highlands (Ballantyne et al. 1998; Ballantyne and Hall 2008) and the upper limit of effective glacial erosion also declines eastwards across much of the Grampians (Hall et al. 2019), such that even the low palaeosurface of the Buchan plain exhibits negligible evidence of erosion (Hall and Sugden 1987; Chap. 21).

During major glaciations, major pre-glacial valleys such as the Great Glen, Strathspey, Strathmore and the Tweed valley formed important arteries for the evacuation of ice from upland sources and guided ice flow towards major ice streams that developed mainly on relatively weak sedimentary rocks or deformable unconsolidated sediments (Bradwell et al. 2019). Activation of ice streams led to draw-down of ice towards the offshore Mesozoic basins of the Moray Firth, Minch, Sea of the Hebrides and North Sea Basin, where ice flowing across weak substrates during Middle Pleistocene glaciations excavated deep basins. Glacial overdeepening over multiple glacial cycles reshaped the Scottish coastline, causing it to recede landwards and severing islands such as Mull, Skye and Orkney from the mainland (Chaps. 8, 9 and 10).

The spatial pattern and depths of cumulative Pleistocene glacial erosion can be estimated by using nonglacial landforms and landsurfaces as reference points. In general, depths of glacial erosion increase westwards across the Grampians. In eastern Buchan, where Tertiary gravel deposits survive on ridges, total erosion has locally been of the order of metres (Hall and Sugden 1987). In the Cairngorms, glacial deepening of pre-glacial valleys probably did not exceed 200–350 m (Hall and Gillespie 2016) and glacial

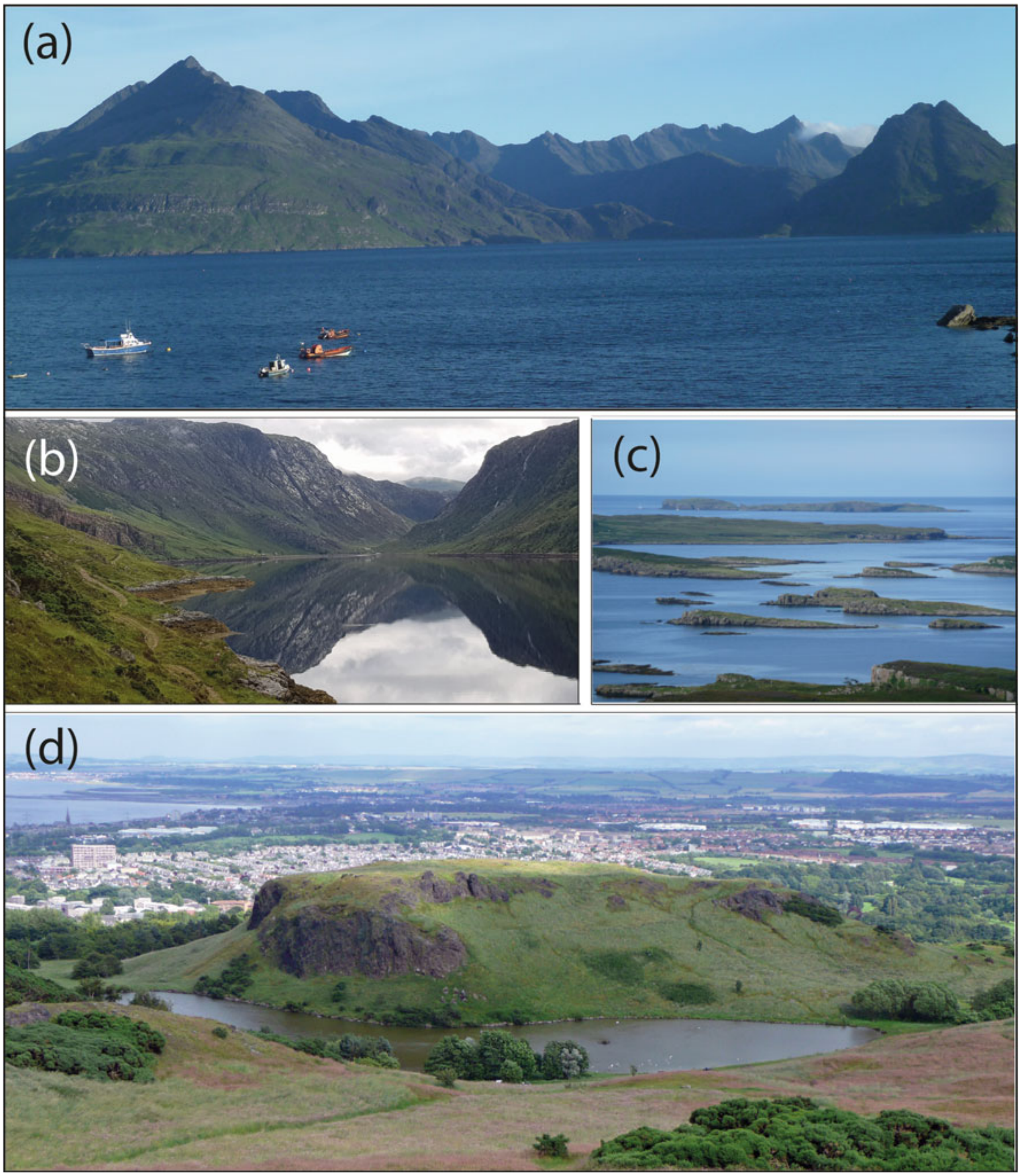

Fig. 4.3 Landscapes of high-intensity glacial erosion. **a** The Black Cuillin, Skye: an alpine glacial landscape fashioned mainly in Palaeogene gabbros by Pleistocene cirque and valley glaciers. **b** Gleann Dubh, Assynt: a mountain glacial landscape and fjord cut mainly in Archaean gneisses. **c** Ulva, Mull: glacially roughened surface of gently inclined Palaeocene basalt lavas; ice movement was left to right. **d** Dunsapie Hill, Edinburgh: part of a resistant Carboniferous basic igneous vent picked out by glacial erosion and standing ~100 m above the adjacent Carboniferous sedimentary rocks. (Images: Adrian Hall)

erosion of plateaux was extremely limited. Average Pleistocene erosion depths across the Dee catchment south of the Cairngorms have been estimated at 30–60 m, with a substantial contribution from the removal of saprock and saprolite (Glasser and Hall 1997). An eastward decline in the intensity of cumulative glacial erosion is also indicated by landform evidence. Glacially streamlined terrain across much of Caithness and along the lower Dee valley merges eastwards with non-streamlined terrain with similar ridge-top elevations, where glacial erosion has been restricted mainly to the excavation of valleys and depressions.

Along the western seaboard, focusing of glacial erosion is also evident. On the open Lewisian gneiss terrain of the

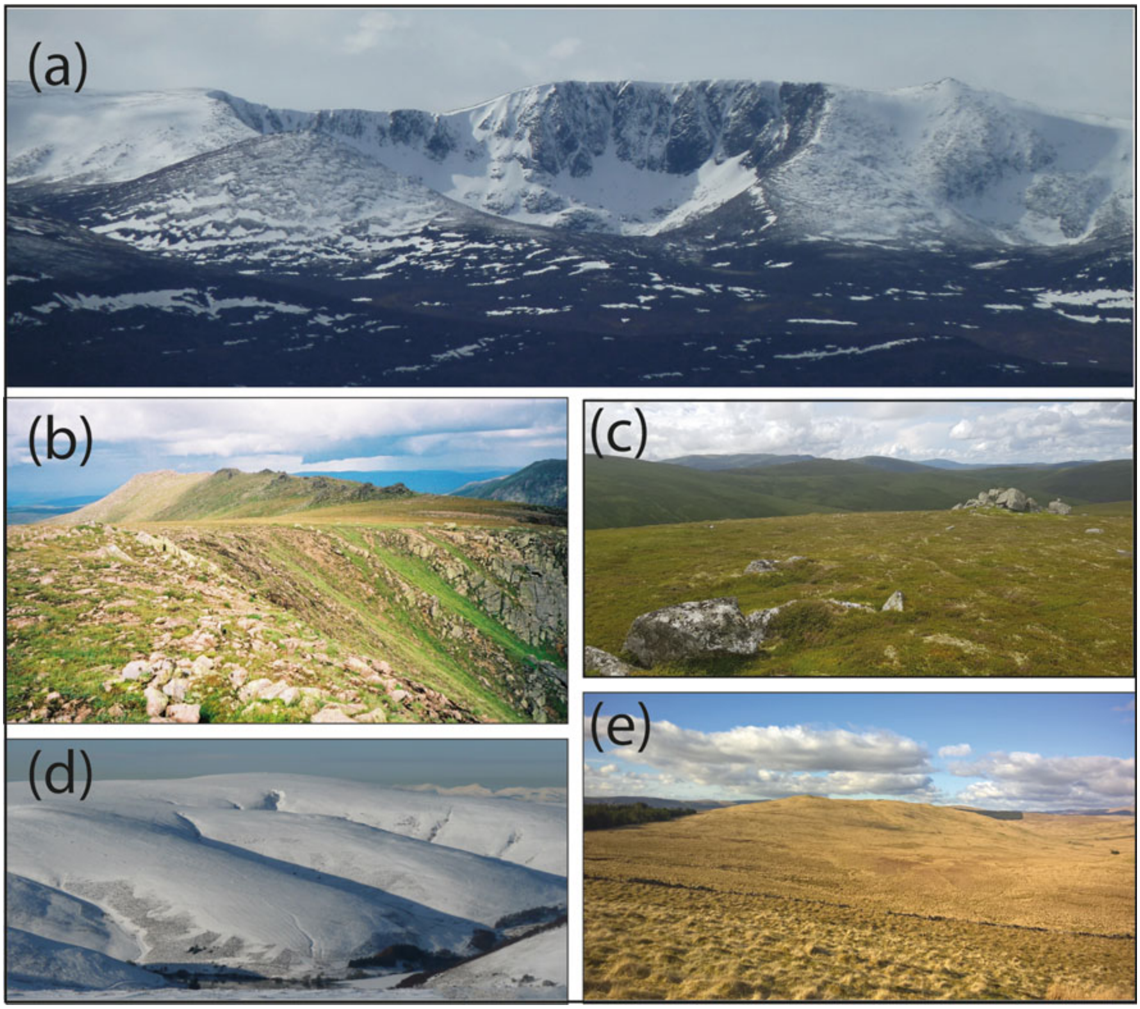

Fig. 4.4 Landscapes of relatively low-intensity glacial erosion. **a** Selective glacial erosion on the north flank of Lochnagar, Grampian Highlands, showing cirques cut into the granite plateau; Meikle Pap on the left is festooned with bouldery solifluction lobes. **b** Mountain-top regolith and granite tor landscape on Beinn a' Bhuird, Cairngorms. **c** Ridge-top tors and fluvial valleys in upper Strathdon. **d** The eastern Lammermuir Hills, Southern Uplands: smooth slopes and fluvial valleys cut in Silurian metasediments. **e** Undulating tableland of Silurian metasediments on the Tweed watershed, Southern Uplands. (Images: Adrian Hall)

Outer Hebrides and Assynt, glacial erosion has apparently been largely limited to the removal of pre-glacial regolith cover (Godard 1965), with deep erosion of basement rocks only in fracture-controlled basins and valleys (Krabbendam and Bradwell 2014). In contrast, mega-grooves up to 6 km long, 10–100 m wide and 5–15 m deep on the streamlined bed of a former ice-stream onset zone in Assynt (Bradwell et al. 2008b) and on the floors of The Minch (Bradwell and Stoker 2015a) and the Sea of the Hebrides (Dove et al. 2015) demonstrate highly effective glacial erosion of hard crystalline rock in zones of focused fast ice flow.

On land, depths of cumulative Pleistocene glacial erosion are greatest in the troughs of the western Highlands, particularly where convergent ice flow resulted in the excavation of deep rock basins. More than 50 trough-floor basins over 40 m deep occur in this region and the deepest (those occupied by Lochs Morar, Ness and Lomond) reach depths of 190–310 m. Similarly, several of the fjords that indent the western coastline descend to depths of over 100 m before shallowing at their outlets, and narrow straits (sounds) that focused ice flow (such as the Inner Sound between Skye and the mainland) exceed 200 m in depth. Even in the Midland Valley, where differential erosion of igneous and sedimentary rocks dominates the landscape (Fig. 4.3d), glacially excavated trenches and rock basins extend below valley floors, locally to depths below sea level (Kearsey et al. 2019).

4.2.4 Interglacial and Interstadial Processes

The nonglacial processes of weathering and erosion that operated in Scotland during and after the retreat of the last sheet (Sects. 4.7 and 4.8) can be considered representative of those that operated during interglacials and interstadials following the MPT. Prior to the expansion of the last ice sheet and again during ice-sheet retreat, permafrost developed down to sea level in terrain not occupied by glacier ice and periglacial processes of frost weathering, ice-wedge formation and solifluction affected extensive areas. That such processes also operated during the Early and Middle Devensian is demonstrated by stratigraphic evidence in England (Murton and Ballantyne 2017) and by the stratigraphic record at Kirkhill in Buchan (Sect. 4.2.2).

Retreat of the last ice sheet and its predecessors was accompanied by a range of paraglacial processes: failure or deformation of critically stressed rock slopes, accumulation of rockfall debris as talus, reworking of glacigenic sediment by rivers, debris flows and coastal erosion, and incision and terracing of glacigenic valley fills. Such paraglacial landscape modification was most rapid during the millennia immediately following deglaciation and declined thereafter as rock slopes stabilized and reserves of glacigenic sediment accessible to reworking processes declined (Ballantyne 2002, 2019b; Ballantyne et al. 2014). Paraglacial sediment stores that accumulated following each episode of deglaciation then contributed to the sediment entrained by subsequent glaciers and ice sheets. The only paraglacial landforms known to have survived at least one ice-sheet glaciation are the scars of major rockslides (Cave and Ballantyne 2016). The preservation of such features indicates that paraglacial rock-slope failure has, over multiple glacial-paraglacial cycles, made a major contribution to trough widening and the lateral extension of cirques (Jarman 2009; Ballantyne 2013a, 2019a; Chap. 14).

The role of weathering during successive interglacials in producing readily erodible soils and regolith that contributed to the sediment transported during subsequent periods of glacial expansion may have been considerable, particularly in the long periods of interstadial and interglacial conditions of the Early Pleistocene. In Buchan, where successive ice sheets were mainly cold-based and glacial erosion was minimal, pre-Quaternary and interglacial saprolites were preserved and frost-shattered rock occurs under glacial deposits; similar weathering covers were once widespread elsewhere but have been removed by glacial erosion.

Similarly, the blockfields and other forms of periglacial regolith that mantle summits and high plateaux provide evidence of long-term weathering of bedrock, mainly under periglacial conditions. Since the downwastage of the last ice sheet, only thin, immature covers of frost-weathered debris have accumulated on ice-scoured summits, implying that mature blockfields have evolved over more than a single glacial-interglacial cycle, possibly over timescales of 100 ka (Marquette et al. 2004). Preservation of Scottish blockfields under cold-based ice during at least the last (and probably earlier) ice-sheet glaciation(s) has been demonstrated by cosmogenic ^{10}Be exposure dating of erratics resting on blockfield surfaces (Fabel et al. 2012). Some periglacial regolith covers may have evolved throughout the Quaternary, with formation of detritus by weathering being balanced by gradual nonglacial erosion (Ballantyne 2010), though the most recent generation of extant blockfields probably formed during the Late Pleistocene (Hopkinson and Ballantyne 2014). Although the blockfields on Scottish mountains represent mainly breakdown of well-jointed rock by frost weathering, they contain assemblages of clay minerals, including gibbsite and kaolinite, that indicate long-term chemical alteration of blockfield debris (Ballantyne 1998a).

The most specular evidence for long-term subaerial weathering and erosion during the Quaternary is provided by the emergence of tors through preferential weathering and erosion of the surrounding surfaces. Terrestrial cosmogenic nuclide (TCN) data indicate that the granite tors of the Cairngorms and conglomerate tors in Caithness pre-date the last interglacial (Phillips et al. 2006; Ballantyne and Hall 2008); the oldest dated tor surface in the Cairngorms has a minimum modelled (exposure and burial) age of ~675 ka. Collectively, the TCN data obtained for the tors on the Cairngorms indicate that nonglacial denudation of rock surfaces and regolith covers operated at rates of 2.8–12.0 m Ma^{-1} during successive interglacials. Palaeosurfaces and high plateaux in Scotland are therefore not pristine pre-Quaternary landforms but experienced up to a few tens of metres of surface lowering under nonglacial conditions during the Quaternary.

4.3 Late Pleistocene

The Late Pleistocene comprises the last (Ipswichian) interglacial (MIS 5e; ~130–116 ka) and the last (Devensian) glacial stage, the latter being equivalent to the Weichselian glacial stage in western Europe. The Devensian is subdivided into the Early (MIS 5d–4; ~116–57 ka), Middle (MIS 3; ~57–31 ka) and Late Devensian (MIS 2; ~31–11.7 ka) substages. The Devensian was a period of marked climatic fluctuations. The synchronised Greenland ice-core oxygen isotope records for this period identify 26 stadial events and 25 intervening interstades (Rasmussen et al. 2014), though none of the latter indicate warming equivalent to that of interglacial periods. The most protracted cold periods identified in the ice-core record occurred during MIS 4 (~71–57 ka) and MIS 2; the latter coincided with the expansion of the last Scottish Ice Sheet (Sect. 4.4) and the former was probably also associated with development of an ice sheet in Scotland, though confirmation remains elusive. Briefer stadial events during MIS 5d–5a and MIS 3 were almost certainly accompanied by the development of ice caps and icefields in upland areas, and limited ice cover may have occupied high ground during some Devensian interstades. The total duration of glaciation in peripheral lowlands was comparatively brief, perhaps amounting to less than 25–30 ka of the last 130 ka.

4.3.1 The Last Interglacial and Devensian Interstades

The Late Pleistocene terrestrial stratigraphy of Scotland has been reviewed by Merritt et al. (2019). Stratigraphic evidence of last interglacial age is represented at a few sites where till-covered peat beds or palaeosols contain thermophilous pollen assemblages or other organic material that indicate a climate as warm as, or warmer, than now. These include sites at Fugla Ness in NW Shetland (Hall et al. 2002), Dalcharn east of Inverness (Walker et al. 1992), Teindland near Elgin (Hall et al. 1995) and Kirkhill in Buchan (Hall et al. 2019). Other till-covered palaeosols or peats have been attributed to formation during cool Early Devensian interstades (MIS 5c or 5a), notably at Sel Ayre in west Shetland, Allt Odhar southeast of Inverness and at various sites in Buchan. Weathering of basal tills of probable Middle Devensian age in northern Lewis, the Nairn valley, the central Grampians and Midland Valley has also been attributed to temperate conditions during the last interglacial or Early Devensian interstades (Merritt et al. 2019).

4.3.2 Early and Middle Devensian Glaciations

During the Early Weichselian (Early Devensian), an extensive ice sheet occupied Fennoscandia (Kleman et al. 1997). In areas that remained beneath cold-based ice in the Late Weichselian, Middle Weichselian glacigenic sediments and depositional landforms such as moraine ridges and eskers are widely preserved (Kleman et al. 2020). Similar ice-sheet growth is likely to have occurred in the British Isles, but few landforms and sediments in Scotland can be attributed with confidence to Early and Middle Devensian glaciations. At various sites, there are two or more till units of probable Devensian age. Some of these contain erratics or display fabrics indicating switches in ice-flow direction, and others are separated by periglacial facies or outwash sands and gravels (Merritt et al. 2017, 2019). Given a lack of dating evidence and the paucity of recognized Devensian interstadial organic sediments, it is uncertain whether stacked till units reflect deposition during successive Devensian glaciations, or episodic changes in the extent or direction of flow of the last ice sheet alone. In the Clyde and Ayrshire basins, however, the lowermost till underlies MIS 3 deposits and is possibly of MIS 4 age though probably older (Finlayson et al. 2010).

Some light is cast on the extent of Early and Middle Devensian glaciation by $^{230}Th/^{234}U$ disequilibrium ages for speleothem formation in limestone caves at the foot of mountains in Assynt (Lawson and Atkinson 1995). Speleothem growth is negligible under ice cover and ceases in permafrost. A cluster of 15 overlapping ages for the period ~95–75 ka (MIS 5c–5a) implies mainly glacier- and permafrost-free conditions within this interval. Conversely, MIS 4 is represented by just two ages, both with large uncertainties (63.6±6 ka and 56±13 ka), suggesting that part or all of Assynt was ice-covered during this period. Lithological evidence for an Early Devensian Scottish ice sheet is also present in a deep-ocean core from the Rockall Trough, 300 km west of Scotland, which appears to indicate expansion of an ice sheet across the Hebrides Shelf during MIS 4 (Hibbert et al. 2010). Much research remains to be done to unravel the terrestrial environmental history of the Early and Middle Devensian in Scotland.

4.3.3 The Late Devensian

The Late Devensian (≈ Late Weichselian) substage is temporally equivalent to MIS 2 (31–11.7 ka) and encompasses Greenland Stadials 1–5 (Rasmussen et al. 2014). In Great Britain, it is subdivided into the Dimlington Stade (~31–14.7 ka), the Lateglacial (or Windermere) Interstade (~14.7–12.9 ka) and the Loch Lomond (≈Younger Dryas) Stade (~12.9–11.7 ka). The Late Devensian incorporates the advance of the last Scottish Ice Sheet to its maximum extent, subsequent shrinkage of the ice sheet, and readvance of glaciers during the Loch Lomond Stade following complete (or nearly complete) disappearance of glacier ice from Scotland during the intervening Lateglacial Interstade.

4.4 The Last Scottish Ice Sheet

The last Scottish Ice Sheet (SIS) was the dominant component of the British-Irish Ice Sheet (BIIS) and for much of its existence reached far beyond the present land area of Scotland. During the Last Glacial Maximum, it extended westward and northwestward across the Atlantic shelf, met the Fennoscandian Ice Sheet (FIS) in the North Sea Basin and was confluent to the south with ice nourished in England, Wales and Ireland

(Fig. 4.5). The ice sheet exhibited highly dynamic behaviour during its lifetime, with radical changes in its configuration and flow patterns, periods of sustained ice streaming and multiple readvances that interrupted its retreat. Ballantyne and Small (2019) have reviewed the pattern and chronology of the expansion and retreat of the SIS and a map and comprehensive database of over 170,000 glacial landforms associated with the BIIS have been published by Clark et al. (2017).

Fig. 4.5 The maximum extent of the last British-Irish Ice Sheet, showing the location of independent centres of ice dispersal and generalized directions of offshore ice movement when the ice sheet reached its maximum extent

4.4.1 Ice-Sheet Expansion and Maximum Extent

The presence of ice-rafted detritus of Scottish provenance in deep-water cores from the Atlantic Ocean indicates that marine-terminating glaciers were at least intermittently present in Scotland during ~43–35 ka (Hibbert et al. 2010), suggesting that an icefield or ice cap of fluctuating extent occupied the Scottish Highlands during most or all of this period. Subsequent expansion of the SIS is recorded by radiocarbon ages for organic material buried under till. These imply that Highland ice did not expand across adjacent lowlands until ~35 ka and possibly not until ~32 ka (Bos et al. 2004; Brown et al. 2007). Similarly, radiocarbon ages obtained for shells in marine sediments overlain by till in the northern North Sea Basin imply ice-free conditions until ~34 ka or later (Graham et al. 2010).

A generalized pattern of ice-sheet expansion has been proposed by Hughes et al. (2014) through the identification of sequential (cross-cutting) flowsets represented by the alignment of glacial bedforms and meltwater channels. Stage 1 in their reconstruction (Fig. 4.6) approximates conditions around 35–32 ka, when ice enveloped the Midland Valley. Stage 2 depicts build-up of an ice divide over SW Scotland and extension of Scottish ice across Ireland and northern England. By stages 3 and 4, thickening and expansion of ice nourished in the Southern Uplands and Ireland had created an ice divide between SW Scotland and Ireland, diverting Scottish ice westward across the Atlantic shelf. Stage 5 depicts the SIS terminating westward at or near the Atlantic shelf break and eastwards confluence with the Fennoscandian Ice Sheet; it represents the maximum extent of ice sourced in Scotland. Southward ice expansion, however, continued during stage 6, by which time ice margins in the north had begun to retreat.

More detailed local evidence complements this general model. In central and southern Scotland, the evidence provided by sequential flowsets and erratic dispersion indicates that after ~35 ka ice from the SW Highlands moved eastwards across the Midland Valley and southwards down the Firth of Clyde, eventually meeting ice advancing from the Southern Uplands and extending into the Irish Sea Basin and up to 200 km into Ireland (Greenwood and Clark 2009; Finlayson et al. 2010, 2014). Subsequent thickening of the Southern Uplands ice mass led to its confluence with the Irish Ice Sheet and formation of a persistent ice divide between Scotland and Ireland, re-routing ice from the SW Highlands westwards across the Malin Shelf. Only ice from the Galloway Hills in SW Scotland continued to flow south into the Irish Sea Basin, where it joined ice from England, Wales and Ireland to feed the Irish Sea Ice Stream. Development of a north–south ice divide east of the Galloway Hills caused ice from the rest of the Southern Uplands to be evacuated eastwards via the Firth of Forth, Tweed valley and Tyne valley into the North Sea Basin, where it was diverted southwards to impinge on the east coast of England (Livingstone et al. 2012; Busfield et al. 2015).

In the west of Scotland, the evidence provided by striae and erratics indicates that the islands of Skye, Mull and Arran developed ice caps that diverted the flow of ice from the mainland and probably remained centres of ice dispersal throughout the lifetime of the SIS (Chap. 10). Similar evidence indicates that an independent ice cap formed and persisted over the Outer Hebrides, feeding ice southwestwards to the shelf edge via the Hebrides Ice Stream (Barra Fan Ice Stream), westwards across the Hebrides Shelf and northeastwards into the Minch Ice Stream (Bradwell and Stoker 2015a; Dove et al. 2015; Callard et al. 2018; Bradwell et al. 2019; Fig. 4.5). Termination of the Barra Fan Ice Stream at the shelf break is indicated by glacimarine deposits on the Barra-Donegal Fan, a major depocentre abutting the Malin Shelf. Farther north on the Hebrides Shelf, however, the ice sheet failed to reach St Kilda, 40–60 km east of the shelf break (Ballantyne et al. 2017; Chap. 9). The maximum reach of the Minch Ice Stream is uncertain: some accounts place its terminus at the shelf edge, but others propose mid-shelf termination (Bradwell and Stoker 2015a; b; Bradwell et al. 2019).

The evolution of ice movement across Caithness, Orkney and Shetland was complex. In Caithness, initial ice flow was northeastwards and was succeeded by: (i) retreat of the ice margin to the north coast; (ii) northwestwards movement of ice from the Moray Firth; (iii) a second retreat of the ice margin from the north coast; and (iv) persistent northwestwards flow of Moray Firth ice towards the Atlantic shelf edge (Hall et al. 2011; Hall and Riding 2016; Merritt et al. 2019; Chap. 8). Analysis of tills on Orkney suggests that initial northwards ice movement was succeeded by northwesterly ice flow across the islands towards the shelf edge (Hall et al. 2016a). Interpretation of the pattern of ice movement on Shetland has polarised around two views: (i) that the archipelago was initially over-ridden by ice moving northwestwards from the North Sea Basin, but later became an independent centre of ice dispersal; and (ii) that Shetland supported an independent ice cap throughout the lifetime of the SIS. Assessment of the field evidence by Hall (2013) supports the second interpretation, as does reconstruction of the pattern of ice retreat in the northern North Sea Basin and across the Orkney-Shetland Platform (Bradwell et al. 2020; Chap. 7).

The pattern of ice flow across eastern Scotland from the Moray Firth to Angus was also complex. From the evidence provided by striae, erratic transport, lithostratigraphy and sequential flowsets, Merritt et al. (2017, 2019) inferred that an initial southeastwards ice flow from the Moray Firth

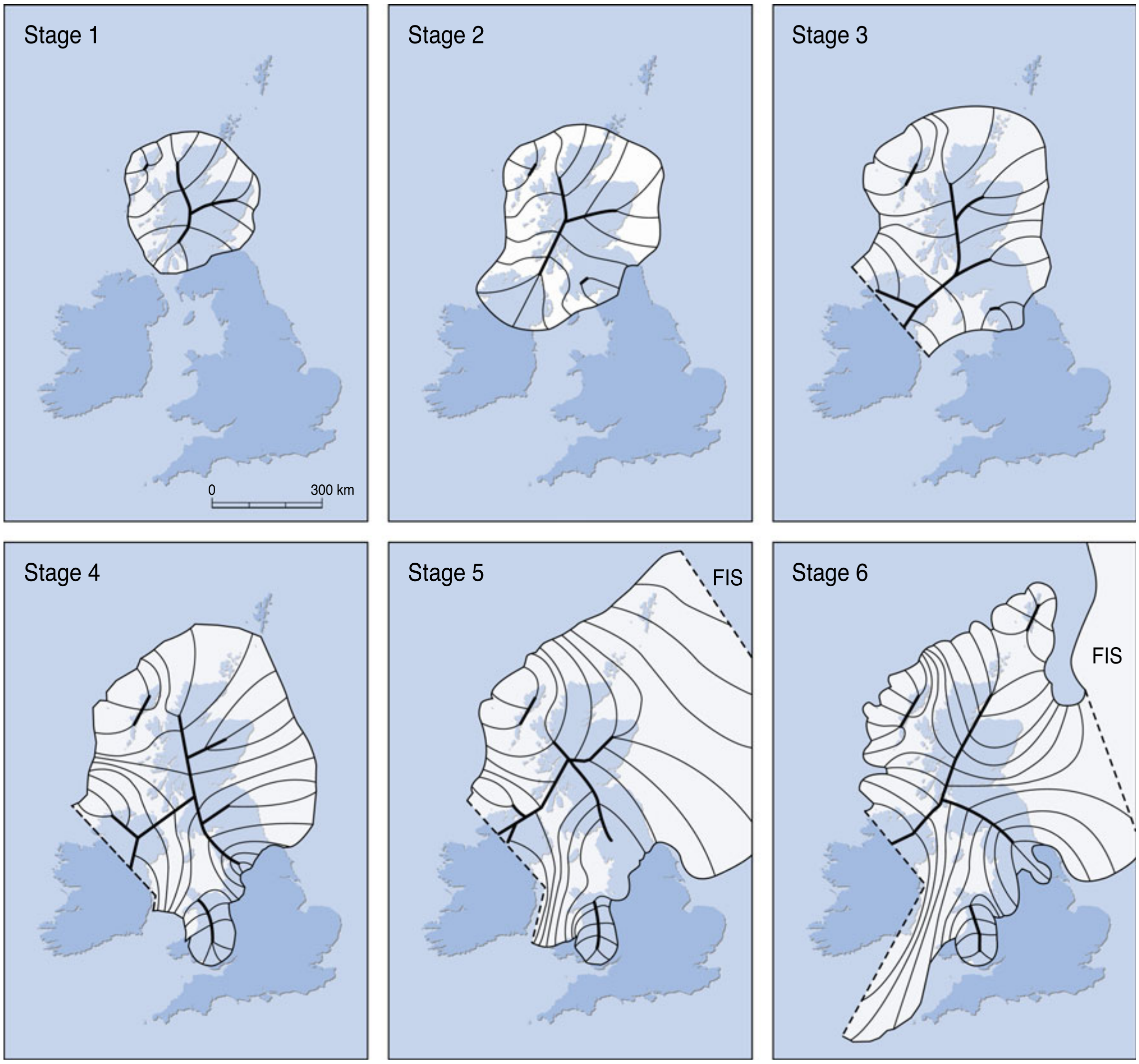

Fig. 4.6 Stages in the growth of the last British-Irish Ice Sheet reconstructed by Hughes et al. (2014). Thick lines represent major ice divides and thin lines represent flowlines interpreted from flowsets and offshore landforms. Smaller centres of ice dispersal are not represented. (Adapted from Hughes et al. (2014) Quat Sci Rev 89:148–169 © 2014 Elsevier Ltd.)

across Buchan was confluent with ice moving eastwards and southeastwards from the Eastern Grampians. At a later stage, ice flowing east from the Moray Firth apparently curved northwestwards across Caithness and Orkney, and ice flow across much of Buchan, Angus and Fife was subsequently dominated by eastwards to southeastwards movement of ice from the Grampians. To explain the switch in the direction of flow of ice from the Moray Firth, Merritt et al. (2017) invoked eastward migration of a major ice divide during ice-sheet build-up, from a westerly axis between Caithness and Shetland, to one in the central part of the northern North Sea Basin, arguing that the latter is necessary to explain northwesterly movement of Moray Firth ice across Caithness and Orkney as the ice sheet expanded.

Radiocarbon ages obtained for marine fauna from sediment cores indicates that all of the central and northern North Sea Basin was over-run by confluence of the SIS with the FIS by ~27 ka (Hughes et al. 2016). There are, however, conflicting interpretations regarding the position of the suture zone, the location of the major ice divide(s) and the changing directions of ice movement in this region. One interpretation is that a broad ice divide eventually became established between NE Scotland and SW Norway, driving ice movement northwestwards across Caithness and Orkney and southeast towards the southern North Sea Basin (Sejrup et al. 2016; Bradwell et al. 2020; Fig. 4.5). Progressive eastward migration of an ice divide to a similar position (Merritt et al. 2017) represents a refinement of this model with the advantage of explaining the radical shift in the direction of flow of ice emanating from the Moray Firth.

The maximum extent of the BIIS is well-constrained for most sectors (Fig. 4.5), though uncertainties remain regarding the limit of grounded ice west of the Outer Hebrides. The SIS had a complex and shifting surface configuration. The dominant ice centres were those located over the Scottish Highlands, Southern Uplands and over or east of the Orkney-Shetland Platform, but the associated ice divides migrated during ice-sheet expansion, causing shifts in ice-flow directions. Smaller independent centres of ice dispersal (ice domes) persisted over the Outer Hebrides, the Cairngorms, the Galloway Hills and the islands of Skye, Mull and Arran. TCN dating of erratics resting on blockfields on mountains in the NW Highlands, and of tors in the Cairngorms, has demonstrated that the last ice sheet over-topped all of the highest ground in Scotland and that the ice was cold-based and frozen to the underlying substrate over most summits and plateaux (Phillips et al. 2006; Fabel et al. 2012; Hopkinson and Ballantyne 2014). Hughes et al. (2014) calculated that the altitude of the ice divide over northern Scotland must have exceeded ~1400 m and models based on inferred glacio-isostatic depression indicate ice-surface altitudes of up to 2500 m.

The advance of the BIIS to its maximum extent was asynchronous. On the Shetland shelf, the ice sheet probably achieved its maximum extent at ~26–25 ka, and the Minch Ice Stream appears to have reached its limit by ~30.2 ka and was retreating by ~27.5 ka (Bradwell et al. 2019, 2020). The Hebrides Ice Stream reached the Atlantic shelf break at ~26.7 ka and withdrew from the shelf edge by ~25.9 ka (Callard et al. 2018). Advance of the grounded ice margin to its limit west of Ireland occurred after ~24.7 ka. TCN and optically stimulated luminescence (OSL) dating suggest that the southern margin of the Irish Sea Ice Stream impinged on the Scilly Isles at ~26–25 ka (Smedley et al. 2017a), but other evidence suggests that it achieved its maximum southerly reach at the edge of the Celtic Shelf between ~24.3 and ~23.0 ka (Chiverrell et al. 2013). The lobe that extended down the east coast of England may not have achieved its maximum extent until ~21.6 ka (Bateman et al. 2017).

4.4.2 Ice Streams and Ice-Sheet Modelling

Ice streams are corridors of fast-flowing ice that discharge much of the ice from ice sheets. Palaeo-ice streams that drained the SIS have been identified from a range of diagnostic features, notably convergent flow patterns, usually sourced in upland areas, a distinct flow track with well-defined lateral margins and markedly elongate bedforms (Fig. 4.7).

On the basis of such evidence, four east-flowing ice streams that drained the SIS have been identified: the Tyne Ice Stream, which evacuated ice from the Southern Uplands, Cheviots, Lake District and Pennines; the Tweed Ice Stream, which flowed eastwards between the Lammermuir and Cheviot Hills; the Strathmore Ice Stream, which flowed eastwards between the Highland edge and NE Fife; and the Moray Firth Ice Stream, a major artery that drained ice from the Great Glen towards the North Sea Basin (Everest et al. 2005; Golledge and Stoker 2006; Livingstone et al. 2015; Merritt et al. 2017). In the western North Sea Basin, coalescence of ice from the Forth and Tweed valleys (and probably the SE Grampians) fed the North Sea Lobe, a major ice stream that moved southwards along the east coast of England to terminate on the north coast of East Anglia (Busfield et al. 2015; Bateman et al. 2017; Fig. 4.5).

To the west of Scotland, ice from the Galloway Hills fed the Irish Sea Ice Stream, which drained at least 17% of the BIIS and terminated at the edge of the Celtic Shelf (Praeg et al. 2015; Fig. 4.5). Farther north, the Hebrides Ice Stream was fed by ice from the Western Highlands and Hebrides that flowed southwestwards across the Sea of the Hebrides and ice flowing westwards from northern Ireland and SW Scotland (Howe et al. 2012; Finlayson et al. 2014; Dove et al. 2015; Callard et al. 2018). Ice from NW Scotland, Skye and eastern Lewis fed the Minch Ice Stream, which drained an area of 10,000–15,000 km^2 and followed a 40–50 km wide trough that extends northwestwards to terminate at the Sula Sgeir Fan (Bradwell and Stoker 2015a; Bradwell et al. 2019; Fig. 4.8).

It is not known whether the ice streams draining the SIS operated throughout much of its existence or only at particular times. Numerical models of the expansion and contraction of the BIIS driven by a scaled NGRIP oxygen-isotope record indicate that the ice sheet experienced 'binge and purge' cycles, with prolonged phases of thickening of predominantly cold-based ice alternating with 'purge' episodes triggered by rapid warming (Hubbard et al. 2009). These models suggest that purge phases were accompanied by the development of ice streams that drew down the ice-sheet surface over its source areas. The best-fit model of Hubbard et al. (2009) generated transient but recurrent ice streaming at all the Scottish ice-stream locations described above. Ice-stream activation probably played a major role in ice-divide migration on land: rapid evacuation of ice by the Minch Ice Stream, for example, may have driven eastward relocation of the ice divide across much of the Northern Highlands.

4.4.3 Deglaciation of the Offshore Shelves

The pattern of deglaciation on the shelves adjacent to Scotland has been reconstructed from bathymetric data depicting seafloor moraines and grounding-zone wedges, complemented by acoustic profiles of sediment facies and depths (Bradwell et al. 2008a, 2020; Clark et al. 2012;

Fig. 4.7 Megagrooves in the onset zone of the Tweed Ice Stream. Ice flow was from bottom left (west) to top right. (Image: Google Earth™)

Bradwell and Stoker 2015b; Callard et al. 2018; Fig. 4.9; Chap. 6). The offshore moraines delimit readvances that interrupted ice-margin retreat and the grounding-zone wedges represent locations where retreat of grounded ice margins slowed, paused or oscillated, depositing thick ridges of sediment, often tens of kilometres long and sometimes several kilometres wide.

For the three major ice streams that drained ice from the southwest and west of Scotland, the timing of ice-margin retreat is constrained by radiocarbon ages of shell fragments and foraminifera in sediment cores, complemented by TCN and OSL ages from terrestrial sites. Rapid retreat of the Irish Sea Ice Stream after ∼24 ka (Smedley et al. 2017b) slowed as the ice margin became pinned against the Isle of Man in the northern Irish Sea Basin at ∼20.8–20.2 ka (Chiverrell et al. 2018). Within the interval 19.3–18.3 ka, retreat was interrupted by a readvance of ice from the Solway Firth (the Scottish Readvance), after which the ice margin underwent oscillatory retreat towards its sources in the Galloway Hills of SW Scotland, where TCN ages record deglaciation at ∼15.2 ka (Ballantyne et al. 2013). Retreat of the Hebrides Ice Stream was underway by ∼25.9 ka, the outer Hebrides Shelf was ice-free by ∼23.2 ka and glacimarine conditions were present in the Sea of the Hebrides by ∼20.2 ka (Callard et al. 2018). Thereafter, retreat slowed as streaming behaviour ceased and the ice margin oscillated over a ∼3 ka period: Barra was deglaciated at ∼17.1 ka, south Harris by ∼17.3 ka, southern Skye at ∼17.6 ka, SW Mull at ∼17.5 ka and western Jura at ∼16.6 ka (Small et al. 2017; Fig. 4.10). The chronology of retreat of the Minch Ice Stream is based mainly on TCN exposure ages and inferred connections with a sequence of 17 grounding-zone wedges that straddle The Minch and the trough followed by the ice stream towards the shelf edge (Bradwell et al. 2019). TCN ages from North Rona suggest that retreat of the ice margin was underway before ∼27.5 ka and by ∼23.3 ka the ice margin lay at the northern exit of The Minch. After ∼18.5 ka, retreat was apparently

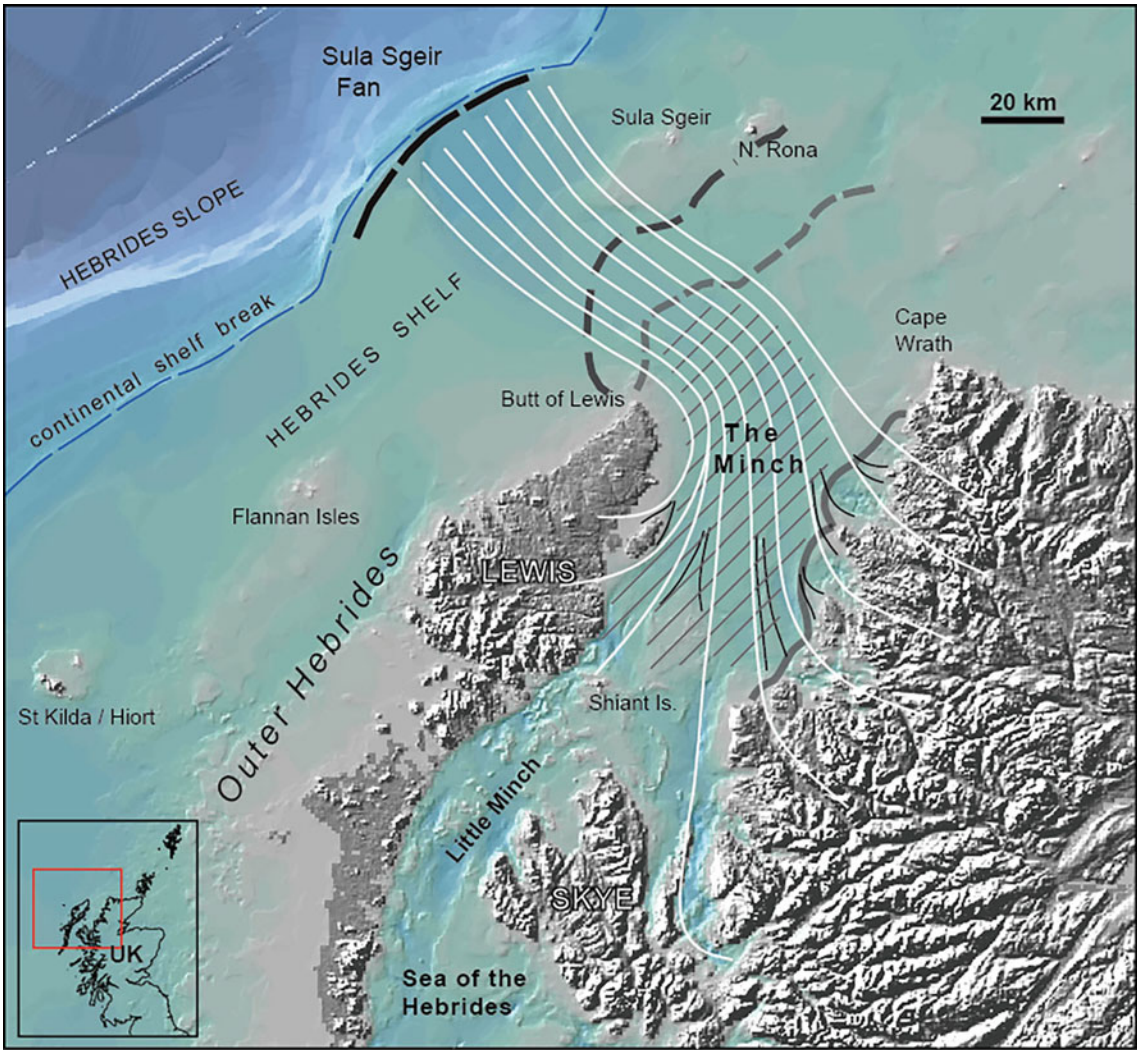

Fig. 4.8 The track of the Minch Ice Stream. White lines are ice-stream flowlines, the thick lines are terminus positions and the hatching indicates an area of streamlined subglacial bedforms and iceberg scours. (From Ballantyne and Small (2019) Earth Env Sci Trans R Soc Edinb 110:93–131 © 2018 The Royal Society of Edinburgh)

accelerated by opening of a calving bay east of the Outer Hebrides, before the ice margin became grounded on Skye and much of the northwest mainland by 16.2–15.4 ka.

The retreat of all three western ice streams therefore occurred under stadial conditions prior to the onset of the Lateglacial Interstade (~14.7 ka) and may have been initiated by enhanced calving losses at marine-terminating ice margins due to rising relative sea levels caused by glacio-isostatic depression of the shelf. Rates of ice-stream retreat were also affected by subglacial topography, accelerating over reverse bed slopes and in deep troughs or basins, and slowing at topographic constrictions. Bradwell et al. (2019) also associated accelerated retreat of the Minch Ice Stream after ~18.5 ka with a change from a 'soft' sediment-covered bed to a 'hard' bed underlain mainly by bedrock.

North of the Scottish Mainland, the pattern of seafloor moraines implies progressive decoupling of mainland ice from a major ice cap centred on the Orkney-Shetland Platform (Bradwell and Stoker 2015b; Bradwell et al. 2020; Fig. 4.9). TCN exposure ages averaging ~23.8 ka for Cape Wrath date the onset of decoupling, and others averaging ~17.5 ka for Dunnet Head on the north coast of Caithness suggest prior severance of the Orkney-Shetland ice centre from the mainland (Fig. 4.10). TCN ages suggest deglaciation of Orkney by ~16.5 ka, isolation of the remnant Shetland ice cap by ~17.0–16.5 ka and complete deglaciation of Shetland within the period 16.5–15.0 ka (Phillips et al. 2008; Bradwell et al. 2020).

Two scenarios have been proposed for severance of the SIS from the FIS in the North Sea Basin. Several accounts suggest that this occurred through southwards extension of a

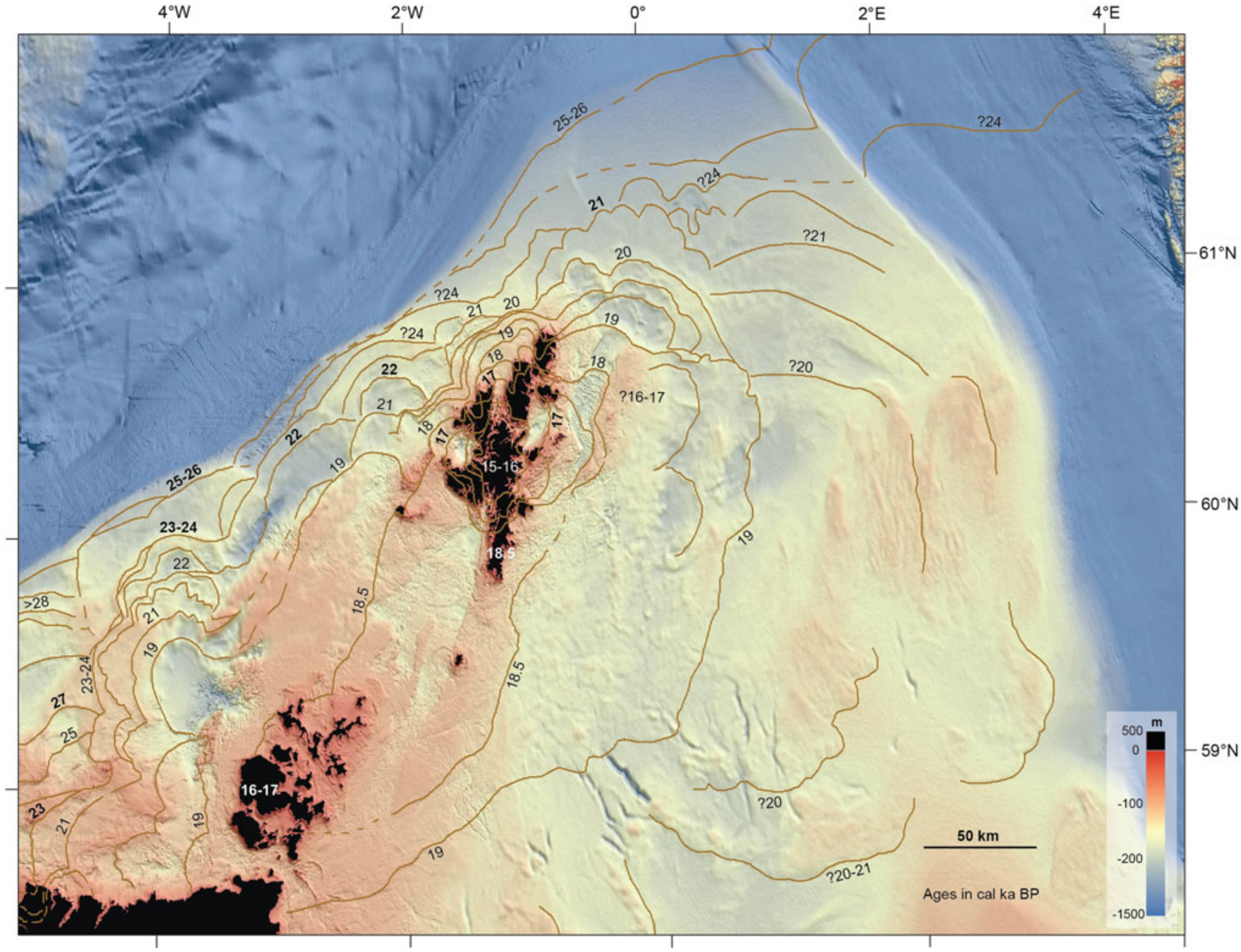

Fig. 4.9 Reconstruction of the pattern and timing of deglaciation in the northernmost sector of the last Scottish Ice Sheet. Ice margin positions (brown lines) are based on the available geomorphological and geological evidence and the numbers are ages (ka). The bold numbers represent the most securely dated ice-margin positions. (From Bradwell et al. (2020) J Quat Sci doi:10.1002/jqs.3163 © The authors)

marine embayment from the shelf edge north of Shetland (Bradwell et al. 2008a; Clark et al. 2012; Merritt et al. 2017). Sejrup et al. (2016) have argued that rapid retreat of the Norwegian Channel Ice Stream (the dominant routeway discharging ice northwards off the west coast of Norway) resulted in decoupling of the two ice sheets at ~18.5 ka and was followed by an eastward readvance of Orkney-Shetland ice. Bradwell et al. (2020) inferred earlier (~23–21 ka) severance of the SIS from the FIS in this sector and subsequent major expansion of the ice cap centred on the Orkney-Shetland Platform. Farther south in the North Sea Basin, there is evidence for at least two later readvances of Orkney-Shetland or Moray Firth ice, at ~17.5 ka and ~16.2 ka (Sejrup et al. 2015). Moraine ridges and grounding-zone wedges indicate oscillation of the margin of Moray Firth ice during its subsequent retreat (Graham et al. 2009).

4.4.4 Deglaciation of Western Scotland

Multiple TCN and radiocarbon ages permit the pattern and timing of deglaciation to be established for western Scotland. The timing of deglaciation at key sites on the Scottish Mainland and the Hebrides is summarized in Figs. 4.10 and 4.11.

In NW Scotland, TCN exposure ages indicate that the ice margin had retreated to the fjords and peninsulas of the west coast by ~16.5–16.0 ka and that mountain summits had emerged from the downwasting ice sheet by ~16.0 ka (Fabel et al. 2012); recessional moraines on the floors of fjords indicate oscillatory retreat of the ice margin in this area (Stoker et al. 2006). Retreat of the ice margin across land was interrupted by the Wester Ross Readvance, which is represented by moraines that can be traced across the peninsulas of Wester Ross from Applecross to Achiltibuie.

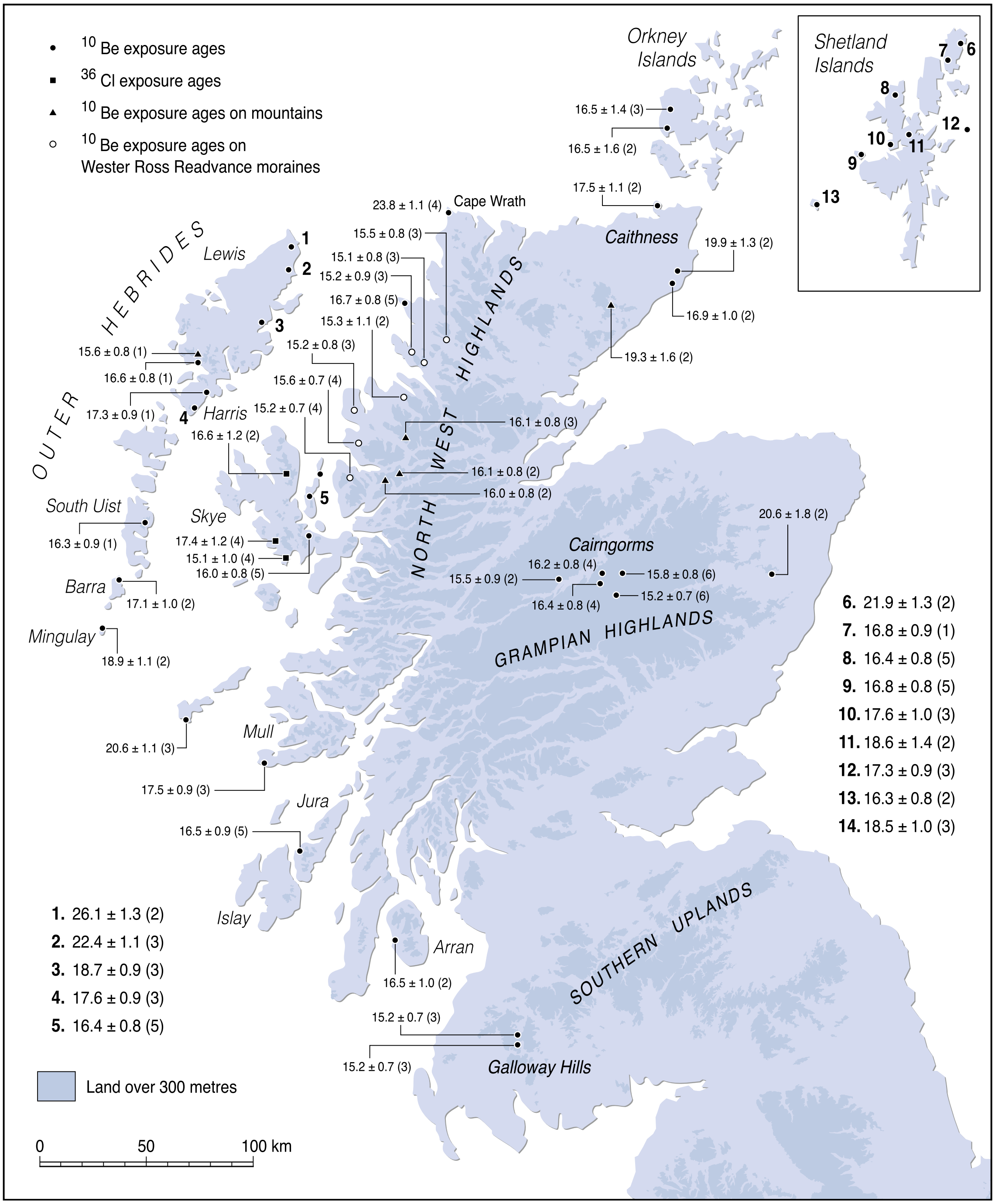

Fig. 4.10 Terrestrial cosmogenic nuclide exposure ages indicating the timing of deglaciation. The dates shown represent uncertainty-weighted means. Statistical outliers are excluded. Each age is followed by the full (± 1σ) uncertainty and the number of individual ages in the sample is shown in brackets. (Based on Ballantyne and Small (2019) Earth Env Sci Trans R Soc Edinb 110:93–131 © 2018 The Royal Society of Edinburgh, updated from Bradwell et al. 2019, 2020)

TCN ages for these moraines indicate that the readvance occurred at 15.3±0.7 ka (Ballantyne and Stone 2012; Ballantyne and Small 2019), possibly in response to cooling evident in the Greenland ice-core oxygen isotope record that culminated at ~15.7–15.3 ka (Fig. 4.11; Chap. 13). Similarly, moraines on the island of Soay that delimit a readvance of the Skye ice cap have yielded a mean TCN age of ~15.2 ka, suggesting that the associated readvance (the Loch Scavaig Readvance) also represents a response to cooling at this time (Small et al. 2016; Chap. 10). South of Skye, the margin of mainland ice had backstepped from most of the Inner Hebrides by ~16.0 ka, by which time the westernmost mainland peninsulas and mountain summits were emerging from the retreating ice sheet (Small et al. 2017). By ~15.0 ka, mainland ice had probably retreated from the coastline of western Scotland, but the subsequent deglacial history of this zone cannot be traced, as the fjords and valleys of western Scotland were reoccupied by glaciers during the Loch Lomond Stade.

The pattern of deglaciation in the Firth of Clyde and adjacent areas has been reconstructed by Finlayson et al. (2010, 2014), who inferred that at ~16.5 ka all of this region was still covered by ice flowing down the Firth of Clyde to impinge on NE Ireland, where there is evidence of a readvance of Scottish ice (the East Antrim Coastal Readvance) sometime after ~17.3–16.6 ka (Ballantyne and Ó Cofaigh 2017). Subsequent retreat resulted in decoupling of Highland and Southern Uplands ice and deglaciation of much of Ayrshire, though a glacier fed by ice from the SW Highlands continued to occupy the Firth of Clyde. The timing of retreat of the Clyde glacier is uncertain, but the inner Clyde estuary and the Glasgow area were almost certainly deglaciated by ~14.7 ka, at the onset of the Lateglacial Interstade (Peacock et al. 2012).

4.4.5 Deglaciation of Eastern and Southern Scotland

During ice-sheet retreat, northwestwards flow of ice from the Moray Firth across Caithness was succeeded by northeastwards ice movement from the mountains of Sutherland (Hall and Riding 2016). TCN ages obtained for sites in Caithness are inconsistent (Fig. 4.10), but the balance of evidence suggests that retreat of Moray Firth ice occurred within the period 17.5–17.0 ka and was succeeded by eastwards expansion of ice from high ground across western Caithness.

In NE Scotland there was major reorganization of ice flow during the period 22–19 ka: ice flowing eastwards from the Grampians and Cairngorms appears to have met ice flowing southeast from the Moray Firth and ice flowing northeastwards from Strathmore. Merritt et al. (2017) concluded that initial deglaciation on land occurred close to the confluence of these three ice masses near Peterhead, where radiocarbon dating of a shell in raised marine silts indicates deglaciation prior to ~17.7 ka. There is evidence for readvances of Moray Firth ice near Elgin and Inverness before the ice front retreated into Loch Ness, then open to the sea (Chap. 15). Recessional moraines in northern Loch Ness indicate punctuated retreat of a grounded ice margin, followed by deposition of a large moraine midway down the loch (Turner et al. 2012).

The timing of deglaciation of the eastern coast of Scotland is controversial. Radiocarbon ages for foraminifera in raised marine muds near Montrose imply very early deglaciation (21–20 ka; McCabe et al. 2007) but are at variance with other evidence indicating much later deglaciation (Ballantyne and Small 2019). Radiocarbon dating of marine shells in ice-proximal estuarine sediments implies deglaciation of the Firth of Tay shortly before ~16.3 ka (Peacock 2003). There is stratigraphic evidence that retreat of the Tay glacier was interrupted by a readvance (the Perth Readvance), though the timing and extent of this event are uncertain. In the eastern Highlands, TCN exposure ages for boulders on moraines indicate retreat of Strathspey ice along the northern flank of the Cairngorms within the period ~16.5–15.5 ka (Hall et al. 2016b; Chap. 18) and are consistent with TCN and radiocarbon ages indicating retreat of ice to the upper Spey valley by ~15.5 ka (Ballantyne and Small 2019; Fig. 4.10).

The timing of ice retreat in southern and SE Scotland has not been established, though a TCN deglaciation age of ~15.2 ka obtained for the Galloway Hills (Ballantyne et al. 2013) suggests that much of the Southern Uplands was deglaciated by that time. Radiocarbon ages from sites near Callander on the Highland boundary imply that all of the central and eastern Midland Valley was deglaciated before ~15.2 ka (Fig. 4.11).

4.4.6 Ice-Sheet Demise

The Greenland ice-core oxygen isotope record provides evidence for rapid warming at ~14.7 ka, the start of the Lateglacial Interstade in Scotland. This warming is captured by subfossil chironomid assemblages at various sites; these indicate a rapid rise in mean July temperatures to 11–13 °C early in the interstade (Brooks and Birks 2000; Brooks et al. 2012, 2016; Fig. 4.11). Rapid warming at this time is generally attributed to northward migration of the North Atlantic Polar Front, resumption of thermohaline circulation in the NE Atlantic and increased penetration of warm airmasses across Scotland.

The chronological evidence outlined above and in Figs. 4.10 and 4.11 suggests that by 15.0–14.7 ka, remnants of the SIS were confined to the Western Grampians and NW Highlands, and thus lay within the limits of the glaciers that formed in Scotland during the Loch Lomond Stade. There has been recurrent debate as to whether glacier ice persisted in the Scottish Highlands during the Lateglacial Interstade (~14.7–12.9 ka). Definitive evidence is elusive. The modelling experiments of Hubbard et al. (2009) suggest that limited ice cover may have survived interstadial warming in the western Highlands and Finlayson et al. (2011) have argued that a thin ice cap survived the interstade on the Beinn Dearg massif in the NW Highlands. Conversely, the TCN ages of erratics on summits in Wester Ross demonstrate that these mountains became ice-free at ~16.0 ka and remained so until the present (Fabel et al. 2012). Ice-rafted debris in a

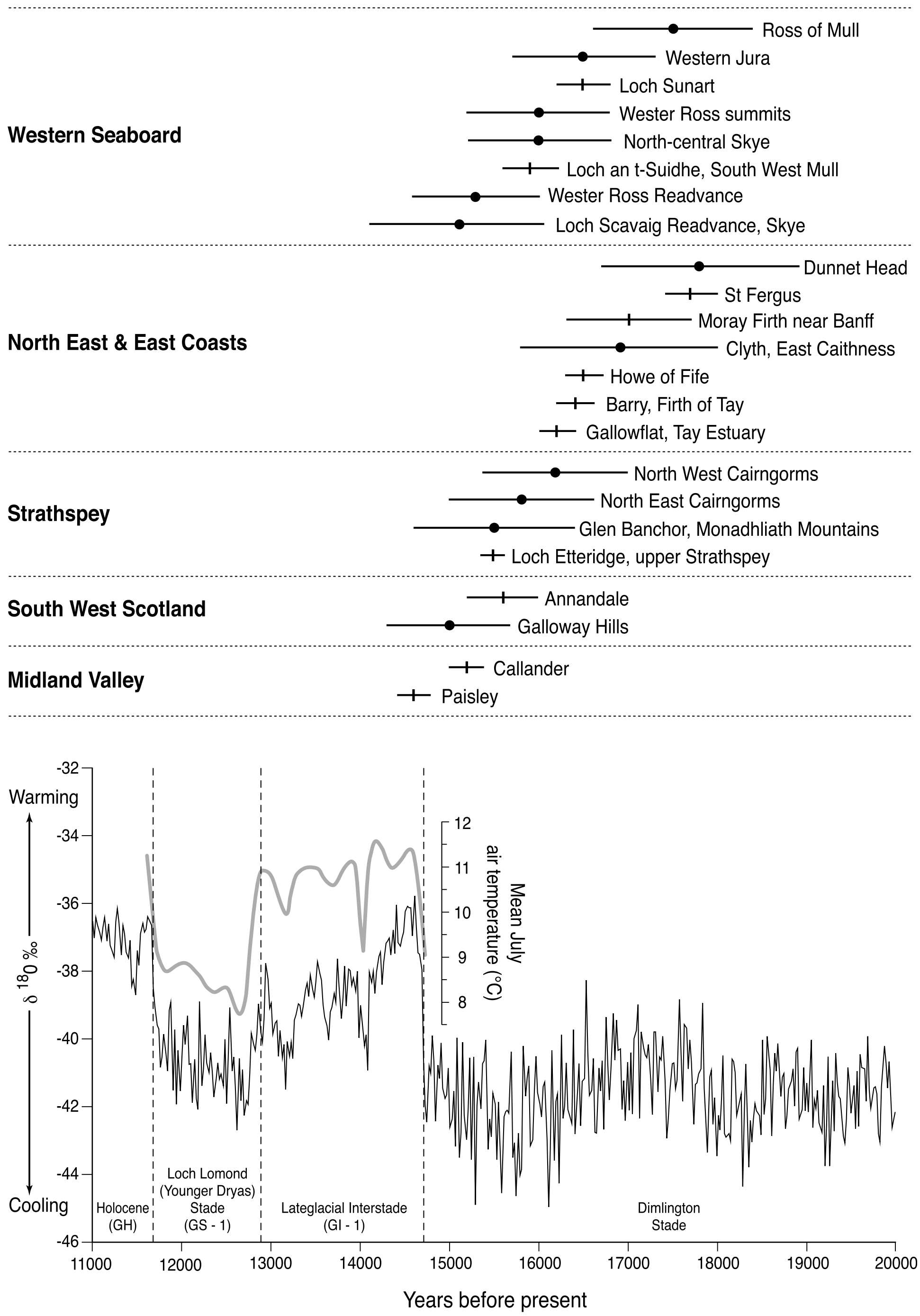

Fig. 4.11 Key TCN and calibrated radiocarbon ages constraining the deglaciation of mainland Scotland and the Inner Hebrides, plotted above the NGRIP ice-core $\delta^{18}O$ record for 20–11 ka (Rasmussen et al. 2014) and mean July temperatures inferred from chironomid assemblages in SE Scotland (Brooks and Birks 2000). Dots are TCN ages and vertical dashes are radiocarbon ages; horizontal lines are $\pm 1\sigma$ uncertainties. TCN ages represent the approximate timing of deglaciation; radiocarbon ages are minimal for the timing of deglaciation. (From Ballantyne and Small (2019) Earth Env Sci Trans R Soc Edinb 110:93–131 © 2018 The Royal Society of Edinburgh)

core from the Hebrides shelf has been dated to ~14.1–13.9 ka, but it is uncertain whether this is of distal (Laurentian) provenance or whether it indicates marine-terminating glaciers in the western Highlands at this time (Small et al. 2013a; b). Given the rapid warming at the onset of the Lateglacial Interstade, it is likely that glaciers occupying the glens and fjords of the Highlands experienced rapid retreat after ~14.7 ka. Survival of high plateau ice caps and cirque glaciers throughout the interstade is possible but remains to be conclusively demonstrated.

4.5 The Loch Lomond Stade (~12.9–11.7 ka) and Loch Lomond Readvance

The Loch Lomond Stade (LLS) is the Scottish equivalent of the Younger Dryas chronozone or Greenland Stadial 1; the term Loch Lomond Readvance (LLR) refers to the growth and culmination of glaciers in Scotland during this period. The stratotype for the LLR is near the southern end of Loch Lomond, where organic detritus radiocarbon-dated to 12.1–11.9 ka is overlain in succession by glacilacustrine deposits, till deposited by a glacier that advanced over the site, then glacifluvial or deltaic sediments deposited during subsequent glacier retreat (MacLeod et al. 2011).

The return of stadial conditions was rapid. Evidence provided by Lateglacial chironomid assemblages implies that mean July sea-level temperatures at the beginning of the Lateglacial Interstade reached ~11–13 °C, then experienced a slight overall decline, punctuated by centennial-scale cold excursions (Brooks et al. 2012, 2016; Fig. 4.11). The onset of the LLS was marked by a fall in mean July temperatures on low ground to ~6.0–8.5 °C. This rapid cooling was caused by disruption of Atlantic meridional overturning circulation (AMOC) and consequent southwards migration of the oceanic polar front, so that polar waters returned to the shores of the British Isles and warm airmasses followed a more southerly trajectory. Disruption of AMOC probably resulted from an influx of freshwater into the North Atlantic Ocean though its source remains contentious (Carlson 2010; Carlson and Clark 2012).

Permafrost features of LLS age in England and Ireland imply that mean annual air temperatures (MAATs) during the coldest part of the stade fell to, or slightly below, −4 to −6 °C (Ballantyne 2018), implying that sea-level MAATs in Scotland during the thermal nadir of the LLS were no higher than about −6 °C to −8°C and mean January temperatures lay within the range –18 to –14 °C, a regime similar to that of present-day Svalbard. The chironomid record for three sites in Scotland suggests that the coldest summer temperatures occurred near the beginning of the stade and were succeeded by gradual, oscillatory warming of 1–2 °C, then rapid warming at the Lateglacial–Holocene transition; the record for a fourth site (on Skye), however, suggests a cooling trend during the LLS that may reflect the proximity of glacier ice.

The extent of the LLR has been determined through mapping of moraines, drift limits, trimlines and meltwater channels, and directions of ice flow have been reconstructed from striae, roches moutonnées, streamlined bedrock outcrops, erratic transport and fluted moraines. Moreover, occupancy of upland terrain by LLS glaciers is often represented by glacial landsystems that differ radically from those outside the LLR limits, which are often dominated by soliflucted till deposits, outwash terraces and relict periglacial landforms (Lukas 2006; Sects. 4.6 and 4.7).

The evidence for the extent of the LLR has been comprehensively summarised by Bickerdike et al. (2016, 2018a). The largest ice mass, the West Highland Icefield (or West Highland glacier complex) straddled the main Highland drainage divide, had a maximum altitude of ~900 m and fed large outlet glaciers westwards into the fjords of western Scotland, southwards to occupy Loch Lomond and the upper Forth valley and eastwards across Rannoch Moor and the eastern glens of the Northern Highlands (Fig. 4.12). Flanking the West Highland Icefield were satellite icefields on Mull and Skye, the mountains of Sutherland, west Drumochter Hills and Eastern Grampians, as well as ice caps on the Beinn Dearg massif and the Monadhliath and Gaick plateaux (Chaps. 10, 12, 13 and 20) and the Tweedsmuir Hills in the Southern Uplands (Chap. 27). Many mountain areas from Orkney in the north to the Galloway Hills in the south and Cairngorms in the east also supported small cirque or valley glaciers; more than 100 small independent glaciers formed in Scotland during the LLS.

Few LLR limits have been dated directly, but the contrasting stratigraphy of sites inside and outside the mapped limits of the readvance has provided evidence for assigning a LLS age (Walker and Lowe 2019) and the radiocarbon ages of basal organic sediments inside mapped readvance limits constrain the timing of subsequent deglaciation (e.g. MacLeod et al. 2011). Such stratigraphic evidence has been supported by the identification of distinctive microscopic shards of volcanic ash in Lateglacial and Early Holocene deposits, notably the Borrobol (~14.1 ka) and Penifiler (~13.9 ka) tephras of Lateglacial Interstadial age, the Vedde Ash (~12.0 ka) of LLS age and the Early Holocene Abernethy tephra (~11.5 ka) and Askja-S tephra (~10.8 ka). Collectively, these represent robust stratigraphic markers for dating deposits relating to the Last Glacial-Interglacial Transition (Timms et al. 2019). At a few sites, the LLS age of readvance moraines has been confirmed by radiocarbon dating of marine shell fragments within morainic deposits (Bromley et al. 2018).

TCN dating of boulders or bedrock surfaces has also been used to confirm a LLS age for mapped readvance limits (e.g. Finlayson et al. 2011; Small et al. 2012). The imprecision inherent in TCN dating, however, has made it difficult to pinpoint the timing of the culmination of the readvance or that of subsequent deglaciation, though TCN ages obtained for boulders on moraines suggest that some glaciers reached their maximum extent before 12.4–12.1 ka (Ballantyne 2012; Ballantyne et al. 2016). Numerical modelling of the expansion and contraction of LLS glaciers based on Greenland ice-core data suggests that icefields in northern Scotland, Mull and Skye reached their terminal limits early

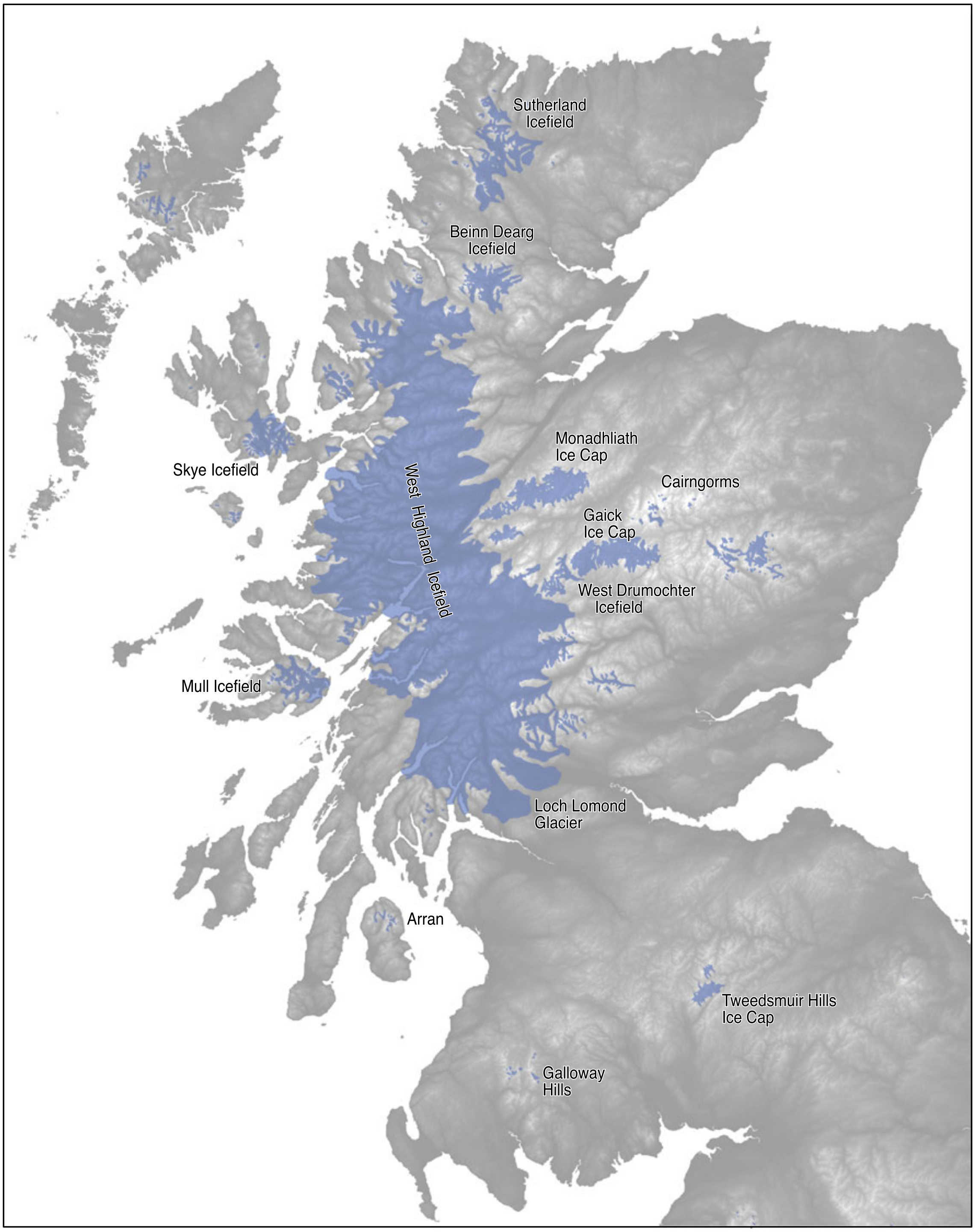

Fig. 4.12 The mapped extent of the Loch Lomond Readvance in Scotland. Numerous summits protruded through the West Highland Icefield as nunataks but are not shown. Recent research has shown that the Gaick Ice Cap was smaller than depicted here (Chandler et al. 2019) and that some glaciers in parts of the southern and SE Grampians were more extensive (Ballantyne unpublished). (From Ballantyne (2019a) Scotland's mountain landscapes: a geomorphological perspective. © 2019 Dunedin Academic Press)

in the LLS ($\sim$12.7–12.6 ka) and that most outlet glaciers of the West Highland Icefield achieved their maximum extent by $\sim$12.5 ka (Golledge et al. 2008). This suggestion must be treated with caution, however, because of disparities between the ice-core record and the summer temperature record inferred from chironomid assemblages, particularly for the final centuries of the Lateglacial Interstade (Fig. 4.11). Early culmination of the LLR has also been advocated by Bromley et al. (2014, 2018) on the basis of radiocarbon dates obtained for marine shells incorporated within coastal moraines and minimum (deglacial) ages obtained for basal organic deposits near the centre of the West Highland Icefield. From these ages they concluded that the West Highland Icefield reached its maximum extent between ~12.8 ka and ~12.6 ka and that subsequent glacier retreat was driven by warming summers, but the validity of their radiocarbon ages has been contested (Lowe et al. 2019). Moreover, their conclusions conflict with: (i) TCN ages indicating deglaciation of Rannoch Moor at ~11.5 ka (Small and Fabel 2016); (ii) a tightly-constrained radiocarbon- and varve-based chronology established for advance and retreat of the southernmost outlet glacier of the West Highland Icefield (MacLeod et al. 2011); and (iii) a varve-based chronology proposed for the glacial lakes dammed by the ice margin in Glen Roy and vicinity (Palmer et al. 2010, 2020). These studies indicate that glacier advance continued until near the end of the LLS and that extensive ice masses persisted (at least locally) until the LLS-Holocene transition (~11.7–11.6 ka). It seems likely that though some small glaciers and icefields culminated in mid-stade at ~12.5–12.4 ka, some of the outlet glaciers of the West Highland Icefield continued to expand until near the end of the LLS.

The abundant geomorphological evidence for the lateral and vertical extent of some LLS glaciers and icefields has permitted contoured reconstructions of their dimensions (e.g. Benn and Ballantyne 2005; Ballantyne 2007a, b; Finlayson et al. 2011; Boston et al. 2015; Chandler and Lukas 2017; Chandler et al. 2019; Fig. 4.13). Equilibrium line altitudes (ELAs) calculated from these reconstructions show two trends: a northwards decrease in ELAs along the west coast from SW Scotland to Lewis (Ballantyne 2007a) that probably represents a northward decline in ablation-season temperatures; and a pronounced eastward rise in ELAs across the Highlands (from 250 m on Mull and 277 m on Skye to 714 m in the Monadhliath Mountains, 751 m on the Gaick plateau and 918 m in the Cairngorms). The latter trend represents an eastwards reduction in LLS snowfall due to the snow-scavenging effects of the West Highland Icefield on westerly airmasses, so that mountains east of the icefield experienced relative aridity (Sissons 1979). Estimating the associated trends in mean annual precipitation (MAP) is difficult, as estimates are strongly influenced by seasonality (Golledge et al. 2010), but by assuming a 'neutral' estimate and mean July sea-level temperature of 8.5±0.3 °C, Chandler et al. (2019) calculated that sea-level MAP exceeded 2000 mm a^{-1} across much of the Hebrides, declining eastwards to <600 mm a^{-1} near the Cairngorms.

4.6 Late Devensian Glacial Landsystems

4.6.1 Glacial Landsystems Associated with Ice-Sheet Retreat

In upland areas of Scotland, major glacial erosional landforms (cirques, glacial troughs, glacial breaches, rock basins, fjords and landscapes of glacial scouring) are the products of multiple episodes of Pleistocene glaciation and it is likely that the last ice sheet accomplished only slight modification of pre-existing erosional features (Sect. 4.2.3). Similarly, the erosional features of the lowlands, such as glacially-steepened scarps and crag-and-tail features, are pre-existing landforms only slightly modified by the last ice sheet. Late Devensian glacial landsystems are therefore represented mainly by glacigenic deposits, depositional landforms and minor erosional landforms superimposed on glaciated landscapes of essentially pre-Late Devensian age.

In most areas, the glacial landsystems (landform-sediment assemblages) that formed during the advance and retreat of the last ice sheet differ strikingly from those associated with the LLR. In lowland areas outside the limits of the LLR, such as much of the Midland Valley, the Solway lowlands and the lower Tweed valley, the landscape is dominated by till sheets locally overlain by outwash deposits, the latter now extensively terraced and quarried for sand and gravel extraction. In many parts of the Midland Valley the till sheets form gently undulating expanses of sediment that blanket the underlying bedrock, burying the rockhead topography except where igneous outcrops protrude through the sediment cover (Fig. 4.3d). Though the lowland till sheets are barely a metre thick in places, glacigenic sediments infilling palaeovalleys in parts of the Forth and Clyde basins reach depths of over 50 m (Kearsey et al. 2019). The deepest deposits in lowland areas of Scotland are locally pre-Late Devensian in age, but the uppermost tills and overlying glacifluvial sediments were deposited by the last ice sheet, although probably containing sediment reworked from earlier glacial deposits (Merritt et al. 2019). High-resolution digital elevation imagery reveals that in many areas the till sheets form broad swells and depressions aligned in the direction of former ice movement. In some areas these grade into drumlin fields or ribbed moraine, notably those of the Glasgow region in west-central Scotland, which record southeastward then eastward movement of the last ice sheet from the SW Highlands towards the Forth valley (Rose and Smith 2008; Finlayson et al. 2010; Chap. 26) and those in the Solway and Galloway lowlands (Chap. 28). In the Tweed valley, eastward-aligned drumlins, megadrumlins and megaflutes represent bedforms in the

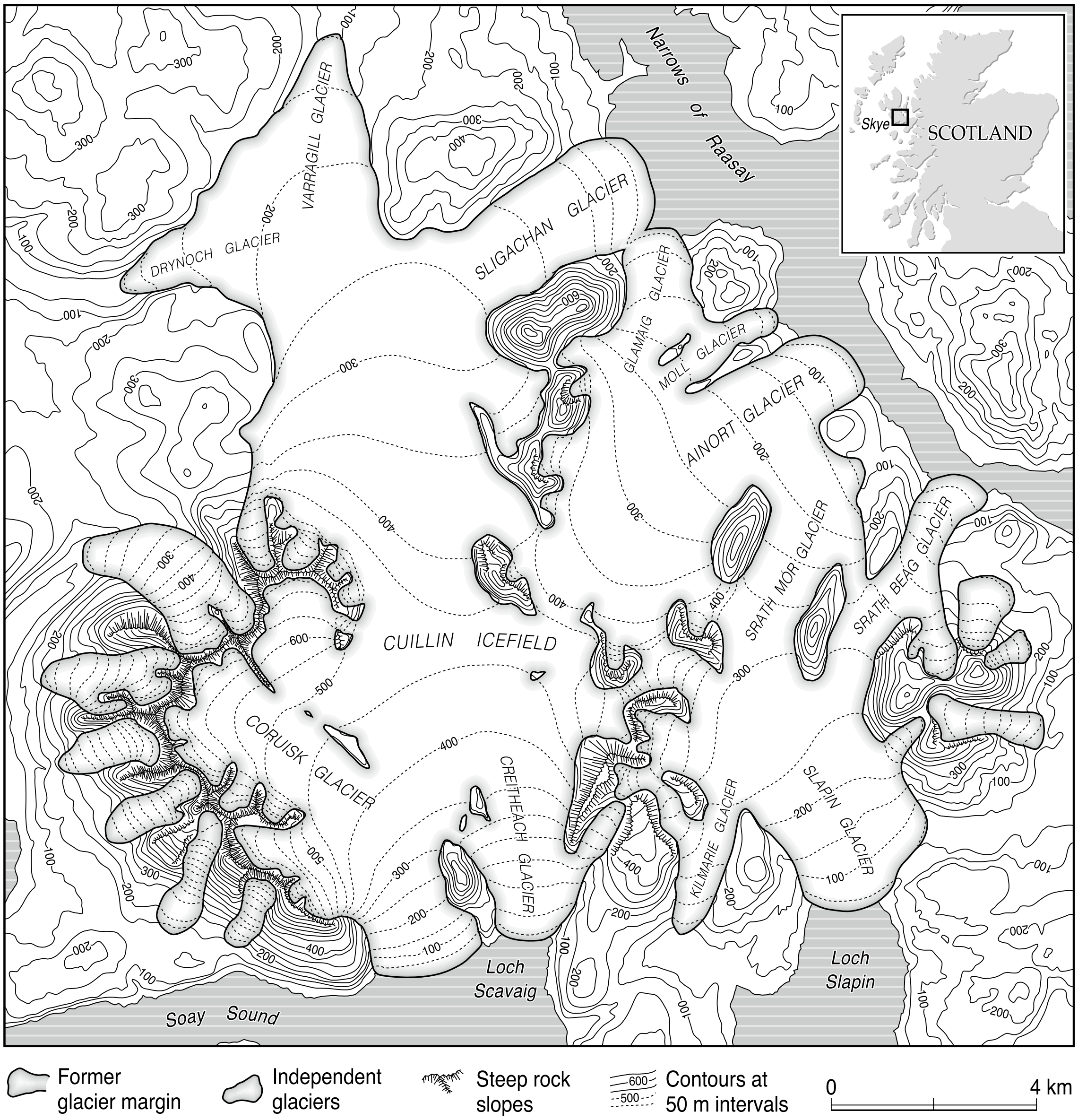

Fig. 4.13 Contoured reconstruction of the Loch Lomond Readvance glaciers that occupied the mountains of south-central Skye. (Adapted from Ballantyne (1989) J Quat Sci 4:95–108 © 1989 Longman Group UK Ltd.)

onset zone of the Tweed Ice Stream (Everest et al. 2005) and in Strathmore, streamlined bedforms and bedrock outcrops define the track of the ice stream that flowed eastwards between the Highland boundary and Fife (Golledge and Stoker 2006).

Throughout lowland Scotland, there is abundant evidence that vigorous meltwater activity accompanied ice-sheet retreat. Networks of meltwater channels mark the routeways of former subglacial and proglacial meltwater channels, some of which are now occupied by underfit streams, whilst others form dry valleys. Numerous meltwater channels aligned obliquely across hillslopes record the locations of ice-marginal and submarginal meltwater flow as the last ice sheet retreated (Evans et al. 2017) and some connect with subglacial esker systems (Chaps. 15 and 25). Deposition of sediment by proglacial meltwater rivers is represented by outwash plains of stratified sands and gravels, particularly in the Moray Firth coastlands, Strathmore, Fife, the Lothians, upper Clyde valley and lower Nith valley; many form flat sandur plains, others form mounds and depressions due to melt-out of buried ice and some have been terraced by later fluvial incision. Where tributary meltwater streams discharged into main valleys, large outwash fans were deposited that merge with sandur plains (Marren 2001) and large palaeosandar composed of coarse, bouldery outwash sediments indicative of high palaeo-discharges accumulated along the Highland boundary in eastern Scotland (Russell 2019). Readvance and recessional moraines are rarely represented in the lowland glacial landsystem, though discontinuous moraine systems are present in the western Midland Valley and Moray Firth coastlands (Finlayson et al. 2010; Merritt et al. 2017, 2019).

The characteristic glacial landsystems of upland areas outside the limits of the LLR, notably the Eastern Grampians and Southern Uplands, are dominated by two elements: thin deposits of till, smoothed by Lateglacial solifluction, that mantle lower slopes; and valley-fill deposits of till, often overlain by thick outwash deposits, as in major upland valleys such as those of the Spey, Dee and Tweed and the glens of the Eastern Grampians. The evidence of powerful meltwater activity as the last ice sheet downwasted is also evident in the form of high-level meltwater channels cut across cols and along hillslopes, and kame terraces stranded high above valley floors (Brazier et al. 1998; Hall et al. 2016b). Most river terraces in this zone are outwash terraces that represent abandoned proglacial sandur surfaces. Such terraces typically occur up to 30 m above present floodplains, are underlain by stratified sands and gravels and support a microtopography of braided palaeochannels; the highest terraces are often pitted with kettle holes, confirming a glacifluvial origin (Robertson-Rintoul 1986; Aitken 1998; Chap. 20). During ice-sheet retreat, large outwash fans developed at the confluence of major upland rivers and at the mouths of tributary valleys, where they grade into the highest outwash terraces (Werritty and McEwen 1997). Such fans often exceed 500 m in width but have been incised and terraced by postglacial rivers. As in lowland areas, there is limited morphological evidence for ice-margin oscillations or readvances. The main exceptions are the moraines delimiting the Wester Ross Readvance (~15.3 ka) in NW Scotland and the Loch Scavaig Readvance (~15.2 ka) on Skye (Sect. 4.4.4), together with evidence for readvance of the ice margin in Strathspey and along the northern flanks of the Cairngorms at ~15.8 ka (Hall et al. 2016b).

4.6.2 Glacial Landsystems of the Loch Lomond Readvance

The landsystems associated with the LLR are typically dominated by glacial rather than glacifluvial landforms. The lateral extent of LLS glaciation is often represented by end and lateral moraines, or by the downvalley extent of glacially-deposited boulders, recessional moraines or thick till deposits (Fig. 4.14). On high ground, the extent of readvance glaciers is locally defined by the upper limit of gullied till or by trimlines. Many upland valleys within the LLR limits are occupied by hummocky moraines in the form of mounds and ridges up to ~30 m high (Fig. 4.14d). Although some of these represent drumlinoid bedforms or ice-stagnation landforms (Benn 1992), most are recessional moraines, many of which originated as ice-contact fans that were modified by ice push and subglacial deformation as ice margins experienced multiple short-lived advances during overall active retreat (Benn and Lukas 2006; Chandler et al. 2020).

Bickerdike et al. (2018b) identified five LLR landsystems. The most widespread of these is the alpine icefield landsystem, which is typical of most of the West Highland Icefield as well as the smaller icefields of Mull, Skye and Sutherland (Fig. 4.15). During the LLS, these areas supported transection glacier complexes, overlooked by nunataks and fed by ice nourished in cirques; directions of ice flow were constrained by topography, along troughs and across cols. The former altitudinal limits of these icefields are marked by trimlines or drift limits that are locally continued downvalley by lateral moraines deposited at the margins of outlet glaciers. End moraines sometimes mark the former termini of icefield outlet glaciers, but in many locations their lateral extent is represented by the outer limit of hummocky recessional moraines.

In the Western Grampian Highlands, however, the ice cover thickened to an altitude of about 900 m, forming a broad dome that covered all but the highest summits. This area contains an ice-cap landsystem characterized by topographically-discordant glacier flow away from the ice centre across cols and locally transverse to the underlying valleys (Golledge 2007). The geomorphological signature of this landsystem includes ice-smoothed bedrock, roches moutonnées, striae and fluted moraines indicating radial ice flow, thick accumulations of till where ice movement was up reverse slopes and recessional moraines aligned obliquely across valleys. Around the margins of this domed ice cap is a transitional zone where the former glaciers became increasingly constrained within major troughs. The limits of these outlet glaciers are sometimes marked by end moraines (such as that which marks the eastern terminus of the Rannoch Moor glacier), or by the lateral extent of hummocky drift.

Distinct from both of the above is the plateau icefield (or plateau ice cap) landsystem that developed where thin, cold-ice-based ice caps on high ground fed outlet glaciers that occupied adjacent valley heads. The main examples are those associated with the high plateaux of the Monadhliath and Gaick in the Grampians (Boston et al. 2015; Chandler et al. 2019), the Beinn Dearg massif in the Northern Highlands (Finlayson et al. 2011; Fig. 4.15) and the Tweedsmuir Hills in the Southern Uplands (Pearce 2014). In these areas, glacial landforms on the plateaux are mainly limited to meltwater channels, and periglacial regolith has been preserved under the ice cover. The limits of some outlet glaciers are marked by moraines, but some outlet valleys contain

Fig. 4.14 Glacial landforms of the Loch Lomond Readvance. **a** End moraine crossing a valley below Beinn Dearg (1084 m) in the NW Highlands. **b** End and lateral moraines and a boulder spread define the limits of a cirque glacier on Beinn na Caillich (732 m), Skye. **c** Boulder limit defining the extent of an outlet glacier of the Beinn Dearg ice cap, NW Highlands. **d** Hummocky moraines in Drumochter Pass, central Grampian Highlands. (Images: **a**, **c**, **d** Colin Ballantyne; **b** Google Earth™ image)

little or no evidence of LLS glaciation, suggesting that the glaciers flowing into these valleys were cold-ice-based or starved of sediment.

The cirque/niche glacier landsystem of Bickerdike et al. (2018b) occurs at the sites of small LLR glaciers peripheral to the main ice masses (Fig. 4.15), such as those of An Teallach and Ben Mòr Coigach in NW Scotland, the Cuillin Hills on Skye, the Isle of Arran, the NW Cairngorms and the Galloway Hills (e.g. Ballantyne 2007b; Chandler and Lukas 2017; Chap. 10; Fig. 4.14b). Landforms characteristic of this landsystem include glacially-abraded bedrock, roches moutonnées and whalebacks in cirque source areas, and glacial limits defined by massive end moraines (where glaciers were fed by rockfall or rockslide debris), arcuate spreads of boulders or broad outer belts of closely-spaced end and recessional moraines.

Examples of the lowland piedmont-lobe landsystem are limited to the distal zones of the glaciers that advanced down Loch Lomond and the upper Forth valley to terminate on low ground in the Midland Valley. The limits of these piedmont lobes are characterized by glacitectonic landforms (thrust-block moraines) and lateral moraines, inside which are chaotic hummocky moraines, eskers, kames and kame terraces (Chap. 24).

Additionally, ice-dammed lake landsystems characterized by varved glacilacustrine deposits, deltas and shorelines occur at locations where LLR glaciers impounded substantial water bodies. The most famous of these is at Glen Roy in Lochaber, where three pronounced shorelines skirt the lower slopes of the glen (Palmer and Lowe 2017; Chap. 16), but shorelines, deltas and glacilacustrine deposits also indicate the former existence of ice-dammed lakes at various other locations, notably east of Loch Tulla in the central Grampians, at Loch Garry in the Drumochter area and at Achnasheen and in Glen Doe in the northern Highlands. Advance of the Loch Lomond glacier also dammed a lake near its terminus (Glacial Lake Blane), depositing a proglacial delta that was glacitectonised as the glacier advanced over it, depositing subglacial till (Macleod et al. 2011; Chap. 24).

The evidence provided by recessional moraines suggest that the LLR glaciers experienced three styles of retreat

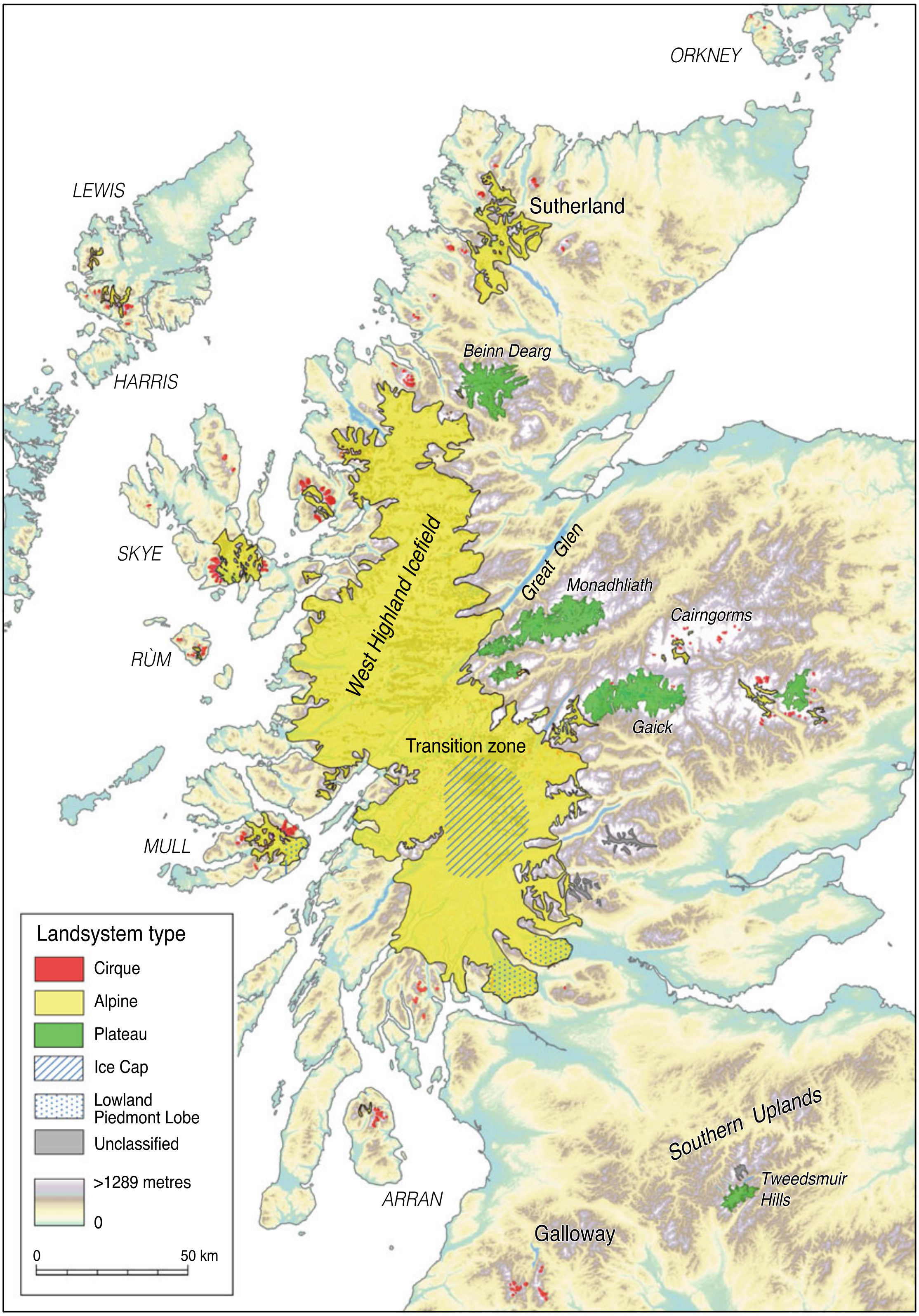

Fig. 4.15 Landsystems of the Loch Lomond Readvance. The main ice mass, the West Highland Icefield or West Highland glacier complex, comprises an ice-cap landsystem that transitions radially into an alpine icefield landsystem and terminates to the south in the piedmont-lobe landsystems south of Loch Lomond and in the upper Forth valley. (Adapted from Bickerdike et al. (2018b) Boreas 47:202–224 © 2017 The authors)

(Bickerdike et al. 2018b). In many locations, recessional moraines extend from former glacier termini to (or almost to) source areas, implying climatically active retreat of oscillating ice margins and suggesting that retreat occurred mainly under stadial conditions. In a few locations, particularly on Skye, initial oscillatory retreat appears to have been succeeded by uninterrupted retreat of the ice margin, or in situ ice stagnation (Benn et al. 1992). At some other locations, particularly at the margins of small cirque glaciers, the glacier terminus appears to have oscillated within a narrow zone, building massive end moraines or multiple moraine ridges, then experienced uninterrupted retreat (Fig. 4.14b). Whether these contrasts indicate different responses to climatic forcing or reflect progressive exhaustion of sediment transported by some glaciers as they retreated has not been established.

4.7 Lateglacial Periglacial and Paraglacial Landforms

The term 'Lateglacial' as employed here refers to the period between the retreat of the last ice sheet (which varied spatially) and the beginning of the Holocene at ~11.7 ka.

Evidence for the development of continuous permafrost down to sea level takes the form of ice-wedge pseudomorphs and polygonal crop marks representing former frost polygons (Murton and Ballantyne 2017). All recorded examples fall outside the limits of the LLR (Fig. 4.16) and though they are undated, evidence from elsewhere in the British Isles indicates that permafrost developed in Scotland both on terrain exposed by the retreating ice sheet and again during the LLS (Ballantyne 2018, 2019b).

In lowland Scotland, sediment-mantled slopes were affected by widespread Lateglacial solifluction. Soliflucted tills occur both below and above organic layers of Lateglacial Interstadial age, implying that solifluction operated both prior to warming at ~14.7 ka and again during the LLS (Ballantyne 2019b). In some parts of the Southern Uplands and Cheviot Hills, Lateglacial valley-fill deposits of reworked till or weathered rock accumulated on valley floors. These deposits have been incised by rivers to form steep-fronted terraces 20–300 m wide, and OSL dating suggests that they mainly accumulated during the LLS (Harrison et al. 2010). Although initially attributed to Lateglacial solifluction over permafrost, their thickness at some sites suggests that they could have been emplaced by debris flows or active-layer failures (Harrison 2002; Chap. 27).

As noted earlier, many Scottish mountain summits and plateaux are mantled by periglacial blockfields, typically up to a metre deep, and in a few areas (mainly the Cairngorms) tors protrude through the blockfield debris (Sect. 4.2.4). Although studies of Scottish blockfields have shown that they developed primarily through frost weathering of bedrock (Ballantyne 1998a, 2010; Hopkinson and Ballantyne 2014), TCN dating of erratics resting on blockfield debris has shown that mature blockfields are of pre-Late Devensian age, and survived beneath the last ice sheet because the ice was cold-based and frozen to the underlying substrate (Fabel et al. 2012); only thin covers of frost-weathered regolith formed on glacially-scoured summits during the Lateglacial period. Some blockfields support large relict sorted circles, nets and stripes, typically 2–4 m in width, that represent frost sorting of the active layer above former permafrost (Fig. 4.17a). Nonsorted patterned ground features on some mountains (earth hummocks and relief stripes) have also been interpreted as representing differential frost heave of frost-susceptible soils during the Lateglacial (Ballantyne and Harris 1994). The most common Lateglacial periglacial landforms on the upper slopes of Scottish mountains, however, are large 'stone-banked' terraces and lobes of bouldery debris that has been rafted downslope through solifluction of underlying fine sediment or (less plausibly) permafrost creep (Fig. 4.17b).

Most talus accumulations in Scotland are relict, vegetation-covered landforms and many are scarred by gullies eroded by debris flows (Fig. 4.17c). Those outside the limits of the LLR accumulated mainly during the Lateglacial period; enhanced rockfall activity at this time has been attributed to a combination of paraglacial stress release and frost wedging of rockwalls (Hinchliffe and Ballantyne 1999, 2009). Lateglacial snow avalanche activity is represented by relict avalanche boulder tongues at a few locations; it is likely that snow avalanches were common under stadial conditions during the Lateglacial, but their geomorphological effects have often been obliterated by Holocene debris flows (Luckman 1992).

Paraglacial rock-slope failures (large-scale rockslides, large rockfalls, rock avalanches and rock-slope deformations) are common features of the Scottish landscape. Over 900 probable sites (mainly on schist mountains) have been identified in the Highlands alone, including over 170 with areas exceeding 0.25 km^2 (Jarman and Harrison 2019; Chap. 14). Many others occurred along scarps in the Midland Valley and Inner Hebrides, and on steep slopes in the Southern Uplands (Chaps. 10, 24 and 27; Fig. 4.17d). TCN dating of the runout debris from 20 rock avalanches or fragmented rockslides outside the limits of LLR glaciers in the Highlands and NW Ireland has shown that all but one occurred during the Lateglacial period, with peak activity 1.6–1.7 ka after ice-sheet deglaciation. Ballantyne et al. (2014) inferred that this time lag represented time-dependent reduction of rock-mass integrity (progressive failure) following deglaciation, but also noted that the peak in landslide activity coincided with maximum rates of glacio-isostatic uplift, suggesting that failure in some cases was triggered by

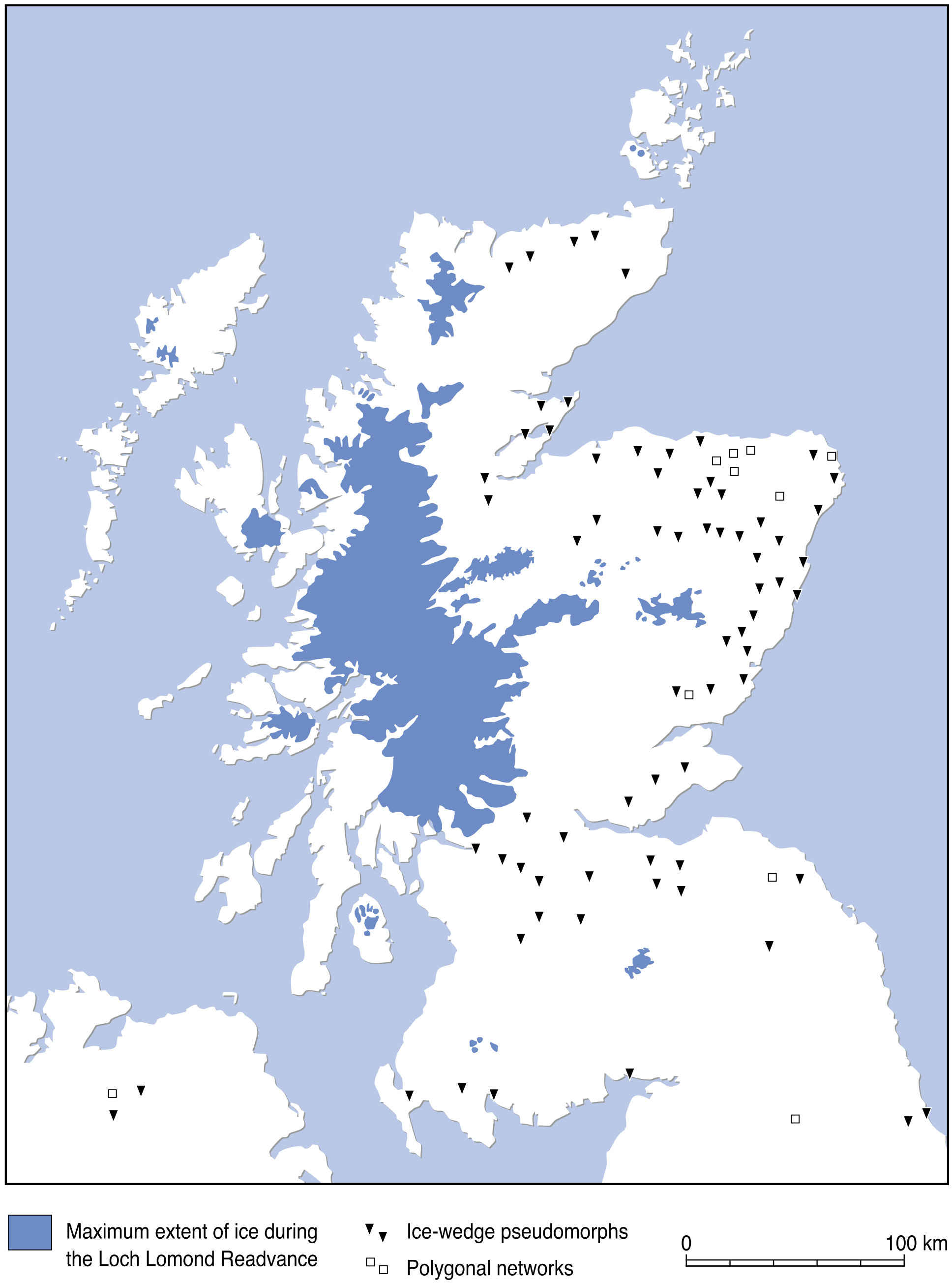

Fig. 4.16 Distribution of documented ice-wedge pseudomorphs and polygonal networks in Scotland. These features indicate that permafrost formed down to sea level following retreat of the last ice sheet. (From Ballantyne (2019a) Scotland's mountain landscapes: a geomorphological perspective. © 2019 Dunedin Academic Press)

uplift-driven seismic activity. Inside the limits of the LLR, numerous rockslide scars represent the sites of Lateglacial landslides where the runout debris was subsequently removed by glacier ice (Ballantyne 2013a; Cave and Ballantyne 2016).

In many major valleys in Scotland, river terraces flank active floodplains. All but the lowest of these are probably outwash terraces that represent abandoned proglacial sandur surfaces (Fig. 4.17e). The limited dating evidence suggests that fluvial incision of palaeosandar began soon after deglaciation as upvalley sources of sediment diminished, and continued throughout most or all of the Lateglacial period (Robertson-Rintoul 1986), and it is notable that the highest terraces inside the limits of the LLR rarely occur more than a few metres above active floodplains. The only evidence for floodplain aggradation during the LLS is from the Kelvin valley in central Scotland (Tipping et al. 2008). In upland areas, however, net incision of outwash sediments appears to have predominated throughout the Lateglacial period (Ballantyne 2019b).

Lateglacial alluvial fans are common throughout Scotland. Large, low-gradient fans, now incised by their parent rivers, occur at the confluences of major upland rivers, such as the Feshie and Spey (Werritty and McEwen 1997;

Fig. 4.17 Lateglacial landforms. **a** Large, relict sorted circles on a summit blockfield, SE Grampians. **b** A stone-banked solifluction lobe, Cairngorm Mountains. **c** Relict, vegetated talus slopes, Wester Ross. **d** Lateglacial rock-slope failure, Lomond Hills, Fife. **e** River terraces, Glen Garry, central Grampians; the higher terraces are outwash terraces formed during retreat of the last ice sheet. **f** Lateglacial fan in upper Glen Roy, Lochaber. (Images: Colin Ballantyne)

Chap. 19). Large Lateglacial alluvial fans also occur in lowland areas, though some, such as the Lindores fan in Fife (Marren 2001) were fed by glacial meltwater streams and should strictly be regarded as outwash fans. Most large upland fans at the mouths of steep tributary valleys are probably paraglacial landforms that accumulated through fluvial reworking of glacigenic deposits; these commonly terminate on high-level outwash terraces, indicating a Lateglacial origin, and have subsequently been incised by their parent streams. Much of the research on Lateglacial fans in Scotland has focused on those associated with the ice-dammed lakes in Glen Roy (Chap. 16; Fig. 4.17f). These were initially interpreted as paraglacial landforms, but stratigraphic investigations by Cornish (2017) suggest that they represent sublacustrine outwash fans deposited as lake levels rose, capped with a layer of gravel deposited when the higher lakes drained. The Glen Roy fans are now deeply incised, with nested lower-level fans inset within their Lateglacial predecessors.

4.8 The Holocene

The onset of the Holocene Epoch at ~11.7 ka was marked in Scotland by rapid warming, degradation of permafrost and the disappearance of glacier ice; temperate conditions similar to those of the present were established by ~11.0 ka (Brooks et al. 2012; Fig. 4.11). Much of the earlier Holocene was characterized by relatively warm summers (the Holocene thermal maximum), followed by a very gradual cooling trend over the past five millennia, during which average temperatures oscillated by up to 2 °C over decades or longer. Apart possibly from a brief cooling event at ~8.2 ka, the most severe period of Holocene cooling in Scotland was the Little Ice Age of the sixteenth to nineteenth centuries, a period characterized by decades of cool, wet summers, exceptional storms and perennial snowcover on some mountains. Tree-ring evidence suggests that within the period AD 1580–1810 summer temperatures in the Cairngorms averaged about 1 °C lower than the 1961–1990 mean (Ryvdal et al. 2017). Some authors have argued that niche glaciers briefly formed in the Cairngorms at this time (Kirkbride et al. 2014), though the dating evidence on which this proposition is based is equivocal.

During the Holocene, many Lateglacial paraglacial landforms (talus accumulations, debris cones, alluvial fans and floodplains) experienced net erosion or incision as a result of reduction or exhaustion of sediment supply (Ballantyne 2008), even though glacigenic sediment continued to dominate paraglacial sediment budgets in headwater catchments (Fame et al. 2018). Superimposed on this trend were: (i) the influence of extreme climatic events (possibly related to periods of wetter, cooler or stormier climate); (ii) the effects of woodland clearance, burning and grazing pressure in altering runoff regimes and lowering the thresholds for erosion of sediment-mantled hillslopes; and (iii) autogenic changes related to pedogenesis, peat formation or (in fluvial systems) local changes in base level. Because of the site-specific nature of most research on Holocene landforms, the relative importance of these effects in dictating Holocene landscape change is difficult to assess (Ballantyne 2019b).

4.8.1 Holocene Periglacial and Aeolian Landforms on Scottish Mountains

It is likely that the present maritime periglacial environment on high plateaux and summits in Scotland is reasonably representative of conditions throughout the Holocene. Above 700–800 m altitude the climate is characterized by mean annual temperatures (1981–2010) of 0.5–5.0 °C and winter air temperatures below −10 °C are fairly uncommon. Permafrost is absent, though seasonal ground freezing may reach depths of 0.4–0.5 m. Gale-force winds (>80 km h^{-1}) are frequent, with occasional gusts exceeding 150 km h^{-1} , and all but the most easterly summits receive >2000 mm of annual precipitation. During much of the last century, average snow-lie (>50% cover) exceeded 100 days per year at 600 m altitude, though in recent winters snowcover has been less persistent.

On Scottish mountain summits, Holocene frost weathering has been limited to granular disaggregation and flaking of exposed bedrock and boulder surfaces, resulting in rounding of outcrops and exposed boulder surfaces. Active frost-patterned ground takes the form of small-scale sorted circles, nets and stripes, formed by differential growth of needle ice (Ballantyne 2001a), though such patterns are limited to unvegetated, frost-susceptible soils containing small clasts. The most widespread active periglacial landforms above 700 m on Scottish mountains are vegetation-covered solifluction terraces and lobes (Fig. 4.18a) with steep risers up to a metre high. Radiocarbon dating of organic matter buried by downslope movement of solifluction lobes has shown that these have been at least intermittently active over the past 5500 years, and recent surface movement averaging ~8–10 mm a^{-1} has been measured over a 35-year period (Ballantyne 2013b). Commonly associated with active solifluction lobes are ploughing boulders that have migrated downslope faster than the surrounding soil, leaving a vegetated furrow upslope. Boulder movement has been attributed to the development of high pore-water pressures during thaw of sub-boulder ice lenses, causing boulders to slide downslope on liquified or softened soil at rates of up to 30 mm a^{-1} (Ballantyne 2001b).

The combined effects of frost and wind erosion have created various landforms on exposed plateaux and cols, notably deflation scars, wind stripes and turf-banked terraces (Ballantyne and Harris 1994) . The most impressive manifestation of upland wind erosion, however, takes the form of

Fig. 4.18 Holocene landforms. **a** Active solifluction terraces at 750 m above Glen Strathfarrar. **b** Deflation surface and remnant 'island' of windblown sand on Ward Hill, Orkney. **c** The Hell's Glen rock-slope failure, Argyll. **d** A Holocene debris cone in Glenuaig, NW Highlands; fresh debris cover indicates recent deposition. **e** The River Coe in Glen Coe, a typical wandering gravel-bed river. **f** Holocene alluvial fan, NW Highlands; most of the fan is now inactive due to fan-head trenching by the parent river. (Images: Colin Ballantyne)

deflation surfaces, expanses of bare ground where wind scour has winnowed away soil particles, leaving sterile surfaces covered by boulders and lag gravels (Chap. 13). Some deflation surfaces support remnant islands of vegetation-covered sand or aeolisols, demonstrating that soil cover was formerly more extensive (Fig. 4.18b). A few plateaux, however, support an intact cover of aeolian deposits preserved under a cover of grasses and sedges. The summit of The Storr (719 m) in northern Skye, for example, is mantled by a vegetation-covered sand sheet up to 2.9 m thick. This deposit consists of particles released from an adjacent cliff by weathering and blown upwards to rain down on the summit, where they have been anchored by vegetation (Ballantyne 1998b; Chap. 10). Other windblown sand accumulations on summits and plateaux may have a similar origin.

Most high-level aeolian deposits, however, form vegetation-covered sand sheets up to 4.0 m thick on lee slopes downwind from deflation surfaces. Many contain a lower unit of weathered sand that slowly accumulated throughout much of the Holocene and an upper unit of unweathered sand. OSL dating of the contact between the two units on several mountains has shown that the upper unit began to accumulate between AD 1550 and 1750 as a result of catastrophic erosion of soils and aeolian deposits on plateau areas upwind (Ballantyne and Morrocco 2006; Morrocco et al. 2007). The timing of such erosion coincides with the Little Ice Age, suggesting that it was triggered by vegetation degradation under persistent snowcover and consequent erosion of exposed soil and sand deposits by strong winds (Chap. 13).

4.8.2 Holocene Slope Failures

A few hundred postglacial rock-slope failures (RSFs) occur within the limits of the LLR in the Scottish Highlands. Of these, only eleven rock avalanches or fragmented rockslides have been securely dated: two apparently deposited debris on LLR glacier ice, two failed shortly after deglaciation and the remainder occurred at intervals throughout the Holocene (Ballantyne and Stone 2013; Ballantyne et al. 2014). There is, however, no information regarding the timing of the numerous arrested rockslides or rock-slope deformations that occur within LLR limits, though failure/displacement of most within a few millennia following final deglaciation is likely. Notable Holocene RSFs include the Beinn an Lochain rock avalanche ($\sim$11.7 ka) and Hell's Glen RSF ($\sim$3.8 ka; Fig. 4.18c) in Argyll, the Storr landslide ($\sim$6.1 ka; Chap. 10) on Skye, the Beinn Alligin rock avalanche ($\sim$4.4 ka; Chap. 14) in Torridon, the Coire Gabhail rockslide near Glen Coe ($\sim$1.7 ka; Chap. 17) and the huge rock-slope deformation on Beinn Fhada in Kintail (Chap. 14). There are very few reports of recent RSFs, apart from those involving collapse of coastal cliffs (Ballantyne et al. 2018).

Many talus accumulations and other steep sediment-mantled slopes have been eroded by Holocene debris flows, and debris flow is probably the dominant agent of hillslope sediment transport on such slopes at present. Some debris flows follow gullies, often depositing sediment on slope-foot debris cones (Fig. 4.18d); others begin as translational slides on open hillslopes (Chap. 5). Radiocarbon dating of organic soils and peat horizons buried by successive debris flows extends back to $\sim$7.0 ka, but the thickness of some debris cones implies that they began to accumulate much earlier (Ballantyne 2019b). There is evidence, however, that Late Holocene woodland clearance and other land-use changes have made some sediment-mantled slopes more vulnerable to slope failure and recurrent debris-flow activity (Foster et al. 2008). All documented recent debris flows have been triggered by rainstorms of exceptional intensity (Milne et al. 2009). Upland hillslopes have also been modified by gully erosion by flood torrents, small-scale translational landslides that have left cuspate scars, and peat slides (e.g. Dykes and Warburton 2008). Recent snow-avalanche activity is usually limited to uprooting of turf and deposition of debris downslope, though active avalanche boulder tongues occur in the Cairngorms (Luckman 1992).

4.8.3 Holocene Fluvial Landforms

The limited dating evidence available suggests little consistency in the timing of Holocene floodplain aggradation or incision in different parts of Scotland. Low-level river terraces in Highland glens may represent a widespread pattern of Late Holocene floodplain aggradation then incision, but this scenario is supported by radiocarbon ages from only three sites (Ballantyne 2008). In the Southern Uplands, radiocarbon dating of Holocene terrace sequences suggests that floodplain stability or aggradation was typical of much of the Early and Middle Holocene, but that subsequent alluvial history was catchment-specific, with sediment accumulation on some floodplains coinciding with fluvial downcutting in others. In this region, Late Holocene episodes of floodplain aggradation have been attributed to release of sediment into river systems due to Neolithic woodland clearance, Iron Age settlement expansion or other anthropogenic impacts (Tipping 1995; Tipping et al. 1999; Chap. 27), but the evidence relating alluviation to prehistoric land-use changes (particularly woodland clearance) is inconclusive. Recent fluvial incision and abandonment of some low-level terraces may reflect increased flood discharges during the Little Ice Age (Tipping 1994). Research on the fluvial history of rivers in the Midland Valley is limited to a study by Tipping et al. (2008) of the alluvial deposits of central reaches of the River Kelvin, where net floodplain aggradation appears to have predominated throughout much of the Early and Middle Holocene but ceased at least 2000 years ago.

Recent floodplain changes in the Highlands have been investigated through comparison of historical records with modern maps and aerial photographs. These studies have focused on rivers in ‘piedmont zones’: high-energy, low-threshold fluvial environments where steep mountain streams join major rivers. Over the past two centuries, the floodplains of such rivers have experienced neither net aggradation nor net incision (Werritty and Leys 2001). Many, however, exhibit cyclic changes in channel planform in response to major flood events and gradual inter-flood adjustment, manifest in lateral channel shifts and changes in active channel area and sinuosity. Studies of channel change on the floodplain of the River Feshie, for example, have shown that alluvial reaches are intrinsically unstable, with evidence of avulsion, channel switching, erosion of sediment from bars and channels and flow reorganization operating over annual to decadal timescales (Chap. 19); it appears that the Feshie and kindred gravel-bed rivers in the Highlands are subject to divergent braiding and meandering tendencies. Other Highland rivers have experienced recent channel switching: the River Coe in the Western Grampians shifted laterally by over 200 m between 1968 and 1988 (McEwen 1994; Fig. 4.18e); the main channel of the lower River Spey migrated laterally by up to 400 m between 1968 and 1992 (Riddell and Fuller 1995); and reaches of the River Tay have experienced lateral channel migration of up to a kilometre since 1783 (Gilvear and Winterbottom 1992). Regulation of major rivers over the past 160 years, however, has tended to reduce channel migration (Winterbottom 2000).

Holocene alluvial fans in the Highlands and Southern Uplands include low-gradient fans inset within Lateglacial fans at the confluence of major rivers (Werritty and McEwen 1997) and numerous small fans at the mouths of steep tributary valleys (Fig. 4.18f). The latter represent episodic erosion of hillslopes or moraines by flood torrents, though in some cases debris flows have also contributed to fan accumulation. A small fan in the Edendon valley (central Grampians), for example, accumulated during just three flood events after ~2.2 ka, implying that the fan was built within a few hours over a timescale of more than two millennia (Ballantyne and Whittington 1999). Similarly, a larger fan in Gleann Lichd (Kintail) apparently accumulated in the course of 12–15 flood episodes after ~3.8 ka (Reid et al. 2003). Intriguingly, all small fans dated through radiocarbon assay of buried soils or peat appear to have accumulated within the past four millennia. Foster et al. (2008) have provided compelling evidence that gully erosion and associated fan accumulation in the Southern Uplands followed cycles of settlement expansion, arguing that anthropogenic effects (particularly woodland clearance) made hillslopes more susceptible to gully formation and erosion during rainstorms. They suggested that the earliest gullying events occurred during the Late Bronze to Early Iron Age (~4.0–2.0 ka), with further periods of intensive gully development at ~1.3–1.1 ka, 0.9–0.7 ka and 0.55–0.45 ka. Analyses of buried peat or soil horizons under or within other fans in the Southern Uplands and Highlands, however, provide no evidence for vegetation change prior to fan accumulation, so it is unlikely that all Late Holocene fans represent a response to human activity in destabilizing slopes.

Several recent studies have focused on the Holocene evolution of bedrock channels in the Highlands. Jansen et al. (2011) employed TCN dating of strath terraces (bedrock channel remnants) to show that knickpoint retreat was most rapid in the Early to Middle Holocene and has since slowed by about two orders of magnitude. They attributed this trend to reduction in the availability of fluvially-transported clasts that constitute the 'tools' responsible for bedrock detachment and incision during floods. For rivers draining the Western Grampians, Jansen et al. (2010) found that bedrock channel widths increase nonlinearly with catchment area and that bedrock channel morphology is conditioned by lithology; quartzite outcrops have formed resistant pinning points that slowed knickpoint retreat. Other studies in Highland catchments have shown that the dimensions of river channels in both alluvial and bedrock reaches are related to catchment area (and thus flood discharges), valley-floor gradient and the calibre of bed material (Addy et al. 2011, 2014). From a strong relationship observed between bedrock channel widths and upstream catchment area, Whitbread et al. (2015) concluded that fluvial erosion has reconfigured bedrock channel morphology since deglaciation, resulting in the adaptation of channels to Holocene runoff regimes and sediment supply.

4.9 Relative Sea-Level Changes and Associated Landforms

4.9.1 Introduction

The patterns of relative sea-level change across Scotland during the Quaternary were influenced by the dimensions of former ice sheets and the spatial and temporal patterns of regional deglaciation. The numerous raised, intertidal and submerged shore platforms, raised beaches and raised deltas along the Scottish coastline represent the interaction between vertical glacio-isostatic movements (centred on the axis of uplift in the Western Grampians) and eustatic changes in global ocean volume caused by expansion and shrinkage of the Pleistocene global ice sheets, modulated by changes in the geoidal sea surface, particularly those caused by changes in the gravitational attraction between ocean water and ice sheets (Dawson 2019).

Although very few older raised shoreline deposits have survived the last ice-sheet glaciation, various raised shore platforms of pre-Late Devensian age locally form prominent features of the coastal landscape. The dimensions of the last ice sheet (Fig. 4.5) and its duration (~35–14 ka) defined the amount of glacio-isostatic downwarping of the lithosphere and the development of a crustal forebulge near and beyond the former ice-sheet margins. Deglaciation was accompanied

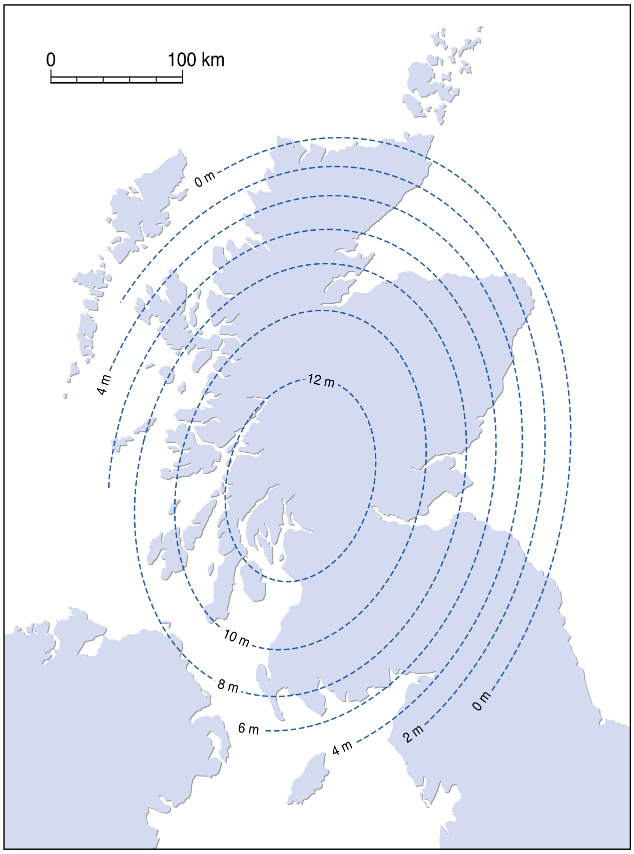

Fig. 4.19 Isobase contours (metres) of the Main Postglacial Shoreline, which formed around Scotland's coasts within the period ~7.8–6.2 ka. Subsequent differential glacio-isostatic uplift has tilted this shoreline, so that it declines in altitude radially away from the uplift centre in the Western Grampians

and followed by glacio-isostatic rebound of areas covered by the ice sheet and forebulge collapse across peripheral areas (Shennan et al. 2018; Smith et al. 2019). Isostatic rebound accounts for the occurrence of Lateglacial and Holocene raised shoreline features at altitudes of up to ~40 m OD (Ordnance Datum). Because the amount of rebound has been greatest near the axis of isostatic uplift, diminishing towards peripheral areas, postglacial raised shorelines decline in elevation away from the uplift centre, passing below present sea level in outlying areas (Fig. 4.19). Moreover, the degree of shoreline tilting has diminished through time, so that the oldest postglacial (Lateglacial and Holocene) raised shorelines exhibit the steepest regional gradients and the youngest have the gentlest gradients. Crustal forebulge collapse partly accounts for the absence of postglacial raised beaches in peripheral areas such as the Outer Hebrides, Orkney and Shetland, where many Quaternary shoreline features are now submerged.

4.9.2 Pre-Late Devensian Coastal Landscapes

Shore platforms (cut in bedrock) that formed prior to the advance of the last ice sheet fall into three categories: strandflat; low-level platforms close to the present tidal range; and high rock platforms. Extensive areas of strandflat occur on several Hebridean islands, notably Benbecula and the Uists in the Outer Hebrides and the outlying islands of Tiree and Coll in the Inner Hebrides (Chaps. 9 and 11) The strandflat consists of subhorizontal, glacially-moulded rock platforms ~3–15 km wide and no higher than ~30 m OD, mainly cut across Lewisian gneiss (Fig. 4.20a). The exceptional width of the Scottish strandflat has prompted suggestions that it may represent a modified subaerial planation surface (Smith et al. 2019) or was formed by coastal erosion during prolonged periods of relative sea-level stability in the Pliocene (Dawson et al. 2013).

Near-horizontal shore platforms at elevations within or close to the present tidal range occur around both the west and east coasts of Scotland (Fig. 4.20b). Some exhibit evidence of glacial abrasion in the form of ice-smoothed rock and others pass below till bluffs, indicating formation prior to the last (and possibly earlier) ice-sheet glaciation(s). Such low-level platforms have been interpreted as inherited landforms that developed through marine erosion during successive interglacials when sea levels were similar to that of the present (Dawson 1980a), locally modified by glacial erosion and postglacial coastal processes.

High rock platform fragments occur along west-facing coastal areas of the Inner Hebrides and the adjacent mainland (Chaps. 10 and 11). The most extensive examples are typically 200–600 m wide and backed by cliffs up to 90 m high. The inner edges of most high platform fragments occur at ~18–35 m OD. On some coasts, platform altitudes vary by several metres over distances of less than 2 km, suggesting that the platform fragments represent more than one former shoreline or tectonic dislocation of a former shoreline. Early accounts attributed an interglacial age to the high platforms of western Scotland, but Sissons (1982) argued that they developed due to periglacial coastal erosion when the margin of the last ice sheet was located along the western seaboard between Wester Ross and Islay. More recently it has been suggested that the origins of these platforms pre-date the Quaternary since their considerable widths imply a prolonged period (or periods) of relative sea-level stability (Dawson et al. 2013). Similar features also occur along parts of the east coast, notably the spectacular high rock platforms at Dunbar, Berwick and Eyemouth. Raised rock platforms partly buried beneath till are also present in some peripheral locations, such as Orkney and the Outer Hebrides (Smith et al. 2019).

Fig. 4.20 Coastlines of Scotland. **a** Strandflat, North Uist, Outer Hebrides. **b** Intertidal rock platform on the Fife coast; the abandoned stack is a volcanic neck. **c** Lateglacial raised delta and modern bay-head beach, Camas Mòr, Coigach. **d** The Main Rock Platform on the islands of Seil and Luing, Firth of Lorn. **e** Sea cliffs and sea caves, Villians of Hamnavoe, Shetland. **f** Sandy beach and eroding dune cordon near the mouth of the Ythan estuary, NE Scotland. (Images: **a**, **e** John Gordon; **b**, **c** Colin Ballantyne; **d** Murray Gray; **f** Google Earth™)

4.9.3 Relative Sea-Level Changes Associated with Ice-Sheet Deglaciation

Retreat of the last ice sheet was accompanied by marine incursion into most coastal areas. In western Scotland, the Lateglacial marine limit reached ~40 m OD across parts of the Inner Hebrides but is lower on the coast of eastern Scotland. At Camas Mòr in Coigach, for example, it is represented by a raised delta at ~15 m OD that merges inland with outwash terraces deposited by the retreating ice sheet (Fig. 4.20c). Relative sea levels subsequently fell as glacio-isostatic rebound outpaced eustatic sea-level rise. The most complete sequence of emerged Lateglacial shorelines representing this lowering of relative sea level is in western

Jura, where extensive spreads of unvegetated Lateglacial gravel beach ridges decline in altitude seaward from the marine limit (Chap. 11). This fall in relative sea level started across the Inner Hebrides at ~16 ka and continued until the onset of the LLS at ~12.9 ka (Shennan et al. 2018).

4.9.4 Shoreline Formation During the Loch Lomond Stade

There is abundant evidence for rapid coastal erosion of bedrock under the severe cold conditions of the LLS, particularly in western Scotland where a conspicuous raised rock platform (the Main Rock Platform or Main Lateglacial Shoreline) and associated backing cliff extend along long stretches of coastline from Ayrshire and Kintyre northwards to Mull and Ardnamurchan (Fig. 4.20d; Chaps. 10 and 11). This shore platform cuts across a range of lithologies and is typically 50–150 m wide with a ~15–20 m high backing cliff indented by numerous relict sea caves. The platform has an altitude of 10–11 m OD in the Oban area and declines in elevation away from the centre of glacio-isostatic uplift along a regional gradient averaging ~0.16 m km^{-1}, passing below present sea level in Ayrshire, southern Arran, Islay and western Mull. Because this platform occurs only outside the limits of the Loch Lomond Readvance, exhibits no evidence of glacial modification and is well developed in areas of limited fetch, it is believed to have formed rapidly under periglacial conditions during the LLS through a combination of frost-wedging of bedrock and removal of debris by wave erosion and sea ice (Dawson 1980b; Sissons 1983; Stone et al. 1996). In eastern Scotland, the Main Lateglacial Shoreline forms a buried gravel layer that occurs under Holocene estuarine deposits in the upper Forth valley. This feature declines eastwards along the uplift gradient; submerged rock platforms at −4 m near Rosyth and −9 m near Cockenzie probably represent the eastward continuation of the same shoreline.

4.9.5 Holocene Relative Sea-Level Changes and Associated Landforms

Along coastlines near the axis of glacio-isostatic uplift, crustal rebound at the start of the Holocene continued to outpace eustatic sea-level riseand relative sea level continued to fall, but coastlines distal to the uplift axis experienced a sustained rise in relative sea level at this time. Numerical modelling of relative sea level at the start of the Holocene suggests that it may have been as low as −20 m to −30 m OD around the Orkney Islands, −15 m OD at Aberdeen and between 0 m and +5 m OD in the Firth of Clyde (Shennan et al. 2018).

During the period ~9.5–7.0 ka, however, rapid glacio-eustatic sea-level rise exceeded rates of glacio-isostatic uplift throughout Scotland, resulting in a relative marine transgression that culminated at ~7.8–6.2 ka. This transgression is represented by the Main Postglacial Shoreline, which constitutes the Holocene marine limit in coastal areas near the axis of uplift. In major estuaries this transgression resulted in the extensive deposition of estuarine sediments (carse clays); subsequent relative marine regression due to continuing (though slowing) isostatic uplift formed the distinctive, low-gradient, emerged carseland surfaces of the lower Forth, Tay and Clyde valleys. Along coastlines exposed to open-fetch conditions, this transgression was associated with the deposition of beach gravels, such as those that mantle the Main Rock Platform along the coasts of Kintyre and the outer Clyde estuary. Continuing glacio-isostatic uplift has elevated such Holocene beach deposits well above present sea level. Near the centre of uplift the Holocene marine limit occurs at ~12–14 m OD, declining radially outwards to intersect present sea level near the north coast of mainland Scotland (Fig. 4.19). However, because of oscillations in eustatic sea level at a time of continuing (though gradual) isostatic uplift, the Holocene marine limit is diachronous: it is represented by the Main Postglacial Shoreline near the uplift axis, but farther from the axis the highest Holocene raised shoreline becomes progressively younger in age (Smith et al. 2012, 2019).

4.9.6 Recent Coastline Changes

Over the past 4000 years, relative sea level around Scotland's coasts has remained fairly stable, with some areas near the centre of isostatic uplift experiencing slow falls in sea level and others distal to the uplift axis experiencing gradual sea-level rise. The present coastline is one of great diversity (May and Hansom 2003). Steep plunging cliffs indented with caves and geos and flanked by stacks and arches occur along several parts of the mainland but are particularly well represented on the coasts of Orkney, Shetland and Caithness, where relative sea level has risen throughout the Holocene. Other rock coasts take the form of bedrock ramps or boulder-strewn intertidal rock platforms (Fig. 4.20b). Such coastlines are relatively stable, though continuing sea-level rise has been implicated in the ongoing collapse of coastal cliffs around Shetland (Ballantyne et al. 2018; Fig. 4.20e) and storm waves on Atlantic coastlines have been shown to overtop coastal cliffs up to 40 m high, quarrying angular boulders from cliff crests and depositing these inland from the cliff top (Hall et al. 2007; Chaps. 7 and 8).

Many low-lying coastlines support a range of depositional coastal landsystems (Chaps. 22 and 23), including beaches of sand or gravel, spits, bars, offshore sandbanks, salt marshes, dune systems and, on some Hebridean islands, coastal plains comprising aeolian deposits of calcareous shell sand, known as machair (Hansom and Angus 2006; Chaps. 9 and 11). Glacigenic deposits constitute the main source of most coastal sediment, though trajectories of sediment transport have been complex: glacigenic sediments may have been directly eroded from the shoreface, transported to estuaries by rivers, derived from the adjacent shelf or eroded from raised beaches, deltas or spits that were originally sourced from glacigenic deposits (Firth et al. 1995). Most present-day coastal deposits therefore represent secondary paraglacial sediment stores; sediment input from eroding rock cliffs is comparatively unimportant.

Hansom (2001) has identified a general decline in sediment supply to depositional coastal systems after the mid-Holocene and a consequent reorganization of sediment transfer into progressively smaller cells and sub-cells separated by headlands. The recent history of such coastal units has been dictated by local circumstances. In the past few decades, net sediment loss has predominated, leading to decreases in beach width, increases in beach slope and erosion of landward dune cordons (Fig. 4.20f; Chap. 5). A survey of the beaches of the Highlands and Hebrides in 1977, for example, found that only 7% were progradational, with widespread erosion of dunes on the Atlantic coastline (Mather and Ritchie 1977). This generalization, however, obscures considerable local diversity. At Tentsmuir in Fife, for example, there has been over 3.5 km of coastal advance in the last 5000 years as relative sea level fell (Ferentinos and McManus 1981; Chap. 23) and historical records show that although the southern part of the Tentsmuir coast has receded over the past two centuries, such erosion has been counterbalanced by shoreline progradation of over 870 m farther north. Nevertheless, projections of future sea-level rise suggest that many Scottish shorelines are vulnerable to increased erosion during this century, with modification or disappearance of sediment-starved beaches and spits, erosion and breaching of dune cordons and reduction in the areas of salt marsh and machair, all which provide important zones of habitat diversity (Chap. 5).

4.10 Conclusion

The Quaternary was a period of recurrent climatic changes, manifest in Scotland by the alternation of cold (glacial) stages and temperate (interglacial) stages, punctuated by stadial and interstadial climatic shifts. Prior to ~0.78 Ma, ice cover in Scotland was mainly limited to mountain areas during successive glacial stages, but up to ten extensive ice sheets subsequently covered most or all of the present land area and extended far out across the adjacent shelves. The cumulative effects of Pleistocene glacial erosion varied spatially. Mountain glacial landscapes developed in the western Highlands; landscapes of areal scouring formed along parts of the western seaboard and Hebrides; upland landscapes of selective linear erosion evolved in the eastern Highlands and Southern Uplands; lowland landscapes of differential glacial erosion and streamlined bedforms developed in the Midland Valley; and landscapes of limited glacial modification were preserved in NE Scotland. Because the last (Late Devensian) ice sheet covered all of the land area, the terrestrial stratigraphic record of earlier events is fragmentary, but the last ice sheet provides a template for understanding the behaviour of its predecessors.

The last Scottish ice sheet expanded after ~35 ka, reached its maximum extent at ~30–27 ka and had retreated to its mountain sources by ~14 ka. To the south it was confluent with ice nourished in Ireland and England, to the east with the Fennoscandian ice sheet and westwards it extended locally to the Atlantic shelf edge. It was polythermal, experienced major changes in configuration and flow patterns and was drained by major ice streams. During the Lateglacial Interstade (~14.7–12.9 ka) ice disappeared or was confined to high plateaux. The Loch Lomond Stade (~12.9–11.7 ka) witnessed a final readvance of mountain glaciers in the form of ice caps, icefields and cirque glaciers. Glacial landsystems associated with the last ice sheet are dominated by till sheets, drumlin fields, ribbed moraine, megaflutes, meltwater channels and outwash deposits. Those produced during the Loch Lomond Stade are dominated by end, lateral and recessional moraines.

Retreat of the last ice sheet was accompanied and followed by the development of permafrost, widespread solifluction and paraglacial landscape modification by rock-slope failure, rockfall activity and reworking of glacigenic sediments by debris flows and rivers. Such processes continued into the Holocene, when periglacial activity became confined to high ground, the incidence of rock-slope failures declined and there was net erosion or incision of talus accumulations, debris cones, alluvial fans and floodplains. Changes in relative sea level during the Quaternary resulted in the formation of shore platforms of varying age and elevation and the formation of Lateglacial and Holocene raised beaches. Many present-day coastal deposits exhibit net sediment loss, which is likely to be exacerbated by future sea-level rise.

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5 Scotland's Changing Landscape

Colin K. Ballantyne, Andrew R. Black, Rob Ferguson, John E. Gordon, and James D. Hansom

Abstract

Landscape changes in Scotland occur in an environment of relative tectonic and climatic stability but widespread human impact. However, climatic trends and projections suggest that gradual warming and loss of snowcover may in the future be accompanied by increases in winter wetness and extreme rainstorm events, particularly in western Scotland. Periodic flood events show no clear historical trend, but are responsible for channel changes in upland rivers, infrastructural damage and urban flooding; increased wetness and ground saturation may increase the frequency of such events. Active rock-slope instability is rare, but there is evidence of increasing debris-flow activity, which has been responsible for disruption of road and rail links. Coastal changes (mainly beach and dune erosion) reflect relative sea-level rise, diminution of offshore sediment supply and the effects of coastal protection structures causing localised erosion and beach steepening. Anthropogenic activity (urban expansion, infrastructure development, land-use change and mineral extraction) has had a major geomorphological and visual impact on the Scottish landscape, at worst resulting in geodiversity loss and landscape blight. Growing recognition of the benefits of geoconservation, however, may help preserve Scotland's outstanding and varied landscapes and landforms.

Keywords

Climate change • Floods • River channel changes • River regulation • Rock-slope stability • Debris flows • Snow avalanches • Coasts • Coastal erosion • Anthropogenic impacts • Geodiversity • Geoconservation

5.1 Introduction

Although the effects of Cenozoic landscape evolution and successive Pleistocene glaciations dominate the geomorphology of Scotland, the postglacial landscape has been modified by a wide range of geomorphological processes. The most dramatic postglacial changes occurred during the Lateglacial and Early Holocene, as mountainsides crumbled, collapsed or deformed, permafrost aggraded then decayed, rivers incised, reworked and re-deposited glacigenic sediments and coastlines receded or prograded in response to changes in relative sea level and sediment supply. In the coastal zone, such changes are recorded in emerged or drowned shorelines; inland, they are represented by a range of relict paraglacial landforms: prehistoric rock-slope failures; gullied talus accumulations and debris cones; and alluvial fans and floodplains now incised and terraced by their parent rivers (Ballantyne 2008, 2019; Chap. 4). Many postglacial landscape elements are therefore relict features formed during rapid landscape adjustment to non-glacial conditions.

As the Scottish pioneers of geomorphology recognised (Chap. 1), however, all landscapes undergo relentless change, albeit very gradual, localised or intermittent.

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Although the era of extensive paraglacial landscape modification and rapid postglacial sea-level and coastal changes had ended by the beginning of the Late Holocene ($\sim$4.2 ka) or earlier, the landscape of Scotland has continued to evolve under conditions of tectonic stability and relative climatic stability. Landscape change during this period has occurred against a background of extensive deforestation and peat growth, and the increasing impact of anthropogenic activity, notably woodland clearance, burning, tillage, overgrazing and, in the past two centuries, urban expansion, infrastructure development, mineral extraction and changes in land use. Recent changes have occurred under a gradually warming climate, from the severe winters and stormy decades of the sixteenth to nineteenth centuries (Chap. 4) to the marked increases in average temperature and precipitation of the past few decades. This chapter summarises the key geomorphological changes that affect Scotland at present, focusing in turn on climatic trends, the recent flood record, changes in fluvial systems, hillslope processes, coastal changes and some key anthropo-geomorphological effects on the landscape. Figure 5.1 shows key sites mentioned in the text.

5.2 Scotland's Changing Climate: Trends and Projections

Climatic trends in Scotland over the century prior to 2004 are summarised by Barnett et al. (2006) and Werritty and Sugden (2012). Mean annual air temperatures across Scotland showed only a modest increase of $\sim$0.5 °C over the period 1914–2004, with marked interannual variability, but a steeper increase of $\sim$1.0 °C over the period 1961–2004. More striking has been a general increase in both precipitation (particularly winter precipitation in northern and western Scotland) and the frequency of heavy rainfall events since 1961. Scotland experienced marked reductions in snowfall and snowcover duration during this period, but several recent winters (2005–2006, 2009–2010, 2010–2011, 2017–2018) have included severe cold spells when temperatures across the country fell to below 0 °C for a week or more due to replacement of westerly flows of warm maritime air by easterly and northerly flows of cold continental air. The frequency of gale-force (>80 km h^{-1}) winds shows no consistent trend over the period 1961–2004.

The effects of recent climate change in Scotland are evident through comparison of data for the decade 2009–2018 with those for the late twentieth century (Kendon et al. 2019; Scottish Government 2019). The average annual air temperature for that decade was 0.67 °C warmer than the 1961–1990 average, annual average precipitation was 15% greater and winter average precipitation was 25% greater. The ten warmest years on record have all occurred since 1997; the warmest year on record was 2014, and the wettest was 2011. Despite such warming, substantial, widespread snow events occurred in 2009, 2010, 2013 and 2018, though the number and severity of such events have declined since the 1960s. Measured by maximum gust speed, there is no persuasive trend in storminess over the past five decades, but total rainfall during extremely wet days (>99th percentile) has increased by 17% since the last decades of the twentieth century over the UK as a whole, and more in Scotland.

Projections of future climatic trends are strongly conditioned by assumed greenhouse gas emission scenarios but have been modelled for the UK for future increases in global temperatures of 2 and 4 °C (Bernie et al. 2018). Under the former scenario, average warming of 1–2 °C is projected across most of the UK, slightly greater in summer than in winter, and there is some indication of a probable shift to drier summers and wetter winters, with significantly increased winter wetness in western Scotland. Under the 4 °C global warming scenario, future trends are similar but more pronounced: average summer temperatures in Scotland are projected to rise by 3–4 °C, winter temperatures by 2–3 °C, with generally drier summers and wetter winters; western Scotland may experience a 10–20% increase in winter precipitation under a regime of strong westerly airflow conditioned by increased frequency of positive North Atlantic Oscillation events. In terms of geomorphological processes, the projected shift to wetter winters in western Scotland is likely to be important in increasing the probability of seasonally high soil moisture conditions, raising water tables and increasing the frequency of soil failures, debris-flow activity and geomorphologically significant flood events should the recent increase in extreme rainfall events continue into the future.

5.3 Floods in Scotland

Floods in Scotland are as diverse as the landscapes on which they occur. Spatial and temporal differences in flood discharge, duration, inundation extent, hydrograph shape, causes and consequences ensure that no two floods are identical. Loss of life due to flooding is rare, but floods often disrupt human activities and infrastructure, such as the evacuation of hundreds of people from their homes in Glasgow in July 2002, following flooding caused by an intense thunderstorm. The causes of river flooding are complex, often depending on the coincidence of multiple factors. Extreme flood events may be caused by rainfall arising from the passage of frontal weather systems that typically extend in swathes over hundreds of kilometres, or from thunderstorms that affect much more localised areas, with much greater rainfall intensities (sometimes briefly exceeding 100 mm h^{-1}) but over durations typically of no

Fig. 5.1 Key sites mentioned in the text. (Map base by Eric Gaba, NordNordWest, Uwe Dedering from Wikimedia Commons CC BY-SA 3.0)

more than a few hours. Irrespective of rainfall type, intensities are normally greater over hills and mountains owing to orographic uplift of air masses and frontal structures. Rapid snowmelt also occasionally generates floods without any rainfall contribution, but more often contributes to flood runoff in combination with frontal rainfall.

Various exacerbating factors can increase flood flows, river levels and impacts (Fig. 5.2). Reduction in infiltration rate due to urbanisation, or parched or frozen ground, may intensify peak runoff rates, although studies of these effects in Scotland are rare. Ground saturation, particularly in winter and spring, is of much more widespread significance. In many Scottish rivers, the highest flood peaks typically occur during passage of a succession of weather fronts. Under these circumstances, temporary storage capacity in soils and shallow aquifers is filled early in a series of rainfall events, such that subsequent events are characterised by higher runoff percentages and often higher peak discharges. Flooding due to ice jams is now rare: historical records document extreme flooding of the River Tay caused by an ice jam blocking the arches of Perth Bridge in 1814, and though ice jams occurred on the River Ayr as recently as 2010, such events are now exceptional.

Changes in land use and land management also influence the propensity for flooding. Compaction of soils due to livestock grazing and the use of agricultural machinery are thought to reduce infiltration and increase runoff rates, while changes in forest cover appear to have less predictable effects. In urban areas, flooding is linked not only to the extent of impermeable surfaces but also the capacity of urban drainage networks. The latter were historically designed to cope with the rainfall of a 30-year storm event but make no allowance for the increase in extreme rainstorm events evident in recent and projected climatic trends (Sect. 5.2), though sustainable drainage systems of ponds, swales and other forms of storage seek to manage flood risks downstream of new developments. Conversely, however, the building of dams for water supply and hydro power, together with a recent shift towards natural flood management, may now be leading to reductions in flooding in some river systems.

Rivers integrate climatic inputs with catchment effects, both natural and human. Given the complexity of runoff-generating processes and the types of change that can occur in river catchments, together with the paucity of systematic hydrological records prior to 1950, no clear pattern emerges of temporal changes in flood magnitudes and frequencies across Scotland during recent decades. Moreover, flood-rich and flood-poor periods tend to alternate at irregular intervals, frustrating attempts to identify trends in river flow records. Research on the flood record of the lower River Tay based on the record provided by flood deposits (Werritty et al. 2006) and the instrumental and historical record of flooding at Perth (MacDonald et al. 2006) exemplifies the challenges involved in identifying trends from the gauged flood record alone. Both studies provided evidence of variability in flood magnitude and frequency over the past 200 years and warned of the dangers of estimating present flood risk using gauged records covering only decadal timescales. A catastrophic flood event on the lower Tay in January 1993, the largest since the ice-jam flood of 1814, resulted in flooding of hundreds of homes, inundation of 52 km^2 of farmland and ~£30 M in direct losses. A flood on this scale could not have been anticipated from the gauged flow record prior to 1993, but analysis of historical data (including flood marks etched on Perth bridge) suggests a recurrence of flood events of this magnitude of ~0.5–1.0% probability in any year.

The recent history of flooding in Scotland has been dominated by a flood-rich period during the late 1980s and early–mid 1990s (Black and Burns 2002). In terms of direct impacts, the greatest losses came from the Strathclyde flood of December 1994 (Fig. 5.3), when widespread, steady

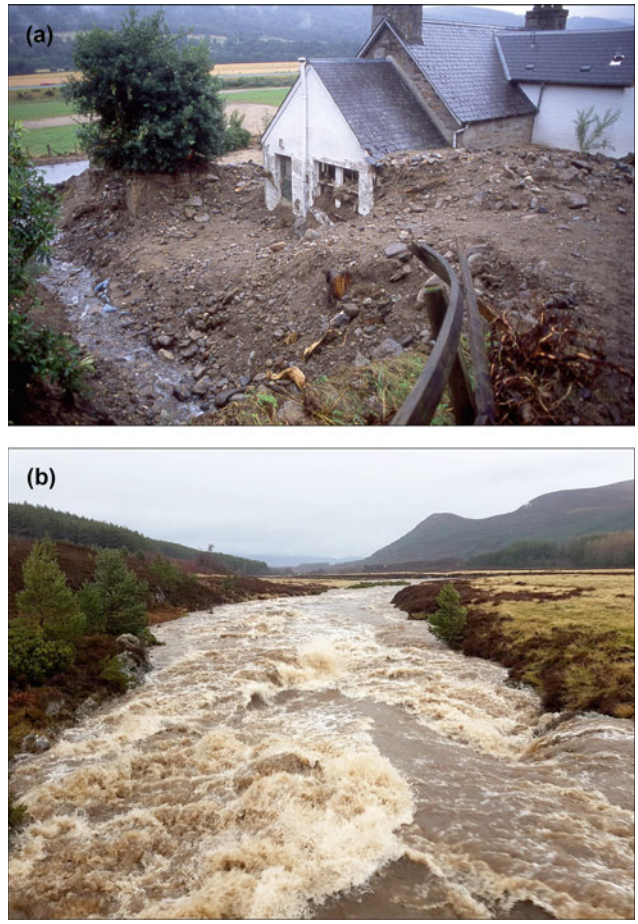

Fig. 5.2 **a** Flood damage caused by coarse sediment transported by a steep hillslope tributary of the River Tummel, August 2002. **b** The River Feshie at the peak of its second-highest flood event in 28 years, December 2018; snowmelt contributed to peak discharge, but the river is contained within its banks. (Images: **a** Andrew Black; **b** Julian Scott)

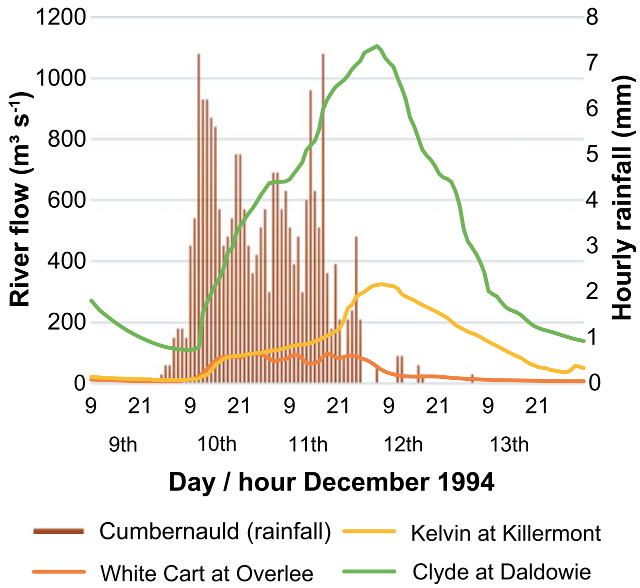

Fig. 5.3 Fairly steady rainfall of 3–4 mm h^{-1} over a 48 hour period during the Strathclyde flood of December 1994 generated contrasting hydrograph responses of the River Clyde and its tributaries, the River Kelvin and White Cart Water

rainfall over 48 hours generated extreme runoff over western Scotland and flooding of areas of high population density, resulting in loss of life, flooding of 700 homes and ~£100 M in direct damages. The scale of losses in this case was probably exacerbated by urban design, diminishing watercourse maintenance effort and social factors. The scale of human impact of this and the January 1993 Tay flood prompted changes in national flood policy, initiated by the 1997 Flood Prevention and Land Drainage (Scotland) Act.

As outlined in the previous section, recent decades have experienced a general increase in wetness in most parts of the UK (Kendon et al. 2019), and future projections indicate further increases of 10–20% or more in annual precipitation in western Scotland, focused in the winter half-year (Bernie et al. 2018), with a possible increase in localised thunderstorm activity. In consequence, river systems across Scotland are likely to experience future increases in flood magnitudes and frequencies, but given the uncertainty in existing estimates of flood risk and the complications of catchment hydrology, it is difficult to predict the scale, distribution and frequency of future flooding events, their geomorphological consequences or their impact on society and infrastructure.

5.4 River Channel Changes

Scotland is a hilly country with a wet maritime climate, so stream power is high in many headwater streams and larger rivers, but sediment supply is generally low as a result of tectonic stability and resistant lithologies. Nearly all Scottish rivers consequently have coarse beds which are fully mobile only in flood conditions. Many reaches are confined by valley sides, glacigenic terraces or flood embankments, and in the Highlands there are many local bedrock knickpoints with very low channel gradients immediately upstream. Active channel change is therefore quite localised and intermittent (Werritty and Hoey 2003) and generally involves reworking of floodplains or removal of sediment from glacial or paraglacial landforms.

5.4.1 Natural Channel Change

Many gravel-bed rivers within or at the edge of the Highlands have wide, shallow channels with alternate bars, and a low-sinuosity 'wandering' pattern. In the absence of confinement, this involves gradual bend development punctuated by chute cutoffs and occasional avulsion. In a few places, the pattern is more or less braided, as in parts of Glen Feshie and near the mouth of the Spey (Chap. 19). In some other places, lateral migration and channel switching have been reduced, but not eliminated, by the construction of flood embankments to facilitate agricultural use of the floodplain, as in the Rivers Tummel and Tay (Winterbottom 2000; Fig. 5.4a). Failure of these embankments in major floods can lead to the deposition of sand/gravel fans on farmland.

Other valleys contain sinuous rivers with abandoned oxbows, but freely meandering reaches are now rare because of river training measures. Bend growth, migration and neck cutoffs in a reach of the Clyde that was not trained until the 1930s have been described by Brazier et al. (1993; Fig. 5.4b).

Many headwater streams and smaller rivers are too confined for much planform change to occur, but they do intermittently remove sediment from valley sides and terraces by basal erosion that triggers shallow landslides. Examples from the Allt Mòr and River Feshie in the Cairngorm Mountains are illustrated in Fig. 5.5. Where tributaries emerge into wide valleys, they deposit fans on which avulsion occurs from time to time in the absence of training works. One of the biggest and most active examples is the Feshie–Spey confluence (Chap. 19).

5.4.2 Effects of, and on, Human Activity

Progressive deforestation throughout the later Holocene, together with climatic fluctuations, must have affected Scotland's rivers through changes in the balance between flood magnitude and sediment supply, but dated evidence is scarce (Chap. 4). The first reliable mapping was General

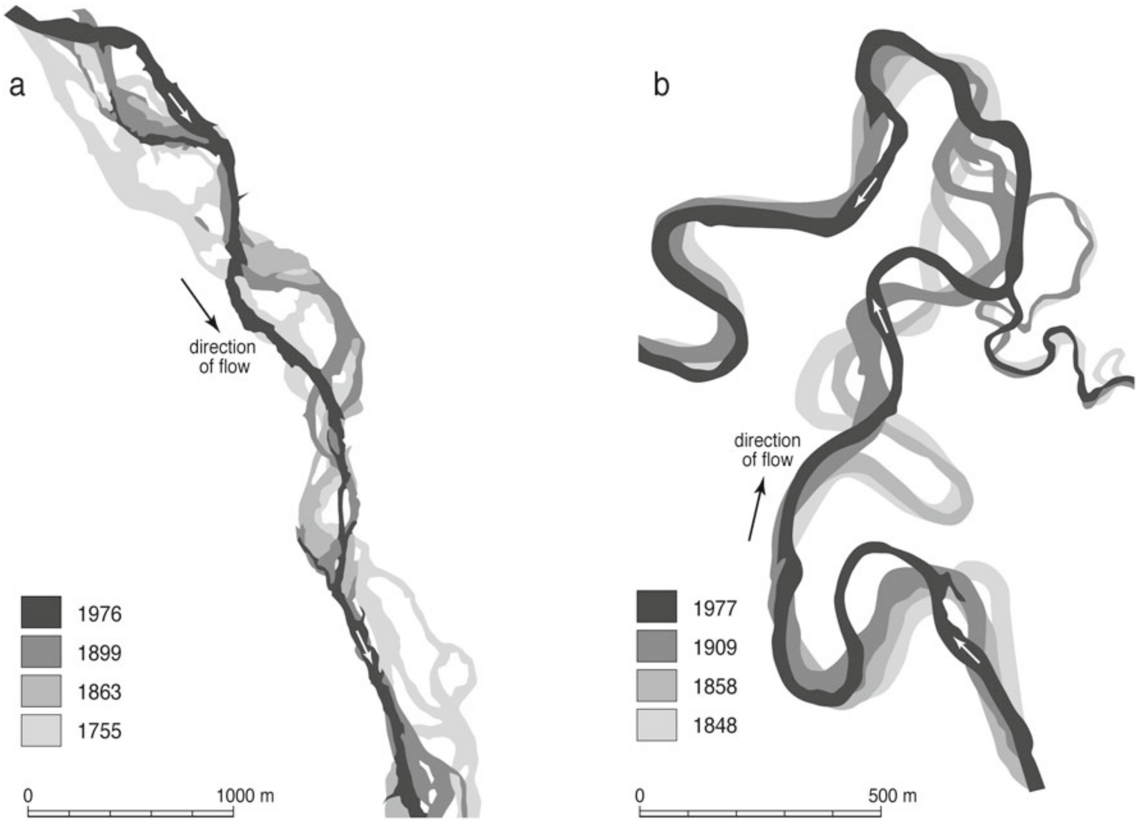

Fig. 5.4 Historical channel changes: **a** the wandering River Tummel (after Winterbottom 2000); **b** the meandering River Clyde (after Brazier et al. 1993). (From Ballantyne (2019) Earth Env Sc Trans R Soc Edinb 110:133–171 © 2018 The Royal Society of Edinburgh)

Roy's 1747–1755 survey of Highland communication routes, followed by the first nationwide Ordnance Survey in the 1860s. The Roy map shows many rivers and streams as having multiple channels, but over the next 100 years or so nearly all rural floodplains with agricultural potential were protected by flood embankments or other interventions to restrict lateral migration. In many cases, channel stability has subsequently been enhanced by riparian tree growth. Urban and industrial expansion was accompanied by hard engineering works on many lowland rivers.

More recently, many Highland rivers and a few in the Southern Uplands have been affected by flow regulation. Scotland has 85% of the UK's installed hydroelectric power capacity involving about 80 main dams, the first built in the 1890s to power aluminium smelting but most between 1930 and 1960 to provide domestic electricity. Many of the schemes divert streamflow from one catchment to another. The receiving streams presumably enlarged their channels but no pre-dam comparative surveys are available. Rivers immediately below dams have conspicuously armoured beds and at least one, the upper Spey, is continuing to modify its channel by marginal aggradation and vegetation colonisation on bars and benches (Gilvear 2004; Chap. 19). Many small run-of-river schemes that involve localised damming along river channels have been constructed in the Highlands in recent years to take advantage of subsidies, but they are unlikely to have more than very local consequences. The same is true of rewilding efforts such as the reintroduction of beavers in some forested lowland areas, restoration of previously channelised reaches and encouragement of floodplain tree growth by controlling deer numbers in Highland glens.

As noted in Sect. 5.3, the frequency of major floods has fluctuated over the last century without any clear overall trend, but it may be starting to increase given the expectation of higher winter rainfall in a warming climate. Much of the ensuing disruption and economic damage results from overbank flooding in built-up areas, but some is directly due to erosion and sediment transport by swollen rivers. If flooding is due to brief convective storms, it has only local effects, as in 1914 when a washed-away railway bridge near Carrbridge caused a train crash in which three people died. Small-scale damage to roads and railways happens somewhere in Scotland several times a year on average, often associated with blocking of bridges or culverts by coarse sediment (as in the 1978 Allt Mòr flood mentioned above; Chap. 19). But flooding caused by heavy frontal rainfall affects much more extensive areas, as in the case of the 1829

Fig. 5.5 Undercutting of glacigenic sediments during major floods: **a** the Allt Mòr shortly after its 1978 flood; **b** the confluence of the Allt Garbhlach (background) with the River Feshie (entering from right in middle distance) eight months after Storm Frank in 2015. (Images: Rob Ferguson)

'Muckle Spate' (Great Flood) in the NE Highlands (McEwen and Werritty 2007). The most recent widespread channel change occurred during Storm Frank at the end of 2015. As well as much small-scale damage, the floods generated by this storm caused the closure of the main London–Glasgow railway line for several weeks, when the River Clyde partly undermined the foundation of a bridge; they also destroyed a 100 m length of the A93 main road inland from Aberdeen when the adjacent River Dee eroded its banks.

5.5 Hillslopes and Mountainsides

5.5.1 Rock-Slope Failures and Rockfall

Although catastrophic rock-slope failures (RSFs) have occurred in Scotland throughout the Holocene (Ballantyne et al. 2014), there are few records of recent large-scale rock-slope instability. There is evidence for a rockslide in Glen Kinglas (Argyll) in the seventeenth century, and twentieth-century RSFs include a rockslide on Raasay and another in Lochaber (Ballantyne 1986). Recent collapse of cliff faces has occurred at a few sites in the Highlands (Fig. 5.6a), notably in Glen Ogle (Perthshire), where a large rockfall caused the premature closure of a railway line in 1965, and from the headwall of Corrie Brandy in the Eastern Grampians, where rockfalls since 2017 have accumulated as talus (Carter and Everest 2019; Chap. 20). The dearth of evidence for recent large-scale slope failures suggests that paraglacial RSF activity is approaching exhaustion: most critically stressed slopes have either failed or regained stability. Only in the coastal zone is there more abundant evidence for recent rock-slope failures, as along cliffed coasts of the Inner Hebrides and Shetland (Ballantyne et al. 2018). There is also little evidence for active rock-slope deformation, though gradual recent movement has been detected on a rockslide above Loch Monar and on a rock slope above Loch Long (Ballantyne 1986) and appears to be occurring on the headwall of Corrie Brandy (Carter and Everest 2019). This paucity of evidence, however, may reflect absence of monitoring; ruptures in the vegetation mat at the crest of a few rock-slope deformations may indicate that not all are completely dormant. Rockfall activity is also much less now than in the Lateglacial and Early Holocene. Most talus slopes in Scotland are relict, vegetated landforms on which erosion by streams and debris flows is now the dominant mode of geomorphological activity (Hinchliffe and Ballantyne 2009; Chap. 4), though isolated examples of larger recent rockfalls are evident at a few Highland locations (Fig. 5.6b).

5.5.2 Debris-Flow Activity

By contrast, evidence of recent debris-flow activity is widespread on steep sediment-mantled slopes, and debris flows (the rapid downslope flow of poorly sorted debris, water and mud) represent the dominant form of recent mass movement in upland areas. These are of two types: hillslope flows initiated by shallow translational failures on open hillslopes (Fig. 5.6c), and channelised flows that follow gullies, often feeding debris cones at the slope foot (Fig. 5.6d), though flows initiated in gullies can encroach on open hillslopes (Fig. 5.6e), and flows originating on open slopes are often captured by gullies, becoming enlarged by addition of sediment from gully walls and floors (Milne et al. 2010). Though hillslope flows are more common, channelised flows deliver much greater volumes (>500–1000 m^3) of debris; hillslope flows usually carry less than 100 m^3 of sediment. Most hillslope flows are initiated on sediment-mantled slopes of 28–42°, though flow-generating failure of soils on slopes as low as 20–22° has been recorded in runoff-focusing hollows (Milne et al. 2009). The spatial density of debris flows tends to be greater on cohesionless sandy soils derived from coarse-grained lithologies, where

Fig. 5.6 Recent mass movement activity. **a** Recent rockslide, Mamore Mountains, Lochaber. **b** Recent rockfall onto a talus cone, Beinn Dearg, NW Highlands. **c** Recent shallow translational landslides and debris flows on a 30° sediment-mantled slope, Eastern Grampians. **d** Gullies excavated in drift-covered hillslopes by repeated debris flows, Drumochter Pass; the sediments have accumulated as debris cones. **e** Debris flows on Beinn Eighe, Wester Ross. **f** Cornice-collapse dirty snow avalanche on Meall Buidhe, central Grampians. (Images: Colin Ballantyne)

debris-flow track densities often exceed 20 per kilometre of slope, a pattern attributed by Milne et al. (2015) to the lower critical-state friction angles in sand-rich sediments.

All recorded debris flows in Scotland have been triggered by prolonged, intense rainstorms, typically exceeding 60 mm in 24 hours with peak intensities >15 mm h^{-1}, often after periods of wet weather or thaw of frozen ground so that antecedent soil moisture conditions are high (Ballantyne 2004; Milne et al. 2009; Winter et al. 2010). During such rainstorms, the rising phreatic surface causes an increase in pore-water pressures and reduction in effective normal stress, leading to localised soil failure as shallow (typically <0.6 m thick) translational slides. The transition from sliding to flow results from apparent liquefaction and remoulding of the mobile soil. Less commonly, debris flows may be generated through mobilisation of sediment on gully floors by flood torrents or by rock-slope failure: toppling failure above the Loch Quoich dam in November 2018, for example, resulted in a flow of ~9000 t of debris that extended downslope for nearly a kilometre, blocking the roadway, destroying a pylon and cutting off the supply of electricity to 20,000 homes.

Debris flows represent a significant hazard for infrastructure, particularly roads and railways routed along the slope foot. In January 2018, a debris flow near Glenfinnan caused derailment of a passenger train, and road closures by debris flows occur annually. The trunk road through Glen Docherty in Wester Ross was blocked by debris in 1957 and again in 1968, and at least three debris-flow events have occurred since then (Strachan 2015). Similarly, the A83 road between Arrochar and Inverary in Argyll has been closed repeatedly by recent debris flows, causing loss of utility and necessitating expensive repairs and remedial measures (McMillan and Holt 2019; Finlayson 2020; Winter 2020). The hazard presented by debris flows was brought into focus in August 2004, when two debris flows in Glen Ogle trapped 20 vehicles on the A85 trunk road, necessitating the airlift of 57 people to safety (Winter et al. 2006; Milne et al. 2009, 2010).

At sites that are prone to recurrent debris-flow activity, return periodicity is decades or less. On the Red Hills of Skye, at least four debris-flow events occurred within the period 1981–2016, and the gullies of An Torc in Drumochter Pass (Fig. 5.6d) have sourced at least seven events in the past 40 years. The frequency of recent debris-flow events has renewed speculation that these are becoming more prevalent (Ballantyne 2004), reflecting an increase in the recurrence of prolonged high-intensity rainstorms of a type that is likely to become more frequent in the future as a consequence of climate change (Winter et al. 2010; Winter 2020).

5.5.3 Snow Avalanches

Although well over 100 snow avalanches are recorded in Scotland in most years, most are small slab avalanches or cornice falls. Where the rupture surface is confined within the snowpack and runout is onto snow-covered slopes, little or no debris is entrained. Even the effects of full-depth 'dirty' avalanches are usually confined to uprooting of turf, erosion of soil and deposition of a thin spread of soil and debris downslope (Ward 1985; Fig. 5.6f). Active avalanche landforms are limited to a small number of high-level sites, notably in the Lairig Ghru (Cairngorms), where intermittently active avalanche boulder tongues extend across the valley floor (Luckman 1992). Although occasional substantial snowfall events associated with easterly or northerly airflow are expected to continue into the future, a general increase in winter warming and the frequency of winter thaw events is likely to continue the current trend of reduced snowfall and snow-lie over the Scottish mountains, so that the frequency of geomorphologically effective snow avalanches is likely to diminish during the present century.

5.6 Coastal Changes

The position of the coast has rarely been stable, and erosion is part of normal coastal dynamics (Duck 2011), with short-term changes superimposed on underlying long-term trends. Rapid postglacial sea-level change, modulated locally by land uplift or subsidence, has driven dramatic movements in the position of the coastline, with the magnitude of lateral shifts influenced by local variations in coastal gradients and sediment supply. The position of modern Scottish shorelines is a legacy of the past and, despite a slowing rate of relative sea-level change over the Late Holocene, present coastal changes may now be gathering pace in response to sediment scarcity and rising relative sea level, together with variation in wave dominance due to fluctuating weather patterns. There is general agreement that the present supply of sediment to beaches both from offshore and fluvial sources is much reduced compared with the past, and this has clear implications for coastal erosion (Komar 2011; Masselink and Russell 2013); sediment-starved coasts are more vulnerable to erosion due to storm events or sea-level rise than coasts where sediment is plentiful (Hansom 2001; May and Hansom 2003). Artificial defence structures also interfere with sediment movement and often exacerbate erosion of adjacent shores (Duck 2011). In a major assessment of sediment availability and movement on the Scottish coast, Ramsey and Brampton (2000) identified eleven discrete coastal sediment cells with negligible sediment exchange between them (Fig. 5.7). With only a few local exceptions, the overall sediment delivery to the coast in these cells is now depleted or negligible, a situation attributed to changes in sea level, wave climate and the impact of artificial defences.

Storm-wave heights are a major factor affecting beach, dune and cliff erosion (Earlie et al. 2015), especially in winter. Bacon and Carter (1993) showed that wave height in the North Sea increased over the 1980s, and the North Atlantic saw increased storm wave activity at the end of the twentieth century (Wolf et al. 2020); in the Northeast Atlantic, winter mean wave height increased over the period 1949–2017 (Castelle et al. 2018). Masselink et al. (2016) demonstrated that the 2013/2014 winter wave conditions on the Atlantic coast of Europe were the most severe since at least 1948, and Brown et al. (2016) have suggested that this has resulted in enhanced coastal vulnerability. Projections for waves in the North Atlantic anticipate that mean wave height may reduce but that the most severe wave heights will increase, with a likelihood of larger wave heights to the north of the UK (Wolf et al. 2020).

While major storms may boost erosion rates over the short term, it has long been known that sea-level rise is an underlying parameter driving coastal erosion, with accelerations in the rate of sea-level rise being accompanied by enhanced erosion (Leatherman et al. 2000; Masselink and Russell 2013). Global average sea levels rose by ~10–15 cm over the twentieth century and by 5.2 cm between 1993 and 2008 (Nicholls and Cazenave 2010). Around the British Isles, the average rate of twentieth-century sea-level rise was ~1.4 mm a^{-1}, again with higher rates more recently. Despite land-uplift rates on much of the Scottish coast being ~1 mm a^{-1} (maximum ~1.7 mm a^{-1}), tide gauge trends suggest relative sea-level rise (RSLR) to have averaged ~3 mm a^{-1} between 1992 and 2013 (Hansom et al. 2017). Ball et al. (2008) investigated three of Scotland's longest-running tide gauges; they attributed flood-frequency changes within the tidal record to underlying changes in mean sea level rather than meteorological effects, suggesting that increased coastal flood risk (and thus coastal erosion) is largely driven by sea-level rise. It is increasingly likely that all Scottish coasts will undergo enhanced RSLR in the future, and a precautionary approach should be informed by the 2018 UKCP high emission (RCP 8.5) scenario (Horsburgh et al. 2020). Under this scenario, Palmer et al. (2018) anticipated a national average RSLR of 0.60 m above 1990 levels for Scotland by 2100, with an expected range of between 0.36 m and 0.96 m. It follows that coastal erosion rates in Scotland are also likely to increase.

Although the Scottish coast is 21,300 km long, only 19% is 'soft' (beaches, dunes and saltmarsh) and potentially erodible; 78% is 'hard', rocky and unlikely to erode at meaningful rates, while 3% is artificial (Fig. 5.7). However, up to 50% of all coastal buildings, roads, railways and water infrastructure (assets worth £13bn at 2017 costings) lie within 50 m of potentially erodible sections of soft coast, with artificial structures protecting £5bn of assets; prior to the analysis provided by the Dynamic Coast project, no national overview existed of the rates and extents of coastal erosion, and coastal erosion-related flooding (Hansom et al. 2017). Dynamic Coast compared the positions of mean high water spring tide for the 1890s, 1970s and the modern period, allowing the extent and rate of shoreline migration to be established, with the more recent rates used to project future erosional loss (Hansom et al. 2017). Since the 1970s, 11% (423 km) of the Scottish coastline has advanced seaward (accreted), 12% (442 km) has retreated landward (eroded) and 77% (2936 km) has remained approximately stable. However, compared with the historic period (1890s to 1970s), and standardised for the difference in time periods, the proportion of retreating coast has increased by 39% since the 1970s, whereas the proportion of advancing coast fell by 22%, with further changes to be anticipated if these recent rates are projected forward to 2050 (Fig. 5.8).

These figures, however, conceal important regional trends. The characteristically open-coast beaches and dunes of the east coast of Scotland have suffered increased erosion extents, dune loss and beach steepening (where low-water positions migrate landward faster than high-water positions) as, for example, at Golspie and Montrose (Chaps. 22 and 23); conversely some east-coast firths and estuaries have experienced increased deposition and accretion, as at Belhaven Bay (Chap. 23; Fig. 5.9). The trends on the west coast and in the Outer Hebrides, Orkney and Shetland are less marked, largely due to the protection from storm-wave activity offered by a crenulate coastline of rocky headlands, islands and inlets. An exception is the outer coast of the southern Outer Hebrides, where erosion and erosion-related flooding of sand-dune dominated coasts place large areas of low-lying agricultural hinterland at risk (Chap. 9). Losses of machair area in the Outer Hebrides and saltmarsh area in the Solway Firth have been linked to sea-level rise (Chaps. 9 and 28). Overall, Scottish sand dune, saltmarsh and machair habitats are anticipated to further reduce from 1900 levels by 36%, 25% and 8%, respectively, by 2060 (Beaumont et al. 2014). In saltmarshes, frontal edge erosion, together with rising ground or defence structures that restrict landward migration, has contributed to 'coastal squeeze' and habitat loss.

A key finding from this national assessment is that compared with the period from the 1890s to the 1970s, average erosion rates have subsequently doubled to 1.0 m a^{-1}, while accretion rates have increased to 1.5 m a^{-1} (Hansom et al. 2017). It seems probable that the recent increase in Scottish coastal erosion rates represents a

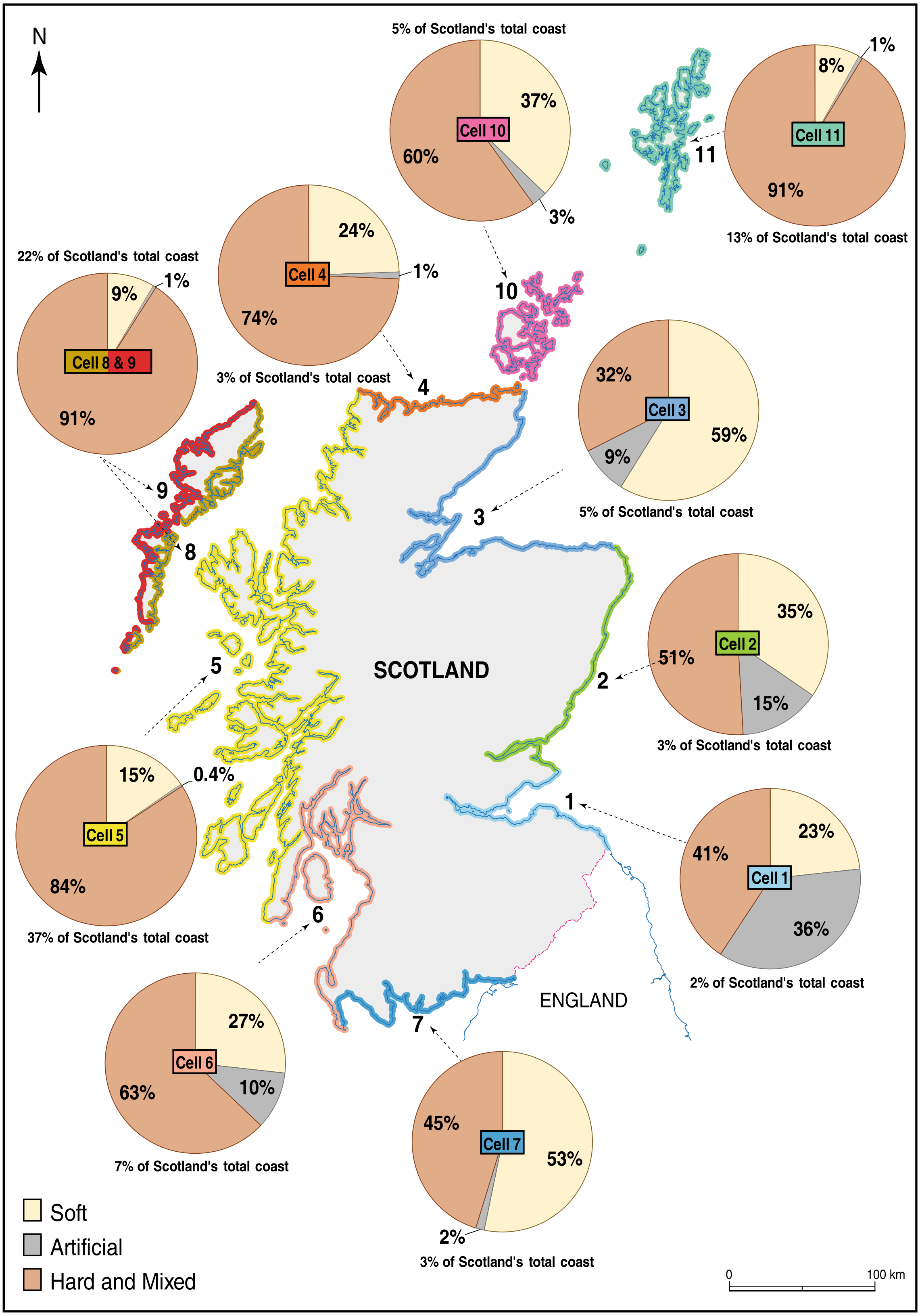

Fig. 5.7 Distribution of coastal types in Scotland within each numbered and coloured coastal sediment cell. (After Hansom et al. 2017)

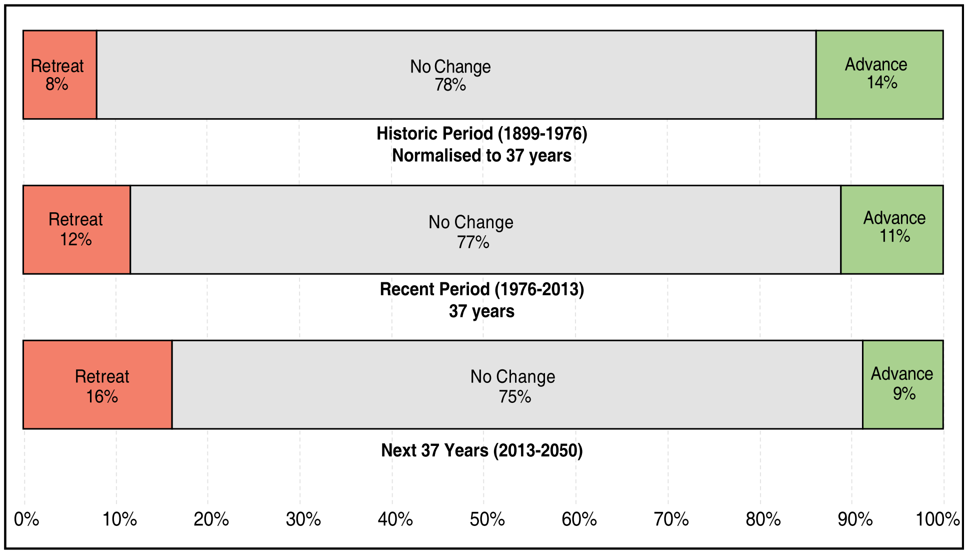

Fig. 5.8 Changes in the amount of coastal retreat (erosion) and advance (accretion) for 1899–1976, 1976–2013 and, based on a continuation of historic and recent trends, projected forward from 2013 to 2050 (normalised for time periods). (After Hansom et al. 2017)

response to changes that are already occurring, particularly rising relative sea level and higher winter-wave impacts at a time when sediment delivery to the coast has diminished. If the observed changes are consistent with climate change expectations (a warming climate, higher relative sea level and increasingly damaging wave events), then a conservative position is to assume that these rates will persist into the future (Fig. 5.8). However, the Dynamic Coast erosion assessments (and the assets affected) are acknowledged to be underestimates, because future changes in RSLR and wave climate are not included. The impact of these changes on assets sited close to potentially erodible coasts will depend on future rates of erosion (Rennie et al. in press). A key factor for future coastal landform and landscape development on erodible coasts is how the risk posed to assets and communities by coastal change will be addressed. There is growing acceptance that hard-engineered solutions may prove unsustainably expensive in the future and that nature-based solutions may offer a viable alternative, together with an anticipation that loss of some coastal land may prove unavoidable and relocation of some coastal assets and infrastructure inevitable (Angus and Hansom 2021). Nevertheless, sandy beaches can survive sea-level rise if they have the accommodation space to retreat landward (Cooper et al. 2020); the Outer Hebrides shoreline, for example, has retreated several hundred metres in places, yet wide sandy beaches survive (Chap. 9). Sea-level rise will inevitably result in shoreline recession, but the future for beaches remains uncertain at locations where their ability to migrate landward is restricted by coastal defence structures.

5.7 Anthropogenic Impacts

Both globally and in Great Britain, humans are now major ‘geomorphological agents’ (Price et al. 2011; Hooke et al. 2012). In Scotland, human impacts on landforms and landscapes have arisen principally from mineral extraction, landfill and land restoration of mineral workings, changes in land use, coastal protection, river engineering and offshore activities (Gordon and Barron 2011). The effects have been both direct (affecting specific landforms) and indirect (such as land-use changes altering catchment runoff or soil erosion). Prehistoric impacts were largely localised, although more pervasive forest clearance and cultivation resulted in accelerated soil erosion and gullying on hillslopes and sediment accumulation in lakes and on river terraces, and incision of alluvial fans and river terraces (Foster et al. 2008; Edwards et al. 2019; Chap. 4). Impacts have accelerated over the last ~200 years through re-shaping the land surface as a consequence of mineral extraction, urban expansion and infrastructure development. Natural geomorphological processes have been altered through engineering works for coast protection and river management, and by dam construction for hydro-electric power. Overgrazing pressure has contributed to accelerated soil erosion and debris-flow activity. These activities have resulted in localised degradation of Scotland's geomorphological diversity through: (i) complete destruction of landforms or partial loss or fragmentation of landform systems; (ii) loss of access, visibility or naturalness of physical

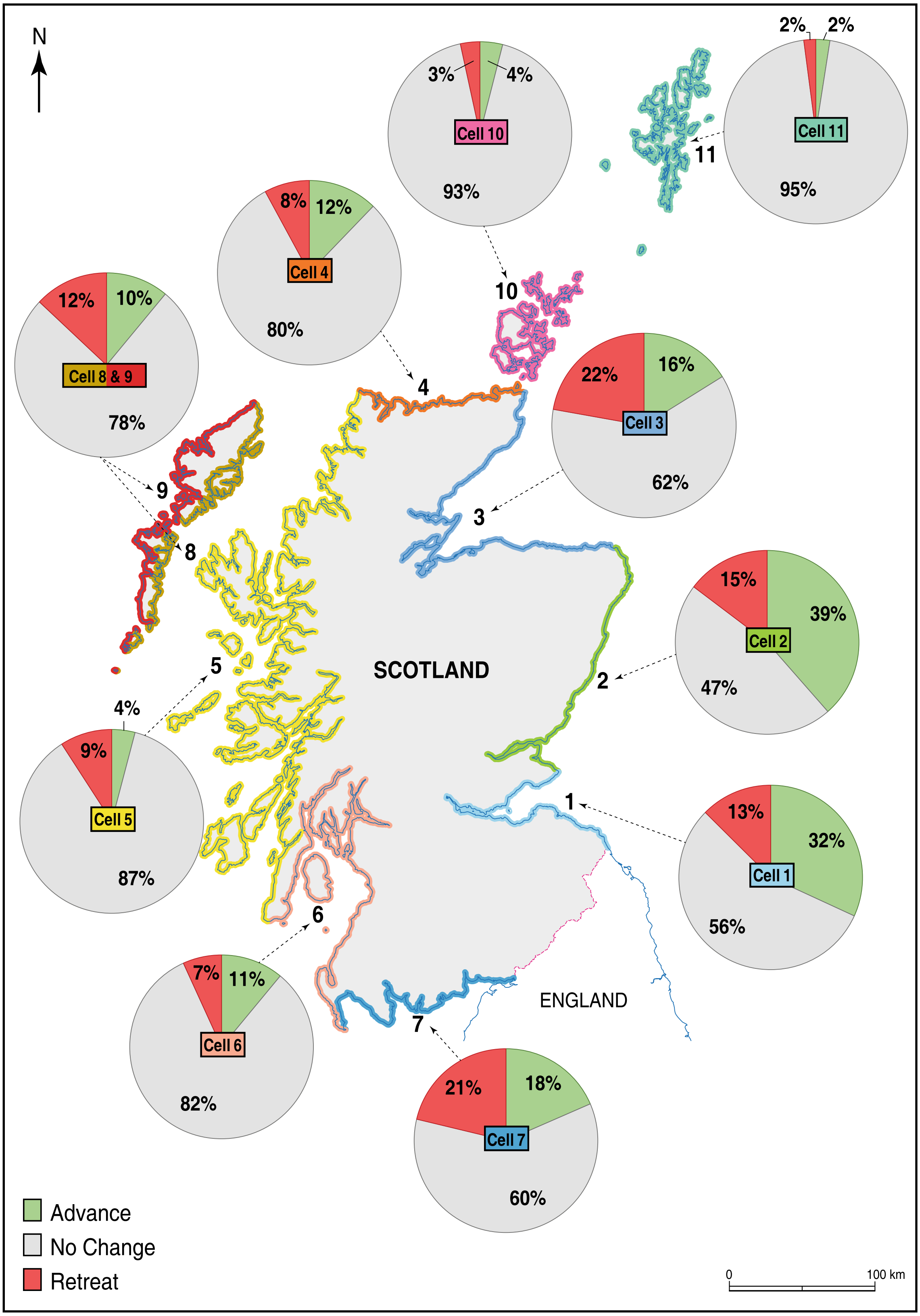

Fig. 5.9 Spatial variation in the amount of retreat (erosion) and advance (accretion) for 1970–2013 by coastal sediment cell. Retreat dominates the outer coast and advance occurs mainly within inlets and estuaries. Note the disparity between the east coast and west coast data. (After Hansom et al. 2017)

features; and (iii) disruption of natural processes and consequent off-site effects such as enhanced erosion or sedimentation elsewhere in a catchment or along the coast.

Human exploitation of geological resources has modified the natural landscape, particularly of the Midland Valley, through quarrying for rock aggregate and building stone, sand and gravel, opencast coal and oil shale. Opencast coal production, in particular, has significantly modified the general form of the topography in the Midland Valley coalfields, resulting in loss of naturalness, while sand and gravel extraction has resulted in the extensive destruction of some glacifluvial landforms (e.g. at Polmont, in West Lothian, in parts of NE Fife and in the Teith valley near Callander; Lowe and Brazier 2021) and frequently in their partial destruction (e.g. at Carstairs in Lanarkshire and Torvean and Strath Nairn, near Inverness; Fig. 5.10a). While there have been short-term benefits in revealing the sedimentary architecture of these landforms, the natural forms have been lost forever. On completion of mineral working, such sites are typically used for landfill or graded and restored for agriculture. However, where many small-scale gravel workings for local purposes have not been backfilled, they provide useful exposures for research and education. Locally, past extraction of beach sand, particularly in the Outer Hebrides, was used for improvement of agricultural land and resulted in beach lowering and exposure of dunes to wind erosion. Also, removal of gravel from riverbeds has locally disrupted channel bedforms, leading to channel process adjustments and sometimes causing accelerated bank erosion downstream.

Mineral extraction has locally led to the creation of new landforms such as the tips (bings) consisting of waste material from the past extraction of oil shale in West Lothian. These form local topographic landmarks exemplified by the Five Sisters Bings, now protected as an Industrial Heritage Site (Fig. 5.10b). An example of imaginative land restoration is the remodelling of one of the bings, Seafield Law, in the form of a crag-and-tail feature that resembles the natural landforms of the area (Gray and Jarman 2003). In a similar creative vein, examples of 'land art' involving earth movement include Charles Jencks' *Landform* installation at the Scottish National Gallery of Modern Art in Edinburgh

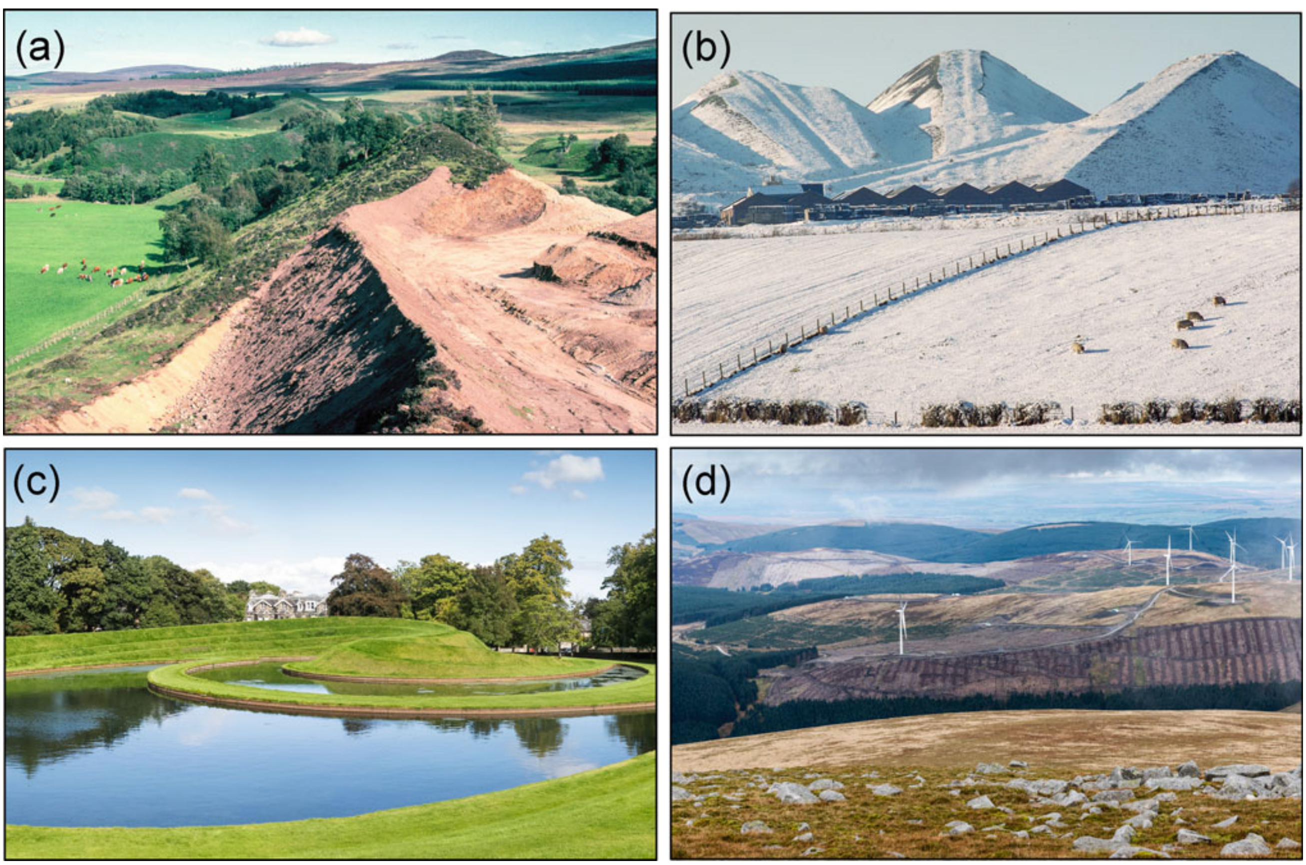

Fig. 5.10 **a** Large-scale sand and gravel quarrying and permanent loss of natural landforms. **b** Part of the 'Five Sisters' oil shale bings, West Lothian. **c** Charles Jencks' *Landform* installation at the Scottish National Gallery of Modern Art, Edinburgh. **d** Landscape impacts from afforestation and construction of wind farms in the Southern Uplands. (Images: **a**, **c**, **d** John Gordon; **b** © Lorne Gill/NatureScot)

(Fig. 5.10c), and his *Crawick Multiverse* in Dumfries and Galloway, the latter developed on the site of a former opencast coal mine.

Changes in land use in the last 250 years have included the introduction of sheep grazing and red deer management for hunting, with historic overstocking considered by some researchers to have contributed to erosion of mineral soils and peat in upland areas, and to enhanced debris-flow activity in conjunction with extreme climatic events (Chap. 4). Construction of access tracks to support upland land-use activities, micro-hydropower developments and wind farms has proliferated, with a strong visual impact but limited damage to landforms (Fig. 5.10d). In the twentieth century, large-scale afforestation has reduced the visibility of many landforms, for example of the eskers at Littlemill in Strath Nairn near Inverness (Chap. 15), meltwater channels at Struie north of Inverness and the emerged coastal plain at Tentsmuir in Fife (Chap. 23). Bulldozed timber extraction roads also have a strong visual impact and local impacts on stream sedimentation. Afforestation was successfully employed to stabilise the formerly mobile sand dunes at Culbin Sands during the first half of the twentieth century (Chap. 22). Proposals to afforest parts of Glen Roy (Chap. 16) were, however, blocked on grounds of their geoheritage value, but neighbouring areas in Glen Spean were planted and are now being harvested.

Coast protection in the form of concrete walls, rock armouring and gabion baskets has obscured parts of important sections in Quaternary deposits (e.g. at Bay of Nigg, near Aberdeen, and Rhu Point, near Helensburgh), and has elsewhere interrupted sediment movements leading to displacement of erosion downdrift (e.g. at Montrose; Chap. 23). Similarly, construction of dams for hydro-electric power generation and river engineering for flood protection and salmon fishing improvements have disrupted natural flow regimes, river-channel landforms and gravel delivery to the coast (Sect. 5.4.2), as on the River Spey (Chaps. 19 and 22), while the River Feshie–Spey confluence fan has a history of channel modification for flood alleviation dating back to the mid-nineteenth century (Brazier and Werritty 1994). Nature-based solutions, however, are now being adopted with potential benefits for biodiversity, flood management and carbon storage, and several river restoration schemes have been implemented, for example on the Eddleston Water in the Southern Uplands and the Water of Mark in the Eastern Grampians (Werritty et al. 2010). Recovery of the Evan Water, a small gravel-bed river in the Southern Uplands, following its diversion to accommodate upgrading of the adjacent A74 road, has demonstrated the value of geomorphologically informed restoration (Gilvear and Bradley 1997).

Flat land on emerged beaches, floodplains or river terraces provides a convenient platform for building and infrastructure, and urban expansion and commercial and industrial developments have all modified such landforms. For example, outwash terraces and emerged beaches in the Forth valley west of Falkirk mapped by Sissons (1976, p. 122) have been heavily obscured, as has the drumlin field on which parts of Glasgow are built (Chap. 26).

Recreation and tourism activities have also involved local infrastructure construction. Unplanned footpath development and resulting soil erosion have proliferated on popular mountains, but the effects are locally mitigated by footpath repair programmes. Construction of golf courses on coastal links and sand dunes has involved re-modelling natural landforms and in some cases stabilisation of active sand fields, as at Menie Links north of Aberdeen. The golf course at Gleneagles, north of Stirling, is unusual in that the fairways have been modelled around the topography of an esker system (Chap. 25). Historically, many sand dunes were destabilised by military use, footpaths and uncontrolled vehicular access (Ritchie and Mather 1984), but generally these impacts have now been moderated by fixed pedestrian access routes and local controls on vehicle access. Paradoxically, ongoing military use of some sand dune sites, such as Morrich More and Barry Links (Chaps. 22 and 23), has resulted in protection of landforms by restricting public access and its associated impacts.

In the marine environment, human impacts on the geodiversity of the seabed have arisen from fishing activities, aggregate extraction, hydrocarbon exploration, renewable energy installations, cables and pipelines, navigational dredging, dredge waste disposal and military activity (Brooks 2013; Chap. 6). As well as direct impacts, these activities have the potential to cause wider indirect effects through the disruption of sediment transport and the need for coastal defence structures where associated installations are sited in areas of potential coastal erosion.

Although many of the pressures on the geoheritage interests of Scotland's landscapes and landforms have been addressed through a range of conservation measures (Chap. 29), these principally apply to nationally protected areas and to a lesser extent to features of regional significance where protection is more discretionary. The wider values of geodiversity at a landscape scale have hitherto received relatively little recognition, but the pressures and threats described above are likely to continue, with the potential for increased rates of geomorphological change in the landscape and their effects compounded by human responses (Brazier et al. 2012). For example, climate change and associated risks of flooding, coastal erosion and coastal 'squeeze' (Sects. 5.3 and 5.6) are likely to lead to increased demands for direct interventions to protect property and infrastructure. However, where natural solutions can be adopted, these may enhance both geodiversity and biodiversity. Rapidly expanding renewable energy developments, particularly offshore and onshore wind farms, have a pronounced visual impact on Scotland's landscapes and seascapes, and further run-of-river hydro-electric schemes may be anticipated with local impacts (Sect. 5.4.2). Demand for carbon offsetting may also drive expansion of commercial forestry. However, future changes in environmental policy that incorporate natural capital and ecosystem services, implementation of nature-based solutions and recognition of the wellbeing benefits from outdoor recreation in natural places should have potentially positive outcomes for geomorphology, as should an increased focus on landscape-scale conservation and enhancement of the role of National Scenic Areas. The National Peatland Plan and the Peatland ACTION programme to restore eroding blanket bogs and reduce carbon losses should see improvements to degraded upland landscapes (Scottish Natural Heritage 2015). Natural woodland regeneration, large-scale native woodland planting and restoration of scrub and its extension up mountain slopes as the climate changes will all impact on landscape character and visibility but also have ecological benefits. Consequently, geoconservation cannot develop in isolation but requires wide adoption and implementation of best-practice management guidelines (Crofts et al. 2020) and better integration in policies and strategies for land use, development planning, and landscape and nature conservation.

5.8 Conclusions

Recent and ongoing changes in the landscape of Scotland occur against a backdrop of relative tectonic and climatic stability but widespread and increasing human impact. Climatic records demonstrate that gradual warming has occurred over recent decades and modelled projections suggest that this trend will continue, with a general reduction in winter snowcover but a continuing likelihood of extreme snow events. More significantly, winter wetness and extreme precipitation events are likely to increase, particularly in western Scotland. Periodic flood events occur due to intense but localised convectional rainstorms or the passage of broad frontal systems, and may be exacerbated by ground saturation or urbanisation, though the historic flood record exhibits no clear temporal trend. Recent floods have caused marked changes in channel configuration in upland rivers. On hillslopes, the dominant effect of rainstorm events has been the generation of debris flows. Recent flooding and debris-flow events have been responsible for significant infrastructural damage, particularly to roads and rail links in upland areas. In coastal environments, the dominant recent landscape changes are twofold: (i) engineering works (coastal defences) that have altered or re-routed sediment supply, causing erosion of some coasts but accretion elsewhere; (ii) diminution of offshore sediment supply, which is locally responsible for erosion of beaches and dune cordons even on coasts unaffected by human activity.

Anthropogenic activity, principally in the form of urban expansion, infrastructure development, land-use change and mineral extraction, is arguably the main driver of change on the Scottish landscape, and likely to continue to be so. Locally, this has had detrimental effects on Scotland's geomorphological diversity through loss or partial loss of landforms or landsystems, regulation of rivers or coastal engineering. Infrastructural developments such as onshore and offshore windfarms, hydro-electric schemes, bulldozed access roads, commercial forestry plantations and clear-felled forestry have a major visual impact on the landscape. Increasing recognition of the benefits that accrue from protection of natural landscapes and their enhancement through geoconservation management, however, offers the possibility of reconciling economic developments with preservation of Scotland's outstanding and varied landscapes and landforms.

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Part II

Landscapes and Landforms

6 Geomorphology of the Continental Shelf

Heather A. Stewart, Tom Bradwell, Gareth D. O. Carter, Dayton Dove, and Joana Gafeira

Abstract

The continental shelf around Scotland covers an area of ~286,500 km^2, around 3.5 times the size of the Scottish landmass. This relatively shallow underwater realm (mainly <200 m water depth) boasts extremely varied geomorphology: from small individual landforms (<100 m^2) to large extensive landsystems (>1000 km^2). These landforms and landsystems relate to both past terrestrial processes, when global sea levels were >100 m lower than at present, and more recent marine processes, active since sea levels rose. This chapter outlines the main geomorphological landsystems found on the shelf, highlighting notable landform examples imaged in high-resolution digital bathymetry data. Many of the landforms have remained exceptionally well preserved since deglaciation, unlike on land, having not been subject to significant disturbance by human activity. The uniquely preserved submarine landforms and landscapes in the shallow seas around Scotland should be protected where possible, especially where they host rare or valuable ecosystems.

Keywords

Glaciation • Multibeam bathymetry • Seabed landforms • Sea-level change • Marine geoconservation

6.1 Introduction

The seabed around Scotland comprises an extremely wide range of geomorphic features and landforms that reflect a variety of environments and processes, past and present, operating over various temporal and spatial scales. These landforms have resulted from tectonic, glacial, paraglacial, mass-movement, fluid-escape, coastal, marine, biogenic and anthropogenic processes. Much of the continental shelf has been periodically covered by ice sheets and other smaller ice masses throughout the Quaternary Period (the last 2.59 Ma), with the last ice sheet reaching its maximum extent at ~30 to 24 ka (Bradwell et al. 2008, 2019a; b; Ballantyne and Small 2019; Chap. 4), during or slightly before the global Last Glacial Maximum (LGM; ~27 to 22 ka). Following ice-sheet retreat, widespread marine inundation of the continental shelf took place with the establishment of marine hydrodynamic conditions. These marine processes continue to modify the seabed environment today.

Formally defined by The Continental Shelf (Designation of Areas) Order 2013, the United Kingdom Continental Shelf (UKCS) encompasses areas of the seabed and sub-surface giving the UK exclusive rights of exploration and exploitation of natural resources within this zone. This Exclusive Economic Zone (EEZ) extends up to 200 nautical miles (370 km) offshore and includes all of the waters adjacent to the UK irrespective of depth (Fig. 6.1). Of the UK EEZ area, Scotland's seas cover ~462,263 km^2 from Mean High Water Springs out to the UKCS limit, an area 6 times the size of Scotland's landmass (Baxter et al. 2011; Scottish Government 2015). Of this vast area, the continental shelf around Scotland covers an area of 286,547 km^2, with

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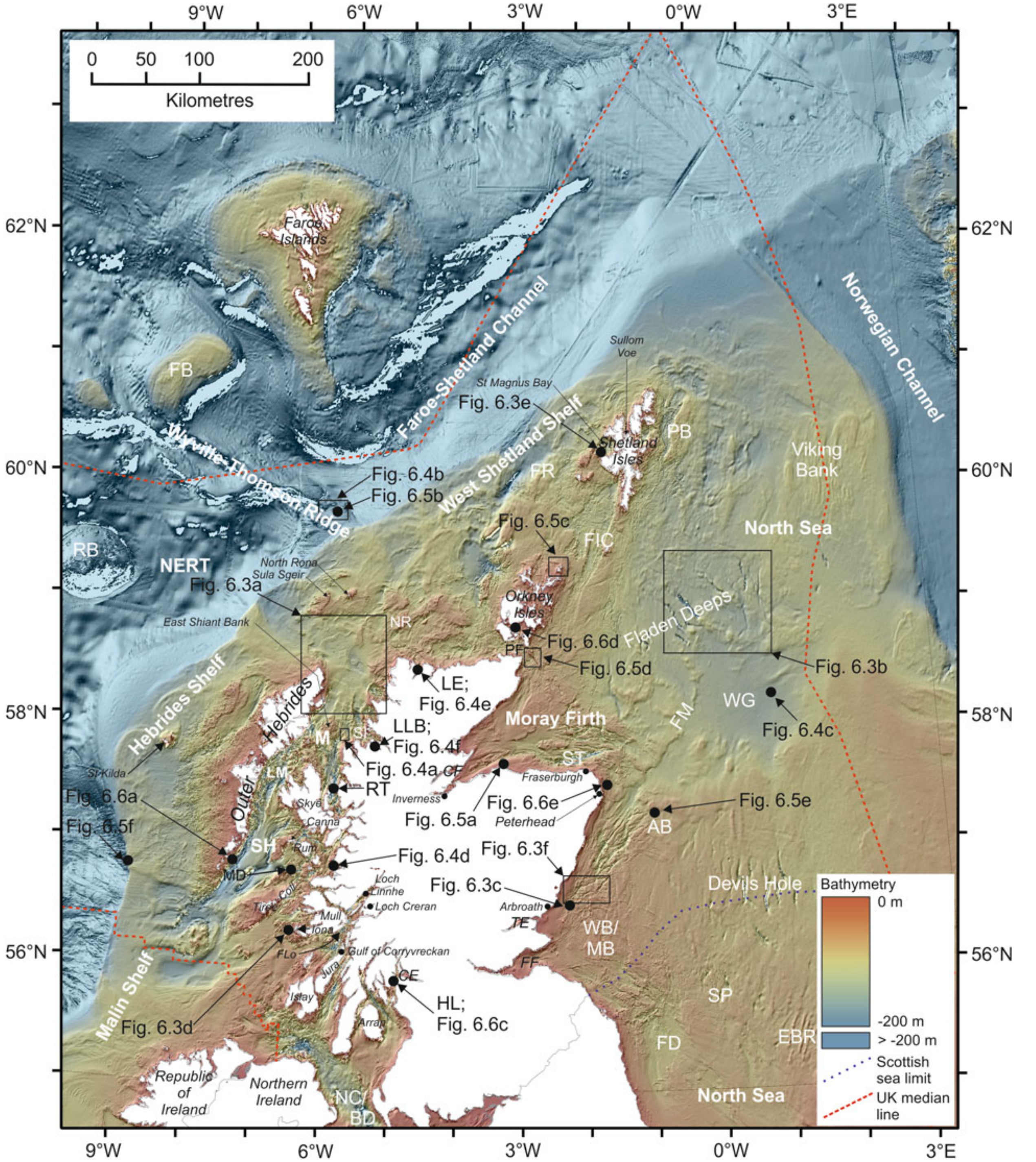

Fig. 6.1 Present-day bathymetry of the continental shelf around Scotland. AB: Aberdeen Bank; BD: Beaufort's Dyke; CE: Clyde Estuary; CF: Cromarty Firth; EBR: East Bank Ridges; FB: Faroe Bank; FD: Farne Deep; FIC: Fair Isle Channel; FM: Fladen Moraine; FF: Firth of Forth; FLo: Firth of Lorne; FR: Foula Ridge; HL: Holy Loch; LE: Loch Eriboll; LLB: Little Loch Broom; LM: Little Minch; M: The Minch; MB: Marr Bank; MD: Muck Deep; NC: North Channel; NERT: North East Rockall Trough; PB: Pobie Bank; PF: Pentland Firth; RB: Rosemary Bank; RT: Raasay Trench; SH: Sea of the Hebrides; SI: Summer Isles; SP: Swallow Pit; ST: Southern Trench; TE: Tay Estuary; WB: Wee Bankie; WG: Witch Ground

Scotland's territorial waters defined as the area within 12 nautical miles (22 km) of the coastline.

The increasing acquisition and availability of high-resolution marine geophysical data over the last two decades have enabled researchers to accurately map seafloor geomorphology, characterize seafloor substrate, identify vulnerable marine ecosystems and manage marine resources. Multibeam bathymetry echosounder (MBES) datasets, in particular, have resulted in a step-change in our understanding of NW European glacial history, and a new understanding of former ice-sheet extent and decay on the continental shelf (e.g. Bradwell et al. 2008; Dunlop et al. 2010; Clark et al. 2012, 2017; Howe et al. 2012; Bradwell and Stoker 2015a, b; Dove et al. 2015). These and other studies co-register high-resolution bathymetry datasets with legacy 2D seismic and geological (core) data to offer new insights on seafloor geomorphology, sub-seabed Quaternary geology and marine ice-sheet behaviour (Ó Cofaigh 2012; Dowdeswell et al. 2016). This chapter highlights the diverse geomorphology of the seabed around Scotland, providing exemplary landforms as case studies.

6.2 Geology, Setting and General Bathymetry

The seabed landscape of the continental shelf comprises striking differences in large-scale geomorphology and smaller-scale landform assemblages east and west of Scotland (Fig. 6.1). The North Sea occupies a shallow epicontinental basin with present water depths typically less than 100 m, generally deepening towards the northern shelf break (at ~200 to 240 m) and eastwards towards the Norwegian Channel (which descends steeply to 550 m). Seafloor depths to the east of Scotland gradually increase with distance offshore, save for a few notable localized deeps in the Moray Firth, the Fladen Ground in the northern North Sea, and offshore Shetland. The east coast of mainland Scotland is notable for the near absence of islands. By contrast, the Hebrides Shelf, to the west of Scotland, has strongly undulating relief culminating in numerous islands, skerries and submerged banks. Furthermore, water depths close to shore west of Scotland often exceed those on the open continental shelf, due to selective glacial erosion exploiting topographical, structural and lithological controls. Good examples of this occur in the Inner Minch and Sea of the Hebrides, where inshore deeps (such as the Muck Deep and Raasay Trench) locally exceed 300 m water depth, greatly exceeding water depths on the Outer Hebrides Shelf (typically 50–130 m below sea level).

The Inner Hebrides Shelf and Minch are confined by the Outer Hebrides in the west and the long, glacially carved coastline of the Scottish mainland to the east. The region incorporates hundreds of islands which, combined with the incised fjordic coastline, corresponds to an extremely complex offshore physiographic environment, characterized by a number of large structurally controlled basins, upstanding volcanic platforms, and glacially over-deepened sea lochs or fjords (McIntyre and Howe 2010). Water depths are variable across this high-relief area, ranging from very shallow (<10 m) to locally very deep (320 m), with a high degree of geomorphological diversity and seabed heterogeneity.

Extensive areas of bedrock occur at seabed around the islands off western Scotland (Fig. 6.1). The well-expressed structural fabric in the Inner Hebrides and The Minch is associated with a series of dominant Late Palaeozoic NNE-trending tectonic faults and basins (Fyfe et al. 1993; Smith 2012; Chap. 2), as well as a number of older and younger intersecting and conjugate faults. The submarine landscape is also punctuated by numerous elongate intrusive igneous structures (e.g. dykes), which most commonly strike NW–SE. Development of NNE-trending basins in both the Inner Hebrides and The Minch modified the pre-existing Precambrian and Early Palaeozoic terrain, with long fault networks along the eastern margin of the Outer Hebrides (Minch Fault) and the eastern margins of Tiree, Coll and Rùm controlling half-grabens that subsequently filled with Mesozoic sedimentary rocks (Fyfe et al. 1993; Stoker et al. 1993; Howe et al. 2015). Due to this tectonic complexity, a range of bedrock strata from Precambrian to Palaeogene age are now exposed at seabed. Broadly speaking, the relatively weak rocks of Mesozoic and Cenozoic age that were predisposed to erosion by repeated Pleistocene glaciation are associated with areas of increased water depth (such as the Little Minch Trough). In contrast, harder crystalline Precambrian basement rocks (e.g. Lewisian Complex), ancient (?meta)sedimentary strata (e.g. Torridonian Group) and volcanic rocks (e.g. Palaeogene volcanics), being more resistant to erosion, remain upstanding (e.g. East Shiant Bank, Nun Rock and North Rona).

Less extensive areas of bedrock crop out at seabed along the North Sea coast, generally only within 5 km of the coast away from the major estuaries around Fife (Carboniferous strata), and from Arbroath to Peterhead (predominantly Dalradian and Devonian strata). There are very few islands offshore eastern Scotland; the small cluster of islands in the Firth of Forth, the subaerial expressions of igneous intrusions, being the most conspicuous. A number of roughly NE–SW trending igneous intrusions crop out at seabed, reaching within 30 m of sea level, ~50 km off the present coastline of SE Scotland (Gatliff et al. 1994).

Holocene sediment cover is patchy and discontinuous on the shelf to the north and west of Scotland. The relative absence of marine (sediment) bedforms in these areas have been attributed to the present hydrodynamic regime, with relatively high current velocities and strong tides removing loose sediment cover and restricting the formation of modern bedforms (Fyfe et al. 1993; Stoker et al. 1993). By contrast, most of the central and northern North Sea is not subject to strong tide- or surge-dominated hydrodynamic conditions; rather, sediment mobility is through wave disturbance, with the exception of the coastal areas of the mainland where tide-dominated currents are typical (Owens 1981). In open ocean settings, such as on the continental shelf west of Scotland, surge-dominated currents predominate with complex semidiurnal tidal cycles and patterns around the intricate, deeply embayed coastline, such as the intense tidal race generated by the Gulf of Corryvreckan (Howe et al. 2015). Stronger current regimes close to the coast and between the Orkney and Shetland islands are responsible for ensuring bedrock remains exposed at seabed.

6.3 Glacial Geomorphology

The onset of the Quaternary Period (2.59 Ma) was marked by intensification of Northern Hemisphere glaciation which strengthened further during the Mid-Pleistocene Transition (~1.25 to 0.7 Ma; Chap. 4). The North Sea Basin preserves evidence for multiple glacial and interglacial cycles within thick sedimentary sequences that capture a detailed record of Northern Hemisphere environmental change (e.g. Graham et al. 2011; Lamb et al. 2017). By contrast, the shelf to the west of Scotland was dominated by glacial erosion during the Mid- to Late Pleistocene and consequently preserves only a discontinuous or partial sedimentary succession in deeper-water basins on the Hebrides Shelf (e.g. Stoker et al. 1993; Stoker and Varming 2011). The largely erosional seabed landscape offshore western Scotland is thought to reflect the impact of fast-flowing ice streams that operated within the British-Irish Ice Sheet (BIIS) on numerous occasions over the last ~1 Ma. These palaeo-ice streams, and their tributaries, dominated the flow pattern and discharge flux of the BIIS (as in present-day Greenland and Antarctica) with the distribution of streaming and non-streaming areas governing the style of subglacial erosion (and/or preservation) of the underlying bedrock landscape (e.g. Hubbard et al. 2009; Bradwell 2013).

Numerous studies have presented and reviewed the evidence for ice-sheet glaciation of the North Sea Basin during the Late Pleistocene (e.g. Gatliff et al. 1994; Graham et al. 2011). High-resolution seismo-acoustic profiles, bathymetric elevation models and shallow marine cores have contributed to a relatively good understanding of the last glacial phase (MIS 2). During this time the BIIS expanded into the central and northern North Sea, and extended to, or close to, the continental shelf edge from the Norwegian Channel and west of Shetland to the Barra–Donegal Fan, NW of Ireland (Bradwell et al. 2008; Graham et al. 2009; Clark et al. 2012, 2017; Sejrup et al. 2016). Detailed seabed mapping of glacial landforms has permitted reconstruction of the pattern of deglaciation following the LGM, at a time of rapidly rising sea levels (Bradwell et al. 2008; Dunlop et al. 2010; Bradwell and Stoker 2015a, b; Dove et al. 2015; Clark et al. 2017; Fig. 6.2). More recently, studies using geomorphological evidence strongly suggest a major dynamic switch in glacial styles on the continental shelf around northern Scotland during deglaciation, from predominantly terrestrial to strongly marine-influenced ice-sheet retreat, resulting in rapid ice-mass losses at key time intervals (Bradwell et al. 2008, 2019a, b; Clark et al. 2012; Sejrup et al. 2016).

6.3.1 Large Moraines

On the mid- and outer shelves, large moraines deposited by the BIIS take the form of broad, curvilinear, sometimes arcuate, occasionally overlapping ridges of glacigenic sediment. These moraines range in length from 5 to 20 km and width from 0.5 to 3.5 km and are typically 10–40 m high (e.g. Stoker and Holmes 1991; Stoker et al. 1993; Bradwell et al. 2008, 2019a, b; Stoker and Varming 2011). The best-preserved examples of these large Late Pleistocene end moraines and glacitectonic push moraines are located on the West Shetland and Hebrides shelves (Figs. 6.2 and 6.3a). These moraines relate to shelf-wide ice-sheet glaciation and many occur adjacent to or directly inshore of large continental slope fans such as the Foula and Rona Wedges, the Sula Sgeir Fan and Barra-Donegal Fan (Stoker et al. 1993; Bradwell et al. 2008; Bradwell and Stoker 2015a). Prominent seafloor ridges 5–40 m high and 1.5–5 km wide extend up to 200 km along the outer West Shetland Shelf (Stoker et al. 2006; Stoker and Varming 2011). Of the large moraines preserved on the Hebrides Shelf, the most distinctive occurs at the shelf edge, is 20–30 m thick, up to 4 km wide and can be traced laterally for around 70 km (Stoker and Holmes 1991; Bradwell and Stoker 2015a). Several ice-marginal moraines on the North Sea Shelf are associated with ice emanating from the Moray Firth (Hall et al. 2003; Bradwell et al. 2008; Sejrup et al. 2009). Perhaps the clearest of these, the Fladen Moraine, forms a ridge 2–6 km wide and 20–30 m high, and is thought to have formed at ~18 to 16 ka (Sejrup et al. 2015; Fig. 6.2). Other large moraines east of mainland Scotland, such as the Wee Bankie and Marr Bank moraine complexes, once thought to mark the LGM offshore ice-sheet limit (Bowen et al. 2002), are now considered to mark significant recessional stages during overall ice-sheet retreat (Golledge and Stoker 2006; Bradwell et al. 2008; Clark et al. 2012; Fig. 6.2).

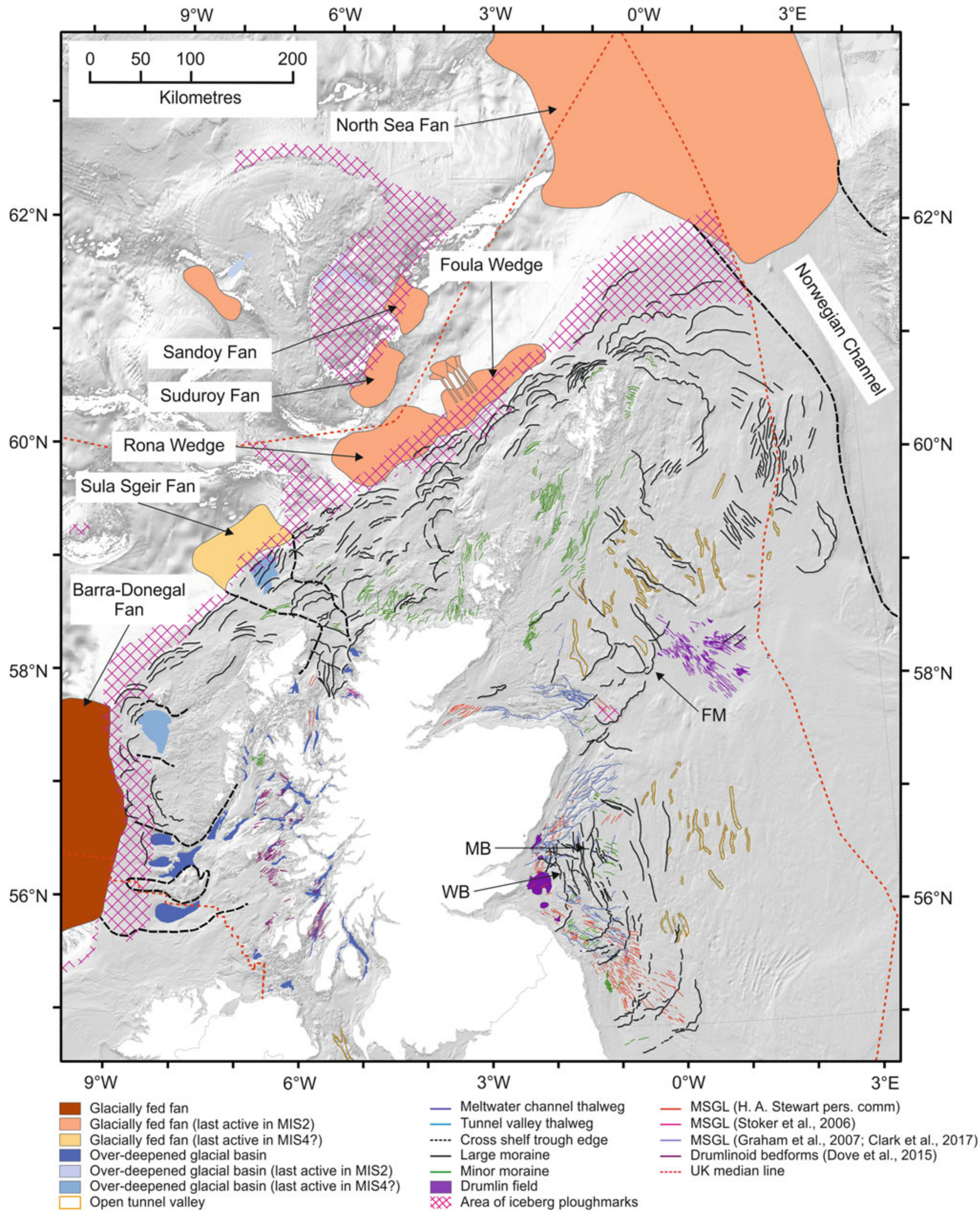

Fig. 6.2 Large-scale subglacial landforms on the continental shelf around Scotland, including: tunnel valleys and streamlined bedforms; ice-marginal landforms represented by large moraines; glacimarine landforms such as trough-mouth fans and iceberg ploughmarks; and glacifluvial landforms such as meltwater channels. MSGL: mega-scale glacial lineation

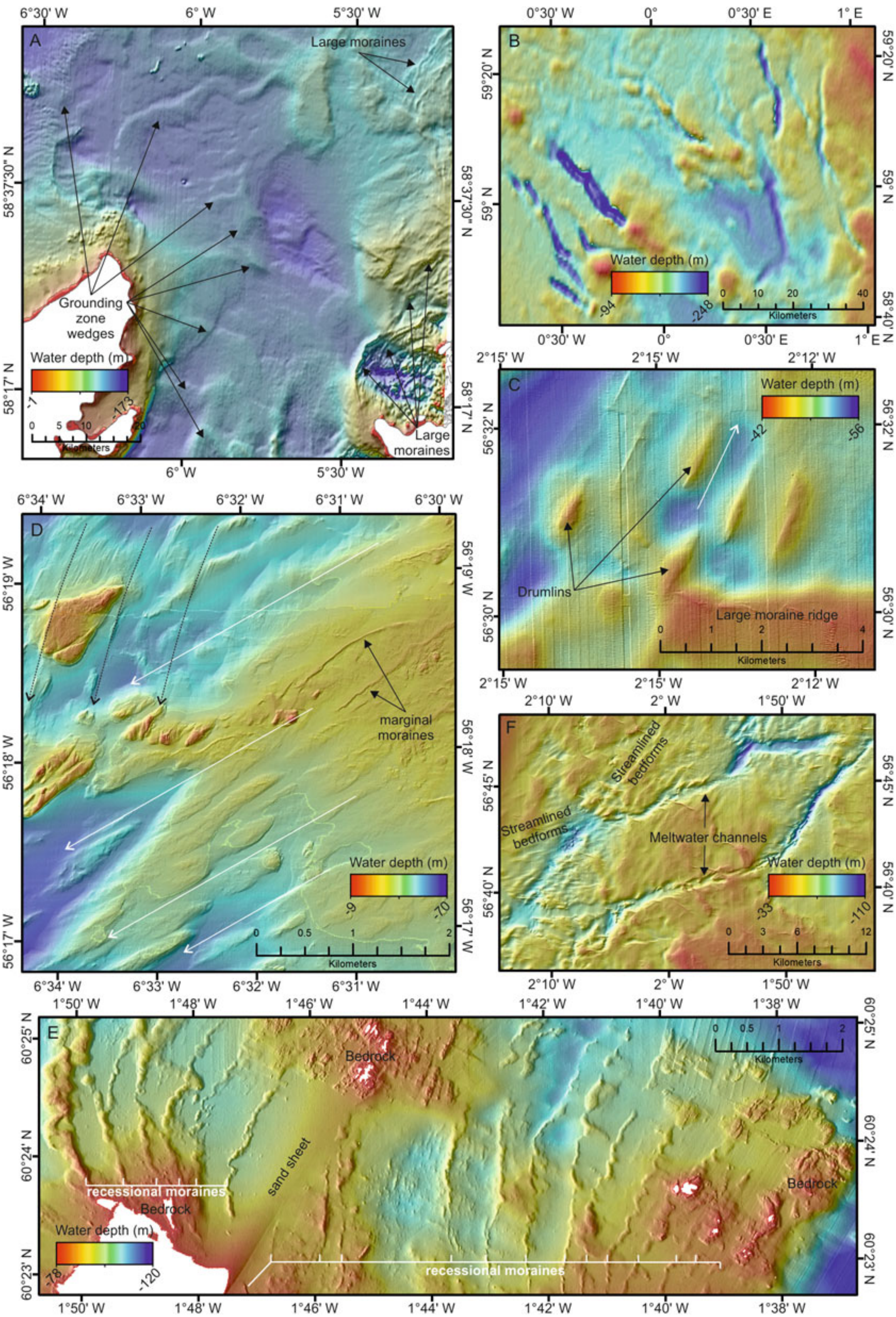

Fig. 6.3 Glacial geomorphology of the seafloor. **a** Example of grounding zone wedges (GZWs) recording grounding-line positions during retreat of the Minch Ice Stream and large moraines mapped in the Minch. **b** The morphology and distribution of a number of large tunnel valleys from Fladen Deeps incised into the seabed, northern North Sea. **c** Drumlin field, part of the offshore expression of the Forth-Tay Ice Stream indicating flow in a NE direction (white arrow). **d** Example of cross-cutting flowsets of streamlined bedforms from the Hebrides Ice Stream, offshore the west coast of Iona. The dominant SW-oriented flow set (white arrows) is superimposed by a later, SSW-directed flow set (black dashed arrows) controlled by local topography. **e** Around 20 recessional moraines demonstrating the stepped eastward retreat of a tidewater ice-margin, St Magnus Bay, west of Shetland. **f** Meltwater channels incised into glacigenic deposits offshore Montrose, eastern Scotland

6.3.2 Grounding-Zone Wedges

Within cross-shelf troughs west of Scotland, broad transverse ridges of glacigenic sediment with very low-angle, asymmetric cross profiles have been mapped as grounding-zone wedges (GZWs; Callard et al. 2018; Bradwell et al. 2019a). GZWs form by the accumulation of glacigenic sediment in the cavity between the ice-sheet grounding line and the projecting (floating or buoyant) ice shelf. Modern examples occur within the tracks of ice streams that have receded, for example on the inner West Greenland Shelf and in the Amundsen Embayment, West Antarctica (Dowdeswell et al. 2016).

Seventeen GZWs (or hybrid forms) have recently been identified within the path of the former Minch Ice Stream off NW Scotland (Bradwell et al. 2019a; Fig. 6.3a). The distribution of GZWs within the Minch ice-stream trough suggests punctuated or episodic recession of the grounding line during overall ice-stream retreat. Bradwell et al. (2019a) related an abrupt reduction in size of GZWs to the geological transition from a soft/weak sediment bed on the outer- and mid-continental shelf to a hard/strong bedrock bed within the Inner Minch. Other GZWs have been identified in the path of the former Hebrides (or Barra-Donegal) Ice Stream on the Malin Shelf, marking episodic retreat of the BIIS from its maximum LGM extent at ~26.7 ka (Arosio et al. 2018; Callard et al. 2018). Fewer GZWs have been identified in the North Sea Basin, possibly owing to the shallower water depths, though a well-developed GZW has been identified by Sejrup et al. (2015) offshore NE Scotland, west of Fladen Ground, indicating ice sourced from the east. Additionally, putative GZWs have recently been identified offshore SE Scotland within the track of the former North Sea Lobe ice stream (Roberts et al. 2019). Although their offshore signature is predominantly beyond Scotland's seas, the most northerly of these GZWs, charting retreat towards the Firth of Forth, is within Scotland's marine realm.

6.3.3 Tunnel Valleys

Tunnel valleys are generally considered to form beneath ice sheets, roughly perpendicular to the ice-sheet margin during deglaciation, and their orientations have been used to infer approximate ice-flow directions (Ó Cofaigh 1996). Tunnel valleys preserved on the seafloor around Scotland (Fig. 6.2) are interpreted as belonging to a single 'generation' relating to LGM ice-sheet deglaciation and MIS 2 flow re-configurations (Stewart 2016).

The most conspicuous bathymetric features within the central and northern North Sea are the Fladen Deeps and Devil's Hole Deeps (Stewart 2016; Fig. 6.1). The Fladen Deeps (Fig. 6.3b) comprise 21 elongate or linear channel-like incisions, the farthest east of which are oriented roughly north–south, whereas the western deeps are oriented roughly NW–SE. The Fladen Deeps range in length from 6.5 to 35 km and are 2–5 km wide; individual channels are incised 50–100 m (maximum 150 m) below the surrounding seabed, attaining maximum water depths of up to 290 m. The Devil's Hole Deeps are located further south and comprise 13 roughly north–south, linear incisions that reach a maximum of ~190 m water depth, incised up to 110 m below the surrounding seabed. The Devil's Hole Deeps range from 10 to 45 km in length and are 2–3.5 km wide.

The largest tunnel valley preserved on the seafloor of the North Sea region is the Southern Trench, in the Moray Firth, which is ~60 km long and up to 210 m deep. The Southern Trench and adjacent east–west oriented deeps close to the Moray and Buchan coasts cut down into Mesozoic bedrock and have been linked to catastrophic meltwater discharge during Late Pleistocene ice-sheet retreat (Long and Stoker 1986). Farther inshore, another seafloor channel oriented roughly NE–SW, located in the Inner Moray Firth, is believed to be an eastward continuation of the Beauly and Ness valleys near Inverness (Chesher and Lawson 1983).

Although clearly defined tunnel valleys are absent from the continental shelf west of Scotland, a number of overdeepened troughs with tunnel-valley-like characteristics are present in the submarine landscape of the Inner Hebrides, most notably west of Muck and north of Rùm (Howe et al. 2012), and Beaufort's Dyke in the North Channel of the Irish Sea between SW Scotland and NE Ireland (Callaway et al. 2011). These tunnel-valley-like glacial overdeepenings have exploited structural weaknesses in the bedrock but, unlike true tunnel valleys, may have formed over more than one glacial cycle.

6.3.4 Streamlined Bedforms

Large, elongate streamlined landforms, comprising sediment- and bedrock-dominated drumlins, crag-and-tail features and mega-scale glacial lineations (also described as mega-flutes and mega-grooves in bedrock) are interpreted to have formed subglacially beneath relatively fast-flowing but firmly grounded ice. The orientation (or long axis) of streamlined forms reflects the former ice-flow direction. Well-preserved examples of streamlined subglacial bedforms offshore Scotland are found in The Minch (Bradwell and Stoker 2015a; Bradwell et al. 2019a), the Inner Sea of the Hebrides (Howe et al. 2012; Dove et al. 2015), offshore eastern Scotland (Golledge and Stoker 2006; Sejrup et al. 2016), in the North Channel (Gandy et al. 2019) and to a lesser degree around Shetland (Bradwell et al. 2019b). Offshore eastern Scotland, highly elongate, flow-parallel, mega-scale glacial lineations and shorter-elongation, flow-aligned drumlins (Fig. 6.3c) are indicative of fast-flowing and persistent ice-sheet flow from both the Forth-Tay and Strathmore Ice Streams during MIS2 (Golledge and Stoker 2006).

Dove et al. (2015) attributed the distribution of streamlined bedforms on the seafloor around the Inner Hebrides to the onset zone of the Hebrides (Barra-Donegal) Ice Stream that drained ~5 to 10% of the BIIS at its maximum extent. The majority of the mapped streamlined bedforms associated with the Hebrides Ice Stream indicate former ice flow to the southwest, although locally bedform orientations can be more variable reflecting topographic and structural influences (Dove et al. 2015). In places, such as off the west coast of Mull and Iona, cross-cutting bedform flowsets suggest ice-sheet re-organization as the dominant SW-directed flow regime evolved into a weaker later-stage SSW-directed flow (Fig. 6.3d).

6.3.5 Smaller (Recessional) Moraines

Suites of similar-sized, relatively small, transverse ridges of glacigenic sediment on the seafloor are interpreted as recessional moraines formed by an intermittently retreating, marine-terminating ice-sheet margin. Typically these more delicate features are <15 m high and 30–200 m wide with fairly regular spacing between ridges (Fig. 6.3e). Often these features are De Geer moraines (a type of recessional moraine), recording the incremental retreat of a grounded tidewater glacier-front over time. The moraines locally overprint streamlined bedforms, thus confirming a deglacial origin and late-stage timing. Where mapped on the continental shelf around Scotland, these suites of recessional (or De Geer) moraines provide important insights into the retreat pattern and style of individual sectors within the last BIIS (Bradwell et al. 2008, 2019b). Comparable modern examples are found at the termini of retreating tidewater glaciers, such as those in the fjords of west Spitsbergen (Ottesen and Dowdeswell 2008; Dowdeswell et al. 2016).

A well-preserved suite of 40–50 recessional De Geer moraines traverse bathymetric highs and deeps and are draped on larger (older) moraines in the vicinity of the Summer Isles in NW Scotland. Identified from high-resolution MBES data (Stoker et al. 2006), these small moraines are indicative of ice-frontal deposition or sediment squeezing during oscillatory retreat of a lightly grounded marine-terminating ice-sheet margin (Bradwell and Stoker 2015a, b, 2016; Chap. 13). Morphologically similar fields of recessional moraines have been identified in the shallow waters around Orkney and Shetland and in the Fair Isle Channel (Bradwell et al. 2008, 2019b; Bradwell and Stoker 2015b). An excellent, well-preserved suite of 20 recessional moraines in St Magnus Bay, west of Shetland (Fig. 6.3e), charts the punctuated retreat of a partially grounded or buoyant tidewater ice-margin during the final stages of ice retreat towards the Shetland mainland at ~18 to 17 ka (Bradwell et al. 2019b). Farther south, discontinuous recessional moraines that occur preferentially on bedrock highs within the Sound of Jura in the Inner Hebrides, are thought to have been deposited following the demise of the Hebrides Ice Stream as tidewater glaciers receded inshore (Dove et al. 2015).

6.3.6 Meltwater Channels

Meltwater channels incised during ice-sheet retreat are relatively common on the Scottish North Sea continental shelf but are less common offshore western Scotland. Such meltwater channels differ from tunnel valleys (Sect. 6.3.3) in that they are subaerial networks of anastomosing channels formed by high water and sediment discharge from a former ice margin, causing incision into both soft/weak sediment beds and hard/strong bedrock beds. Some of the most striking examples trend roughly parallel to the Scottish east coast from Arbroath in the south to Peterhead in the north (Fig. 6.3f). Many are incised into the Marr Bank and Aberdeen Bank glacigenic formations and relate to melting during the later stages of ice-sheet retreat (Golledge and Stoker 2006). This extensive meltwater channel network covers an area that extends ~100 km from south to north and ~85 km from west to east; individual channels range from 200 m to 2.5 km in width, with depths ranging from 10 to 75 m into the surrounding seabed. Other prominent dendritic meltwater channel networks are present 27–40 km northwest of Fraserburgh and elsewhere in the outer Moray Firth.

Within the Sea of the Hebrides, around Rùm and Canna, meltwater channels of complex or composite origin are eroded into both bedrock and sediment. These features, interspersed with drumlinoid bedforms, take the form of weakly sinuous to curvilinear channels 1–6 km long and 200–500 m wide, and are incised 10–50 m below the surrounding seabed. They are thought to have been carved by the flow of both subglacial and proglacial meltwater (Howe et al. 2012). Similar channel features, with complex genesis, occur on the seabed in the vicinity of the Summer Isles and in the eastern Minch (Stoker et al. 2006).

6.4 Glacimarine Geomorphology, Fluid-Escape and Mass-Movement Landforms

Glaciation has left clear geomorphological hallmarks on the seafloor around Scotland (Sect. 6.3). However, the waning of ice masses has also physically modified the submarine landscape in other ways.

6.4.1 Iceberg Ploughmarks

Iceberg ploughmarks are formed where the keels of drifting icebergs gouge and scour the seabed, and may reflect former ocean currents and/or dominant wind patterns. They are usually randomly organized and overlapping in plan view, and have limited water-depth ranges. Individual ploughmarks typically follow an approximately straight or slightly curved course, although 'wandering', sinuous and contorted forms have all been observed on the seafloor in Scottish seas (Long et al. 2011).

Iceberg ploughmarks are found in deep-water localities such as the Malin Shelf (Dunlop et al. 2010), and elsewhere on the tops of topographic highs, on submarine banks, and on the crests of large moraines and GZWs (Bradwell and Stoker 2015a, b; Fig. 6.4a). Large fields of iceberg ploughmarks have also been identified on the outermost portions of the Hebrides and West Shetland shelves (Stevenson et al. 2011; Cotterill and Leslie 2013; Bradwell et al. 2019b; Fig. 6.4b), but are less common on the seafloor of the North Sea, probably due to a paucity of deep-water basins.

Individual ploughmarks typically consist of raised ridges, or berms, separated by shallow depressions generally 20–80 m wide and averaging 1–3 m deep (Long et al. 2011; Stevenson et al. 2011), though exceptionally ploughmarks can exceed 10 m in depth (e.g. Cotterill and Leslie 2013; Stewart and Long 2016). Ploughmarks varying in width and depth, and locally exceeding 100 m from berm to berm, can indicate where icebergs have lodged in the seafloor, forming meltout iceberg pits (Long et al. 2011). Unusually wide (>200 m) linear ploughmarks with shallow U-shaped (width: depth ratio >100:1) or multi-bermed cross profiles have been taken to indicate very large ice-shelf generated icebergs, thought to result from ice-shelf breakup events during retreat of the Minch Ice Stream (Bradwell et al. 2019a).

6.4.2 Pockmarks

Seabed pockmarks are typically formed by the focused expulsion of biogenic or thermogenic fluids from within sub-seabed sediments, resulting in the removal of fine-grained sediments and the creation of a shallow, circular or elliptical depressions in the seafloor (Judd and Hovland 2009). Pockmarks are found sporadically on the continental shelf and in nearshore waters around Scotland, predominantly in muddy, organic-rich sediments. Good examples of large pockmarks occur in deeper mud-laden basins within west coast fjords or sea lochs, or adjacent to fjord mouths (Stoker et al. 2006; Howe et al. 2012; Audsley et al. 2019). Morphometric analysis has shown these large west coast pockmarks range from 100 to 300 m in diameter and are up to 15 m deep (below the surrounding seabed). Some pockmarks form discrete chains, whereas others are elongate depressions thought to be the result of preferential bottom currents (Fig. 6.4c). The age and activity status of pockmarks in Scottish fjordic settings remains uncertain, although some deep pockmarks (>10 m) appear to have been active over a prolonged period judging by the degree of Holocene hemipelagic sedimentation surrounding them (Audsley et al. 2019).

The mud-dominated sediments of the Witch Ground Formation and the fine-grained Flags Formation occur extensively at seabed across much of the central and northern North Sea, and are known to be significant shallow gas-bearing units (Long 1992). This area of the North Sea Basin, near the eastern margin of the UK EEZ, exhibits an unusually high concentration of small pockmarks (30 km^{-2}), typically less than 50 m in diameter. Mapping studies have shown a reduction in pockmark density (<5 km^{-2}) and an increase in pockmark diameter (100–150 m) close to the basin edge (Gafeira et al. 2018). The vast majority of pockmarks within the Witch Ground Basin are less than 3 m deep (Gafeira et al. 2018) and may represent periodic gas-escape activity over at least the last 8 ka (Judd and Hovland 2009). Additionally, several unusually large pockmarks with diameters of up to 450 m and depths of 15–18 m are found in isolation or form pockmark complexes (Fig. 6.4d). These include the 'Scanner', 'Scotia', 'Challenger' and 'Alkor' pockmark complexes, located close to the margins of the Witch Ground Basin, and are thought to have been created during catastrophic gas-escape events (Long 1992; Gafeira et al. 2018; Böttner et al. 2019).

Pockmarks are rare across the mid- to outer West Shetland and Hebrides shelves, although an area ~55 km NE of North Rona hosts a small number of (possibly relict) pockmarks (Stoker et al. 1993) at 80–150 m water depth, and >50 pockmarks have been identified west of Orkney. Other isolated pockmarks have been observed in sub-bottom profiler data from mud-dominated basins east and west of Shetland.

6.4.3 Submarine Landslides

Submarine landslides are associated with a variety of triggering mechanisms, including dynamic loading of contouritic horizons in deeper water, changes in sedimentation

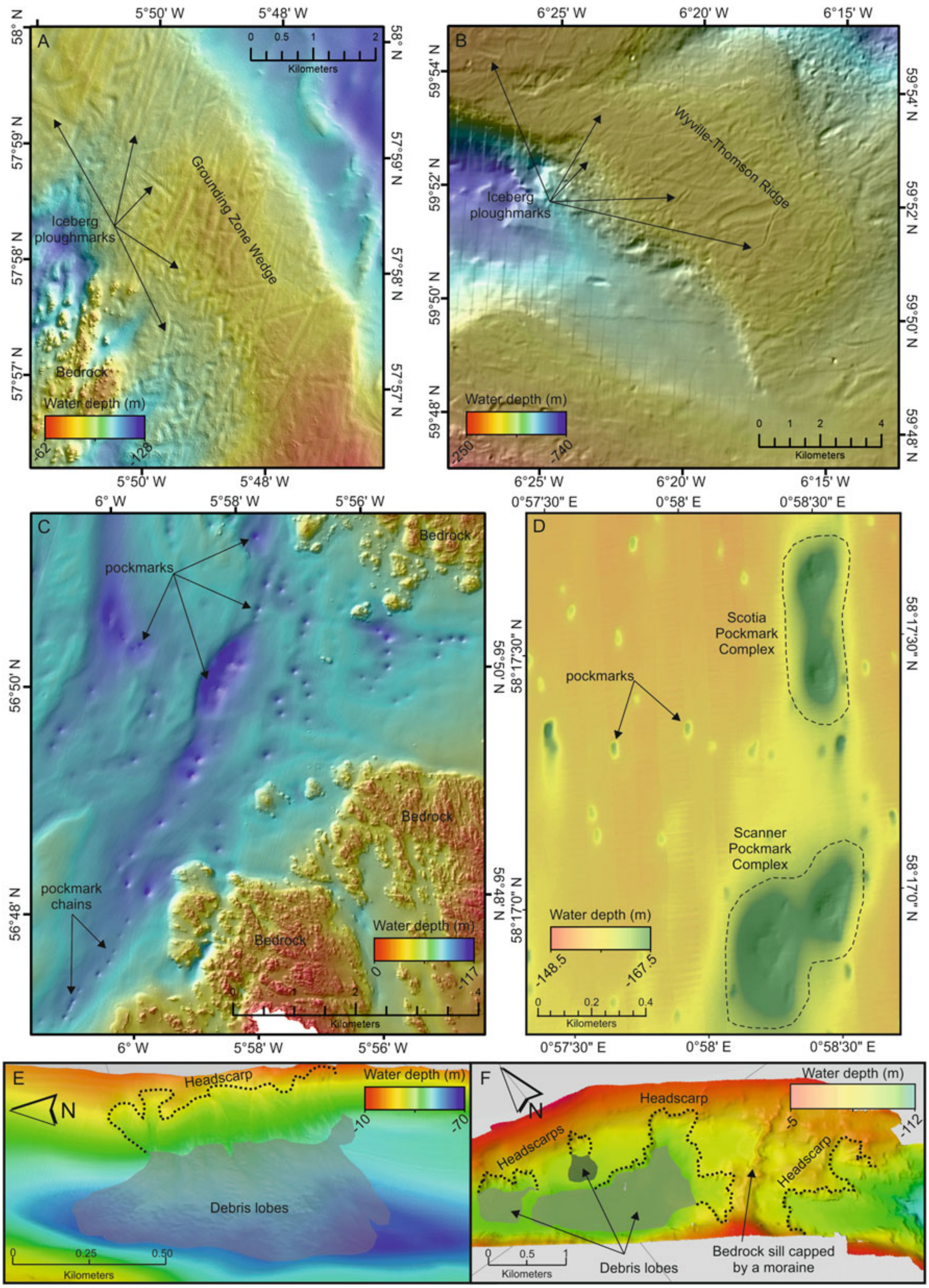

Fig. 6.4 Glacimarine geomorphology, fluid-escape and mass-movement landforms of the seafloor. **a** Iceberg ploughmarks located on the crest of a grounding zone wedge located within The Minch. **b** Iceberg ploughmarks on the Wyville-Thomson Ridge. **c** Example of pockmarks from nearshore fjordic environments and fjord approaches, east of Rùm and near Arisaig. **d** The Scotia and Scanner pockmark complexes. **e** Example of debris lobes, Loch Eriboll. **f** The Little Loch Broom paraglacial slope failures (Scoraig, Carnach, Rireavach and Badcaul slides)

style related to glacial and interglacial cycles, seismicity triggered by glacio-isostatic unloading, and instability due to gas escape.

Submarine slopes have been undergoing periodic readjustment since MIS2 ice-sheet deglaciation, through the Lateglacial period and into the Holocene. The retreat of the BIIS resulted in high sedimentation rates on slopes with thick accumulations of unconsolidated glacimarine and morainic sediments. Ice-sheet loss also radically changed the stress fields in the shallow crust of Scotland and the surrounding continental shelf, resulting in differential unloading and glacio-isostatic uplift rates, with the greatest rebound near the former ice-sheet centre (Smith et al. 2019; Chap. 4). Unsurprisingly, submarine mass-movement features on the continental shelf are mainly confined to steeper slopes within the more neotectonically active areas, such as the fjords of western Scotland.

In NW Scotland, along the eastern submarine slope of Loch Eriboll, a series of submarine debris flows have resulted in multiple scars in soft glacimarine sediments and the build-up of debris lobes up to 4 m thick (Fig. 6.4e). The lack of post-failure sediment cover on the debris lobes indicates a relatively recent event, probably Holocene in age (Carter et al. 2020). More complex submarine mass-wasting features have been identified farther south in the fjords of Wester Ross. Carter et al. (2020) have described several subaqueous mass-movement scars and deposits exhibiting translational, rotational and planar debris-flow failures in the steep-sided fjord of Little Loch Broom (Fig. 6.4f). The large translational and rotational failures of the Little Loch Broom Slide Complex and the Badcaul Slide occurred at $\sim$15 to 13 ka BP, shortly after deglaciation of the fjord (Stoker et al. 2010). Comparable submarine slope failures, evident in MBES data, have been noted in the Sound of Mull, Firth of Lorn, Loch Linnhe and Holy Loch along their steeper glacially modified slope sections. Although undated, these features postdate deglaciation and are very probably Holocene in age (Carter et al. 2020).

In deeper waters, seabed-headwall and slip-surface scars are mainly confined to the continental slopes beyond the West Shetland and Hebrides shelves, and the Rockall Trough. The majority of these mass-movement features have been attributed to the Pleistocene or earlier, including the Sula Sgeir Fan debris-flow deposits along the Hebrides Slope (pre- or Early Devensian), and the multi-event Peach Slide Complex of Late Devensian to Early Holocene age on the northern slopes of the Barra-Donegal Fan (e.g. Evans et al. 2005; Long et al. 2011). On the West Shetland Slope two distinct headwall scars and associated debris-flow deposits are located $\sim$100 km northwest of Shetland. The Afen Slide affects an area of $\sim$40 km^2 and displaced $\sim$0.2 to 0.4 km^3 of sediment; morphological relationships show four stages of failure indicating different modes of sediment transport (hydroplane flows and block slides; Wilson et al. 2004). The Walker Slide is a small ($\sim$0.002 km^3) mass movement scar at a similar water depth (850 m) to the Afen Slide but $\sim$17 km to the northeast (Long et al. 2011).

6.5 Marine Landforms and Bedforms

Erosional submarine landforms generated by the action of the sea in the (former) littoral zone are typically cut into bedrock; whereas marine bedforms in the littoral or sublittoral zone are accretionary, mobile and composed of sediment.

6.5.1 Submerged Platforms

Evidence for relative changes in sea level is preserved at seabed around Scotland, primarily in the form of eroded rock platforms that are broadly horizontal with a steep seaward edge, or gently seaward sloping, in water depths of up to 120 m, and potentially up to 155 m. In contrast to the well-studied terrestrial record of relative sea-level change in Scotland (Smith et al. 2019), there have been relatively few investigations of the now-drowned continental shelf.

Smith et al. (2019) have synthesized geomorphic evidence of sea-level change from around Scotland, with prominent submerged platforms identified within the Firth of Lorn (Hall and Rashid 1977), as well as offshore Shetland and Orkney (Flinn 1969; Chaps. 7 and 8). These features are largely eroded across bedrock, where the base level of erosion is a function of the interplay between eustasy, glacio-isostasy and potentially neotectonics. The identification of these features is difficult without high-resolution MBES data and developing chronological constraints is a considerable challenge. Stoker and Graham (1985) have provided the only relative age control on submerged platforms in Scottish waters by ascribing ages to the underlying and overlying strata (through seismostratigraphic analysis), bracketing formation between MIS11 and MIS2. Smith et al. (2019) predicted that with the increased application of MBES data, further submarine evidence of sea-level change is likely to be identified; indeed, recent work around Orkney has identified a pronounced sequence of bedrock platforms or terraces probably relating to former sea levels (Dove et al. 2021).

6.5.2 Sandy Bedforms

Mobile (and relict) marine bedforms, such as sand waves, megaripples, and sand ribbons, are mainly found in areas of sandy seabed on the inner shelf, in shallow estuarine or nearshore settings (e.g. Fig. 6.5a) where the present hydrodynamic conditions are conducive to the movement of clastic sediments.

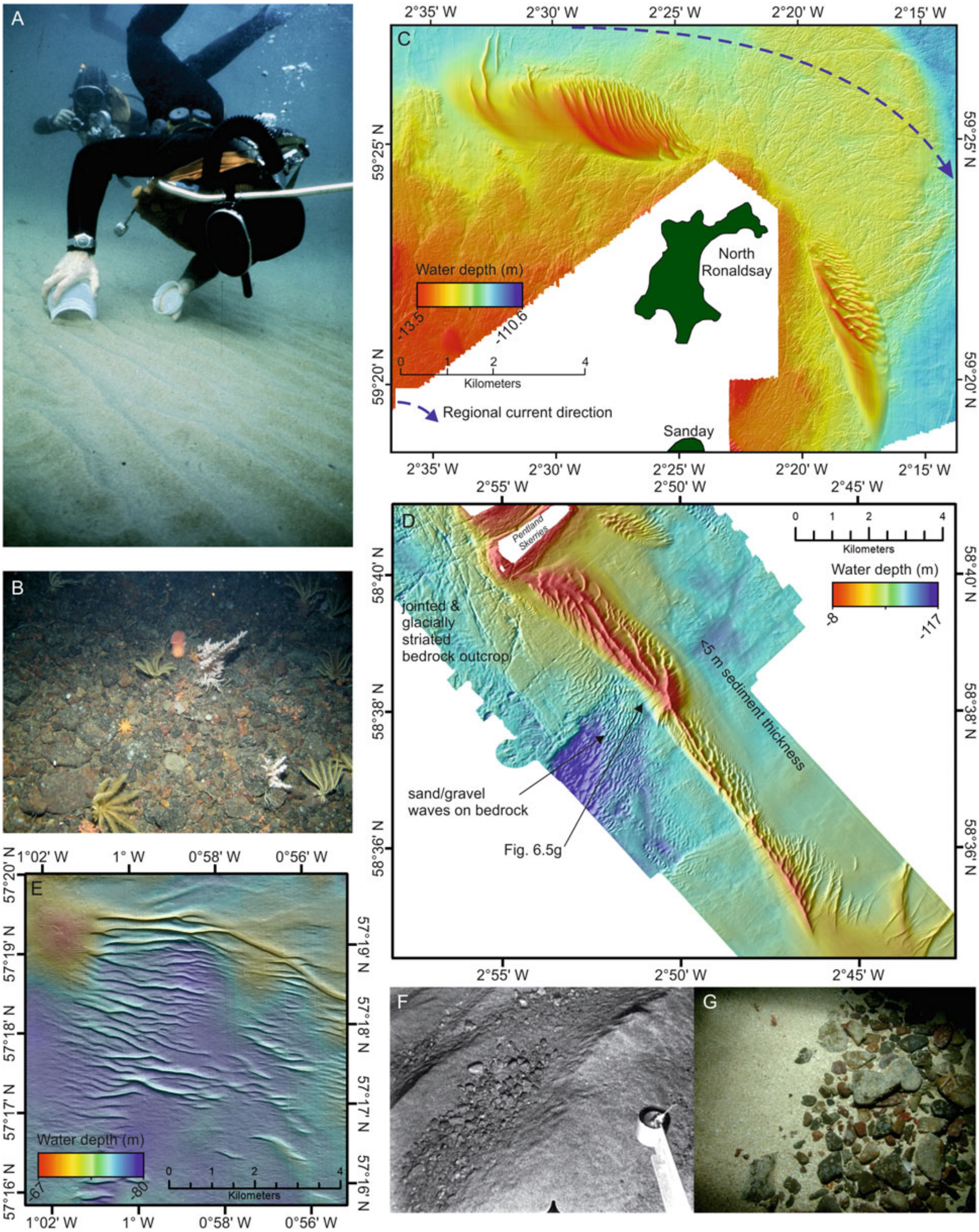

Fig. 6.5 Postglacial marine bedforms of the seafloor. **a** Seafloor photograph of sand ripples offshore Lossiemouth. **b** Cobble gravel on the outer Hebridean Shelf. **c** Well-defined sandbanks located to the north and east of North Ronaldsay, Orkney; the sand waves measure up to ~12 m in height with 200–500 m wavelengths. **d** Sandy Riddle, a large gravel and sandbank located immediately east of the Pentland Firth. **e** Large, relict sand waves in ~75 m water depth, ~50 km offshore eastern Scotland. **f** Parallel linear ripples with gravel-rich troughs on the Outer Hebrides Shelf (after Stow et al. 2002, courtesy of the Geological Society of London). **g** Seafloor photograph showing sediment wave migration, Sandy Riddle. For location see Fig. 6.5d

An abundance of sand waves and megaripples occurs on the seabed in the vicinity of Orkney and Shetland (Stoker et al. 1993). Farrow et al. (1984) reported three large (30 m high, 10 km long and 0.5 km wide) shelly carbonate sand (and gravel) banks northeast of Orkney (Fig. 6.5c). They attributed these conspicuous marine bedforms to the influence of islands and headlands on the local hydrodynamic regime. The orientation of smaller sand waves on these sandbanks, here and in the wider Orkney region, suggests a west to east (broadly clockwise) direction of sediment transport, with storm waves probably playing an important role in sediment mobility. Radiocarbon dates obtained from these sand wave sediments indicate a pre-Holocene age, likely to reflect recycling of older shelly material within the sediments (Farrow et al. 1984).

Farther south, at the eastern entrance to the Pentland Firth, between Caithness and Orkney, is the large 'Sandy Riddle' banner bank (Fig. 6.5d). This large, complex marine bedform consists of an asymmetric sandbank, 10–12 km long, 1–2 km wide and ~60 m high, formed of calcareous coarse sand and gravel mainly deposited during the Holocene. However, the sandbank may have formed atop a core of pre-existing glacigenic sediment (Fairley et al. 2015). Active mobile sand waves and megaripples, up to 10 m high and 80–200 m in wavelength, are superimposed on the flanks of Sandy Riddle. Smaller but equally active sand wave fields, west of Stroma within the Pentland Firth, have wavelengths of up to 400 m, and occur in medium to coarse sands. Tidal currents here average 2.0 m s^{-1} and can exceed 5.0 m s^{-1}, one of the strongest tidal streams in European waters (Fairley et al. 2015).

Large sand waves (with heights of up to 17 m and wavelengths of 200 m) occur in shallow coastal waters around Peterhead, often with smaller sand waves or climbing megaripples on their stoss sides (Gatliff et al. 1994). Farther offshore (~50 km east of Peterhead) there are large sand waves 8 m high, with 160–270 m wavelengths, in water depths of 60–80 m. This area presently experiences surface tidal currents of only 0.3–0.5 m s^{-1}, and the sand waves are thought to be relict features that are now mobile under extreme storm conditions or perhaps relate to a period of low relative sea level (>10 m below present) during the Early Holocene (Owens 1981; Gatliff et al. 1994; Fig. 6.5e).

West of Scotland, on the Hebrides Shelf, the direction of sediment transport is generally from south to north (Inall et al. 2009). Asymmetric sand waves occur as isolated features or clustered in fields closer to shore, for example in the Sound of Harris and off the west coast of Lewis. Long sand ribbons are present around many of the Outer Hebrides islands, where tidal currents are consistently high (>1.0 m s^{-1}) and locally focused (Stoker et al. 1993). Sand streaks, sand ribbons and longitudinal sand patches are widespread on the West Shetland and Hebrides shelves where they range in size from several metres up to several hundreds of metres in width and up to a few kilometres in length (Stevenson et al. 2011; Cotterill and Leslie 2013). In places, these thin mobile bedforms overlie gravel lags or have gravel-rich troughs within them (Fig. 6.5f, g).

6.5.3 Gravel Bedforms

Gravel-dominated bedforms are rarer on the continental shelf, but gravel lags (Fig. 6.5b) are not uncommon across the shelf, especially in shallower waters around large mid-shelf bathymetric highs such as Sula Sgeir, North Rona and the Outer Hebrides Platform. In general, the terrigenous gravelly sediments offshore Scotland have been winnowed from the underlying Pleistocene deposits (e.g. glacial diamict) during the Lateglacial to Early Holocene marine transgression. This has resulted in the formation of lag deposits consisting of poorly sorted, subangular to subrounded lithic clasts of gravel grade (Fig. 6.5b, f). On the outer continental shelves, characterized by iceberg ploughmarks, irregular gravel ridges are prevalent, slightly raised above the general level of the otherwise sandy seafloor.

The seabed on much of the Orkney-Shetland Platform exhibits winnowed gravel lag accumulations, as do the isolated bathymetric highs of the Pobie Bank and Viking Bank where gravel deposits are common. Farther south, in the eastern North Sea, bathymetric highs such as the Marr Bank and Aberdeen Bank are also covered in extensive gravel lag deposits (Johnson et al. 1993; Gatliff et al. 1994). Wewetzer et al. (1999) reported abundant gravelly sand dunes, 0.5 m in height and 2–10 m in wavelength, in the middle Tay estuary in water depths of <10 m.

The North Channel between SW Scotland and NE Ireland is an area of high peak-tidal currents in excess of 1.5 m s^{-1}. The seabed here hosts transverse gravel ridges draped on bedrock. These low-relief gravel bedforms are 1 m wide, up to 0.6 m high, and reach 1.25 km in length (Fyfe et al. 1993).

6.6 Reefs as Biogenic Landforms

Biogenic seabed structures are created by the marine flora and fauna in the benthic zone (Gordon et al. 2016), potentially comprising the organism itself (e.g. maerl beds, horse mussels beds and *Sabellaria* spp.), or arising from the organism's activities (e.g. whale feeding marks) or the effects of biota on other hydrodynamic and sedimentation patterns (e.g. cold-water coral sediment mounds). Very few biogenic structures in extra-tropical oceans are large enough to be classed as submarine landforms in their own right. However, cold-water coral reefs function as 'ecosystem engineers', trapping sediments, altering local oceanographic currents and providing a refuge for other organisms. A rare example of an extensive cold-water coral reef within Scotland's territorial waters is the Mingulay Reef Complex in the Sea of the Hebrides, which comprises several areas supporting reef mounds preferentially located on bedrock highs. De Clippele et al. (2017) delineated over 500 *Lophelia pertusa* 'mini-mounds' between 13 and 60 m wide and up to ~100 m long located on one ridge named 'Mingulay Reef 01' (Fig. 6.6a, b). Cold-water coral reefs are both long-lived and slow-growing, and are therefore vulnerable to damage from fishing activities and climate change.

Large spreads of seabed gravel with abundant sessile (attached) and mobile biota may technically fall within the definition of 'reefs' as specified in the UK, European and International legislation. Areas of coarse gravel, or stony reef, provide important seabed habitats, for example as spawning areas for fish, and form areas for encrusting organisms to attach. Bivalve fragments tend to dominate the biogenic sands and gravels west of Orkney and the Outer Hebrides, whereas barnacles and attached serpulid worms form the larger component in inshore waters around Shetland (Gatliff et al. 1994).

6.7 Marine Geoconservation

Disturbance to seabed geomorphology from human activities takes many forms arising, for example, from fishing, dredging, anchoring of vessels and platforms, hydrocarbon exploration and offshore renewable-energy installations. The impact of recreational moorings on local biogenic reefs is evident in Loch Creran, Argyll, where damage has intensified in recent decades. Industrial platforms used in the hydrocarbon industry are fabricated and serviced in the Cromarty Firth and Clyde estuary, leaving 'spud can' footprints and mooring scours on the seafloor.

Perhaps the best-documented anthropogenic seabed landforms are those within Holy Loch, Argyll. Holy Loch was used as a naval base during World War II, and then as a US Navy submarine base for more than 30 years until 1992 (Miller et al. 2000). Evidence of this naval activity is still visible on the seafloor (Fig. 6.6c), where linear scour marks created by heavy mooring chains secured to the floating naval docks are visible in MBES data. It is reasonable to assume that in a relatively compact, shallow inlet like Holy Loch, vessels would have made contact with the seafloor numerous times, particularly in the nearshore zone.

Physical anthropogenic structures on the seafloor, such as shipwrecks (Fig. 6.6d), pipelines (Fig. 6.6e) and cables, may alter the local hydrodynamic regime, resulting in scour and remobilization of sediments, but can also form a substrate or refuge for marine organisms, such as the artificial reefs in Loch Linnhe. Since hydrocarbon exploration began in earnest in the North Sea Basin in the 1960s, extracted hydrocarbons have been transported to onshore terminals such as at St Fergus, north of Peterhead (Fig. 6.6e), and Sullom Voe in Shetland. At these terminals, a number of pipelines converge and make landfall in the nearshore zone, sometimes resulting in their unplanned burial, exposure or suspension as a result of disturbance to sediment transport pathways.

Scotland's National Marine Plan (Scottish Government 2015) provides an overarching legislative framework for the implementation of the Marine (Scotland) Act 2010 and the Marine and Coastal Access Act 2009, both in protecting the marine environment and enabling sustainable development of existing and emerging marine industries. It requires a holistic approach that values both geodiversity and biodiversity and the interactions between them (Gray et al. 2013). In 2018, the Scottish Marine Protected Area (MPA) Network covered approximately 22% of Scotland's seas, comprising 231 sites that will deliver the Scottish Government's vision of clean, healthy, safe, productive and biologically and geologically diverse marine and coastal environments (Scottish Government 2018).

Increased knowledge of seafloor geomorphology and submarine landscapes around Scotland, coupled with research into seabed substrate integrity and biodiversity, has resulted in marine geoconservation rising in prominence at national, European and international levels (Gordon et al. 2016, 2018). A number of the seafloor geomorphological features described in this chapter reside within the existing MPA network, for example the Firth of Forth Banks Complex MPA (Wee Bankie moraine), Central Fladen MPA (tunnel valleys), the Wester Ross MPA (subaqueous De Geer moraines, submarine landslides and pockmarks) and the Scanner Pockmark Special Area of Conservation (Gordon et al. 2016). However, given the exceptional preservation of this integral part of Scotland's geoheritage and the direct linkages between geodiversity and biodiversity, there is scope for future inclusion of other geoheritage features. This is especially important within Scotland's seas where glacial processes have shaped much of the current seabed topography and glacial deposits generally determine seabed sediment composition, both of which fundamentally impact the benthic community composition and distribution.

6.8 Conclusions

The seabed around Scotland hosts an extremely wide range of landforms, from steep-sided fjords and broad sandy beaches across the highly varied terrain of the continental shelf to the deep water of the continental slope. The broad-scale geomorphology of the submarine continental shelf, 3.5 times

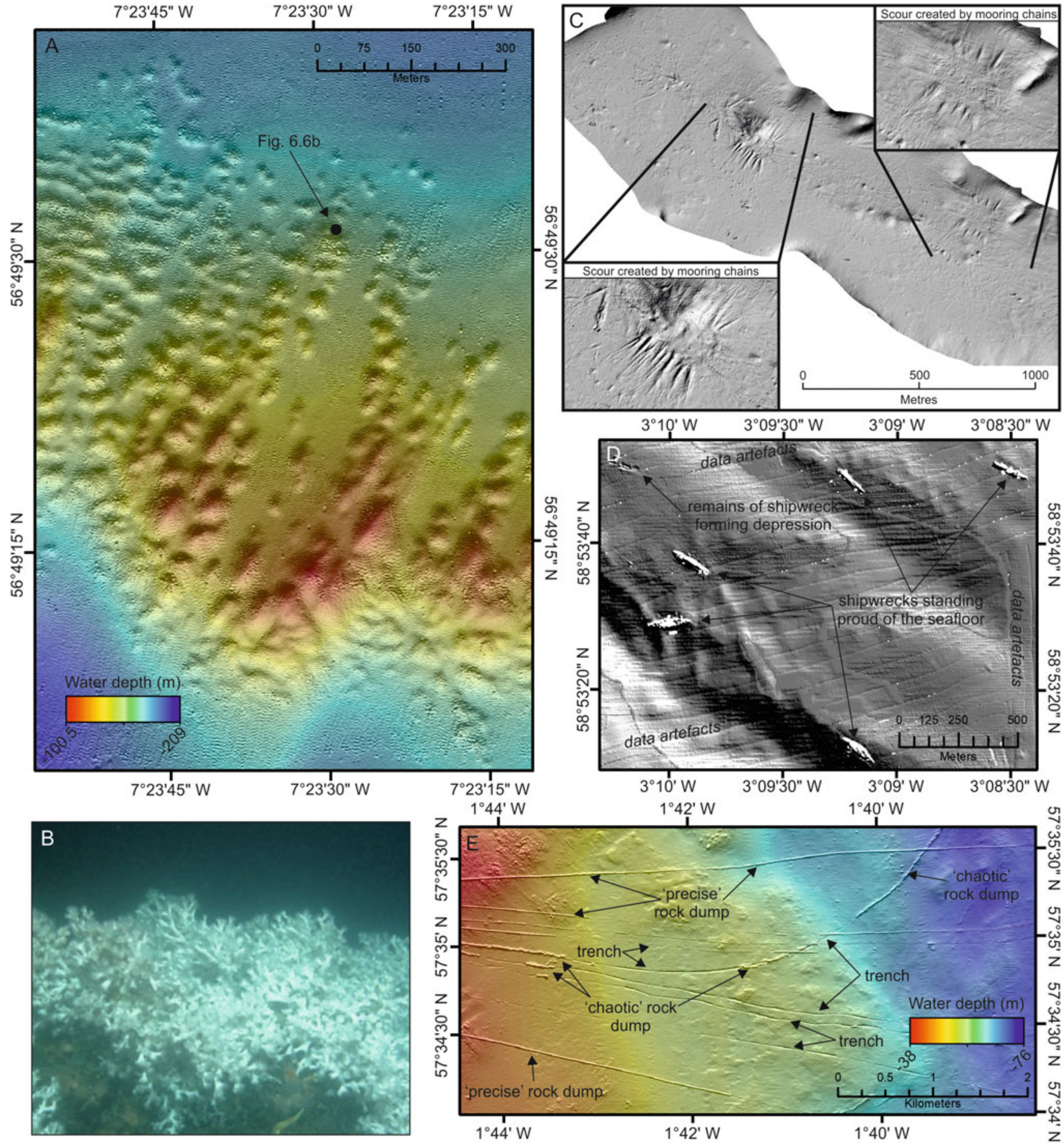

Fig. 6.6 Biogenic and anthropogenic landforms of the seafloor. **a** Part of the Mingulay Reef Complex ('Mingulay Reef 01') showing numerous biogenic sediment mounds. **b** Seafloor photograph of live coral colony from the 'Mingulay Reef 01'. For location see Fig. 6.6a. **c** Image highlighting areas of anthropogenic activity on the floor of Holy Loch (after Carter et al. 2020). **d** A small selection of shipwrecks in Scapa Flow, Orkney. **e** Pipelines making landfall at St Fergus Gas Terminal, NE Scotland

the size of the Scottish landmass, owes much to the Cenozoic geological history of NW Europe. Superimposed on this, at a smaller scale, is a vast, varied and largely well-preserved array of submarine landforms recording key global and regional events spanning the last ~1 Ma. These events include the onset, flow pattern and retreat of the last ice sheets; ice-shelf breakup and iceberg trajectories; Quaternary sea-level change and crustal movements;

earthquake-induced submarine slope failures; temporal changes in hydrodynamic regime; and episodic natural gas venting. Some of the landforms described here are amongst the best examples of their kind in Europe, with many remaining exceptionally well-preserved since deglaciation. However, human disturbance of the seabed has been increasing over the last 50–60 years, with geomorphological evidence of the commercial fishing, hydrocarbon extraction and renewable energy industries all clearly evident on the seabed. In the last decade, offshore activities have been geographically restricted with the introduction of government legislation ('The Marine Acts' (2009, 2010) and the Scottish National Marine Plan (2015)) and the subsequent establishment of the Marine Protected Areas network—now covering around 22% of Scotland's seabed. In the coming decade, further conservation of Scotland's unique submarine geoheritage and seabed geomorphology should be encouraged, especially where these landforms host rare or valuable ecosystems.

Acknowledgements Regional bathymetry data displayed in Figs. 6.1, 6.2, 6.3a, b are derived from the EMODNET Digital Terrain Model for European Seas (www.emodnet-hydrography.eu). MBES data presented in Figs. 6.3c–f, 6.4a, c, 6.5c, e, 6.6d, e were acquired as part of the Civil Hydrography Programme (CHP) on behalf of the Maritime and Coastguard Agency (MCA) and made available to the British Geological Survey (BGS) through an agreement with the MCA. MBES data displayed in Figs. 6.4e, f and 6.6c were acquired by the BGS. Acquisition of MBES data and photographs shown in Figs. 6.4d, 6.5d, g was funded by the Department for Business, Energy and Industrial Strategy's Offshore Energy Strategic Environmental Assessment programme areas 5 and 2; Crown Copyright, all rights reserved. The collection of MBES data and photographs shown in Figs. 6.4b and 6.5b was funded by the Department for Business, Energy and Industrial Strategy's Offshore Energy Strategic Environmental Assessment programme area 7 and the Department for Environment, Food and Rural Affairs through their advisors, the Joint Nature Conservation Committee, and the offshore Special Areas for Conservation programme; Crown Copyright, all rights reserved. MBES data shown in Fig. 6.6a were acquired as part of the MINCH project on board the RV *Lough Foyle*. Photograph Fig. 6.6b was collected during the Changing Oceans Expedition 2012 (Cruise JC073) University of Edinburgh. BGS authors publish with the permission of the Executive Director of the BGS (United Kingdom Research and Innovation).

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Shetland

7

Adrian M. Hall, James D. Hansom, and John E. Gordon

Abstract

The Shetland Isles display a remarkable diversity of geology and landforms. The varied relief and topography and the indented coastline are strongly influenced by the bedrock geology and structure at a variety of scales. During the last glaciation, Shetland supported an independent ice cap that extended across the adjacent continental shelves. Landforms of glacial erosion include glacially eroded valleys, breached watersheds, roughened bedrock surfaces and offshore deeps, but depositional landforms largely lie offshore. The coastal landscape is predominantly rocky, with an outstanding assemblage of eroded cliffs, caves, stacks and arches, with inlets drowned by rising postglacial sea levels. The severe wave-energy environment, particularly on the Atlantic coasts, has produced exceptional examples of cliff-top storm deposits. Inactive and active periglacial landforms occur at a relatively low altitude on Ronas Hill (450 m), reflecting the influence of wind and frost activity.

Keywords

Geology and landscape • Shetland Ice Cap • Periglacial landforms • Wind-patterned vegetation • Peat erosion • Drowned coastline • Rock-coast landforms • Cliff-top storm deposits

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7.1 Introduction

The Shetland archipelago (∼60 to 61°N) is the northernmost part of the British Isles and occupies the northern extremity of the Orkney-Shetland Platform (OSP), the topographic divide that separates the NE Atlantic and the North Sea (Ziegler 1981; Fig. 7.1). Shetland has a small total land area (<1500 km^2) but is extraordinarily rich in its geological and geomorphological diversity. The archipelago spans 120 km along a roughly north–south orientation, ∼170 km north of the Scottish mainland. The largest island (Mainland) is deeply indented by inlets that subdivide it into north Mainland, the Walls peninsula (west Mainland), central Mainland (north of the main town, Lerwick) and south Mainland; northeast of Mainland lie the large islands of Yell, Unst and Fetlar (Fig. 7.2). Numerous smaller islands and skerries lie offshore from Mainland, the largest being Fair Isle, Foula, Papa Stour, Whalsay and Bressay.

The juxtaposition of many metamorphic, igneous and sedimentary rock types on Shetland provides a remarkable array of landscapes in which the imprints of lithological and structural controls are manifest. During the cold stages of the Pleistocene, Shetland supported an independent ice cap that extended across neighbouring shelves. The impact of glacial erosion varied across the archipelago, with deep erosion along dales and *voes* (the local name for coastal inlets), but minor erosion along the former ice-divide and at higher elevations. Postglacial sea-level rise drowned glacial valley floors and the inner parts of the shelves. Rising relative sea levels over the Holocene have progressively exposed the outer coast to erosion under the full force of Atlantic storm waves. The cliff coasts of Shetland represent some of the finest coastal scenery in the world.

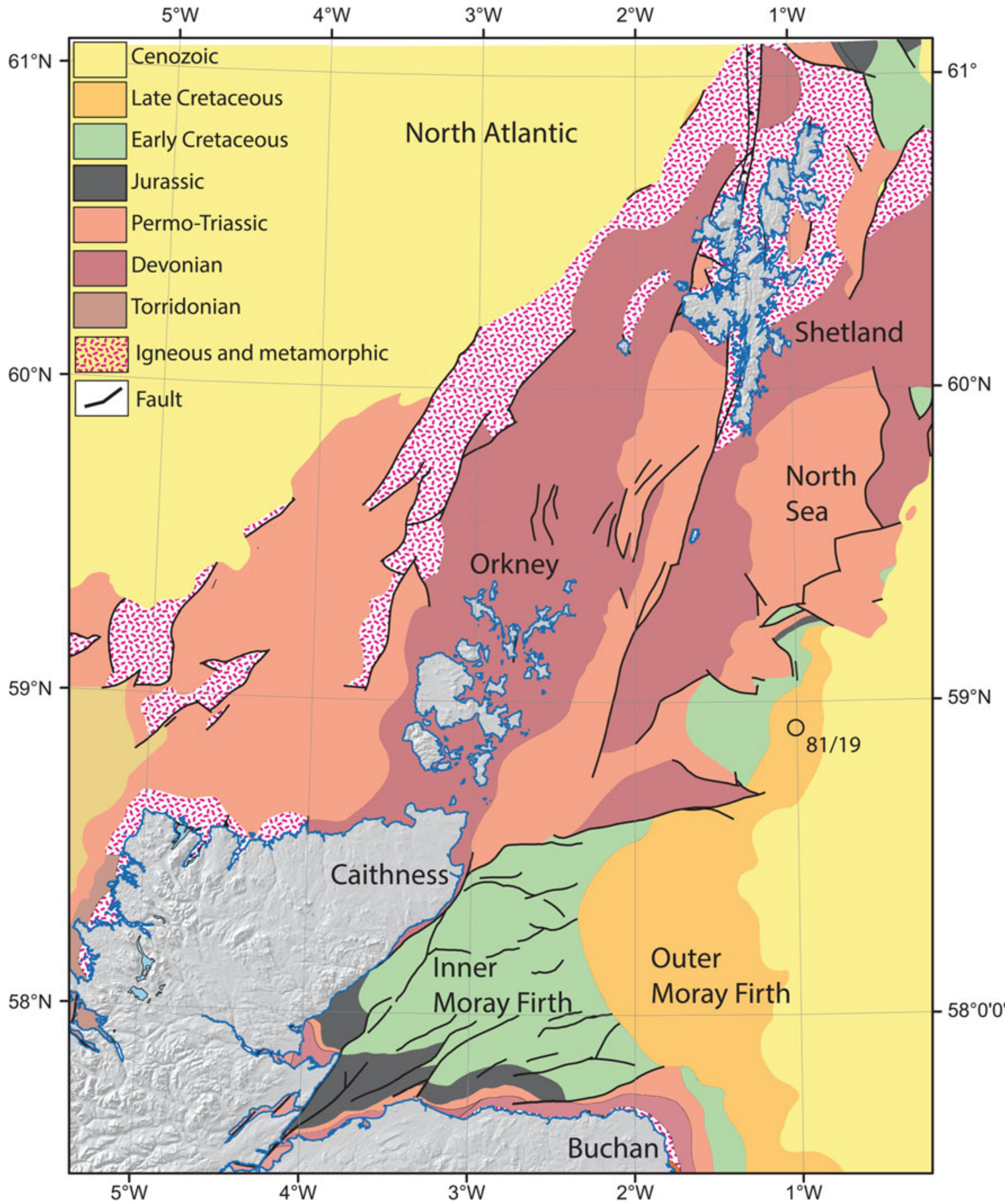

Fig. 7.1 Geology of the Orkney-Shetland Platform. (Simplified from © NERC 2016. All rights reserved. Contains Ordnance Survey Data © Crown Copyright and database right 2016. Produced using Copernicus data and information funded by the European Union-EU-DEM layers. NEXTMap Britain™ elevation data from Intermap Technologies)

7.2 Geology

7.2.1 Geology and Structure

Shetland stands on a part of the Caledonian Orogenic belt and comprises several different crustal fragments, or terranes, assembled by strike-slip movements along three major north–south transform fault zones during the closure of the Iapetus Ocean in the Late Silurian (Chap. 2): the Melby, Walls and Nesting fault zones. The Walls Boundary Fault Zone represents the northern continuation of the Great Glen Fault Zone in Scotland. It was initiated during the Caledonian Orogeny, reactivated and dislocated during late

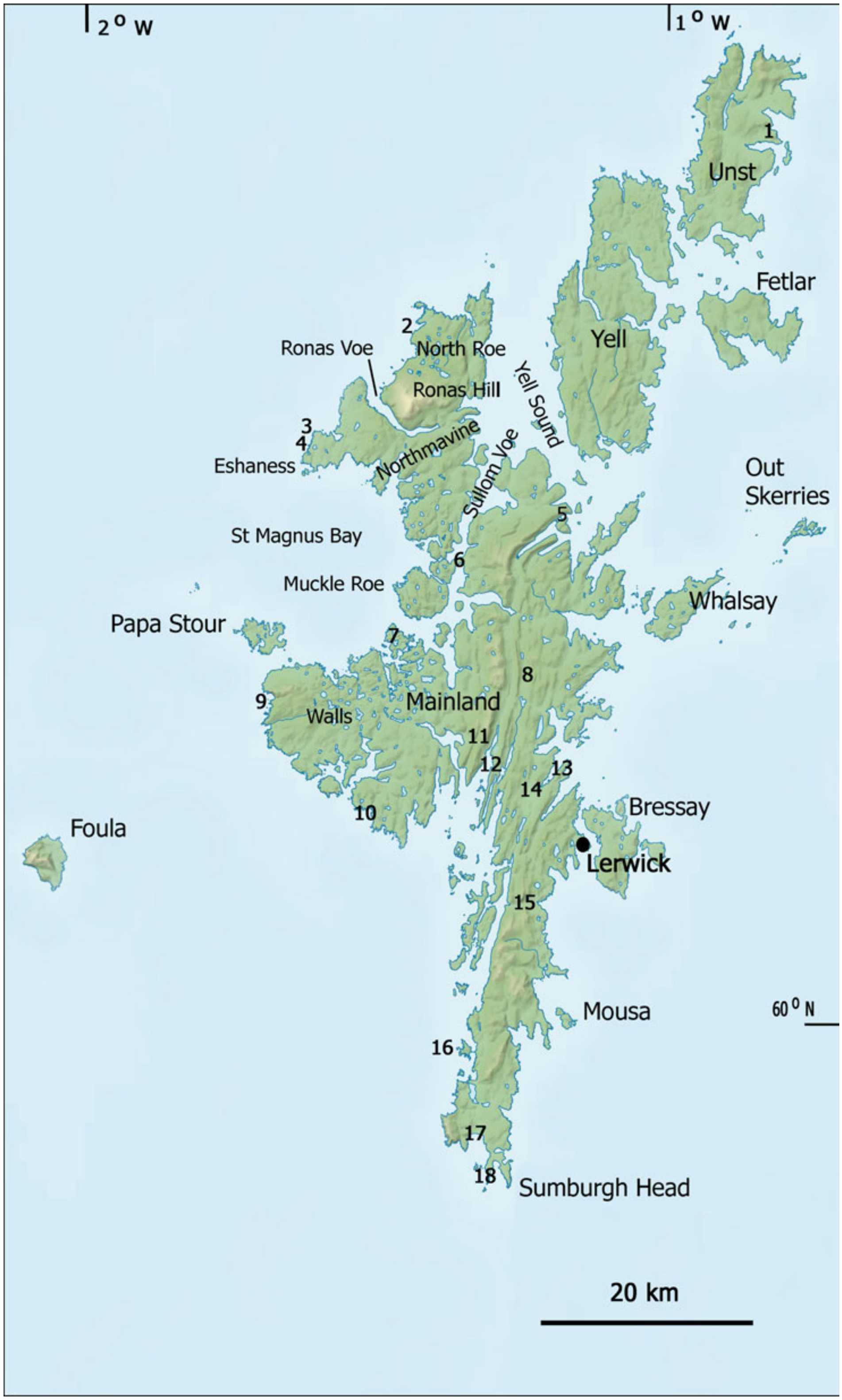

Fig. 7.2 The Shetland Isles: relief and main locations mentioned in the text. 1: Keen of Hamar; 2: Fugla Ness; 3: Grind of the Navir; 4: Villians of Hamnavoe; 5: Ayres of Swinister; 6: Mavis Grind. 7: Vementry; 8: Petta Dale; 9: Sel Ayre; 10: Westerwick; 11: Tresta; 12: Weisdale Voe; 13: Lax Firth; 14: Whiteness; 15: Quarff gap; 16: St Ninian's Isle; 17: Quendale; 18: Ness of Burgi. Fair Isle lies ~38 km southwest of Sumburgh Head. (Base map contains Ordnance Survey data © Crown copyright and database right. Creative Commons Attribution-Share Alike 3.0 Unported license)

Carboniferous inversion of the Orcadian Basin and became active again during the Mesozoic or Cenozoic, with up to 15 km of left-lateral displacement (Watts et al. 2007).

The bedrock geology of Shetland is complex (Mykura 1976; Mykura and Phemister 1976). The oldest rocks form a basement platform of late Archaean (2500–3000 Ma)

'Lewisian' gneiss that crops out along the north coasts of west Mainland and North Roe. Resting on the basement is a succession of Mesoproterozoic and Neoproterozoic metasedimentary rocks which display several phases of deformation and alteration. Banded gneisses, mica schists and psammites of the Moine Supergroup underlie most of Yell and parts of north Mainland, while Dalradian quartzite, limestone, mica schist and metamorphosed lavas form much of south, central and eastern Mainland, western Unst and western Fetlar. Fragments of the floor and underlying crust of the Iapetus Ocean are found in the ophiolite complex of eastern Unst and Fetlar; this complex was obducted westwards on to the edge of Laurentia during the Caledonian Orogeny. The ultrabasic and basic peridotites and gabbros provide a rare insight deep into oceanic crust and upper mantle rocks (Flinn 2014).

Rocks of the Devonian Old Red Sandstone Supergroup formed mainly in the later stages of the Caledonian Orogeny. Sandstones predominate, with volcanic agglomerate and lava confined to Papa Stour and at Eshaness. Caledonian plutonic complexes, predominantly of granite, were intruded into both the metamorphic and sedimentary rocks; granite now underlies the highest hill on Shetland, Ronas Hill (450 m).

The OSP was uplifted in the earliest Palaeocene and large parts remained above sea level thereafter, with varying rates of erosion and sediment supply to flanking sedimentary basins throughout the Cenozoic (Anell et al. 2012). Sediment volumes in the northern North Sea indicate that the depth of erosion was profound, with estimated losses of ~1 km of rock from the OSP in the Palaeocene and Eocene. As the present level of erosion approximates to that of the sub-Devonian and sub-Permian unconformities (Flinn 1977), the missing rock was former Palaeozoic and Mesozoic sedimentary cover (Wilkinson 2017). The clearest evidence of a former land surface on Shetland is provided by the hilly sub-Devonian terrain, a surface preserved where Devonian sandstones and conglomerates rest on the basement in both south and north Mainland. Aside from these patches of exhumed Palaeozoic unconformity surfaces, all present-day landforms on the OSP are of Neogene or younger age. The uplifted OSP was an important source area for sediments in the North Sea from the late Miocene onwards, and the Shetland area continued to provide fluvio-deltaic sediment during the early Pleistocene (Ottesen et al. 2018).

7.2.2 Geology and Landscape

Close links exist between rocks and relief on Shetland. At a regional scale, the landscape and topography are strongly conditioned by the roughly north–south orientation of major faults and the main rock units, which has determined the 'box-shaped' outlines of Unst, Yell, North Roe and south Mainland. Fault movements produced zones of crushed and fractured rock, later exploited by rivers and glaciers to create major valleys and the arms of firths and voes (Mykura 1976). The parallel ridges and valleys of eastern Mainland (Fig. 7.3) reflect near-vertical foliation and differential erosion of psammites, pelites, quartzites and limestones, resulting in a subdued Appalachian-style relief (Coque-Delhuille and Veyret 1988), one of the best examples of its kind in Scotland, comparable to the terrain flanking the Ladder Hills in the northeast Grampians. In contrast, west of the Walls Boundary Fault Zone, more varied relief reflects the diverse geology, which includes granite intrusions and folded Devonian sandstone and lavas.

Three case studies illustrate how different rock types give distinctive terrains at local scales on Shetland. First, weathered bedrock is exposed in many bays and valleys, where it is often associated with zones of hydrothermal alteration. In the Burn of Tactigill, east of Tresta, the alteration zone in mica schist is up to 30 m wide and follows the line of a NNE–SSW fault (Coque-Delhuille and Veyret 1988). The most widespread type of weathering is a sandy regolith found in pockets beneath till. The presence of weathered granite up to 10 m thick also partly explains the erosion of the deep north–south valley of Petta Dale in east Mainland.

Second, the rugged peninsula of Muckle Roe displays many elements typical of granite scenery. The 5 km wide stock-like intrusion of granophyre is of homogeneous mineralogy, but is cut by major crush zones and shows wide variations in fracture spacing (Mykura and Phemister 1976). Ice flow was parallel to NNW–SSE fracture sets and has excavated rock basins and trenches and heightened inland cliffs along this trend. Surficial blocky and granular disintegration of the granite is common along hematite-stained shear zones 1–10 m wide. These vertical zones of weakness have acted as major controls on the locations of geos (narrow, cliffed inlets) and bays at the coast (Fig. 7.4a) and trenches and basins inland. Similar fracture sets form parallel geos in the sandstones of south Mainland.

Third, on Unst, the mineralogy of ultrabasic rocks has strongly influenced slopes, soils and vegetation. Soils are skeletal and plant growth is locally inhibited, exposing the ground surface to disturbance by frost action and slopewash. The Keen of Hamar National Nature Reserve has many plant rarities on its extensive areas of serpentinite rubble (Carter et al. 1987; Fig. 7.4b). High alkalinity has also prevented peat development but made possible the local cementation of screes with travertine (Flinn and Pentecost 1995) and the accumulation of Late Holocene spring-fed diatomite (Flinn 1996).

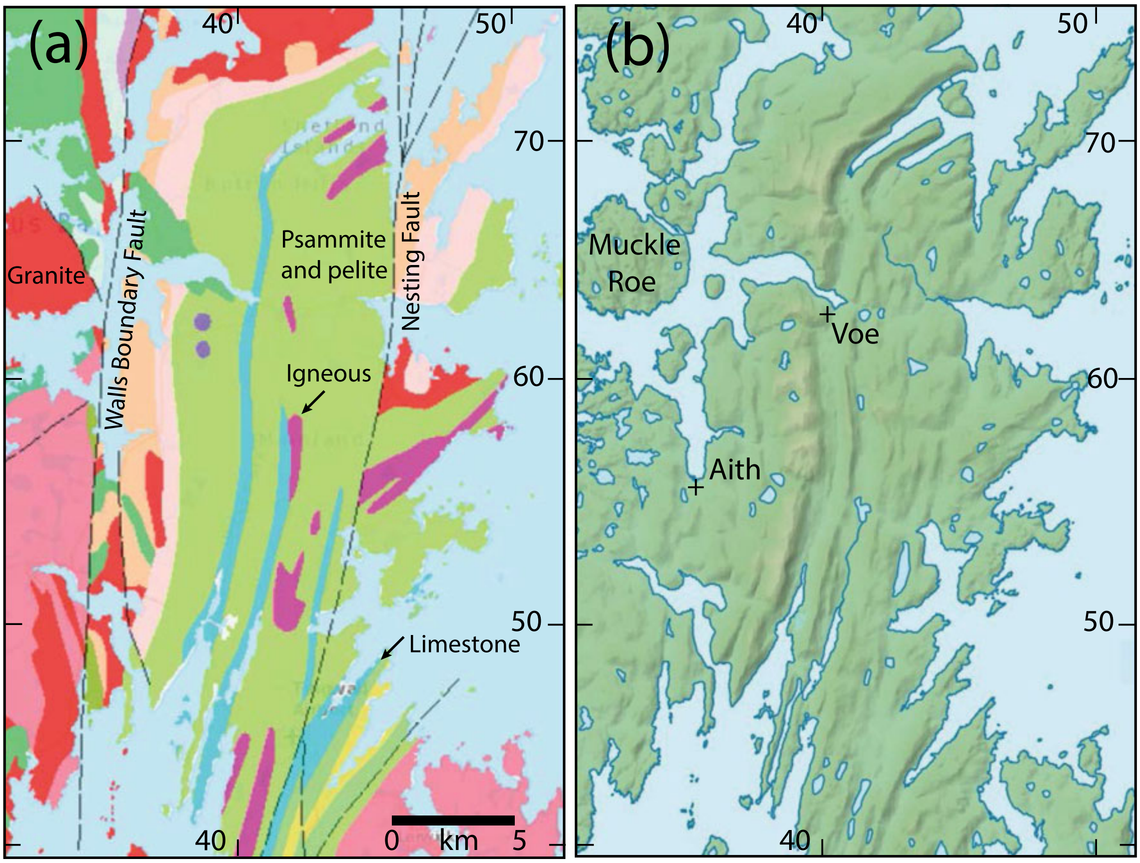

Fig. 7.3 Geological controls on Appalachian-style relief, eastern Mainland. **a** Geology. (Based on British Geological Survey mapping (© UKRI)). **b** Relief. (Based on Ordnance Survey data © Crown copyright and licensed under the Creative Commons Attribution-Share Alike 3.0 Unported license)

7.3 Pleistocene Glacial History and Landforms

The maritime setting of Shetland, its generally low topography and position at ~60° N strongly influenced the regional impacts of Pleistocene fluctuations in temperature and sea level. During cold stages, sea level fell as low as −130 m and an ice cap developed over Shetland which, at its maximum extent, advanced to the edge of the Atlantic continental shelf (Chap. 4). During warm stages, sea level rose close to its present level and a maritime climate, similar to today's, was established.

7.3.1 Landforms and Landscapes of Glacial Erosion

Shetland is a narrow archipelago with a linear ice-shed zone that exhibits little or no glacial modification, flanked to the west and east by belts of more pronounced glacial erosion (Fig. 7.5). There is a discordance, however, between the radial pattern of ice flow established during the last cold stage and the NNE alignment of valleys on Mainland. A possible explanation is that the valleys were eroded mainly in periods when Shetland was occupied by an icefield of limited extent, with topographically constrained ice flow.

The axial ice-shed zone is evident from the distribution of striae and ice-roughened bedrock. Ice roughening is not widely developed along the former ice shed, and remnants of weathered and altered bedrock occur frequently (Fig. 7.5), both features indicating limited glacial erosion. The intensity of glacial erosion also decreases with altitude. On Ronas Hill, there is a sharp upper limit to ice roughening at ~200 m, with glacially disturbed granite regolith at lower altitudes. Close to the summit, granite surfaces carry weathering pits up to 15 cm deep that likely relate to weathering before the last glaciation. The hill summits on Foula also show few signs of vigorous glacial erosion (Bradwell et al. 2019).

In western Mainland, ice-roughened scenery is widespread, with striae and roches moutonnées indicating the movement of ice to the west and northwest. Knock-and-lochan terrain is developed on rock types with

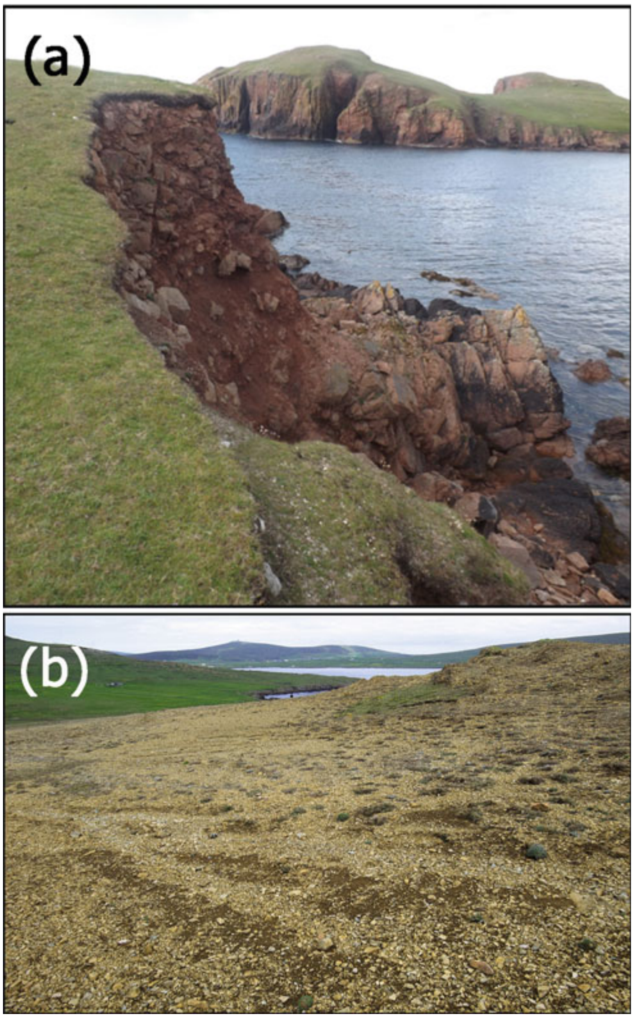

Fig. 7.4 **a** North Ham, Muckle Roe. Blocky disintegration of hematite-stained granophyre in the foreground. In the background prominent NNW–SSE fracture sets form geos on Harri Stack. **b** The skeletal serpentine soils of the Keen of Hamar support nationally rare plant communities. (Images: **a** Adrian Hall; **b** ©Lorne Gill/NatureScot)

variable, but locally wide, joint spacing, such as the gneisses, granites and psammites on Vementry. On the east coast, rock basins and ice-moulded crags indicate ice flow towards the east and northeast (Fig. 7.5). Where sharp boundaries exist between smooth slopes and roughened terrain, for example, north of Ronas Hill (Bradwell et al. 2019), the small elevation differences indicate that the depth of glacial erosion necessary to produce this ice-roughened terrain may be <10 m. Rough, boulder-strewn granite terrain known as the Giant's Garden found northeast of Ronas Hill may have formed under a late phase of glacial ripping (Hall et al. 2020; Fig. 7.6a).

Valleys on Shetland are relatively shallow, reflecting the limited relief of the landmass. The highest cliffs on valley sides occur where the valleys have been cut through a ridge of high ground, as on Ronas Voe beneath the cliffs of The Brough, or where the ice has flowed parallel to the structural grain, as on the flanks of Weisdale (Fig. 7.6b). The degree of valley connectivity is high, with glacial valleys separating hill masses, as on Whiteness. This compartmented relief is perhaps best expressed in the isolation of islands, such as Whalsay and Vementry, from the neighbouring landmass. Important through-valleys exist, notably The Daal on Foula and crossing the isthmus of Mavis Grind. The Quarff gap, southwest of Lerwick, may be a relic of the Devonian landscape (Flinn 1977).

Shetland's shallow voes and firths are valleys that have been drowned by Lateglacial and Holocene sea-level rise (Fig. 7.6b). Many show undoubted evidence of glacial erosion, with ice-moulded bedrock on their flanks, steep sides, a straight or gently-curving course and the presence of enclosed rock basins, but the depth of glacial erosion has often been modest and nowhere exceeds 100 m.

Many voes, such as Sullom Voe, have dendritic valley systems at their heads. Valley incision has been facilitated by the steep gradients for ice flow provided by the drop between the current landmass and the seabed offshore, which falls steeply towards −100 m. Rock basins occur offshore and may lie hidden beneath sediment on valley floors onshore. Yell Sound has deeps down to −90 m. Rockhead in the deepest part of St Magnus Bay is in Permo-Triassic sandstone and lies at −200 m, buried under 80 m of Late Quaternary fill.

7.3.2 Deposits Pre-dating the Last Glaciation

The present landscape of Shetland represents the cumulative imprint of multiple episodes of Pleistocene glaciation. More recent glaciations have largely removed evidence for earlier events, but two critical sites preserve a detailed record of changing environmental conditions on the islands before the last (Late Devensian) ice sheet. At Fugla Ness, a layer of peat resting on till is overlain by periglacial slope deposits and then till deposited by the last ice sheet (Fig. 7.6c). The peat preserves a pollen record and subfossil wood remains, mainly the roots of Scots Pine (*Pinus sylvestris*), that demonstrate a former tree cover in Shetland. The age of the peat is uncertain. Pollen evidence suggests correlation with penultimate or Gortian/Hoxnian interglacial deposits in Ireland (Birks and Ransom 1969). However, a last interglacial age is supported by a Uranium-series date of 110+40/–30 ka on the peat itself (Hall et al. 2002).

At Sel Ayre, organic deposits overlie periglacial slope deposits and contain pollen assemblages that represent vegetation covers ranging from grasses to a rich herb flora (Birks and Peglar 1979; Hall et al. 2002), distinct from those at Fugla Ness. Luminescence ages of 98–105 ka obtained for overlying sands place these deposits in Marine Isotope

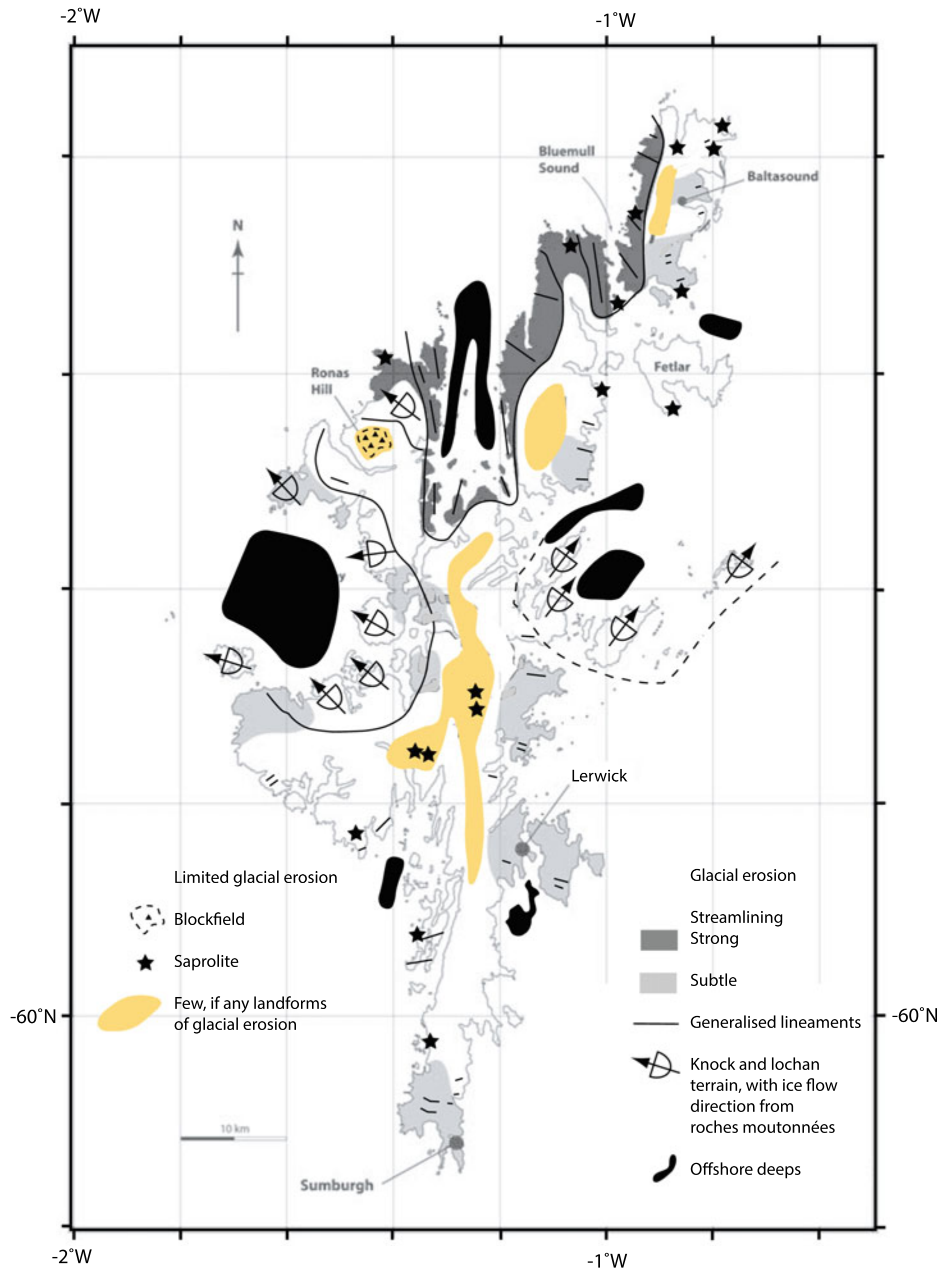

Fig. 7.5 Glacial features of Shetland. (Adapted from Hall 2013; Creative Commons Attribution 4.0 International License)

Substage (MIS) 5c (Brørup Interstade). Overlying deposits indicate the return of periglacial conditions during the cold interval MIS 5b, followed later by full glacial conditions and deposition of till by the last ice sheet.

7.3.3 The Last (Late Devensian) Glaciation

The maximum limits of the British-Irish Ice Sheet (BIIS) during the Middle and Late Pleistocene lie close to the NW Atlantic shelf edge (Chaps. 4 and 6). Offshore data indicate that the Shetland Ice Cap (SIC) extended well beyond the current landmass to the east and north, and close to the edge of the continental shelf west of Shetland (Johnson et al. 1993; Stoker et al. 1993; Ross 1996; Bradwell et al. 2008, 2019; Clark et al. 2012). In the North Sea, the BIIS was at times confluent with the Fennoscandian Ice Sheet (FIS; Graham et al. 2011).

There has been debate since the nineteenth century as to whether Shetland supported a local ice cap during the last glaciation (Flinn 1978), or whether the islands were overwhelmed by the FIS (Peach and Horne 1879). However, firm evidence for the former presence of the FIS (or ice from mainland Scotland) is absent on Shetland. In western Shetland, ice flowsets (Golledge et al. 2008) generally conform to patterns of westward ice flow recognized from the dispersal of

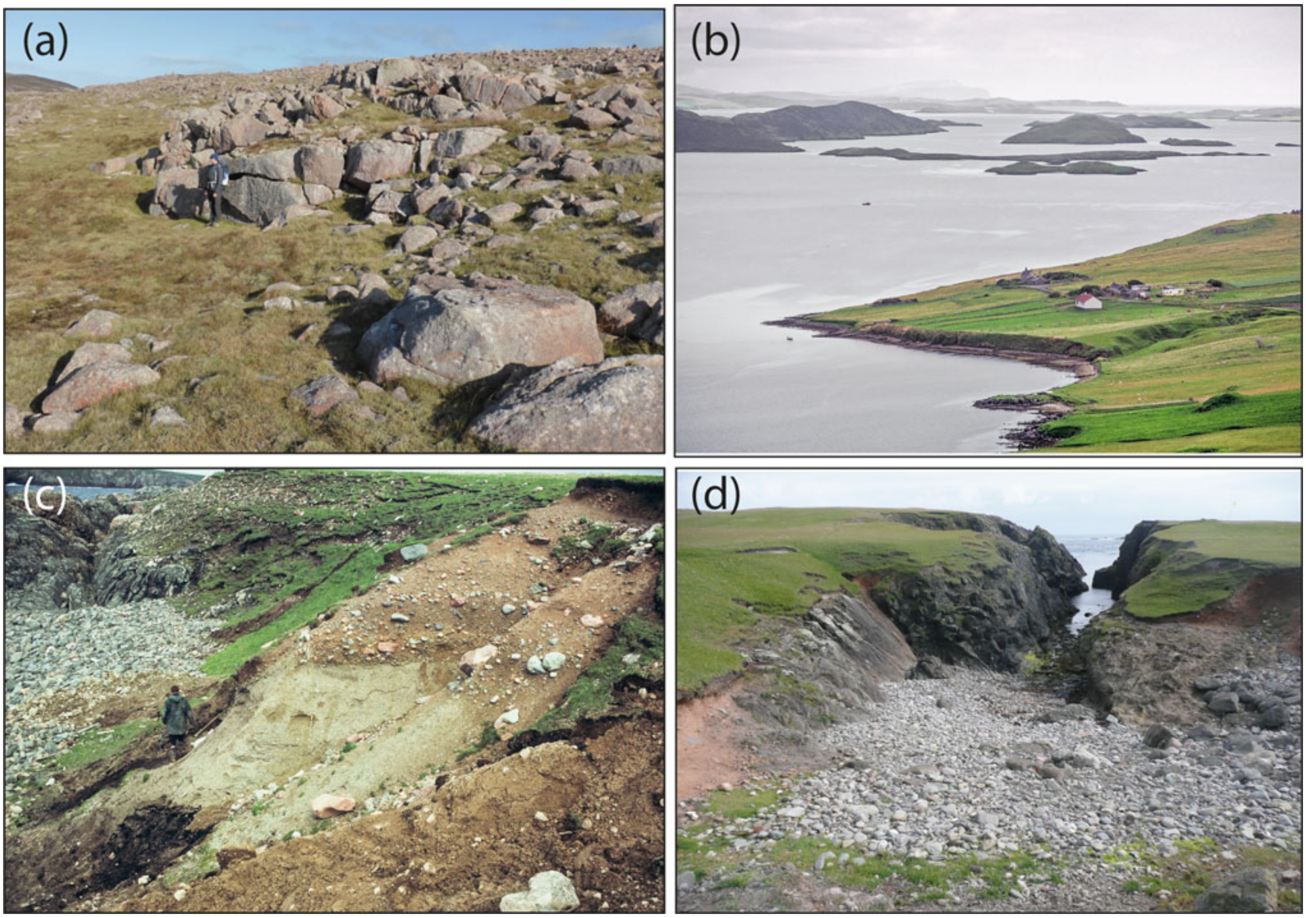

Fig. 7.6 **a** Ice-roughened granite terrain at Beorgs of Housetter, near Ronas Hill. Dilated rock fractures and boulder spreads suggest that glacial ripping operated during deglaciation. **b** Weisdale Voe, a glacially eroded valley drowned by postglacial sea-level rise. **c** Fugla Ness. Interglacial peat is overlain by periglacial slope deposits and till deposited by the last Shetland Ice Cap. **d** Till-filled geo head, South Harbour, Fair Isle. (Images: **a**, **d** Adrian Hall; **b**, **c** John Gordon)

erratics, striae and roches moutonnées (Mykura and Phemister 1976; Flinn 1978; Hall 2013). Evidence from eastern Shetland, however, in the form of mega-scale glacial lineations (Golledge et al. 2008) and glacitectonic structures (Carr and Hiemstra 2013), has been interpreted to suggest westward ice flow. This evidence conflicts with reports of only one supposed Scandinavian glacial erratic on Mainland, the absence of evidence for glacial transport of Permo-Triassic and younger Mesozoic debris across the islands from offshore basins located to the east, and the apparent absence of any glacial deposits from the FIS (Hall 2013). Erratics on Fair Isle, however, include rock types sourced from Shetland, Scotland and, possibly, Norway (Hall and Fraser 2014).

The evidence of radial ice-flow patterns from erratics, glacial bedforms and striae provides little support for flow of external ice across Shetland (Fig. 7.5). Instead, these patterns support the existence of an independent Shetland ice cap throughout the Late Devensian (Hall 2013). Recent results show widespread and significant spatial fluctuations in the extent and flow configuration of the SIC (Bradwell et al. 2019; Chap. 4). At its maximum extent at ~26 to 25 ka, the SIC was confluent with the FIS in the northern North Sea. Subsequent retreat led to parting of these ice masses, with establishment of a seaway to the east of Shetland by ~19 ka. The SIC shrank quickly at 19–18 ka, and eventually disappeared between 16 and 15 ka. The dynamism of the last SIC was a product of high snowfall on the Atlantic margin, limited topographic control and a high susceptibility to changes in sea level and the rapid loss of ice mass through iceberg calving as sea level rose.

7.3.4 Landscapes and Landforms of Glacial Deposition

The depositional zone of the last SIC lies largely offshore. Sets of moraines on the West Shetland Shelf have been interpreted to represent multiple phases of expansion of the SIC during the Middle and Late Devensian (Davison 2004). The pattern of deglaciation around Shetland has become much clearer in recent years with the advent of detailed imagery of bedforms on the adjacent seabed (Bradwell et al. 2008, 2019; Chap. 4). The SIC retreated towards the spine of the islands from both west and east. Moraines on Papa Stour and on the adjacent seafloor mark late stages of ice retreat into St Magnus Bay (Bradwell et al. 2019; Chap. 6), and De Geer moraines on the floor of the Pobie Basin, to the east of Unst, indicate southward retreat of the ice margin (Bradwell et al. 2008). Moraines on Unst (Golledge et al. 2008) mark positions of the western flank of this retreating ice lobe. Clusters of cosmogenic nuclide exposure ages obtained for sites on Papa Stour, Muckle Roe and North Roe shows that the SIC had retreated onto the present land area by 17 ka, with large tidewater glaciers discharging into the main voes and sounds between the islands (Bradwell et al. 2019).

7.4 Periglacial Landforms and Deposits

At several sites on Shetland, till rests on brecciated rocks that appear to have been subjected to significant frost weathering and mass movement under periglacial conditions at elevations down to present sea level. At Fugla Ness and Sel Ayre the organic deposits are overlain by periglacial slope deposits which pre-date the last SIC (Fig. 7.6c).

Upland periglacial landforms and deposits are well represented on the granite regolith at Ronas Hill (Ball and Goodier 1974; Coque-Delhuille and Veyret 1988). They include blockfields and stone-banked terraces at 365–427 m, comparable to those on the Cairngorms at altitudes above 1000 m (Fig. 7.7a) Whilst the mountain-top regolith likely has an earlier origin, these features were probably last active during the Lateglacial.

Other periglacial features on Ronas Hill have been active during the Holocene at elevations that are unusually low for the British Isles. These active forms include solifluction terraces and deflation surfaces, reflecting a prevailing subarctic-oceanic climatic environment of extreme wind speeds, heavy but now short-lived snowfalls, high soil-moisture levels, except in high summer, and frequent freeze–thaw cycles. Also present are horizontal and oblique turf-banked terraces, wind-patterned vegetation (wind stripes) and composite stripe-terrace features (Veyret and Coque-Delhuille 1989; Fig. 7.7b). Active frost-sorted patterned ground even occurs near sea level on Fetlar and Unst, on serpentine soils that are inimical to vegetation colonization (Spence 1957; Ball and Goodier 1974; Carter et al. 1987).

7.5 Peat Erosion and Bog Bursts

Approximately 50% of Shetland is covered by blanket bog several metres deep developed on acid substrates (Veyret and Coque-Delhuille 1993). The bog covers the base of the slopes and ridge and hill tops up to an altitude of about 200 m, including slopes that reach 14°–15°. Areas of actively eroding bare peat, gullies and peat hags are widespread (Birnie 1993). Under extreme rainfall events, the peat may become saturated and unstable, leading to bog bursts and peat slides, as occurred in 1990 and 2003, with deep erosion of the peat, often exposing the bleached bedrock surface underneath (Veyret and Coque-Delhuille 1993; Dykes and Warburton 2008; Fig. 7.8).

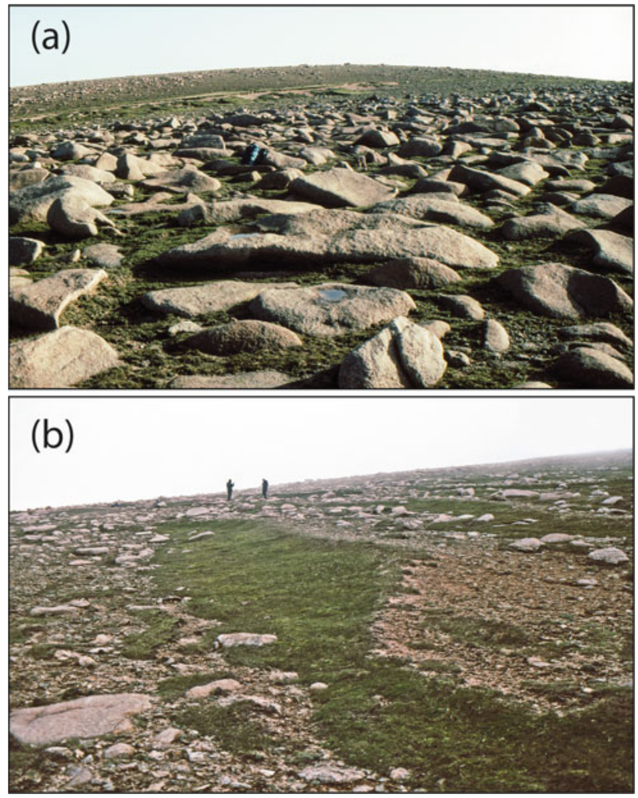

Fig. 7.7 Periglacial landforms on Shetland. **a** Granite blockfield, Ronas Hill. The remnants of an actively eroding aeolian sand sheet occur in the middle distance. **b** Wind stripes, Ronas Hill. (Images: John Gordon)

7.6 Coastal Landforms

The coast of Shetland is predominantly hard and rocky, characterized by strong structural controls and an impressive array of dramatic cliffs, caves, stacks and arches. The archipelago sits between two storm-swept seaboards, the Atlantic and the North Sea. With relatively deep water extending close to the coast, there is often limited wave attenuation, and much of the outer coast is subject to a severe wave-energy environment. The coastline includes sea cliffs up to 275 m high, some of the highest in Europe. The cliffs are the present-day terminations of ridges and hills that have

Fig. 7.8 Peat slide north of Voe, north Mainland. (Image: John Gordon)

been truncated by the retreating coastline. The sea caves on Papa Stour and Eshaness are amongst the longest (up to 300 m) in the world. The occurrence of till deposits within the heads of geos (Fig. 7.6d) indicates that parts of the Shetland coast predate at least the last glaciation and formed when interglacial sea levels were similar to that of the present. Nevertheless, almost all substantial glacial deposits lie offshore (Bradwell et al. 2008) and the lack of onshore glacial and periglacial sediment has restricted sediment availability for beach-building. Beaches and dunes are therefore limited to a few locations, mainly within bays or other sheltered sites where the impact of wave activity is reduced, allowing retention of locally available sediment. In general, the coastal landforms of Shetland reflect the influence of four factors: strong structural controls, the effects of past glaciations, rapid sea-level change over millennia and a relatively severe morphogenetic environment.

7.6.1 Structural Controls on Coastal Landforms

Shetland is dominated by the largely north–south or NNE–SSW orientation of major faults and lithological boundaries. Selective weathering and erosion along faults, most recently by glaciers, has resulted in the formation of several large firths and numerous linear sea lochs (voes) that extend far inland. The overriding control of structure is best seen in southwest and northeast Mainland where rising sea level has flooded low ground to form a series of long parallel inlets controlled by the strike of the underlying structure (Fig. 7.3). At the local scale, changes in lithology, minor faults, fracture zones and dykes have influenced the development of almost all the smaller inlets, geos and other landforms that fret the entire coast (Fig. 7.9a, b). Many voe and bay heads expose rocks that are altered, weathered or otherwise fractured, with low resistance to wave attack. Lithology has also strongly influenced the propensity of cliffed coastlines to landsliding: 72% of 128 recorded coastal rock-slope failures on Shetland are seated on metasedimentary rocks with relatively few on other lithologies (Ballantyne et al. 2018).

At Eshaness, flat-lying and gently-inclined andesitic tuffs, ignimbrites and lavas form almost vertical cliffs that rise from deep water to heights of $\sim$45 m; locally these have stepped profiles, with steep cliff faces in the resistant tuffs and wide terraces cut across blocky, andesite lavas (Hansom 2003a). The coast here is fully exposed to Atlantic storm waves and, in common with Papa Stour to the south (Hansom 2003b), marine exploitation of numerous vertical and horizontal fractures in these extrusive rocks has resulted in the development of numerous stacks, arches, geos and subterranean passages, some of which have extended into spectacular blowholes (Fig. 7.9c).

7.6.2 Glaciation Control on Coastal Landforms

Although the sides of some voes and firths in southwest and northeast Mainland exhibit evidence of ice-moulding and glacial steepening, most Shetland inlets are relatively unmodified, possibly pre-glacial, valleys that have been drowned by sea-level rise. The north–south axis of the Shetland ice cap during successive glacial cycles meant that

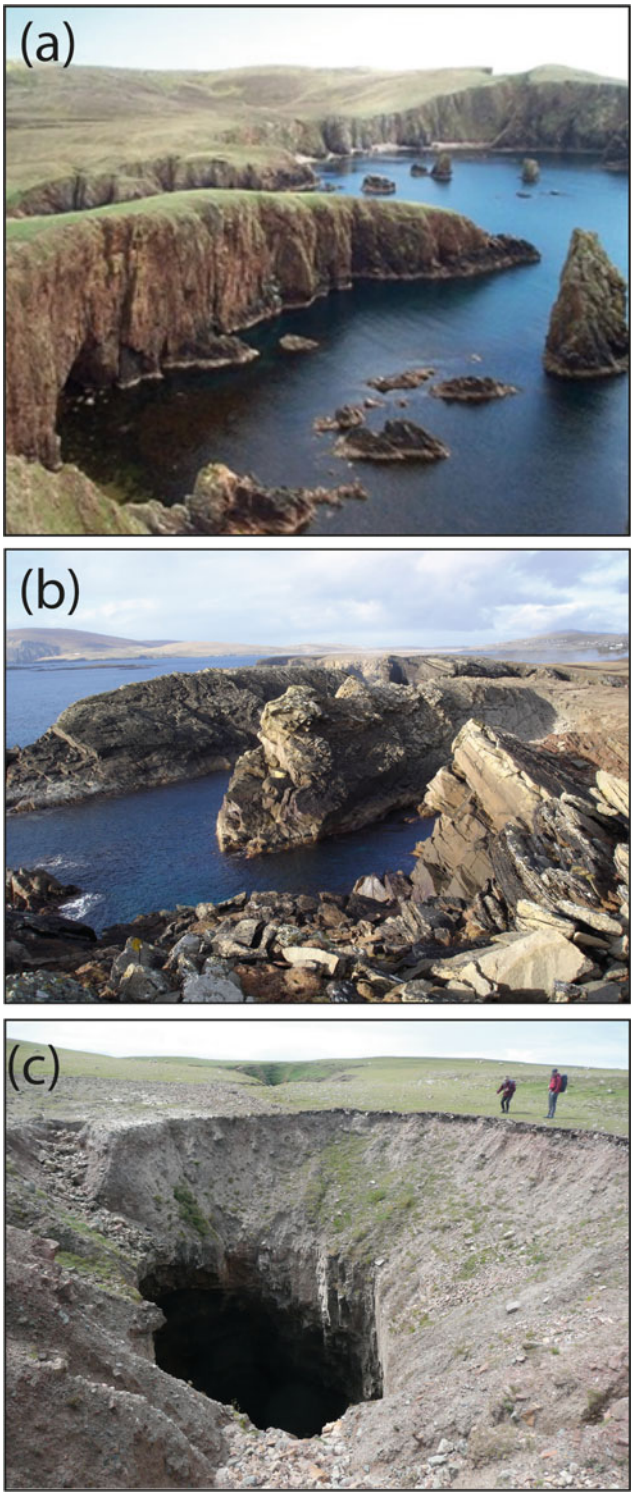

Fig. 7.9 **a** The cliffed coast of Westerwick, central Mainland, is typical of the rocky and fretted Shetland coast with steep cliffs, stacks and skerries. **b** View north over Ness of Burgi, Sumburgh, south Mainland. Steeply dipping sandstone with SSW–NNE orientated fractures has been exploited by waves to form parallel geos. The lower parts of the 10–20 m high cliffs have been cleared of loose rock by storm waves that have also stripped vegetation, soil and regolith from the cliff tops. Cliff-top storm deposits occur on the most exposed parts of the headland. **c** Christie's Hole in Papa Stour is a till-capped cliff, geo and subterranean extension where part of the roof collapsed in 1981. (Images: **a**, **b** James Hansom; **c** Adrian Hall)

ice flowed at right angles to much of the coast, across the structural grain, producing glacially moulded and streamlined cliff-top surfaces but leaving cliff faces relatively unmodified. The preservation of sediments pre-dating the expansion of the last SIC below the west-facing cliff top at Sel Ayre (Sect. 7.3.2) illustrates the lack of glacial modification of lee-side cliffs. In general, only limited amounts of glacial deposition have occurred on land in Shetland; this has resulted in a lack of sediment available for substantial beach development.

7.6.3 Sea-Level Control on Coastal Landforms

The present form of the Scottish coastline reflects the interaction of eustatic sea level rise since the Last Glacial Maximum ($\sim$26 to 19 ka) and glacio-isostatic uplift caused by the shrinkage and eventual disappearance of the last ice sheet. The amount of glacio-isostatic uplift declines with distance from the axis of uplift in the western Grampian Highlands. On Shetland, >450 km from the uplift axis, eustatic sea-level rise has outpaced uplift throughout the Lateglacial and Holocene (Shennan et al. 2018), leading to progressive drowning of the coastline; postglacial emerged beaches are therefore absent. A relative sea-level (RSL) curve for Shetland constructed by Bondevik et al. (2005) indicates that at $\sim$8 ka RSL stood at −30 to −20 m and by $\sim$4 ka it lay at −4 m, thereafter gradually rising to present levels. At several locations, terrestrial peat deposits of mid-Holocene age now occur in the intertidal and subtidal zones due to subsequent inundation (Mykura 1976; Birnie 1981; Bondevik et al. 2005).

Although many of the cliffs on the outer coasts of Shetland plunge to depths of up to 80 m, there is in some locations a distinct break of slope at or just below the present intertidal zone, with nearshore rockhead sloping gently away from the base of cliffs, though intertidal or sub-tidal rock platforms produced by postglacial cliff retreat are narrow or localized (Flinn 2014). Small, structurally-controlled bedrock ramps also exist at present sea level in several locations, such as on Foula, Fetlar and north of Eshaness. Intriguingly, small, glacially eroded, low-relief bedrock benches at $\sim$20 m in NE Unst, at Ness on North Roe and at Eshaness have been interpreted as emerged shore platforms (Coque-Delhuille and Veyret 1988), possibly implying interglacial sea levels markedly higher than those of the present. The sea floor adjacent to Shetland displays steps and nearly horizontal surfaces, commonly at depths of $\sim$24 m, $\sim$46 m and $\sim$82 m below present sea level (Flinn 1964). The $\sim$46 m and $\sim$82 m surfaces have parallels with similar submerged platforms around St Kilda (Chap. 9), and probably indicate marine plantation during multiple periods of low sea level.

Sea-level control may also have influenced the pattern and timing of coastal landslides on Shetland. Cosmogenic ^{10}Be exposure ages obtained for two coastal landslides on Fetlar demonstrate that these occurred at $\sim$4.5 ka and $\sim$4.8 ka, at a time when rising sea levels allowed storm waves to access the cliff foot and initiate instability (Ballantyne et al. 2018). Moreover, 24% of 128 recorded coastal landslide sites on Shetland exhibit evidence of recent activity in the form of fresh scarps, areas of stripped bedrock and accumulations of debris at the base of cliffs. The chasm of The Sneck on Foula is an outstanding example of an impending rock-slope failure.

7.6.4 Morphogenetic Environment and Coastal Landforms

The Shetland coast is subject to strong winds with a high frequency of stormy conditions. Mean annual wind speed is 6.5–7.5 m s^{-1}, with 58 days of gales per year, mainly from the southwest, northwest and north (Barne et al. 1997). During a storm in January 1992, gusts reached 34.7 m s^{-1} (125 km h^{-1}) at Lerwick, with an unofficial gust of 67 m s^{-1} (241 km h^{-1}) recorded at Muckle Flugga, Unst. The Shetland wave climate is equally severe, with waves at the Schiehallion oil platform, 160 km west of Eshaness, regularly exceeding 20 m and with the highest individual wave in January 1992 reaching 28 m, only 1.1 m less than the highest individual wave ever recorded, at Rockall, 250 km west of Scotland (Holliday et al. 2006). Foley (2019) has suggested that between 1900 and 2009 there may have been an increase in the intensity, but not the frequency, of cyclonic weather associated with severe gales and associated storm waves. Deepwater waves generally attenuate as they approach shore and this is the case within the Shetland voes and inlets, where small beaches and spits have developed. Along much of the outer Shetland coast, however, steep or vertical cliffs plunge into deep water; wave attenuation is therefore often very limited at such locations and the shore itself can be exposed to exceptionally high wave energies (Fig. 7.9a).

Large unbroken incident waves regularly impact steep cliff faces on Shetland with enough force to quarry large blocks from the upper cliff face and top surface, before decoupling into green-water bores capable of transporting the detached blocks landward over the cliff-top surface and depositing them in cliff-top ridges (Hall et al. 2006; Fig. 7.10a). Well-developed examples include Grind of the Navir and nearby Villians of Hamnavoe, at Eshaness (Hall et al. 2008) and the cliffs of Out Skerries and Mousa on the North Sea coast (Hall et al. 2006). At these locations, cliff-top storm deposits (CTSDs) of large boulders up to 2 m long have accumulated in multiple ridges up to 3 m high and 50 m from the cliff edge (Fig. 7.10). Modelling by Hansom et al. (2008) suggests that cliff-top bore velocities of up to 7 m s^{-1} and bore depths of up to 5 m are required to effect boulder transport at Grind of the Navir, a conclusion supported by video recordings during storms in 2011 and 2013.

Fig. 7.10 **a** Six cliff-top storm deposit (CTSD) ridges at Grind of the Navir lie at up to 20 m above sea level and were emplaced and regularly reorganized by overtopping storm waves. **b** The wave-eroded basin at Grind of the Navir is accessed by bores of overtopping water that have transported large boulders (in the foreground) landwards to accumulate in CTSD ridges, tens of metres from the cliff edge. (Images: James Hansom)

Repeat surveys (Fig. 7.10b) and modern human-derived flotsam trapped within the ridges demonstrate that CTSDs are subject to frequent reorganization during major storms (Hall et al. 2008). However, radiocarbon dating of marine shells within the ridges and of peat above and below storm-deposited boulders, together with optically stimulated luminescence ages of intercalated sand, provide approximate ages of enhanced CTSD emplacement during periods of increased storm activity during AD 400–550, AD 700–1050, AD 1300–1900 and since AD 1950 (Hansom and Hall 2009). Despite claims to the contrary, CTSDs do not appear to be tsunami deposits and no records of significant tsunami impact on Shetland coasts exist for the past ~1500 years (Hall et al. 2010). As elsewhere on North Atlantic coasts (Suanez et al. 2009), stormwave impact appears to have been responsible for both the origin and ongoing modification of CTSDs.

However, three possible older tsunami layers (mainly sand) have been recorded at various sites in Shetland (Bondevik et al. 2005). The earliest, dated to ~8.1 ka, is attributable to the well-documented Storegga submarine landslide west of Norway (Dawson et al. 1988). The two younger 'tsunami' events are estimated to have occurred at ~ 5.5 ka and ~1.5 ka (Dawson et al. 2006), but their origin is unknown and evidence for post-Storegga tsunami events has yet to be found outside of Shetland (Smith et al. 2019).

In locations sheltered from the most severe storm waves, beaches, dunes and tombolos have developed where sand and gravel are locally available. Minor sand, gravel and boulder beaches are common at cliff-foot and bayhead settings, but extensive beaches are mainly limited to the southwest coast of Mainland, where the St Ninian's Isle tombolo (Fig. 7.11a) is the largest geomorphologically active sand tombolo in Britain. Despite ongoing rising sea levels in Shetland, the 500 m long tombolo has been in existence since at least AD 1700. Strikingly symmetrical in

Fig. 7.11 **a** The strikingly symmetrical St Ninian's Isle tombolo in south Mainland is the largest geomorphologically active sand tombolo in Britain. **b** Ayres of Swinister. Only the South Ayre (right) is a true tombolo; the others are spits that enclose The Houb tidal lagoon. The central spit is younger, less well developed and formed of smaller and more angular sediment than the outer ayres that are exposed to ocean waves. (Images: **a** © Colin Smith, Creative Commons Licence; **b** © Robert W. Watt, Creative Commons Licence)

plan, it is subject to wave activity from opposing north and south directions. With waves breaking simultaneously along the entire length of the two flanking beaches, the planforms of both beaches appear to be in dynamic equilibrium, despite overtopping during storms. With 50% shell-sand content, the tombolo is currently nourished by nearshore shell banks and more locally from thin layers of till- and rock cliff-sourced gravels (Hansom 2003c). Excess sand is blown from both ends of the tombolo into climbing dune systems, the largest of which lies at the mainland end.

Twin sand beaches backed by dunes join Sumburgh Head to Mainland and bear all the hallmarks of a wide tombolo with a low-lying, sand-infilled centre that now forms the foundations of an airport runway. Other sandy beaches and dunes occur nearby in relatively sheltered sites. At Quendale, close to Sumburgh Head (Kelley et al. 2018), dune systems became mobile during the exceptionally stormy conditions of the Little Ice Age and buried settlements and farmland (Bampton et al. 2017). Significant changes to beaches on Unst also occurred at this time (Preston et al. 2020). Elsewhere in Shetland, many short spits and tombolos (locally called *ayres*) jut out from sheltered coasts, locally linking individual islands or connecting islands to Mainland. Many are found some way along, and often on both shores of, Shetland voes, at points where the transport capacity of waves has diminished sufficiently to allow sediment deposition. In such conditions, the ayres have built out from each shore and some have joined to form gravel tombolos. An outstanding example is the Ayres of Swinister in north Mainland, where three gravel spits are fed by angular and subangular gravels eroded from the adjacent rocky shore (Fig. 7.11b). They all extend from the north shore although only one, the South Ayre, forms a complete tombolo connecting an offshore island to Mainland. The other two extend out from the Mainland coast as spits that have rotated over time to become mid-bay barriers orientated to face incoming wave directions, and enclose a tidal basin called The Houb (Hansom 2003d). Subtidal peat dated to ~5.3 ka (Birnie 1981) partly floors The Houb and appears to underlie only its central barrier, suggesting that marine flooding arrested further peat development. No peat outcrops can be seen eroding out of the foreshores of the North and South Ayres, which suggests the central barrier formed later and was fed by locally derived, wave-winnowed sediments once The Houb had formed.

The past impact of sea-level rise on Shetland beaches has been partly documented by the Dynamic Coast project but since the coast is mainly rocky, only limited erosion has been identified since the 1970s (Hansom et al. 2017). However, it is likely that as sea-level rise accelerates, more widespread erosion will affect Shetland's beaches.

7.7 Conclusions

The Shetland archipelago is outstanding for the scenic quality of its terrestrial and coastal landscapes. The underlying geology and rock structure have been exploited by long-term differential weathering and erosion to shape the dramatic coastline and the varied topography and landforms of the islands. Glacial erosion during the Pleistocene emphasized pre-existing valley patterns, the lower parts of which were drowned by rising postglacial sea level to form the intricate, strongly indented coastline. Offshore, in the bays and on the continental shelves, recessional moraines record the landwards retreat of the Shetland Ice Cap during the last deglaciation. Geomorphological highlights include the Appalachian-style relief in east Mainland, the assemblage of spectacular rock-coast landforms, storm-generated cliff-top deposits, coastal spits, bars and tombolos (including the largest active sandy tombolo in Britain), active wind- and frost-related periglacial landforms at relatively low altitudes, and peat slides and bog bursts on the blanket bog. The entire archipelago is included within Shetland UNESCO Global Geopark, a fitting accolade for a remarkable landscape fashioned by its diverse geology and geomorphology.

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8 Orkney and Caithness

Adrian M. Hall and James D. Hansom

Abstract

The landscapes and landforms of Orkney and Caithness have been profoundly influenced by the varied lithology and structures of the Devonian sedimentary cover. During the Neogene, an extensive planation surface developed across Caithness, with sandy weathering covers, which was subsequently uplifted and dislocated by fault movements. In the Pleistocene, ice flowed out of the Moray Firth, overdeepening the marine basin, and crossed northern Caithness and Orkney. The land surface was locally roughened and streamlined but in southwest Caithness and on some hill summits during the most recent ice-sheet glaciation(s) the ice cover remained cold-based and non-erosive. The last ice sheet left complex glacigenic deposits and formed sequences of recessional moraines during its retreat from near the Atlantic shelf edge. The coast is predominantly rocky, with an outstanding assemblage of cliffs, caves, stacks, arches and shore platforms reactivated by rising post-glacial sea levels. The severe wave-energy environment has produced remarkable examples of cliff-top storm deposits. In contrast, in sheltered, low wave-energy environments in the northern isles of Orkney, glacigenic and biogenic sediments have been reworked into extensive beaches, tombolos and dune systems. The exceptional landscape of the Flow Country of Caithness developed through the formation of extensive blanket bog during the Holocene.

Keywords

Palaeosurfaces • Moray Firth ice stream • Cliffs • Shore platform • Cliff-top storm deposits • Beaches and dunes • Flow country • Blanket bog

8.1 Introduction

Orkney and Caithness are located on the Orkney-Shetland Platform (OSP), flanked by the sedimentary basins of the Moray Firth and northern North Sea and by the West Shetland shelf. The Orkney Isles archipelago has a total land area of $\sim$1000 km^2, with over 70 islands and skerries. Most of the terrain is rolling and low-lying, seldom higher than 100 m OD (Ordnance Datum), apart from hills on Hoy and Rousay and in the central basin of west Mainland. Caithness is significantly larger in area, $\sim$1750 km^2, and dominated by the plain of Caithness at 50–200 m OD, with hills rising above 300 m along its southern fringe. Lower and middle Devonian flagstones dominate the scenery of Orkney and Caithness to form one of the largest areas of sedimentary rock terrain in Scotland.

The geomorphology of Orkney and Caithness displays several features of exceptional interest. The plain of Caithness largely conforms to a Neogene planation surface, with remnants of sandy weathering covers. The extent and lithological diversity of the Devonian sedimentary sequence, especially in Caithness, has allowed the development of a range of structural landforms. During the Pleistocene, Caithness and Orkney were overrun by successive Scottish ice sheets, which locally deposited distinctive shell-rich tills derived from erosion of Mesozoic and older rocks from the Moray Firth Basin. In northern Orkney, till contains scarce, far-travelled glacial erratics from Scotland, the North Sea and Scandinavia. The Atlantic coast of Orkney and much of the Caithness coast is precipitous, with cliffs up to 340 m high on Hoy. Along these exposed coasts, storm-wave

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erosion has produced many striking physical features. These include sea stacks; long narrow inlets or *geos* excavated along joints, faults and dykes; blowholes or *gloups*; and cliff-top storm deposits. Elsewhere in Orkney, distinctive depositional forms include sand and gravel bars and spits in more sheltered locations. In western and central Caithness, the Flow Country is one of the largest expanses of blanket bog in the world.

8.2 Geological Framework

The geology of Caithness and Orkney is dominated by a metamorphic and igneous basement with a Devonian sedimentary cover. The varied suite of gneisses, mica-schists and quartzites on the western and southern borders of Caithness belongs to the Moine Supergroup. These metamorphic rocks now lie in Caledonide thrust nappes that were later intruded by granite and diorite (Kocks et al. 2006; Lundmark et al. 2019). In Caithness, shear zones and major faults associated with thrusting have a generally north–south orientation and guide the courses of rivers draining to the north coast, and the detailed configuration of bays and headlands on that coast (Fig. 8.1a). The Great Glen Fault system, with its splays and extensions in the Helmsdale and Wick Faults, forms a major crustal discontinuity along the edge of the Mesozoic Moray Firth Basin. The fault system was initiated in the Caledonian Orogeny, reactivated during the Palaeogene, with significant differential movements between fault blocks in the inner Moray Firth (Argent et al. 2002; Richardson et al. 2005), and probably remained active into the Neogene.

During the Devonian, the Orcadian Basin was occupied by a vast freshwater lake, Lake Orcadie, which extended from north of the Highland Boundary Fault through Shetland to Greenland. The basin fill is dominated by laminites, mudstones, limestones and sandstones (known locally as *flagstones* where thinly bedded) derived from lacustrine, fluvial and aeolian sediments deposited in and around the lake (Trewin and Thirlwall 2002). The Caithness and Orkney Devonian sediments have a total thickness of ~2 km (Marshall and Hewett 2003). In Orkney, the beds are often folded into gentle anticlines and synclines with axes trending between NNW and NNE (Fig. 8.1b). In Caithness, a major anticline brings to the surface the basal Lower Devonian conglomerates at Sarclet. North and south of this arch, the strata lie in the Ackergill and Latheron Synclines (Crampton and Carruthers 1914).

Upper Devonian rocks occur on Hoy and at Dunnet Head in northern Caithness. On Hoy, the eroded surface of the faulted and gently folded Middle Devonian sandstone is overlain by tuffs and basalt lavas and mainly aeolian sandstones (Tang et al. 2018). The Devonian sequence in Caithness and Orkney is cut by prominent dolerite and lamprophyre dykes emplaced during regional extensive magmatic activity in the late Carboniferous and Permian (Lundmark et al. 2011). The dykes are often eroded out as geos in cliffs (Wilson et al. 1935).

Rocks at the present land surface have been exhumed from beneath a formerly thick cover of Palaeozoic and younger rocks. In Caithness, the maturity of Devonian oil-bearing laminites indicates that uplift and erosion in the Permo-Triassic removed 2–3 km of Late Devonian and Carboniferous cover (Hillier and Marshall 1992). In Orkney, the Devonian sequence was less deeply eroded in the Variscan uplift phase (Brown et al. 2019). Overlying Palaeozoic and younger strata were removed when Orkney and Caithness were uplifted in the Palaeogene (Morton et al. 2004; Wilkinson 2017). The volumes of Palaeocene–Eocene sands shed to the outer Moray Firth Basin, which reach maximum thicknesses of 2 km (Liu and Galloway 1997), imply deep erosion of source areas on the OSP. Hence, the landforms in Caithness and Orkney date mainly from the Neogene and Pleistocene.

8.3 Pre-glacial Relief Development

8.3.1 Exhumed Relief

The sub-Devonian unconformity has influenced the present topography along the edges of the Devonian cover in southern and western Caithness (Godard 1965; Fig. 8.1a). Numerous outliers of Devonian conglomerate and sandstone indicate former extension westward into Sutherland. Spectacular inselbergs are developed in resistant conglomerates and quartzites in SE Caithness (Fig. 8.2a). On the flanks of Scaraben, lithified Devonian scree plasters ancient hill slopes, and locally sourced Devonian conglomerates fill the northern end of the Berriedale gorge cut through the quartzite outcrop (Crampton and Carruthers 1914). Yet nearby, granitic breccias rest on a much gentler surface at the sub-Devonian unconformity (Godard 1965). Similar transitions between exhumed and present low-angle surfaces occur on the northern part of the Strath Halladale Granite. The proximity of the sub-Devonian unconformity to the present erosion level is a widespread feature in the eastern Scottish Highlands.

8.3.2 Neogene Landform Development

The drainage pattern in Caithness and neighbouring parts of Sutherland includes major rivers that drain to the north and southeast. The lower courses of the Helmsdale River and Berriedale Water, which reach the Moray Firth via narrow valleys, predate the 400–700 m uplift of the coastal hills

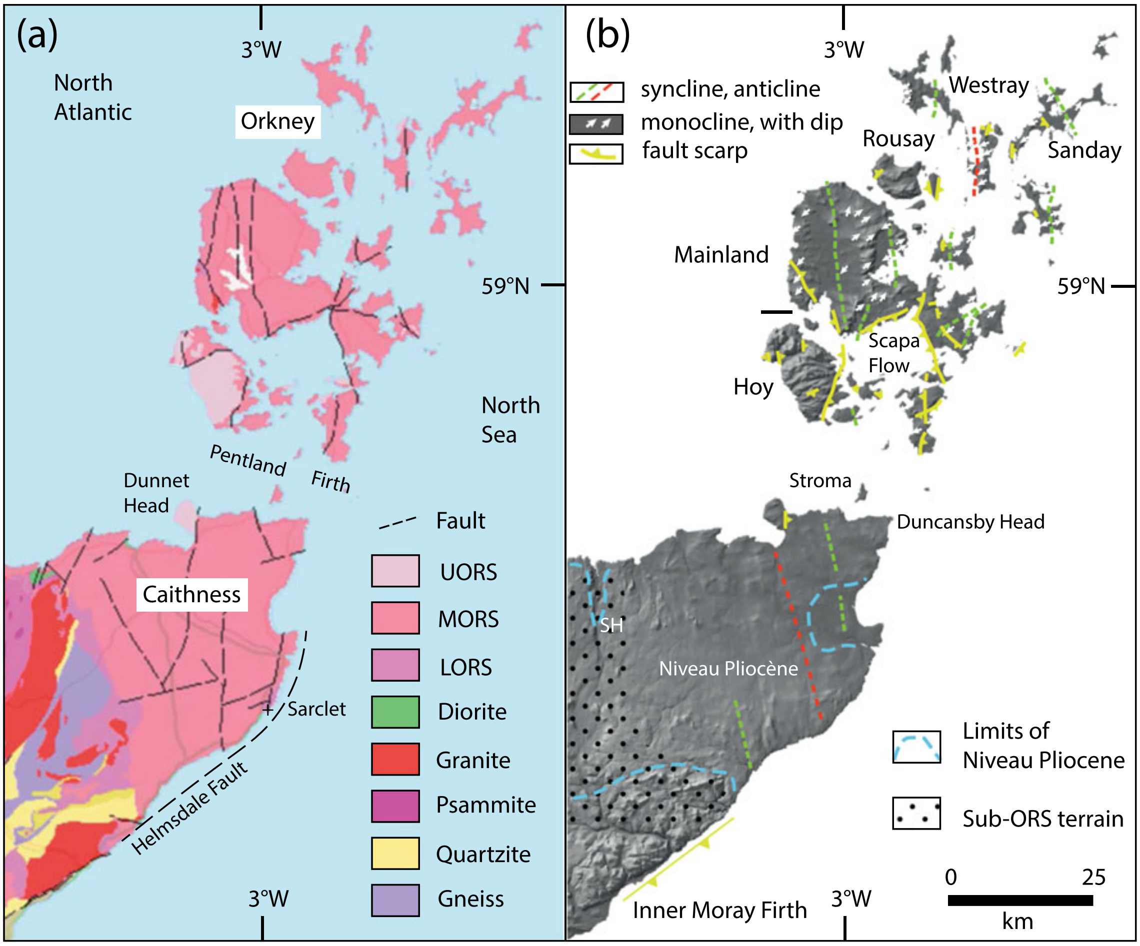

Fig. 8.1 Geology and relief in Caithness and Orkney. **a** Geology. Simplified Devonian sedimentary sequence: UORS (Upper Old Red Sandstone); MORS (Middle Old Red Sandstone); LORS (Lower Old Red Sandstone). **b** Relief. Outline of the main structural features on Orkney (after Godard 1965) and palaeosurfaces in Caithness. SH: Strath Halladale

along the Helmsdale Fault (Read 1931) and so are a superimposed part of an ancient centripetal drainage system of the inner Moray Firth (Hall 1991).

The largest and likely oldest group of landforms in Caithness are the erosion surfaces which cross-cut the sub-Devonian unconformity. Godard (1965) recognised fragments of an erosion surface on the Knockfin Heights at ~430 m OD and in the hills of southern Caithness at 250–320 m OD. The dominant relief feature, however, is his 'Niveau Pliocène' (Pliocene palaeosurface), which extends as a monotonous rolling plain across much of Caithness at an elevation of 80–210 m. It is cut across metamorphic rock types and tilted Devonian strata (Fig. 8.2). This palaeosurface continues westward into Sutherland on the north coast (Fig. 8.1b) and fragments survive at 65–90 m in east Mainland, Orkney. Godard pointed out that the Niveau Pliocène is a relatively young feature because of its low elevation, limited dislocation by faults, abrupt termination at the coast in high sea cliffs, extension inland along the main valleys, and preferential development across rocks of low resistance to chemical weathering.

Remnants of former weathering covers occur widely in SW Caithness. Depths of weathering locally exceed 5 m and the weathering is of grus type, with low clay contents (Omand 1975). Rock structures are preserved, including quartz veins, foliation and joints. Only the most vulnerable primary minerals have broken down, so that the saprolite is formed of blocks, granules and grains of broken rock and minerals. Impressive sections in deeply weathered rock occur along the margin of the Helmsdale Granite at Suisgill in Helmsdale (Zauyah 1976). The age of the weathering is uncertain, but it is likely that this grus-type weathering developed under temperate conditions in the Pliocene and Early Pleistocene.

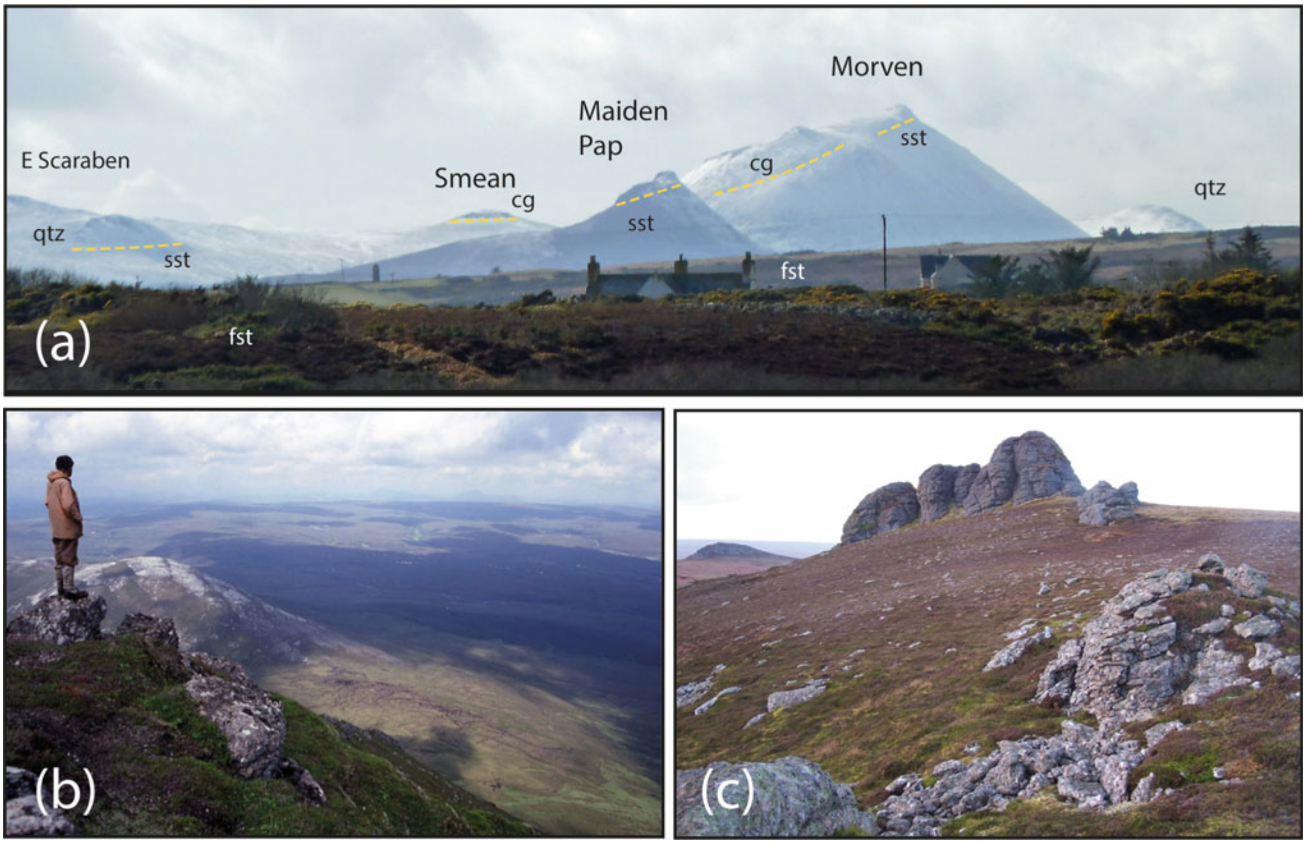

Fig. 8.2 Landscapes of southern Caithness. **a** View southwest from Lybster across the elevated Caithness plain. In the foreground, the Niveau Pliocène is developed in Devonian flagstones (fst). An abrupt break of slope is developed in sandstone (sst), with hill summits in conglomerate (cg). The exhumed sub-Devonian topography is seen in the Dalradian quartzite (qtz) hill of Scaraben which rises out of the Devonian cover. **b** View west from Morven (706 m) over the quartzite summit of the Small Mount, stripped bare by inland ice moving across it. In the background is the Niveau Pliocène. **c** Tors in Lower Devonian conglomerate at 509 m on Smean, SW Caithness. (Images: Adrian Hall)

8.3.3 Rocks and Relief

Whilst the Caithness relief is locally inherited from the sub-Devonian unconformity, there are clear relationships between topography, rock type and fracturing produced by differential weathering and erosion through the Neogene. In the Appalachian-type topography of the Langwell Forest, the ridges follow resistant bands of quartzite and the valleys have been opened out in more rapidly weathered, often biotite-rich igneous and metamorphic rocks. On the subdued Moine plateau in SW Caithness, low hills in granite and Devonian conglomerate rise above the surrounding sandstone terrain (Futty and Dry 1977). Topographic basins occur in rocks with low resistance to weathering, such as the Reay Diorite and the fractured and veined biotite gneisses west of Borrobol (Read 1931).

On the Caithness plain, the dips of the flagstones and sandstones are generally low and control the detail of the many gentle slopes. In Orkney, Godard (1956) identified structural landforms where hills and valley slopes locally conform to anticlines, monoclines and synclines in the Devonian sedimentary sequence (Fig. 8.1b). He noted also that the rounded whaleback forms of many hills have convex upper slopes and concave lower slopes and are abruptly cut off at the coast. Rousay and western Westray display a strikingly rugged topography; hard flagstones and soft laminites form stepped hillsides, often with craggy risers known locally as *hamars* (Dry and Sinclair 1986). At numerous locations on Orkney, fault scarps result from differential neotectonic movement of rock blocks (Godard 1956), with the fault-bounded and flooded basin of Scapa Flow providing several examples. The high elevations of the Late Devonian sandstones on western Hoy and at Dunnet Head also appear anomalous and may indicate Neogene uplift.

8.4 Pleistocene Landforms and Sediments

8.4.1 Ice-Sheet Configuration

During the Pleistocene, Caithness and Orkney lay peripheral to the main ice accumulation centres in Scotland. During successive glaciations, ice moving out of the inner Moray Firth crossed Caithness and Orkney and was confluent for long periods with inland ice emanating from the Caithness-Sutherland border. The Moray Firth Ice Stream (MFIS) expanded into the North Sea and, at the Last Glacial Maximum (LGM), was confluent with the Fennoscandian Ice Sheet and the Shetland Ice Cap east of Orkney (Merritt et al. 2019).

Glaciological models for the last British-Irish Ice Sheet indicate a limited duration of ice cover in Caithness amounting to a total of <50% of the modelled 40–10 ka period (Hubbard et al. 2009). A local ice cap in Orkney is a feature of these models, but no stratigraphic evidence has been identified for associated ice movements. These models suggest that maximum ice thickness was <1500 m over Caithness at ~20 ka and that the ice emanating from the Moray Firth was warm-based for long periods, but ice to the west and ice over the hills of southern Caithness and Hoy remained cold-based for all or most of the last glaciation. The models also suggest that the MFIS and ice in adjacent parts of the North Sea were highly dynamic in their configuration and activity, with long periods of ice build-up and short phases of ice streaming.

8.4.2 Patterns of Glacial Erosion

The inner Moray Firth Basin is a major regional feature of glacial erosion. Pliocene deltaic or lacustrine sediments, with kaolinitic clays and lignite, occur in the outer Moray Firth ~100 km east from the present shoreline (Andrews et al. 1990), and the present marine embayment did not exist at this time. The seabed off Helmsdale lies at −40 m and drops to −60 m off Dunnet Head and east of Orkney. A set of linear deeps occurs at ~200 m depth towards the Buchan coast. The deeps likely represent the westernmost tunnel valleys in the central North Sea (Ottesen et al. 2020; Chap. 6). Excavation of tens of metres of Mesozoic and Palaeogene sedimentary rocks from the inner Moray Firth is implied, likely mainly a product of erosion by the MFIS over the last million years, a period of deep erosion of inner shelves and sedimentary basins in NW Europe (Hall and van Boeckel 2020). Glacial erosion also resulted in overdeepening of the Pentland Firth, likely along a pre-existing fluvial valley (Godard 1956), and the separation of Orkney from Caithness. The firth was a pathway for ice streams during the last glaciation (Hall et al. 2016). In the Orkney archipelago, however, inter-island water depths rarely exceed 20 m, except in a few areas, such as the NW–SE oriented Stronsay Firth. Isolation of the individual islands is a product of limited glacial deepening of pre-existing hollows and valleys along Hoy Sound, Eynhallow Sound and Westray Firth, combined with postglacial submergence.

Ice-roughened terrain is restricted in extent, in part due to the control of the Devonian sedimentary rocks on glacial erosion (Fig. 8.3). The jointed and often horizontally bedded sandstones and flagstones were readily quarried by glacier ice but are compact, strongly cemented and relatively resistant to abrasion. However, high terrain roughness is present locally at the coast (e.g. between Clyth and Ulbster) where ice from the Moray Firth was forced to rise ~180 m from the adjacent sea floor. The ice exploited weaknesses in the Berriedale Sandstone and formed rock ridges and furrows that now hold several shallow lochs along north–south aligned faults (Crampton and Carruthers 1914). Stepped slope profiles have lost sandstone blocks to lateral and lee-side plucking. On the plain of Caithness itself, the terrain is weakly streamlined (Fig. 8.3) but the density and depth of lake basins are low. On the western or inland edge of the plain, few glacial bedforms occur and weathering is preserved widely on exposed basement rock, indicating limited glacial erosion. Only a few tens of metres height difference exist between these erosion zones, thereby constraining the likely depth of Pleistocene glacial erosion. The pattern of erosion is largely an expression of the longer duration of erosive, warm-based basal ice conditions in NE Caithness.

On the mountains of southern Caithness and eastern Sutherland, geomorphological evidence for the upper limit of glacial erosion implies that, during the LGM, warm-based erosive ice extended no higher than 410–600 m (Fig. 8.2b). Sites below these altitudes have yielded cosmogenic ^{10}Be apparent exposure ages for deglaciation of 16.6 ± 2.1 ka to 18.5 ± 2.4 ka (Ballantyne and Hall 2008; Chap. 4). At higher elevations, a cover of frost-weathered detritus occurs, with summit tors on conglomerate bedrock (Fig. 8.2c) that have yielded ^{10}Be apparent exposure ages of 146.6 ± 15.1 ka and 178.2 ± 18.1 ka. These ages are minima but indicate that erosion rates on the exposed conglomerate were low. Limited glacial erosion is also consistent with a ^{10}Be exposure age of 29.3 ± 2.9 ka obtained for bedrock on Ward Hill, Hoy. These ages and the preservation of frost-weathered debris on summits indicate that during the last glaciation the ice cover over the highest ground in Caithness and Orkney remained cold-based and accomplished very limited erosion.

Well-developed cirques are absent on the Caithness hills and only three are present in Orkney, all on north-facing locations in northern Hoy (Fig. 8.4a). All three are notable in

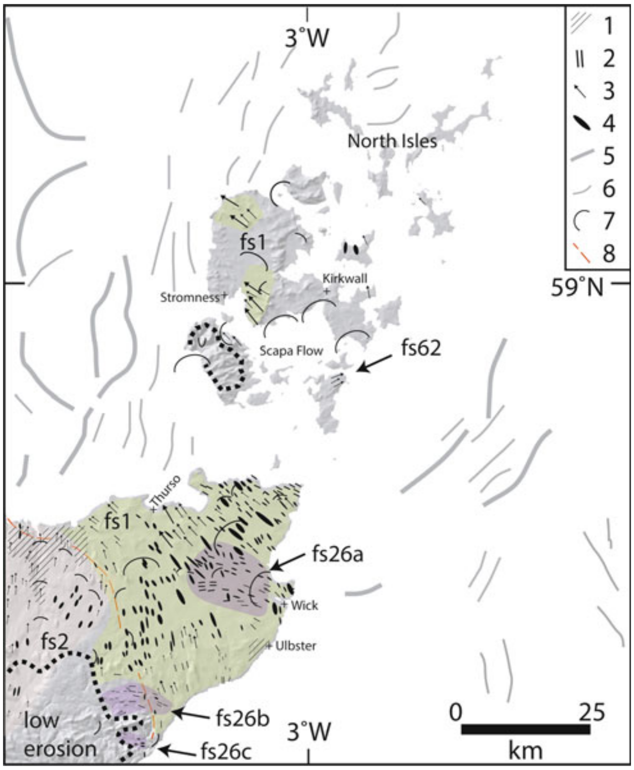

Fig. 8.3 Glacial landforms and flow sets (fs) on Caithness and Orkney. 1: ice-roughened terrain. 2: mega-scale glacial lineation. 3: crag and tail. 4: drumlin. 5: major offshore moraine. 6: minor offshore moraine. 7: former ice margin position on land. 8: western edge of till deposited by ice from the Moray Firth. (Modified from Hall et al. 2016. Data for landforms 1–3 modified from Hughes et al. 2010; data for landforms 4–5 modified from Bradwell and Stoker 2015)

having cirque floors at unusually low elevations (75–130 m OD) indicating high snow accumulation rates near present sea level during Pleistocene cold stages (Barr et al. 2017).

8.4.3 Glacial Deposition

A series of moraine ridges close to the NW Atlantic shelf edge marks the maximum westward extent of Scottish ice during the Middle and Late Devensian (Bradwell et al. 2008; Clark et al. 2012). During the LGM, the British-Irish Ice Sheet was confluent with the Fennoscandian Ice Sheet in the North Sea (Graham et al. 2011). Earlier models showing ice-free areas at low elevations during the LGM in peripheral parts of Scotland, including Caithness and Orkney (Sutherland 1984; Bowen and Sykes 1988), are inconsistent with stratigraphic and dating evidence for LGM ice cover in these areas (Hall and Whittington 1989; Ballantyne and Hall 2008; Phillips et al. 2008; Hall and Riding 2016). Nonetheless, the stratigraphic record in Caithness and Orkney is key to understanding the timing, sequence and pattern of the last ice-sheet glaciation, and its associated landforms (Merritt et al. 2019).

In Caithness and Orkney, glacial till is generally thin and rests locally on fragmented and weathered flagstones and sandstones, though in Caithness, till thicknesses reach 30 m in valleys and in other accommodation spaces. At Leavad, a remarkable glacitectonic raft occurs, 220 m long, 49 m wide and 8 m thick, composed of Lower Cretaceous sandstone that has been transported for over 15 km from its source by ice moving out the Moray Firth. Tills derived from the Moray Firth are typically dark in colour, stiff and mud-rich with many far-travelled erratic clasts; often conspicuous shell fragments led historically to these deposits being termed *shelly till* (Omand 1982).

The preservation of stacked till sequences has allowed the development of a three-stage model for the last glacial cycle in Caithness and Orkney (Hall and Riding 2016; Hall et al. 2016). From Orkney Mainland northwards, a lower brown till with few far-travelled clasts occurs locally and records a first stage of ice movement (Rae 1976). Striae orientations indicate deposition by ice that flowed to the northwest from the platform east of Orkney. In Caithness, aside from older glacial deposits found at depth, the oldest till units attributed to the Late Devensian are found around Dunbeath and Melvich (Gordon 1993; Hall et al. 2011). These tills record an early expansion of ice from the NW Highlands at least as far east as Watten (Omand 1973). This early flow stage is associated with bedrock and till lineations in SW Caithness and is provisionally assigned to the 32–28 ka interval.

In Orkney, a second stage is represented by younger tills that contain marine shell fragments and far-travelled clasts derived mainly from around the Moray Firth Basin and beyond. Scandinavian erratics occur on the northern isles of Orkney (Hall et al. 2016) and on Fair Isle (Chap. 7), but till palynomorph content suggests reworking into Scottish tills from older sediments in the outer Moray Firth and central North Sea; there is no firm evidence that the Fennoscandian Ice Sheet flowed over northern Orkney during the last glaciation. In Caithness, shelly tills are associated with mega-scale glacial lineations (Hughes et al. 2010, 2014), but till stratigraphy indicates that these subglacial landforms are composite products of multiple ice-flow events that followed similar paths (Hall et al. 2016; Hall and Riding 2016). At Wester Clett, for example, an inland till is overlain by cryoturbated gravel and three calcareous, mud-rich till units deposited by ice flowing northwestwards from the Moray Firth Basin (Hall et al. 2011). Glacial erratics, striae and other ice-flow indicators also indicate powerful flow of ice out of the Moray Firth Basin and across Caithness and Orkney, likely at 24–18.5 ka.

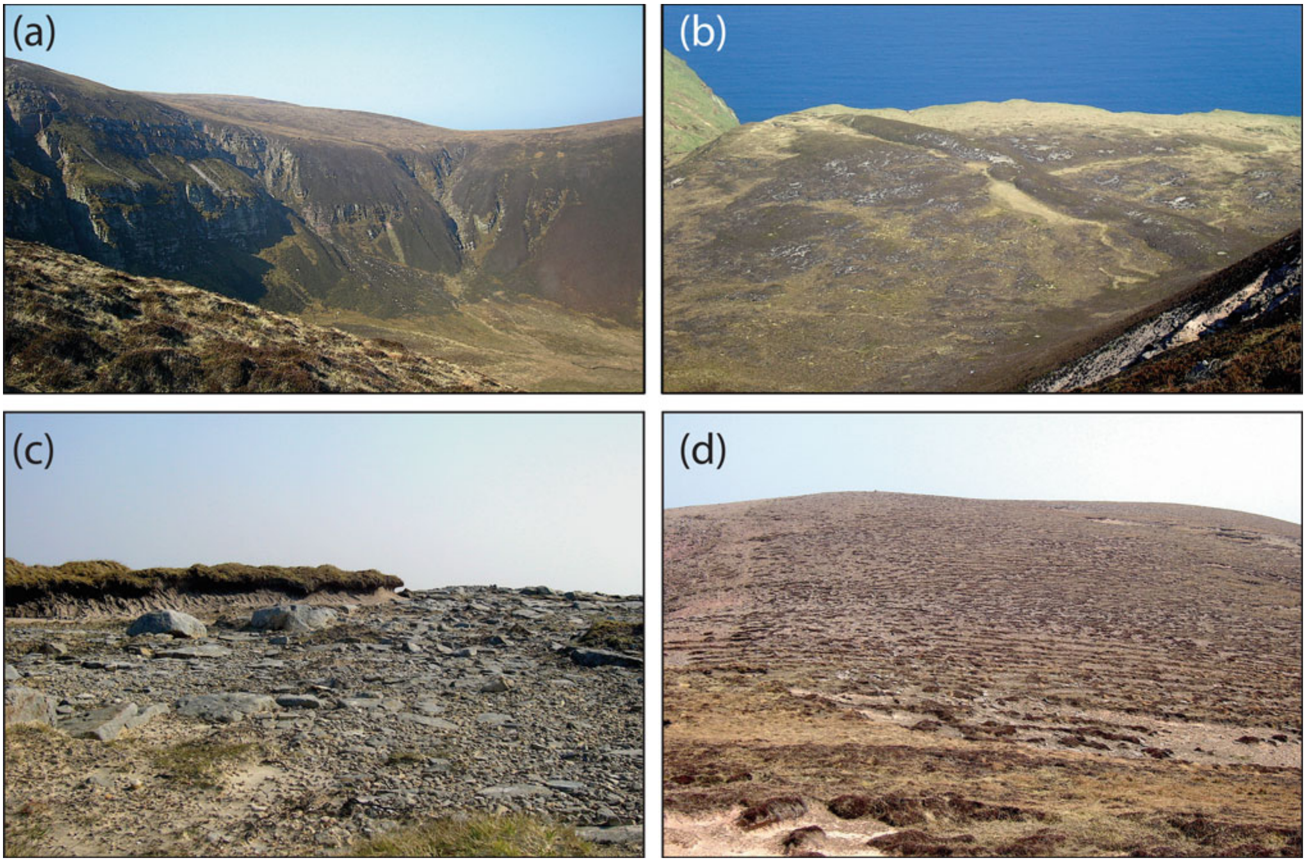

Fig. 8.4 A Enegars corrie, one of three low-level cirques in northern Hoy; the cirque floor is only 110–130 m above present sea level. **b** Loch Lomond Stadial end moraine on the lip of Enegars corrie. **c** Deflation surface and remnant 'island' of aeolian sand at 450 m on Ward Hill, Hoy. **d** Wind-patterned ground and turf-banked terraces on Ward Hill, with a remnant of formerly more extensive aeolian sand cover in the foreground. (Images: Colin Ballantyne)

The final stage marks the retreat of ice flowing from the Moray Firth Basin. Moraine systems occur on the Atlantic shelf (Bradwell and Stoker 2015), on west Rousay and in central Orkney Mainland (Wilson et al. 1935), and moraines and glacilacustrine sediments are present in the valleys of western Hoy (Hall et al. 2016; Fig. 8.3). The moraine ridges indicate southeastwards retreat across Orkney, into Scapa Flow and the Moray Firth. An uppermost rubble till was deposited in this late ice-flow phase (Hall et al. 2016). Cosmogenic exposure ages indicate that low ground on west Mainland became ice-free by ~6 ka (Ballantyne and Small 2018; Chap. 4). A final advance of Moray Firth ice deposited till and formed moraine ridges on low ground bordering the Pentland Firth. Final deglaciation across low ground in Caithness was at ~15 ka (Merritt et al. 2019).

On Orkney, two small (~0.7 km^2) glaciers subsequently reoccupied low-level cirques in northern Hoy, depositing well-defined end moraines (Fig. 8.4a, b). Cosmogenic ^{10}Be exposure ages obtained for boulders on the moraines averaged 11.7 ± 0.6 ka, confirming that that these glaciers formed under the renewed cold conditions of the Loch Lomond Stade (~12.9 to 11.7 ka). Reconstructed equilibrium-line altitudes for these glaciers (~100 m and ~180 m above present sea level) are the lowest obtained for Loch Lomond Stadial glaciers in Britain, reflecting their northerly location and high snowfall input augmented by the contribution of snow blown from adjacent high ground (Ballantyne et al. 2007).

8.5 Lateglacial and Holocene Landforms

8.5.1 High-Level Aeolian Features

The present climate of Caithness and Orkney is modulated by proximity to the North Atlantic. Annual temperature ranges are small and precipitation is moderate (750–1000 mm a^{-1}), but gale-force winds (>80 km h^{-1}) are common over high ground. Ward Hill (479 m) on Hoy in Orkney supports some of the most outstanding assemblages of high-level aeolian landforms in Scotland (Goodier and Ball 1975; Sutherland 1993b). The summit plateau is dominated by an extensive deflation surface developed on periglacial regolith, interrupted by isolated ‘islands’ of vegetation-covered aeolian sand sheets (Fig. 8.4c). The latter resemble those on mountains in the NW Highlands (Chap. 13), where they have been shown to represent the remnants of more widespread sand cover that was extensively eroded under the stormy conditions of the sixteenth to nineteenth centuries (Morrocco et al. 2007). The surrounding slopes are decorated with wind stripes (bands of alternating vegetated and bare ground aligned normal to the dominant wind direction; Fig. 8.4d). On steeper slopes, the vegetated stripes have arrested the downslope creep of regolith to form turf-banked terraces with vegetated treads up to 1.5 m high and bare treads (Ballantyne and Harris, 1994). The erosive potential of strong winds, salt spray and hillslope runoff is also seen on the scarred and stripped hillslopes of the western seaboard of Orkney around Yesnaby, on Fitty Hill on Westray and on Mull Head, Papa Westray (Dry and Sinclair 1986).

8.5.2 Blanket Bog

Covering about 4000 km^2, the Flow Country of inland Caithness is amongst the largest expanses of blanket bog in the world (Joosten et al. 2016). It is a substantial terrestrial carbon store and has outstanding nature conservation value (Chapman et al. 2009; Andersen et al. 2018). The Flows include an exceptionally wide range of landforms, vegetation and surface pattern types, including numerous pool systems (Lindsay et al. 1988). The range of mire types varies from those of the lowland Caithness plain in the east, with continental affinities, through to those on the Knockfin Heights more typical of the mountainous oceanic west (Fig. 8.5). Extensive areas of active blanket bog, where *Sphagnum* mosses and other bog species ensure continuing peat accumulation, occur in intimate association with a range of open-water, wet heath, grassland and fen communities. Similar blanket peat developed extensively over low-lying parts of the Scottish Highland landscape during the Middle to Late Holocene, either as an inevitable but rapid end-stage to soil development in this generally cold and wet climate or as a result of climatic change (Tipping 2008).

8.6 Coastal Landforms

The west coast of Orkney and most of the Caithness coast is rocky, steep and cliffbound (Fig. 8.6a, b). At St. John’s Head, Hoy, the cliffs reach 340 m in height, but most do not exceed 100 m. Exposure to storm waves has produced spectacular cliff forms, including world-famous sea stacks on Hoy and at Duncansby Head, arches, geos cut along joints, faults and dykes, blow holes (locally known as *gloups*), shore platforms, and cliff-top storm deposits. Elsewhere, gravel bars, spits and small sandy beaches and dunes occur where glacigenic or fluvial sediments are locally available and there is shelter from severe wave activity. In the northern Orkney Isles, beaches and dunes are extensive on the islands of Westray, Stronsay and Sanday. The development of the Orkney and Caithness coast has been controlled by the same factors common to the Atlantic and North Sea coasts of northern Scotland, including strong structural controls, a legacy of glacial erosion and deposition, changing sea levels and, in exposed areas, a relatively severe morphogenetic environment.

8.6.1 Structural Controls

The 20 km of cliffs stretching north from the west coast of Hoy to Mainland Orkney provide some of the best examples in Europe of structural control of Devonian sedimentary rocks on coastal landform development (Hansom 2003a). At St John’s Head on Hoy, alternating beds of relatively weak, pebbly sandstone with occasional beds of harder grey flagstone bestow a slab-like, notched and often overhung profile cut by deep, steep-sided geos that often bisect headlands to produce sea stacks. Variations in hardness and near-horizontal bedding, combined with multiple joints, cracks and faults, are important factors explaining the spectacular cliffs, caves and stacks of this coastline.

The 137 m high Old Man of Hoy is one of the tallest and most spectacular sea stacks in Britain (Fig. 8.7a). Its vertical and overhanging walls fall sheer on all sides to a base of resistant basalt lava (Kellock 1969). Marine undercutting of less resistant and friable Devonian sediments at the base of the stack has resulted in sequential failure of the harder beds above. The stack is separated from the adjacent cliffs by a 60 m wide gap strewn with debris fallen from a collapsed arch that once connected it with the adjacent cliff (Hansom and Evans 1995). Historical records show that the arch failed between AD 1750 and 1819, leaving the Old Man as a twin-legged stack with a central arch. One leg was lost in an early nineteenth-century storm, creating the present monolith. At Yesnaby (Fig. 8.7b) the stacks are blocky and rectangular in plan but narrow downwards with arches eroded through their bases, indicating marine undercutting leading to the collapse of overlying beds.

On the North Sea coast of Caithness, one of the finest stretches of cliff coastline in mainland Britain extends south from Duncansby Head. Like Orkney, this coast is dominated by high cliffs, deep geos and narrow shore platforms, arches, blowholes, stacks and skerries. Cited frequently in the international literature (e.g. Trenhaile 1987), the impressive Stacks of Duncansby rise more than 60 m as castellated

Fig. 8.5 The extensive blanket bog of the Flow Country, Caithness. View northwest across patterned blanket bog with pools, hummocks and ridges near Forsinard. (Image: © Steve Moore/NatureScot)

pyramids (Hansom 2003b; Fig. 8.7c). The pyramidal shape of the stacks and a limited basal shore platform imply that wave erosion of the tightly packed lower flagstones has been less effective than subaerial weathering. This coast is characterised by steep, narrow and long geos extending inland, often with vertical or overhanging cliffs and basal notches. South of the Stacks of Duncansby, the 250 m long Wife Geo forms one of the finest compound geos in Scotland, displaying a stunning assemblage of caves, arches, plunging vertical cliffs, rock pinnacles and buttresses.

8.6.2 The Legacy of Glaciation

The erosional impact of glaciation on coastal landforms in Orkney and Caithness is related to the direction of former ice movement in relation to coastal orientation. Some cliff tops on the Caithness North Sea coast have a bevelled bedrock edge resulting from glacial erosion as ice moving out of the Moray Firth Basin surmounted the pre-existing cliffs and high ground (Fig. 8.6a, b). A striking example occurs on a cliff face near Latheronwheel, where '…the face of the cliff is finely polished and striated; the striae beginning near the water-level, and ascending the cliff obliquely' (Peach and Horne 1881, p. 324). In contrast, lee-side locations in northern Caithness and western Orkney suffered less erosion, and cliffs, geos and shore platforms were locally buried beneath till (Hall et al. 2011; Fig. 8.6b, c). Erosional overdeepening occurred in the Pentland Firth but in Orkney glacial deepening was generally restricted (see Sect. 8.4.2) with inter-island water depths rarely exceeding 20 m, despite postglacial submergence.

The legacy of glacial deposition on coastal development relates to glacigenic and glacifluvial sediments deposited both at the coast and inland in Orkney and Caithness (Sect. 8.4.3). Few substantial rivers occur in Orkney, so the main sediment source for beach building has been glacigenic sediment deposited in the nearshore zone and at the coast itself. In Caithness, these sources are augmented by river-borne sediment derived mainly from glacigenic and glacifluvial deposits inland. The impact of glacial erosion and deposition on coastal landform development has been modulated by the effect of subsequent changes in sea level.

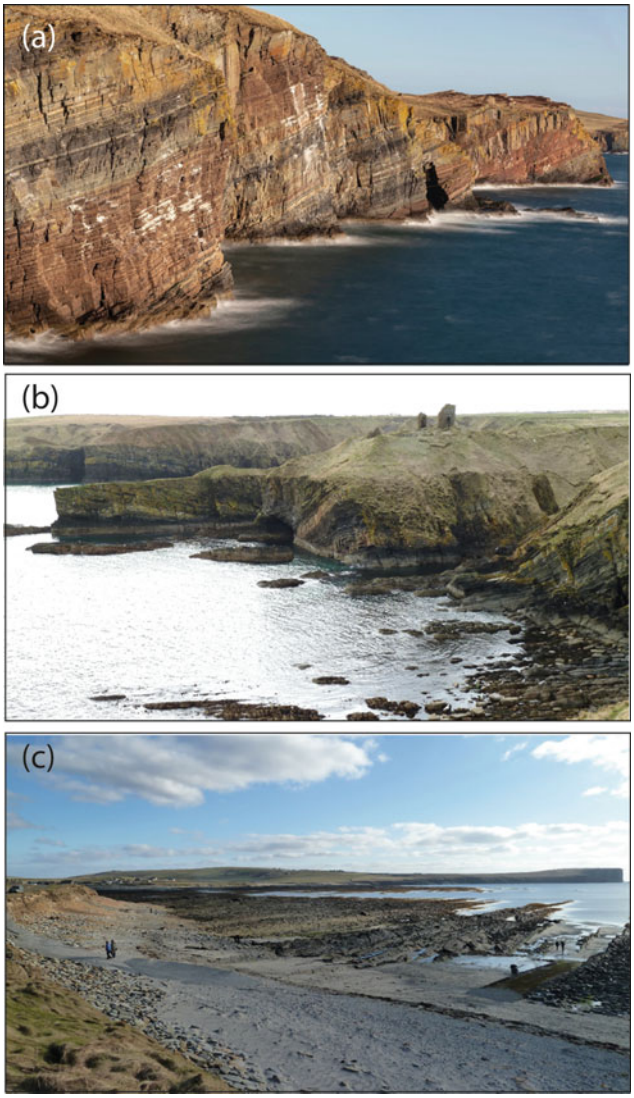

Fig. 8.6 A Steep cliffs of Lower Devonian sandstone north of Whaligoe, Caithness, show glacially bevelled cliff tops in the foreground and quarried steps in the distance. **b** Cliffs at Achastle, Caithness, and a possible emerged shore platform at ~15 m OD, cut across dipping sandstones and overlain by till. **c** Birsay, NW Mainland, Orkney, where the modern shore platform, adjusted to present-day marine processes, continues beneath a cover of glacial till, demonstrating inheritance from former sea levels. (Images: **a** Ken Crossan; **b**, **c** Adrian Hall)

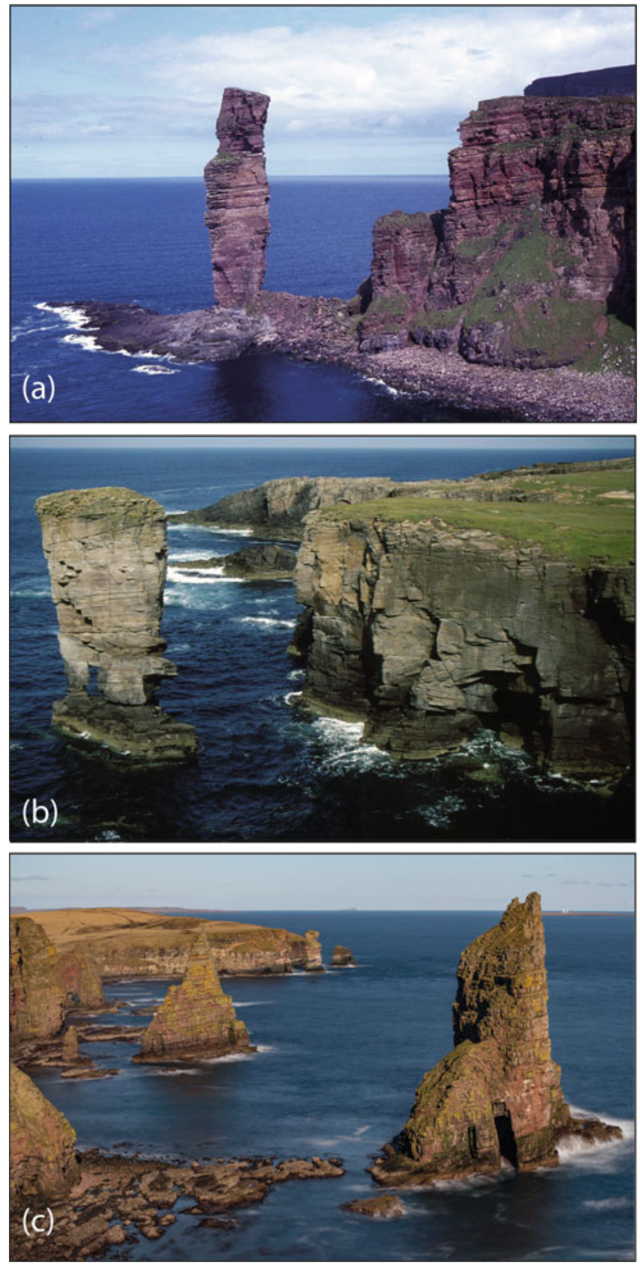

Fig. 8.7 Stacks and cliffs. **a** The Old Man of Hoy, Orkney, is less than 300 years old. **b** Yesnaby Castle, Orkney; gently-dipping Lower Devonian sandstones are undercut by waves to produce a stack that tapers downwards. **c** The Stacks of Duncansbay, Caithness; the upward taper of both stacks suggests that subaerial erosion exceeds marine erosion at the base. (Images: **a** © Colin Park, licenced for reuse by CC-BY-SA/2.0; **b** Ken Crossan; **c** © Russel Wills, licenced for reuse by CC-BY-SA/2.0)

8.6.3 Sea-Level Change

Sea-level change during the Quaternary has fundamentally influenced coastal configuration and position, particularly in those areas where widespread postglacial submergence of low-lying land resulted in the creation of islands and skerries, such as in the northern isles of Orkney. However, closer to the centre of isostatic uplift in Scotland in southern Caithness, isolated emerged beaches at 1–2 m OD suggest the postglacial rise in relative sea level was outpaced by land uplift (Omand 1982; Johnstone and Mykura 1989). On the cliffed coasts of Orkney and Caithness, sea-level rise following deglaciation reactivated many pre-existing coastal landforms, including cliffs, shore platforms and geos (Fig. 8.6b, c). More subtle impacts on the development of all Scottish coasts, including those of Orkney and Caithness, is

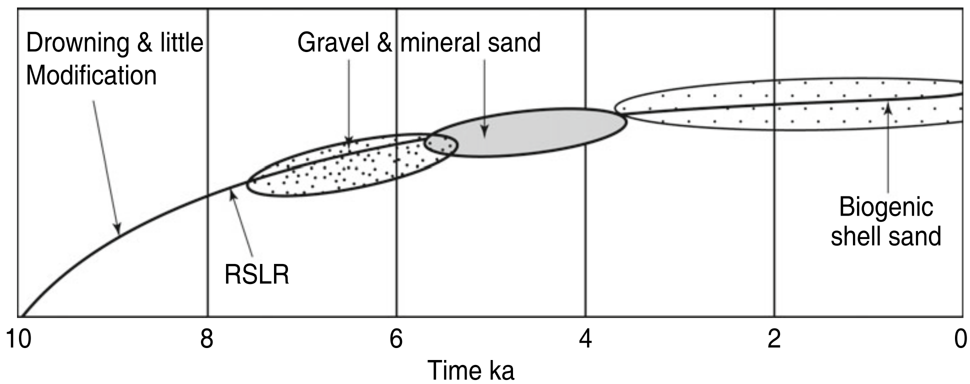

Fig. 8.8 Diagram illustrating the sequence of sediment arrival on Sanday beaches. Early rapid RSL rise mainly resulted in shoreface drowning, with onshore migration of sediment and beach building increasing as RSL rise slowed. Farrow et al. (1984) calculated rapid biogenic accumulation rates of 540 g m^{-2} a^{-1} (up to 64 cm ka^{-1}) on the Orkney Shelf. (Modified after Rennie 2006)

related to the effects of Holocene sea-level change on coastal sediment budgets (Hansom 2001).

Beyond the present coastline, submerged features have been attributed to former lower sea levels. For example, Flinn (1969) identified rock platforms at −60 m off Orkney. Bathymetric contours and recent multibeam sonar mapping around the coast of Orkney and north coast of Caithness show submerged cliffs at −10 m and −45 m water depths. Whilst it is tempting to suggest these relate to Holocene stillstands, no direct evidence of age exists. Several elevated surfaces identified as marine features (Godard 1965) may be of glacial or structural origin, but there is more convincing evidence of pre-last glaciation shorelines at several locations. On Hoy, periglacial slope deposits and till bury beach gravels resting on a shore platform at 6–12 m OD (Wilson et al. 1935; Sutherland 1993a), and on Sanday, a 600 m wide intertidal rock platform extends beneath till at Start Point. Similarly, at Scara Taing on Rousay and at Birsay on Mainland Orkney, till rests on top of pre-last glaciation striated rock platforms (Fig. 8.6c). In Caithness, both striated and till-covered intertidal and emerged rock platforms occur (Hall and Riding 2016). The glacially bevelled cliff tops in east Caithness indicate that the cliffs predate at least one glaciation and are inherited from earlier periods when sea levels were close to those of today (Fig. 8.6a).

Following deglaciation, falling relative sea level (RSL), indicated by Lateglacial shoreline fragments at 27 m and 15 m OD on the north Caithness coast (Auton et al. 2005), was replaced by a generally rising trend over the Early and Middle Holocene. In a review of sea-level data and modelling, Shennan et al. (2018) suggest that the falling RSL in Orkney and northern Caithness was replaced by a rapid rise after ~11 ka to a level close to present by ~6 ka. De la Vega Leinert et al. (2012) show the rising RSL to have reached −0.6 m by ~5.5 ka at Scapa Bay in Mainland Orkney, and at Otterswick, in Sanday, a forest at about −1.5 m OD became submerged at ~6.5 ka (Rennie 2006). Several undated emerged coastal features at ±2 m OD in Orkney and the north coast of Caithness are inferred by Brown (2003) to represent a higher relative sea level during the Viking Period about 1000 years ago. Summarising the available data, Shennan et al. (2018) concluded that RSL in Orkney, and at Wick in Caithness, has lain within ± 2 m of present sea level over the past ~6 ka.

Holocene sea-level rise transformed the shorelines of Orkney and Caithness. In Orkney, low-lying necks of land became inundated to create numerous islands and skerries, and the slowing rate of sea-level rise and progressive submergence resulted in large volumes of sediment being driven onshore from relatively shallow nearshore and offshore surfaces. Initially, glacigenic sources dominated, but as these became progressively depleted, biogenic carbonate shell-sand became an increasingly important sediment source for beach development (Hansom 1999; Rennie 2006; Fig. 8.8).

8.6.4 Morphogenetic Environment and Coastal Landforms

The Orkney coast shares many of the high-energy morphogenetic conditions found in the Shetland Isles (Chap. 7). The most characteristic feature is the frequency of strong winds and severe storm-wave conditions. Hourly mean wind speeds exceed ~30 km h^{-1} for over 30% of the year and gales occur on average for 29 days per year. Along the west (Atlantic) coast of Orkney, the sea floor falls steeply to 60 m depth, and exposure to westerly and northerly storms produces a high-energy wave climate (Hansom 2003a), whereas the North Sea coasts of Orkney and Caithness are more frequently exposed to severe northerly and easterly storms (Fig. 8.9), with occasional southerly storms. However,

Fig. 8.9 Easterly storm waves at Latheronwheel, Caithness. The North Sea coasts of Caithness and Orkney are exposed to severe northerly and easterly storm waves capable of cliff-foot quarrying and deposition of cliff-top storm deposits. At Wick, Caithness, a 1245 t foundation block was quarried by storm waves from the harbour breakwater in 1872 and its 2396 t replacement received similar treatment in 1877. (Image: Ken Crossan)

within inlets and between islands, such as Hoy Sound and Scapa Flow, wave conditions are generally less severe than on the open coast. Reconstructions of 8 ka of North Atlantic storminess, derived from peat bog data at Shebster in Caithness, suggest that over the past few thousand years storm frequency has decreased, whereas intensity has increased (Stewart et al. 2017). This is consistent with phases of enhanced cliff-top storm deposit (CTSD) activity in both Shetland and Orkney (Hansom and Hall 2009). Data for the period 1900–2009 also suggest that intensity rather than frequency of severe gales and associated storm waves may be increasing (Foley 2019).

Despite highly effective wave erosion on the west coast of Orkney, the rate of modern erosion at the cliff base and cliff top is both intermittent and site specific. As on the coast of Shetland, the occurrence of CTSDs in Orkney and Caithness indicates that relatively unattenuated storm waves readily access cliff tops in places where deepwater occurs close inshore (Hall et al. 2006). During a great storm in 1862, wave water overtopped 60 m high cliffs at the north end of the island of Stroma in the Pentland Firth. On the Atlantic coast of Orkney, CTSDs occur at 10–20 m OD where water depths reach −30 m within 500 m of the shore and intertidal rock platforms are narrow or absent. The 14 m high vertical cliff at The Nev, on Westray, is backed by a 75 m wide wave-washed platform whose landward limit is marked by ridges of CTSDs (Fig. 8.10). Where CTSDs occur, and along other stretches of cliff-line, a stripped zone is common where vegetation, weathered regolith and friable rock have been denuded by wave wash, spray and wind, with seaward-returning wave water forming dendritic systems of shallow cliff-top gullies. Below cliffs on low-elevation coasts, well-developed shore platforms provide evidence of recent marine erosion. Some platforms are boulder-abraded, some have steps and sockets from which joint-bound blocks have been removed, and others have sloping ramps that induce wave-breaking and notch cutting of the backing cliff. The sea stacks at Yesnaby, Qui Ayre, Duncansby and elsewhere lie tens of metres offshore, suggesting that wave erosion at their bases is now reduced due to retreat of the adjacent cliffs, loss of cliff-foot abrasives and a change from breaking waves (stack-foot erosion) to reflected waves associated with limited erosion (Fig. 8.7b, c).

Well-developed sandy beaches, some with dunes, occur locally in bays in Orkney, with the largest sandy beaches on the low-lying islands of Sanday and Stronsay. In general, the seabed around these two islands is relatively shallow and whilst the Holocene rising sea level created islands by

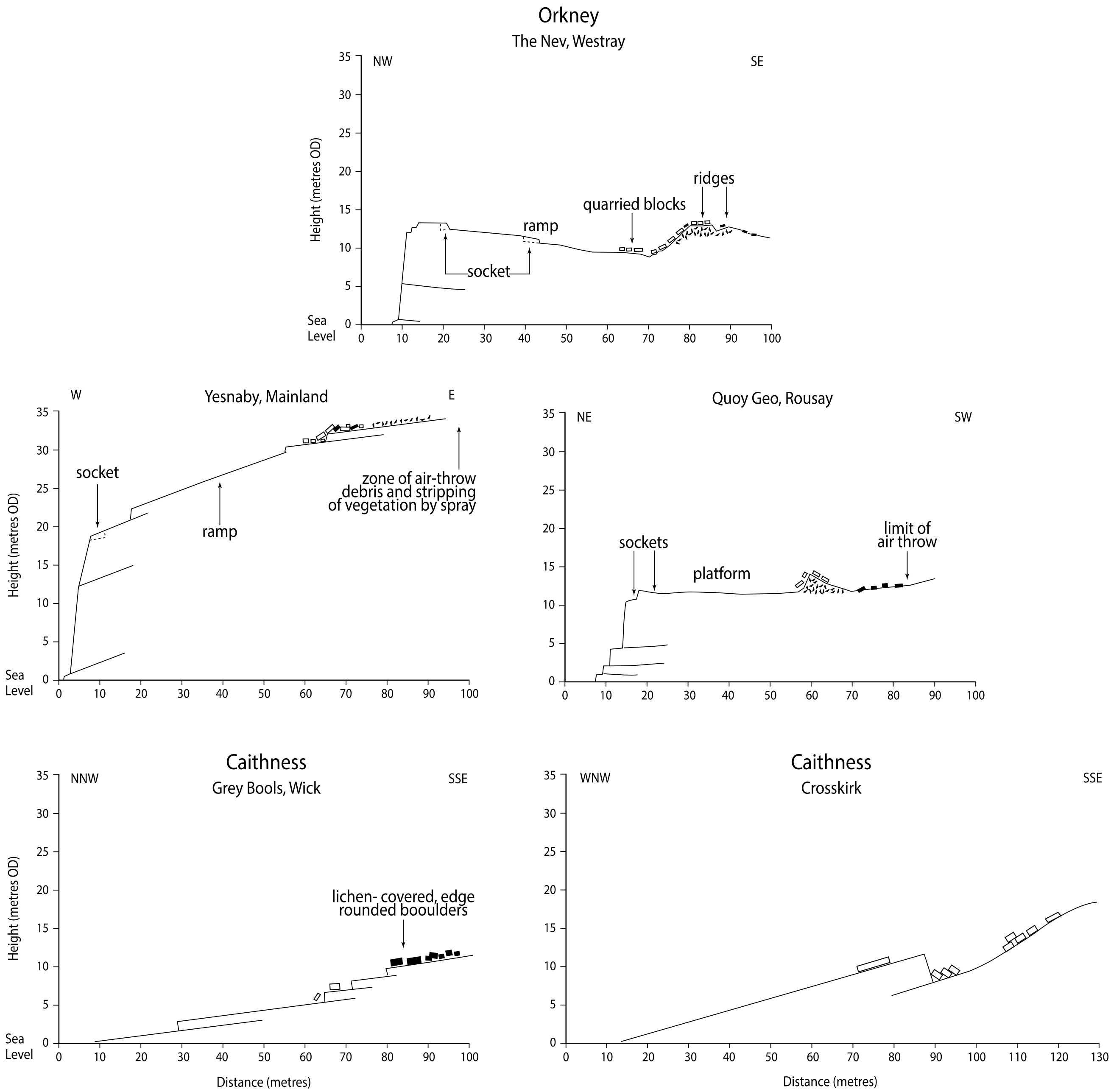

Fig. 8.10 Cliff-top storm deposits (CTSDs) occur in Orkney and Caithness at sites where deep water lies close inshore to steep cliffs, allowing unattenuated storm waves to overtop, quarrying large blocks from the cliff-top edge and transporting them tens of metres inland, where they accumulate in ridges or spreads of imbricated boulders. (Redrawn after Hall et al. 2006)

flooding low-lying areas, it also delivered large amounts of onshore-moving sediment, resulting in island-tying (Rennie 2006). The combined minerogenic and organic sediment supply has resulted in an outstanding assemblage of coastal depositional features on Sanday, including shell-rich tombolos, spits, sandflats and dunes. Island-tying is also represented by the spectacular 2 km long, gravel-cored sandy tombolo connecting Sanday to the former offshore island of Tres Ness and enclosing the intertidal Cata Sand and the Plain of Fidge sand dunes (Fig. 8.11), and the former islands of Els Ness, Tofts Ness and Lop Ness have similarly been subject to island-tying and now form promontories. However, where sediment supply is restricted, islands continue to be isolated by sea-level rise. For example, rocky Start Point in Sanday was mapped as a headland in 1822 (Fig. 8.12), but is now an island at all but the lowest of tides (Rennie 2006).

Fig. 8.11 Tres Ness tombolo now encloses the former marine inlet of Cata Sand, Sanday, and was formed by northward extension of gravel bars that progressively joined Tress Ness to Sanday. (Image: Google Earth™)

Although several small beaches and dunes occur along the crenulated north coast of Caithness, substantial beach development is limited to the two large embayments of Torrisdale Bay and Dunnet Bay. At Torrisdale Bay, well-developed dunes have climbed to 110 m OD over a glacially moulded rock ridge that lies landwards of a wide intertidal sand beach. At Dunnet Bay, a wide sand beach is backed by a massive, sharp-crested, coastal dune ridge with a gently sloping dune plain on its landward side. The 3 km wide bay is a sediment trap with almost complete refraction of the unidirectional incoming waves when they arrive at the beach. Only 20% of the sediment at Dunnet Bay is biogenic, so most is sourced from onshore and offshore glacigenic material (Hansom 2003c). Between 1968 and 1998, beach erosion and frontal dune recession resulted in sand loss at an average rate of 20 m a^{-1}, mainly within the steep-sided erosion corridors that punctuate the high dune ridge. However, phases of beach and dune erosion followed by accretion are not uncommon; by 2013 new foredune development had allowed partial stabilisation of the frontal dune ridge (Fig. 8.13).

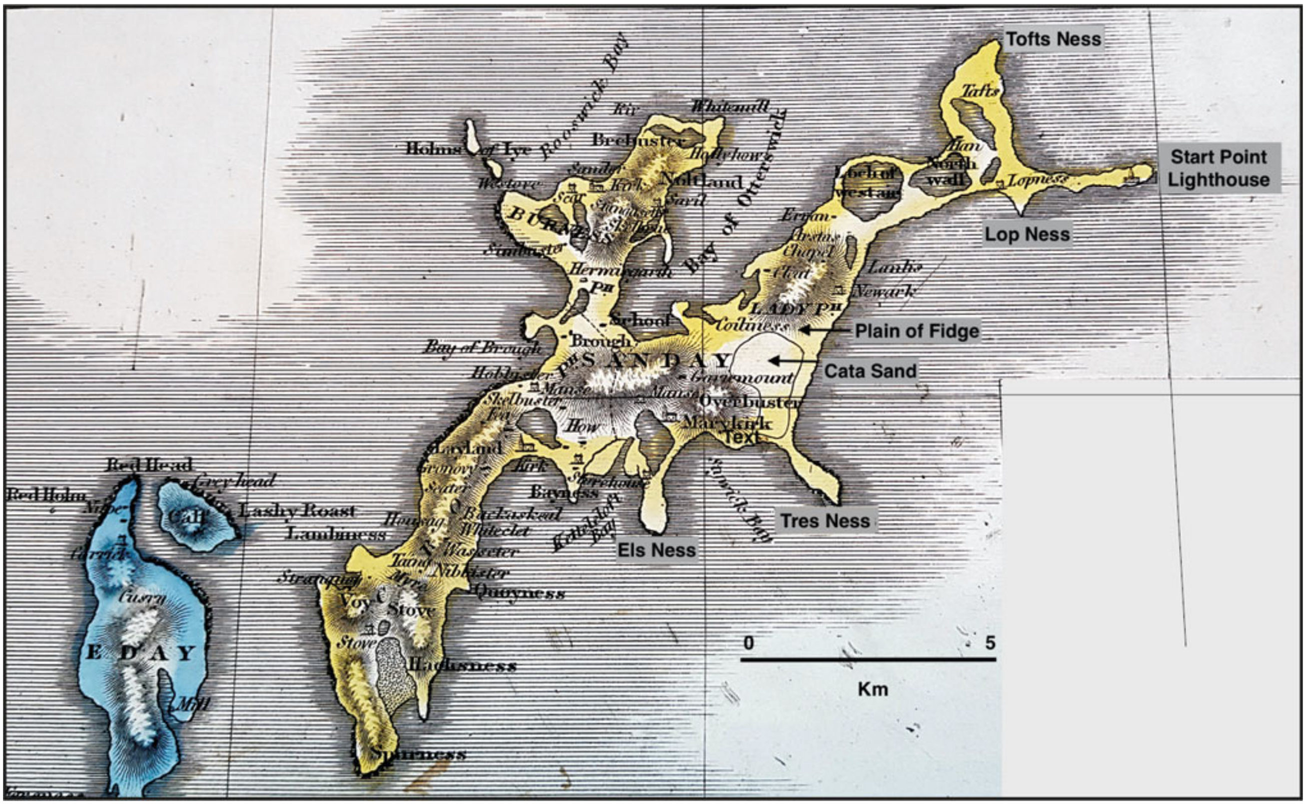

Fig. 8.12 Extract of Sanday from John Thomson's map of 1822. It depicts the intertidal area of Cata Sand as low-lying land behind the Tres Ness island-linking tombolo. The now-flooded low-lying areas behind the former islands of Els Ness, Tofts Ness and Lop Ness have also been enclosed by island-linking sand and gravel barriers. The present island of Start Point with its lighthouse is depicted as a headland

8.7 Conclusions

The geomorphology of Orkney and Caithness is dominated by the varied lithology and structures of the Devonian sedimentary sequence. This is reflected in the generally subdued landscapes truncated at the coast in spectacular cliffs. The low relief of central and northeast Caithness is emphasised by the presence of an uplifted and dislocated Pliocene planation surface at 80–210 m altitude. The excavation of inter-island firths and the overdeepening of the inner Moray Firth are products of effective ice-sheet erosion. Pleistocene glacial modifications on land are relatively subtle, but the interactions between ice masses from different sources produced complex patterns of striae, till deposition, erratic distribution and moraines both onshore and offshore; these are key to understanding the timing and flow patterns of the last Scottish Ice Sheet. The area contains outstanding rock coast landforms, including some of the most spectacular sea stacks in Britain, and cliff-top storm deposits, the latter produced under the severe wave-energy environment, particularly on the Atlantic coasts. In sheltered, low

Fig. 8.13 The beach and dunes at Dunnet Bay in 2013. Following a period of erosion that undercut steep frontal dune faces, new foredune formation resulted in partial, probably temporary stability to the dune faces behind. (Image: © Bill Boaden licenced for reuse by CC-BY-SA/2.0)

wave-energy environments in the northern isles of Orkney, where glacigenic and biogenic sediments have been reworked, extensive beaches, tombolos and dune systems add to the geomorphological diversity. During the Holocene, the development of extensive blanket bog has formed the exceptional landscapes of the Flow Country, internationally important today for their nature conservation value.

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9 The Outer Hebrides and St Kilda

Adrian M. Hall, Colin K. Ballantyne, and James D. Hansom

Abstract

The Outer Hebrides Platform extends west from the present island chain towards the Atlantic continental shelf edge and represents a fragment of Archaean crust (Lewisian gneiss) that was differentially uplifted during the Palaeogene and tilted westwards during the Neogene. An extensive planation surface developed close to sea level in the Pliocene and was subsequently modified by Pleistocene marine and glacial erosion to form an extensive, partly submerged strandflat. During Pleistocene cold stages the Outer Hebrides supported an independent ice cap, but mainland ice periodically over-ran the extremities of the island chain and flowed through the sounds separating individual islands. Ice-roughened knock-and-lochan terrain is extensive across lowland areas, but during the last (and probably earlier) glacial stage(s) the Outer Hebrides Ice Cap remained cold-based and non-erosive above 450–700 m and on lower ground in NW Lewis, permitting the preservation of summit blockfields and over-ridden raised beach gravels. During the last glacial stage (~35 to 14 ka) the ice cap fed the Minch and Hebrides Ice Streams and extended westwards across the shelf but did not reach St Kilda, which supported only small cirque glaciers. Readvance of glaciers during the Loch Lomond Stade (~12.9 to 11.7 ka) was limited to the mountains of North Harris and SW Lewis, and resulted in the deposition of end, lateral and multiple recessional moraines. Postglacial sea-level rise drowned the structurally guided and glacially modified coastline, creating numerous rocky inlets; on Atlantic coasts, large volumes of sediment were moved shoreward onto beach, dune and machair (calcareous shell sand) land systems. The beach, dune and machair features of the Outer Hebrides and the exceptionally high sea cliffs and stacks of St Kilda represent some of the most iconic coastal scenery in Scotland.

Keywords

Outer Hebrides platform • Planation surface • Strandflat • Outer Hebrides Ice Cap • Knock-and-lochan terrain • Loch Lomond Readvance • Moraines • Beaches • Dunes and machair

9.1 Introduction

The Outer Hebrides form a 210 km long chain of islands (the 'Long Island') off the northwest coast of mainland Scotland (Fig. 9.1) and represent some of the most remote and beautiful landscapes in the British Isles. The scenery is unmistakably Hebridean in character, with craggy mountains, hills and ridges rising from low, rough, rock platforms that are partly buried beneath blanket bog in the interior and by the sandy plains of the coastal machair. Water is visible everywhere in inland lochs and lochans, and sea lochs extend deep into the islands.

The islands represent the narrow, emergent part of the much larger Outer Hebrides Platform (OHP), a mainly submerged basement shelf that extends westwards beyond St Kilda to the edge of the North Atlantic continental shelf (Fig. 9.2a). The OHP is a buoyant fragment of Archaean crust, with some of the oldest rocks in Europe. At the onset of the opening of the North Atlantic in the Mesozoic, the

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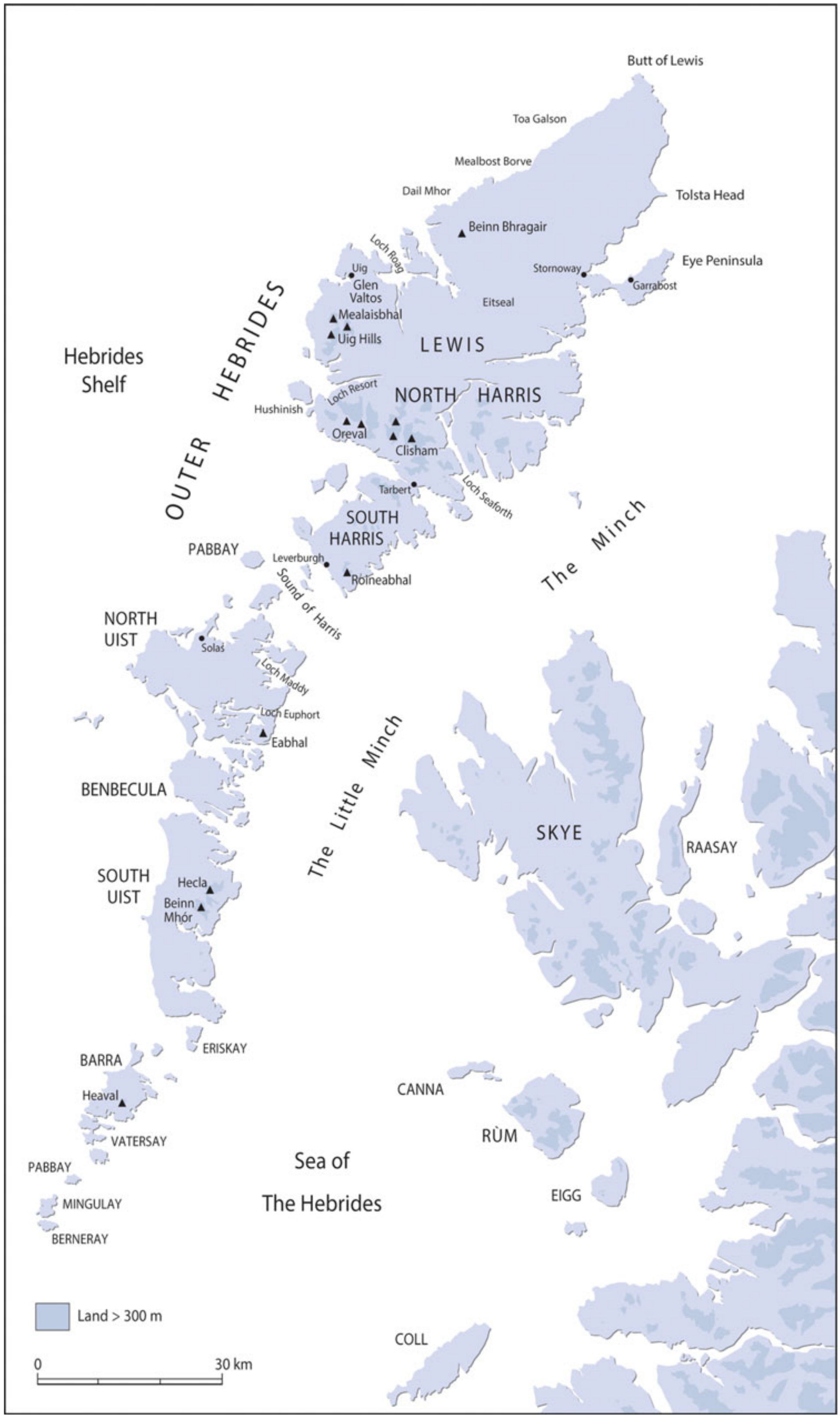
Fig. 9.1 The Outer Hebrides, showing key locations mentioned in the text



OHP formed the westernmost part of the Scottish land area. The OHP was tilted westwards in the Neogene, leading to its burial along the continental edge (Stoker et al. 2010) and has been shaped by subaerial and coastal weathering and erosion over the past 50 Ma.

9.1.1 Geological Foundations

Lewisian gneisses underlie most of the Outer Hebrides and are rich in biotite and hornblende, but felsic in composition, with well-defined foliation (Fettes et al. 1992). Basic and ultrabasic inclusions are common. Folded within the gneiss are metasediments and metavolcanic rocks exposed in the Langavat and Leverburgh Belts, which flank the gabbro, anorthosite, norite and diorite of the South Harris Igneous Complex. The gneisses were subject to several phases of granulite and amphibolite facies metamorphism and dyke injection from the Late Archaean into the Late Palaeoproterozoic. The OHP is bounded to the east by the Outer Hebrides Fault Zone (OHFZ) , which extends from northern Lewis to Barra and forms the eastern seaboard of North and South Uist and parts of South Harris. Initiation of the OHFZ on Lewis dates from ~1700 Ma or earlier, but thrusting recurred during the final part of the Caledonian Orogeny (425–395 Ma).

The separation of Greenland, Fennoscandia and northern Scotland was underway by the Late Carboniferous. During the Permo-Triassic (~250 Ma), extensional basins developed on and around the OHP, including the Flannan trough, the Stornoway basin and the Minch basin. Each basin was filled by red-bed sandstones and conglomerates derived from erosion of upstanding basement blocks under semi-arid conditions. Sediment input to basins declined through the Mesozoic, indicating reduction of relief on the OHP. Acceleration of sea-floor spreading in the nascent North Atlantic from the Late Cretaceous onwards and the passage of the Iceland Plume at 62–58 Ma led to the outburst of magmatic activity that formed the Hebridean Igneous Province (HIP), including the St Kilda and Skye igneous centres (Chap. 10). Buried Cenozoic unconformities on the continental shelf likely record major episodes of uplift and accelerated erosion across NW Scotland during the Palaeocene (~62 to 55 Ma), late Eocene (~35 Ma), late Oligocene (~24 Ma) and early Pliocene (~4 Ma; Stoker et al. 2010). Rocks of the HIP on the Outer Hebrides are now restricted to the Palaeocene dolerite dyke swarms of Barra, south Harris and south Lewis (Fig. 9.2a), a mid-Eocene vent at Loch Roag in western Lewis (Faithfull et al. 2012) and flood basalts that occur in proximity to the Minch Fault and off the coast of NW Lewis (Fettes et al. 1992; Fig. 9.2b).

9.1.2 Cenozoic Landscape Evolution

The large-scale topography on the OHP reflects the interplay through time between tectonics, rocks and erosion. The oldest landscape traces are provided by the hilly sub-Permo-Triassic basement unconformity on the Eye Peninsula (Jehu and Craig 1934; a contemporaneous mountainous terrain to the northwest shed gravel and sand to the Stornoway half-graben (Steel and Wilson 1975) but this upland topography was later erased by erosion. In the Late Cretaceous, chalk deposition across The Minch and the Sea of the Hebrides basins indicates low clastic sediment input and low relief along the spine of the Outer Hebrides. The St Kilda igneous centre (Sect. 9.8) represents a Palaeogene volcanic core, from which all lavas and pyroclastic rocks have been removed by erosion. Igneous activity on St Kilda dates from ~60 Ma; the oldest rocks are represented by gabbro and dolerite, with younger granites, felsites and sets of sheets and dykes (Harding et al. 1984; Ritchie and Hitchen 1996; Sect. 9.8.1). Proximity to spreading sites in the North Atlantic, however, appears to have extended the duration of minor Palaeogene magmatic activity on the OHP until ~43 Ma (Jones et al. 1986; Faithfull et al. 2012).

9.2 Tectonics, Erosion and Landscapes

HIP magmatism was accompanied by kilometre-scale regional uplift in Hebridean Scotland, with major differential tectonic movements around the igneous centres of Skye and Mull (Chap. 3). Thin Palaeogene lavas may have extended locally onto Lewis (Watson 1977). Low-temperature thermochronology suggests the loss of 2–3 km of rock from above the present surface of the Clisham block in North Harris since 60 Ma (Persano et al. 2007; Amin 2020). However, a suite of undated, but late stage, vesicular tholeiite dykes is present down to sea level in south Harris and Lewis (Figs. 9.2b and 9.3a). Large amygdales filled with glass and zeolites, including chabazite, indicate emplacement at maximum depths of 200–800 m (Fettes et al. 1992). As present summit elevations on south Harris reach 459 m on Roineabhal, the land surface at the time of dyke emplacement stood, at most, a few hundred metres above that of the present (Fig. 9.3b). Thermochronology indicates that a large part of this erosion occurred during the Palaeogene (Amin 2020).

A striking feature of the islands (Fig. 9.3c) and the seabed to the west (Fig. 9.2a, b) is the extensive low-relief surface that extends to beyond St Kilda. The gneiss surface is covered by Palaeocene flood basalts on the seabed off western Lewis and by Miocene sediments towards the edge of the continental shelf. Kaolinitic weathering of Oligo-Miocene

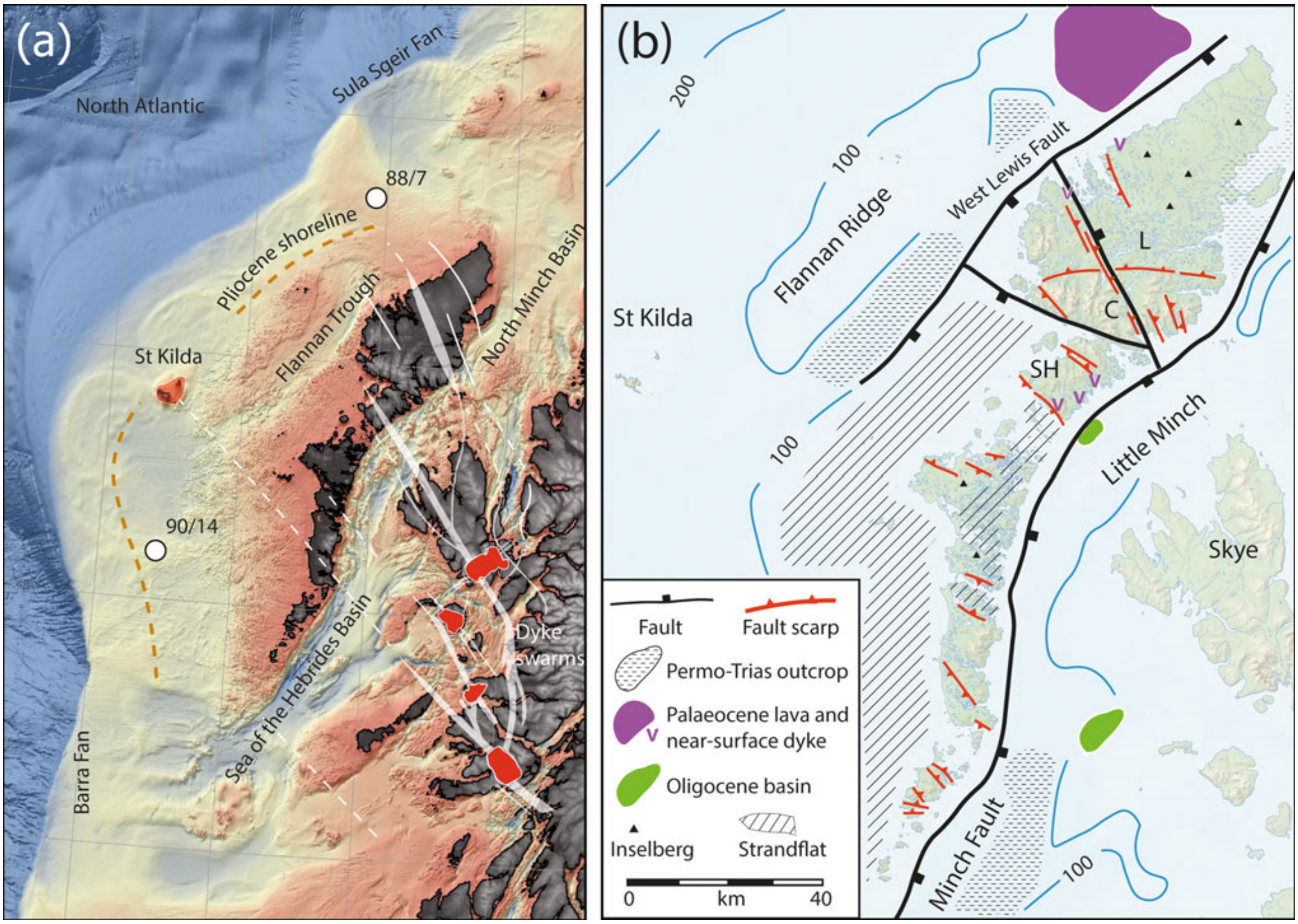


Fig. 9.2 **a** Outer Hebrides Platform (OHP) extends towards the edge of the continental shelf on the North Atlantic passive margin and is covered by Neogene and Pleistocene sediment in its western part. Major igneous centres of the Hebridean Igneous Province are shown in red, with associated dyke swarms in white. Bathymetric data from EMODnet Bathymetry Consortium (2018): EMODnet Digital Bathymetry (DTM). **b** Main elements of the Cenozoic geomorphology of the OHP. L: Lewis, C: Clisham and SH: South Harris rock blocks. (Relief map licence Eric Gaba, NordNordWest, Uwe Dedering/CC BY-SA 3.0)

age has been recorded in a core from the feather edge of the Miocene cover (Stoker et al. 1993). Kaolin- and lignite-bearing Late Oligocene sediments in the Little Minch (Smythe and Kenolty 1975) represent reworking of weathering covers of similar type and age. No comparable kaolinitic weathering has been reported for the present land area. The outer part of the continental shelf was drowned in the Miocene as the OHP was tilted towards the west (Stoker et al. 1993), and uplift of the Outer Isles likely led to erosion and removal of mature weathering covers. Younger, generally shallow, grus-type weathering developed subsequently (Fig. 9.3d), dominated by smectite and illite clays (Godard 1965).

The OHP is crossed by major fault, shear and fracture zones with long histories of movement and mineralisation (Fettes et al. 1992). Fault zones initiated at the onset of brittle deformation in the Archaean were reactivated during the Mesozoic and were later intruded by dykes in the Palaeocene. Palaeocene to Eocene dykes on Lewis (Franklin 2013) and the Main Sill on the Shiant Isles in The Minch (Gibb and Henderson 1996) have been displaced by minor faulting. On Lewis and Harris, major fracture zones separate three main rock blocks: the Lewis, Clisham and South Harris blocks. The edges of the main fault blocks are bounded by scarps (Fig. 9.2b). As no strong lithological contrasts occur in the basement gneisses across the faults, the scarps have been interpreted as Tertiary fault scarps (Godard 1965). The contrast between the high elevation and the deep erosion on the Clisham block (Fig. 9.2) and the low elevation and presence of a vent and vesicular dykes on north Lewis (Fig. 9.3a) indicates major differential movements across the Loch Seaforth Fault during the Palaeogene (Fig. 9.3c). The glacially roughened erosion surface that occurs close to sea level across Benbecula and the Uists reappears northwards across the Sound of Harris at 50–110 m elevation and then continues as a much wider, low platform that extends across

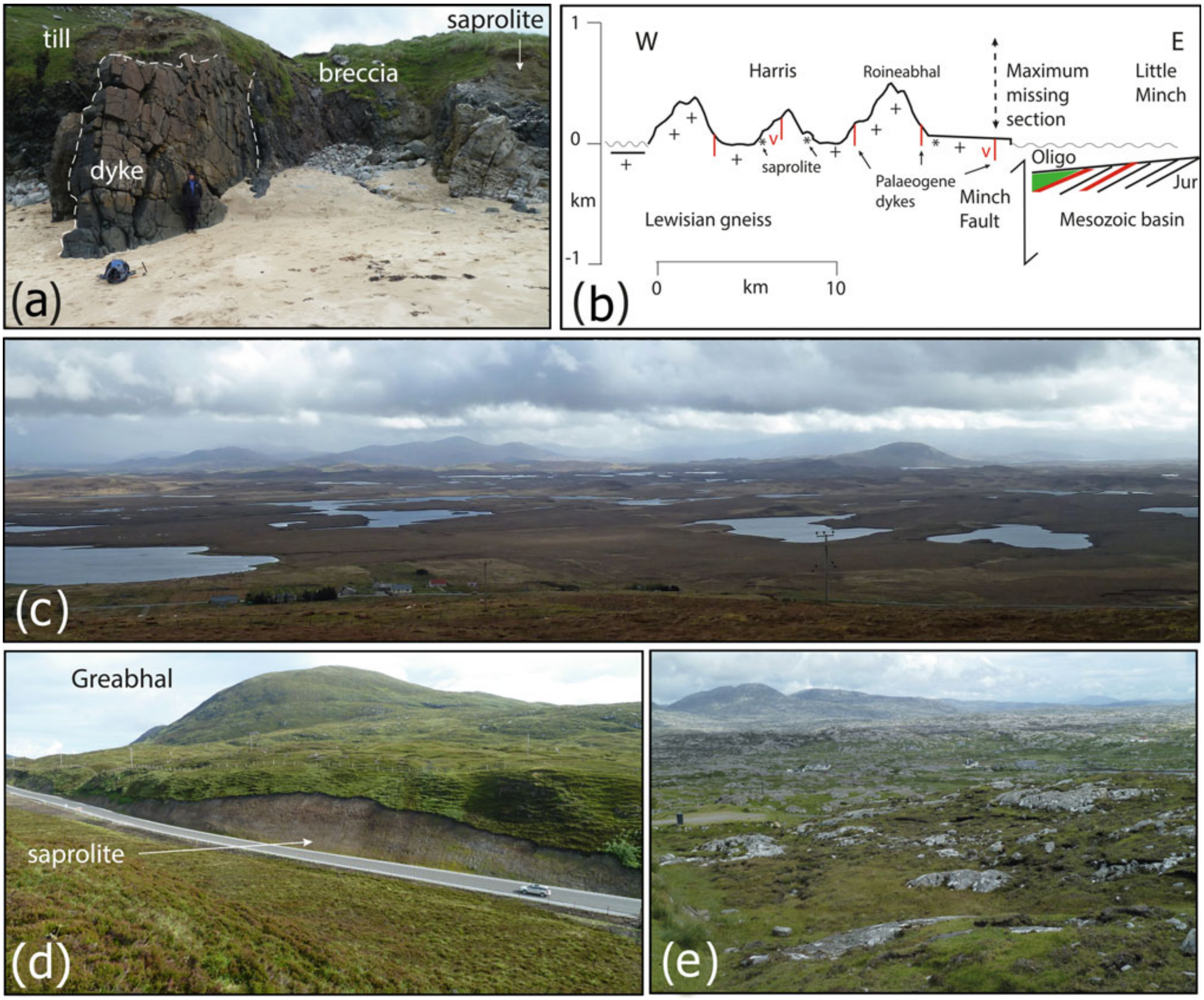


Fig. 9.3 **a** Section in a major NNW–SSE shear zone at Dàil Mòr, Lewis. Fault breccias were intruded by a Palaeocene vesicular dyke, with later fault reactivation (Franklin 2013). The rocks are extensively broken, altered and weathered. In common with other shear zones in the Outer Hebrides, the highly fragmented rocks have been exploited by glacial and marine erosion to form a bay and a backing, steep-sided valley. **b** Schematic W–E section across south Harris to the Little Minch, showing the position of vesicular dykes and the estimated depth of missing section since dyke emplacement. **c** View south from the inselberg, Eitsal (233 m), across the hilly Pliocene surface at ~100 m and the Loch Seaforth and Resort Faults towards the uplifted fault blocks of south Lewis and Harris. **d** Section at Gleann Choiseleitir, south Harris. Sheared metasediments are weathered at depths of >10 m to a granular, sandy saprolite. The NW–SE oriented shear zone extends as a valley to Rodel. **e** The low-elevation surface of east Harris at 50–110 m looking north from Finsbay. The surface is developed in gneissic ultrabasic and acidic intrusions of the Lewisian basement and was shaped in the Pliocene but lowered and roughened by glacial erosion beneath the Outer Hebrides Ice Cap, with excavation of fracture zones. (Images: Adrian Hall)

much of Lewis at altitudes of 80–140 m (Godard 1965). The continuity of the erosion surface, and its grus weathering covers, indicate a geologically recent, likely Pliocene, age. Its dislocation and abrupt termination in sea cliffs in Lewis indicate later, minor uplift, perhaps at ~ 4 Ma. Geological and geomorphological evidence therefore provides support for Neogene fault movements on the OHP, as in the Sea of the Hebrides (Le Coeur 1988).

9.3 Rocks and Relief

The Lewisian gneisses are very hard and resistant to physical weathering and mechanical erosion (Fig. 9.3e). The gneisses are, however, locally closely fractured and frequently biotite- and hornblende-bearing, with only moderate resistance to chemical weathering. Variability in structure and mineralogy has been exploited by differential weathering and erosion to form regional-scale (1–10 km) landforms. Fracture zones are particularly susceptible to erosion due to their high permeability, large surface area available for chemical and physical weathering and small rock block sizes. Fracture zones often control the position and alignment of major valleys, inlets and sounds on the OHP. Major shear zones in metasediments form trenches floored by broken, altered and weathered rocks, as near Leverburgh in south Harris (Fig. 9.3d), and along Loch Euphort in North Uist (Godard 1965). Many of the sounds between the islands are associated with faults, shear zones and fractures. Crossing fracture zones form depressions and basins, as on south Lewis around Lochs Langavat and Resort (Fig. 9.4a).

Gneiss surfaces exposed on hill summits tend to lack zones of intense fracture. Examples include the low inselbergs, such as Beinn Bhragair, that rise above the plateau of NW Lewis, and Heaval and other hills on Barra. On south Harris, each main hill mass is bounded by north–south or NW–SE oriented fracture zones, producing ridges and elongate dome forms. The structural influence of the ESE-inclined, low-angle thrust sheets of the Outer Isles Thrust is evident on Eabhal (347 m) in North Uist, and is also expressed by similar asymmetries of hills in eastern Lewis, Harris and the Uists.

9.4 Quaternary Environments

North Atlantic Ocean temperature records show a long-term cooling trend over the last 5 Ma, with two major steps during the late Pliocene (3.1–2.4 Ma) and at the mid-Pleistocene transition (MPT; 1.25–0.7 Ma; Lawrence et al. 2010). During and after the MPT, North Atlantic sea level fell as low as −130 m during glacial maxima (Chap. 4). During successive Pleistocene glaciations, an independent ice cap, the Outer Hebrides Ice Cap (OHIC), developed on the islands and extended westwards towards the Hebrides

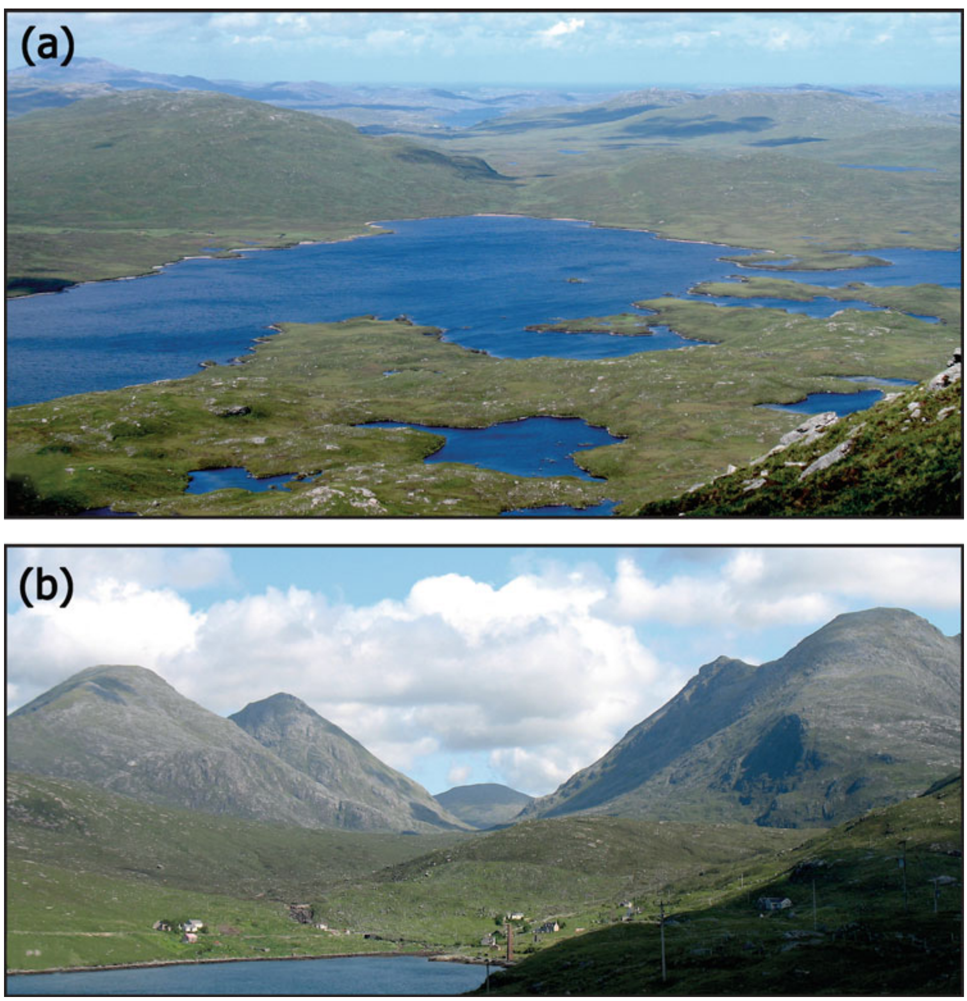


Fig. 9.4 Glaciated landscapes of the Outer Hebrides. **a** Loch Langavat in south Lewis occupies a basin excavated along crossing fracture zones; partly peat-covered knock-and-lochan terrain occurs in the foreground, and the hills of Lewis form the background. **b** The Harris mountains, looking north: the Eader–Langavat trough separates Uisgnaval Mór (729 m) and Teilesval (697 m) on the left from Mulla-fo-dheas (743 m) on the right. (Images: Colin Ballantyne)

shelf edge. Along the east coast of the islands, the OHIC fed ice streams in The Minch and the Sea of the Hebrides (Bradwell et al. 2007; Dove et al. 2015). During interglacial stages, relative sea level rose by up to ~10 m above Ordnance Datum, as indicated by raised beach gravels on Lewis and Vatersay (Peacock 1984); glacier ice was absent (Walker 1984b). Environmental conditions intermediate between full glacial and interglacial modes were, however, cumulatively of the longest duration over the past 800 ka, with eustatic sea level oscillating between –30 and –70 m (Spratt and Lisiecki 2016). Parts of the OHP between these elevations were exposed to multiple marine transgressions and regressions throughout the Pleistocene.

The earliest glaciations of the OHP are recorded in core 88/7 (Fig. 9.2a), which includes ice-rafted debris derived from glaciers in NW Scotland shortly after the Gauss-Matuyama palaeomagnetic reversal at ~2.6 Ma (Stoker et al. 1994). The first glaciation to reach the shelf edge was not until the Middle Pleistocene; a widespread erosional unconformity, overlain by till, is attributed to Marine Isotope Stage (MIS) 12 at ~474 ka, and further expansions of ice sheets onto the shelf occurred during MIS 10, 8 and 6 (Sejrup et al. 2005). During the last glacial cycle, the OHIC expanded onto the shelf in MIS 4 and MIS 2 (Hibbert et al. 2010). Ice-sheet models for the last glaciation suggest that short-lived ice streams developed within the OHIC and flowed through channels between the islands, including the Sound of Harris (Hubbard et al. 2009). Along its eastern margin, the OHIC was confluent with mainland ice streams that fed the shelf-edge Sula Sgeir and Barra Fans (Bradwell et al. 2007; Dove et al. 2015).

9.4.1 Glacial Erosion During the Pleistocene

The pattern of glacial erosion on the OHP provides a window on the thermal regime of the OHIC through the Middle and Late Pleistocene. The preservation of weathered rock and MIS 6/5 sediments at low elevations in NW Lewis and blockfields capping the highest summits indicate areas where the OHIC remained frozen to its bed and non-erosive during the last and probably earlier glacial cycles. Elsewhere, the last ice cap was warm-based and erosive during much or all of the last glacial cycle.

During the Pliocene, the shoreline on the OHP lay west of its present position (Stoker et al. 1993; Fig. 9.2a). Sedimentary evidence is lacking for the existence of the marine basins of The Minch and the Sea of the Hebrides at that time. Much of the seabed today is below –100 m and overdeepened trenches and depressions extend to greater depths, notably along the Minch Fault (Chesher et al. 1983). The Late Pliocene and Pleistocene sediments in the Sula Sgeir and Barra-Donegal Fans, the shelf-edge fans fed by the Minch and Hebrides Ice Streams, respectively, are up to 600–800 m thick. The volume of sediment in the Sula Sgeir Fan (Stoker et al. 1993) exceeds the volume of the present Minch marine basin (Chesher et al. 1983), which itself represents about half of the main source area for the Minch Ice Stream (Bradwell et al. 2007). The marine basins east of the Outer Hebrides are therefore mainly the end-products of Pleistocene glacial erosion, formed through deep excavation of relatively weak Permo-Triassic and Jurassic sedimentary rocks. Separation of the individual Outer Hebridean islands is likely also a product of glacial erosion by ice from the OHIC and possibly also mainland ice flowing westward through low corridors across the original 'Long Island'.

Lowland areas of the Outer Hebrides are dominated by extensive tracts of ice-scoured gneiss, with classic knock-and-lochan topography where resilient rock kernels form hummocks and hills separated by shallow trenches and depressions excavated along fracture zones (Figs. 9.3e and 9.4a). Geikie (1877, p. 78) captured the essence of such terrain, noting that much of Lewis had '…been swept and ground by a great glacier: round–backed rocks, hummocks and hills bare of drift and soil, with countless pools and lakelets nestling in their hollows, impart to the region a character of great desolation'. Typical examples of such ice-roughened lowland landscapes occur in west Lewis around Loch Roag, east Lewis south of Loch Erisort and across much of south Harris (Fig. 9.3e); elsewhere the bedrock is largely obscured by peat cover. Smoother topography that is lacking in well-developed glacial bedforms occurs along the NW coast of Lewis, where raised beach gravels were preserved beneath cold-based ice during the last (Late Devensian) ice-sheet glaciation (Hall 1995). Where transitions from smooth to rougher terrain occur at similar elevations, glacial erosion appears to have removed only a few tens of metres of rock, an inference supported by the preservation of weathered and highly fractured rock in lee-side locations within ice-roughened terrain (Godard 1965).

Rock has been removed mainly from fracture-guided trenches and basins. Bedrock samples from whalebacks at low and intermediate elevations on south Lewis and in north Harris include sites where cosmogenic ^{10}Be apparent exposure ages exceed local deglaciation ages (Stone and Ballantyne 2006); the presence of inherited cosmogenic nuclides indicates that <2 to 3 m of rock was eroded from resistant bedrock highs beneath the last OHIC. The evolution of knock-and-lochan terrain has been attributed to prolonged pre-glacial chemical weathering focused along fracture zones, followed by stripping of saprolite cover by glacial erosion to reveal an irregular etch surface (Godard 1965; Krabbendam and Bradwell 2014). Glacial modification of rock surfaces by abrasion and plucking has shaped

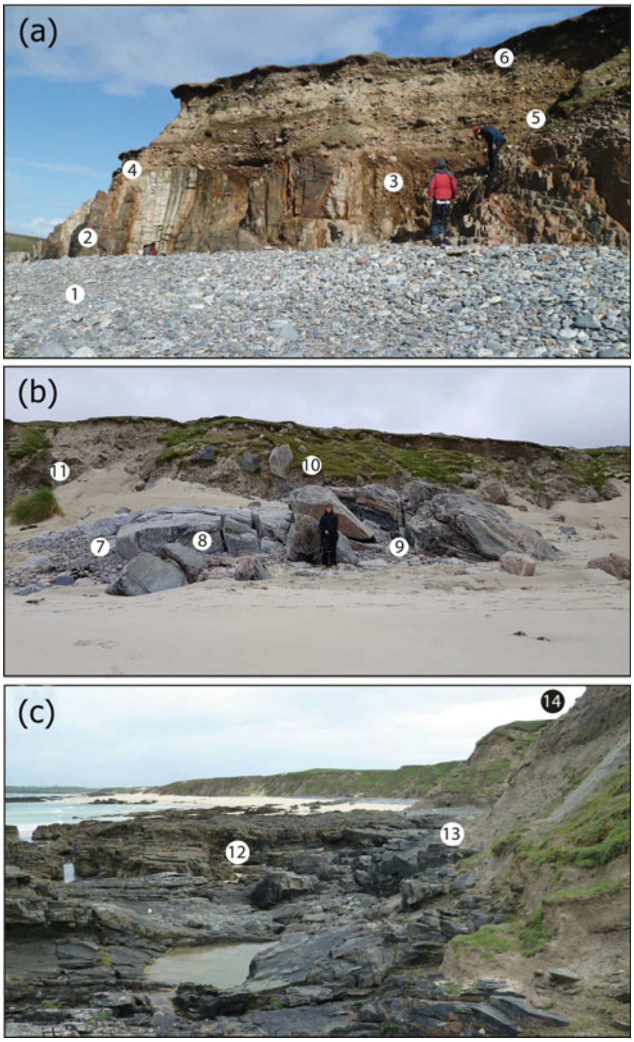


Fig. 9.5 **a** Cliff-section at Melbost Borve in the zone of limited glacial erosion in NW Lewis. 1: Storm beach. 2: Cliff in felsic and basic gneisses with closely spaced vertical fractures. 3: Shallow weathering. 4: Emerged shore platform. 5: Raised beach gravel. 6: Thin layer of Late Devensian till emplaced without significant erosion of the underlying sediment, landforms and bedrock. **b** Roche moutonnée and till cover near Hushinish, Harris. 7: Abraded upper surface. 8: Dilated fractures. 9: Socket from which blocks removed. 10: Angular boulders in overlying sandy till. 11: Gravel- and sand-rich till. The features indicate the operation of glacial ripping at a late stage of the last glaciation by ice moving from right to left. **c** Coastal section at Swainbost, NW Lewis, where a lobe of the Minch Ice Stream ponded meltwater against the cliff line during deglaciation. 12: Emerged rock platform. 13: Glacifluvial sands. 14: Sand- and mud-rich diamicts. The soft sediments are glacitectonically deformed. (Images: Adrian Hall)

whalebacks and roches moutonnées (Fig. 9.3e) but locally these features have been disrupted by glacial ripping (Hall et al. 2020; Fig. 9.5b).

The high ground of north Harris and the Uig Hills of SW Lewis provides evidence of phases of both ice-sheet and mountain glaciation. A trimline representing the upper limit of erosion by the last ice sheet reaches an elevation of 700 m and descends both westwards and eastwards from the highest ground (Ballantyne and McCarroll 1995). The alignments of roches moutonnées and striae at lower elevations also suggest that the mountains were a centre of radial ice dispersal during the local Last Glacial Maximum (LLGM). The main mountain mass of north Harris has four north–south trending ridges separated by deep troughs and culminates in the peaks of Tirga Mòr (679 m), Oreval (662 m), Uisgnaval Mòr (729 m) and Clisham (799 m). All three intervening troughs are through valleys that breach the west–east watershed (Fig. 9.4b) and probably represent trough-head retreat and progressive lowering of the drainage divide during successive periods of radial ice flow. Deep cirques are scalloped into the flanks of the north Harris troughs and along the outer margins of the mountains; these hollows were excavated by mountain glaciers during multiple Pleistocene cold stages. Cirque floors typically lie only 100–250 m above present sea level and are amongst the lowest in Scotland. The South Uist hills, which culminate in Beinn Mhòr (621 m) and Hecla (606 m), constitute a miniature mountain glacial landscape, with two deep troughs draining the massif to the east, and two spectacular cirques on the northern slopes of Hecla.

9.4.2 Other Landforms and Sediments that Pre-date the Last Ice Cap

In addition to landforms of glacial erosion, several other features pre-date the last glaciation. These include high-elevation blockfields and tors (Sect. 9.6), and coastal cliffs, geos and platforms formed when relative sea level was higher or lower than at present (Sect. 9.7).

One or more emerged rock platforms, locally with weathered surfaces, can be traced along much of the coastline of NW Lewis (von Weymarn 1979). The rock platforms are overlain locally by sediments that pre-date till of the last glaciation. A buried, weathered till bed that rests on a platform at Mealbost Borve contains Torridonian erratics, which are also found in the overlying beach gravels (Smith et al. 2019; Fig. 9.5a). At Sgarbh Sgeir (Toa Galson), the weathered platform is overlain by organic mud, with pollen indicative of open, floristically diverse grassland, and the Galson beach gravels (Sutherland and Walker 1984). The beach gravel is cryoturbated in places, and locally overlain by a single till unit deposited by the OHIC, or, farther north, by a complex sedimentary sequence deposited by ice moving out of The Minch (Hall 1995). The sedimentary record indicates that the Atlantic coast of NW Lewis experienced limited glacial erosion beneath at least two phases of OHIC glaciation. Beach gravels also occur below tills on Barra and Vatersay (Selby 1987). The fragments of emerged shore platforms and beach gravels on the Outer Hebrides represent an important record of sea-level change but lack firm dating control.

9.5 The Outer Hebrides Ice Cap During the Last Glaciation

9.5.1 The Late Devensian Ice Sheet

Terrestrial evidence relating to the timing of initial ice expansion during the Late Devensian is limited to sites in northern Lewis. On sea cliffs at Tolsta Head in NE Lewis, thin organic-rich sands and silts below till have yielded ages of ~33.5 to 32.5 cal ^{14}C ka (Whittington and Hall 2002), consistent with the ages of ~43.4 to ~38.5 cal ^{14}C ka for ice-transported marine shells recovered from till in northernmost Lewis (Sutherland and Walker 1984).

Trimlines on the mountains of the Outer Hebrides were initially interpreted as representing the upper limit of the last ice sheet as it advanced westward from the mainland, or the maximum surface altitude of the Outer Hebrides Ice Cap (Ballantyne and McCarroll 1995; Ballantyne and Hallam 2001). However, terrestrial cosmogenic nuclide (TCN) exposure dating of erratic boulders on blockfields above similar trimlines in the NW Highlands has demonstrated that these were emplaced by the last ice sheet (Fabel et al. 2012). Moreover, perched glacially deposited boulders occur on blockfields and rock outcrops on several summits in North Harris, indicating that these were also over-run by glacier ice during the LLGM. The trimlines on the Outer Hebrides are now considered to represent the upper limit of erosion by warm-based, sliding ice, implying that the ice at higher altitudes remained cold-based, frozen to the underlying substrate, and capable of only limited erosion of the blockfields that occur above trimlines. Numerical modelling of the dimensions of the last ice sheet suggests that the ice surface elevation may have exceeded 1250 m over the mountains of Harris (Hubbard et al. 2009).

Mapping of erratic dispersal and the alignment of striae and ice-moulded bedrock has established that the island chain was occupied by an independent ice cap that was confluent with the mainland ice sheet and extended westwards across the Hebrides Shelf. Peacock (1984, 1991) concluded that the main centres of ice dispersal for the last OHIC comprised an elongate ice dome centred over the west coast of the Uists, a second dome over the mountains of Harris and SW Lewis, and a possible third dome centred on north-central Lewis. The absence of mainland erratics throughout much of the Outer Hebrides suggests that this ice cap persisted as an independent centre of ice dispersal during much of the lifetime of the last Scottish Ice Sheet. The localized presence of mainland erratics, however, may indicate that parts of the archipelago were episodically over-run by ice from the east during the last glaciation. For example, clasts of Torridonian sandstone, Cambrian quartzite and sedimentary rocks derived from the floors of the Minch and the Sea of the Hebrides basins are found on the south and west coasts of South Uist, and along the Sound of Harris (Merritt et al. 2019). Mainland ice also impinged on the far north and northwest of Lewis and the Eye Peninsula during deglaciation (Peacock 1991). Sandstone and quartzite erratics on Barra and Vatersay, together with NW-aligned striae on Mingulay, suggest that the southern extremity of the islands was also over-run at some time by ice from the east (Peacock 1984).

The westward extent of the OHIC is uncertain but available information favours mid-shelf termination of ice flowing westwards from the Outer Hebrides during the LLGM, implying that shelf-edge moraines farther west are pre-Late Devensian in age (Bradwell and Stoker 2015; Ballantyne et al. 2017). The Minch Ice Stream reached its limit by ~30.2 ka, but by ~18.5 ka open water was present east of Lewis (Bradwell et al. 2019). The Hebrides Ice Stream reached the shelf break at ~26.7 ka, but by ~20.2 ka much of the Sea of The Hebrides was ice-free (Callard et al. 2018), and Barra, in the south of the island chain, was deglaciated at ~17.1 ka (Small et al. 2017). TCN exposure ages reported by Stone and Ballantyne (2006) for bedrock surfaces in mountainous inland areas of Harris and the Uists fall within the range ~17.3 to 16.3 ka, and a single sample from the summit of Oreval (662 m) yielded a TCN exposure age of ~15.6 ka, implying that the ice had retreated to its mountain source areas before rapid warming at ~14.7 ka heralded the onset of the Lateglacial Interstade.

9.5.2 The Loch Lomond Readvance (LLR)

It is likely that glacier ice disappeared from the Outer Hebrides under the cool temperate climate of the Lateglacial Interstade (~14.7 to 12.9 ka). There is, however, abundant geomorphological evidence (end, lateral and recessional moraines, drift limits, trimlines and the alignment of striae, roches moutonnées, fluted moraines and streamlined bedrock) for the development of cirque and valley glaciers in the mountains of North Harris and the Uig Hills of SW Lewis under the renewed cold conditions of the Loch Lomond Stade (~12.9 to 11.7 ka; Ballantyne 2006, 2007). Such evidence indicates that the Uig Hills supported four valley glaciers with a total area of 11.7 km^2, and that North Harris was occupied by valley and cirque glaciers with a total area of ~35 km^2 (Fig. 9.6), though it is possible that some valley glaciers were fed by thin, cold-based ice caps that occupied adjacent high ground. No unequivocal evidence for a readvance occurs on the mountains of South Uist

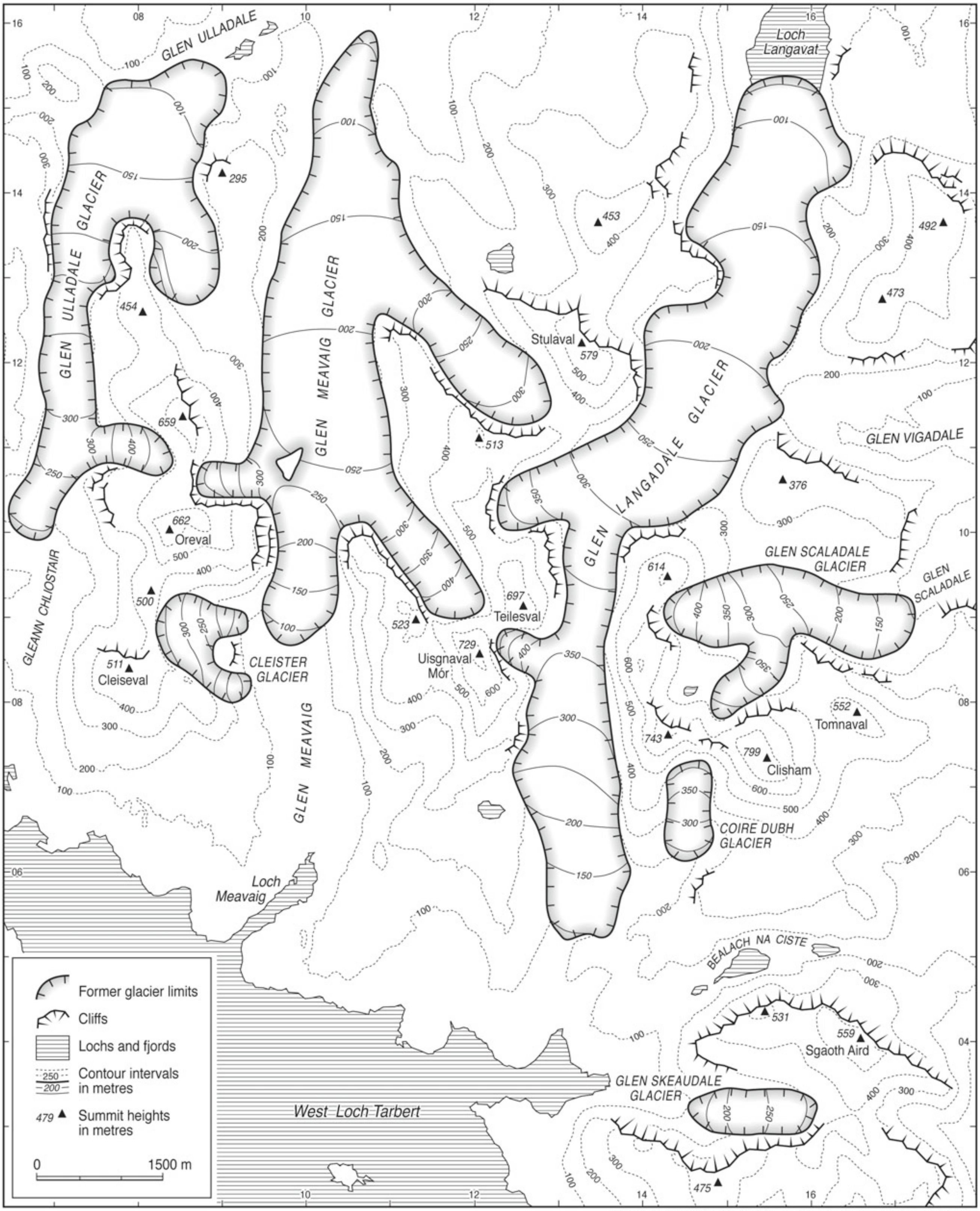


Fig. 9.6 The extent of the main area of the Loch Lomond Readvance in north Harris; outlying cirque glaciers are not shown and it is possible that the extent of ice on cols and high ground was greater than shown here. (From Ballantyne (2007) © 2007 Elsevier Ltd)

(Ballantyne and Hallam 2001). Mean equilibrium-line altitudes (ELAs) calculated for the LLR glaciers on Harris and SW Lewis average 200–215 m and are amongst the lowest in Scotland, reflecting both a northward decline in ELAs along the western seaboard (Ballantyne 2006, 2007) and copious snow inputs along the Atlantic margin during the Loch Lomond Stade (Chandler et al. 2019).

9.5.3 Glacial Deposits and Depositional Landforms

Almost all glacial deposits and depositional landforms hitherto identified on the island chain are of Late Devensian ($\sim$31 to 11.7 ka) age: those outside the LLR limits were deposited by the last OHIC, or by mainland ice at the extremities of the island chain; those inside the LLR limits by the mountain glaciers of the Loch Lomond Stade.

Outside the limits of the LLR, till deposits are patchily distributed throughout most of the Outer Hebrides, with continuous till cover being mainly restricted to peat-covered areas of north Lewis. OHIC tills are mainly of local origin, typically greenish-grey in colour. Some tills are stiff, overconsolidated deposits consisting of subrounded gneiss clasts up to boulder size embedded in a matrix of silt and sand, whereas others are dominated by abundant subangular to angular clasts within a loose sandy matrix (Fig. 9.5b). The mud- and sand-rich reddish tills of the Eye Peninsula and the coastal fringes of northern Lewis are partly derived from Triassic sandstones and often contain shells dredged by ice from The Minch; intercalation with water-lain sediment and glacitectonic disturbance are widespread (Peacock 1984; Fig. 9.5c).

In most areas, the inland till forms a gently undulating till sheet, typically no more than a few metres thick, with restricted areas of weak surface streamlining (Hughes et al. 2010). Moraine ridges have been reported only in northernmost Lewis and on the Eye Peninsula (Peacock 1991), but nested lateral or recessional moraines also occur along the fjord shores of SE Lewis.

Valleys within the LLR limits support a depositional land system dominated by end moraines, lateral moraines and drift limits that enclose suites of multiple, closely spaced recessional moraines that form long continuous nested ridges or chains of hummocks up to $\sim$10 m high (Ballantyne 2006, 2007; Fig. 9.7a–d). Fluted moraines also occur near the sources of some LLR glaciers. Not all hummocks on the floors of the glacial troughs in north Harris are recessional moraines: low, streamlined hummocks near former ice divides appear to represent glacial bedforms that developed in the zone of divergent ice flow (Fig. 9.7e).

The contrast between the thin, patchy till cover outside the LLR limits and the thick hummocky drift on valley floors that were occupied by LLR glaciers is partly attributable to the fact that whereas the last OHIC overtopped all high ground, the LLR glaciers were overlooked by steep rock walls. There is evidence elsewhere in Scotland for enhanced rockfall activity during the Loch Lomond Stade (Hinchliffe et al. 1998), and it is likely rockfall debris falling onto LLR glaciers was reworked into the moraines. The LLR glaciers also advanced across earlier till, outwash deposits and paraglacial landforms (rockslide runout detritus, talus accumulations, debris cones, alluvial fans and floodplain deposits). These all supplied readily entrainable debris to be reworked by glacier ice into the numerous moraines that now occupy valley and cirque floors (Fig. 9.7a–e).

9.5.4 Glacifluvial Landforms

Glacifluvial landforms are poorly represented in the Outer Hebrides, though bedrock channels incised by subglacial meltwater streams are present on low cols between the hills and mountains of Harris, and numerous small meltwater channels cut obliquely across drift-mantled slopes inside the limits of the LLR. Low spreads of coarse outwash deposits occur both outside and a short distance inside the northern limits of the LLR in north Harris (Ballantyne 2007). Eskers, kames, kame terraces and a glacilacustrine delta are present outside the readvance limits around Uig in west Lewis (Peacock 1984). The most impressive glacifluvial feature in the Outer Hebrides is the Glen Valtos meltwater channel, which occupies a 2.5 km long gorge incised up to 50 m into bedrock in west Lewis (Fig. 9.7f). This channel has apparently been cut by subglacial meltwater flow over several successive glacial cycles (Sutherland 1993).

9.6 Periglacial and Postglacial Landforms

Blockfields of bouldery, frost-weathered debris cap mountain summits on Harris and South Uist, and the highest peak (Mealisbhal, 574 m) in the Uig Hills. Most are openwork accumulations of boulders up to a metre long (Fig. 9.8a) but on some summits the boulders are embedded in coarse sand. Protruding through the blockfield debris are occasional outcrops of frost-weathered rock and tors up to 5 m high, particularly along the crests of the South Uist hills (Ballantyne and Hallam 2001). The presence of glacially emplaced gneiss boulders resting on blockfield debris and rock outcrops demonstrates that these features are pre-Late Devensian landforms that, in common with similar features elsewhere in Scotland, were preserved under cold-based glacier ice during the LLGM (Chaps. 4 and 13).

On upper slopes, the boulder cover has migrated downslope to form debris-mantled slopes (Fig. 9.8a) locally

Fig. 9.7 Glacial and glacifluvial landforms. **a** Coire Dubh cirque, north Harris, which was occupied by a small Loch Lomond Readvance (LLR) glacier. Hummocky recessional moraines track the oscillatory retreat of this glacier. **b** Recessional (lateral) moraines mark the pulsed retreat of the LLR glacier that occupied Langadale, north Harris. **c** Hummocky recessional moraines near Loch Bhoisimid, north Harris. **d** Exposure of till in hummocky moraine, Glen Meavaig. **e** Streamlined hummocks, interpreted as subglacial bedforms, near the former LLR ice divide at the head of Glen Langadale. **f** The Glen Valtos meltwater channel, SW Lewis. (Images: **a–e** Colin Ballantyne; **f** John Gordon)

festooned by relict, bouldery ('stone-banked') solifluction lobes with risers up to 2 m high. These are absent inside the LLR limits, suggesting that they developed under permafrost conditions during the Lateglacial period (Chap. 4), and were immobilized as permafrost degraded under the rapidly warming conditions of the Lateglacial-Holocene transition. Holocene periglacial landforms on the mountains of the Outer Hebrides are limited to small-scale solifluction lobes and rare occurrences of small-scale, frost-sorted patterned ground.

Large rock-slope failures are comparatively rare on Lewisian gneiss mountains. Cave and Ballantyne (2016) identified ten failure sites on the mountains of Harris and South Uist, but most are debris-free failure scars in the form of cliffed embayments that represent rockslide release zones

Fig. 9.8 **a** The mountains of north Harris, showing coarse blockfield debris in the foreground and debris-mantled slopes on the upper parts of the adjacent mountains. **b** The Sròn Ard rock-slope failure near Loch Bhoisimid, north Harris. (Images: Colin Ballantyne)

where the associated runout debris has been removed by glaciers. Postglacial rock-slope failures, such as that at Sròn Ard (Fig. 9.8b), are generally small in comparison with similar features on the Scottish mainland. Holocene talus skirts the slopes flanking glacial troughs, notably those in the trough that bisects the Uig Hills. Older talus accumulations flank the eastern end of the Glen Valtos meltwater channel, where stratification of the talus debris probably implies accumulation under periglacial conditions during the Lateglacial period.

9.7 Coastal Landforms

The Outer Hebrides are characterized by a wide range of spectacular coastal landforms (Fig. 9.9). Erosional forms range from near-vertical cliffs that plunge into deep water, as at Tolsta Head on Lewis, to low rocky shorelines with a multitude of inlets and islets, notably the east coast of the Uists. Depositional forms include small bay-head beaches, such as those on the west coast of Lewis and continuous stretches of sand beach, dune and machair (a unique form of calcareous shell sand dune system; Sect. 9.7.5), such as those on the Atlantic coast of North and South Uist. Coastal systems on the Outer Hebrides share a common set of formative influences with the rest of the Atlantic seaboard of NW Scotland, including structural control, glaciation, sea-level change and morphogenetic environment.

9.7.1 Structural Controls

The configuration of the east (Minch) coast of the Outer Hebrides is dictated by the Minch Fault, where the resistant Lewisian gneiss in the west abuts the Torridonian and Permo-Triassic sandstones, Jurassic mudstones and Palaeogene volcanic rocks of the Minch Basin. In consequence, the Minch coast is mainly rocky, with cliffs and narrow intertidal shore platforms. The west (Atlantic) coast is generally low-lying and sediment-covered in the Uists and Benbecula. The northwest coast of Lewis is linear and aligns with a major fault ~5 km offshore. Fault control over coastal orientation is apparent also along the major faults that bound the structural blocks of Harris and south Lewis, resulting in linear coastal segments along the Sound of Harris and the southwest coast of North Harris (Fig. 9.1).

The coastline is extensively broken by inlets that mainly follow faults, shears and fracture zones (Fig. 9.9a, b). The long sea lochs of Lochs Seaforth, Tarbert and Roag follow NNW–SSE oriented faults and fractures (Franklin 2013). In the Uists, Loch Euphort is aligned west–east, but most of the long Minch coast inlets follow NW–SE aligned faults or shear zones (Fettes et al. 1992; Stoker et al. 1993; Figs. 9.1 and 9.2b). The detail of headlands, bays and platforms is widely controlled by fracture orientation and spacing, for example at Loch Roag, and the orientation and morphology of coastal rock skerries, headlands and reefs are linked to regional foliation of the gneiss, as at Loch Maddy in South Uist (Fig. 9.9c).

On the northeast coast of Lewis, outcropping Permian sandstone supports strikingly different coastal features and the coast is low in elevation, with broad shore platforms. Erosion of the sandstones and local glacial deposits provides a source of sand and gravel for beaches linking Stornoway to the Eye peninsula and extending intermittently northwards along the eastern shoreline towards the prominent gneiss cliffs at Tolsta Head.

9.7.2 The Impact of Glaciation

Glaciated coasts are often strongly influenced by inheritance of glacial landforms and sediments. Generally, inlets on the Outer Hebrides coast are not primarily products of wave erosion, but have developed along faults or structural

Fig. 9.9 **a** Indented and glacially roughened rock coast, south Lewis. **b** Steep-sided, fault-controlled Loch Seaforth extending inland from the Minch coast. **c** View west over the partly submerged strandflat coast of North Uist, showing the SW–NE structural trend of Loch Maddy (right) and the fault-controlled Minch coast in foreground. **d** View north towards Solas and Udal, North Uist, typifies the western coast of the Uists; beaches, backed by a narrow cordon of dunes and machair, mask the underlying bedrock. **e** Loch Arnol, NW Lewis, is impounded by a sporadically overtopped gravel barrier that rolls landwards during storms. (Images: **a**, **b**, **e** James Hansom; **c**, **d** © P. and A. Macdonald/Aerographica/NatureScot)

weaknesses exploited by glacial erosion. The northwest coast of Lewis transitions from a straight coast, lacking in large inlets and bays, to a highly indented and island-studded coast south of Loch Roag (Fig. 9.9a), reflecting a southward increase in the intensity and depth of glacial erosion. The straight, northwestern segment experienced limited glacial erosion beneath cold-based ice, allowing the localized preservation of beach gravels below till, whilst the indented southwestern segment was selectively eroded beneath warm-based ice flowing in trenches and valleys. The wider implication is that rising sea level has occupied pre-existing glacial landscapes in the Outer Hebrides.

The Loch Maddy area in North Uist represents a classic example of drowned knock-and-lochan topography (Hansom 2003a). Loch Maddy itself is 7 km long and wide but no more than 20 m deep. Fracture-guided glacial trenches now form long, thin, shallow arms of the sea, or *fjärds* (Earll and Pagett 1984). The low, crenulate shoreline, peppered by islets (Fig. 9.9c), is similar to emergent shorelines around the Baltic Sea (Hall et al. 2019). The fjärd coasts of the Outer Hebrides are part of a more extensive, submarine platform that extends for tens of kilometres to the west (Fig. 9.2b) and resembles the emergent strandflat of the Norwegian coast. The fjärd coasts and submerged platform have developed through multiple cycles of erosion by ice, frost and waves (Dawson et al. 2013), with a general form probably inherited from an earlier subaerial planation surface (Smith et al. 2019). During the Holocene marine transgression, calcareous shell sand accumulated in surface depressions on the platform, subsequently migrating shoreward to form suites of low, dune-capped sand barriers backed by machair grassland (Sect. 9.7.5). These Atlantic beaches have been interpreted as a form of barrier island, uniquely developed atop bedrock (Cooper et al. 2012), and demonstrate the key role of inherited bedrock topography on coastal evolution. Spectacular beaches, backed by machair grassland, occur along almost all the west coasts of the Uists and Barra, as exemplified at Solas and Baleshare in North Uist and Askernish and Kilphedar in South Uist (Fig. 9.9d).

Substantial cover of glacigenic sediment is uncommon at the coast (Peacock 1991) and lack of source sediment has resulted in very few beaches on the Minch coast south of the Eye Peninsula. Where patchy till cover occurs, it has provided sediment for intertidal gravel and sand, such as occurs along the shores of Lochs Maddy and Euphort. On the west coast of Harris and SW Lewis, local glacigenic sediments, augmented by offshore shelly sands, form the source sediments for beach and dune development at several, mainly sheltered, sites, such as Hushinish and Scarasta (Fig. 9.1). Elsewhere, where valleys meet the coast in NW Lewis, short gravel barriers nourished by glacigenic gravels and wave quarrying of local rock impound lagoons, such as that at Arnol (Fig. 9.9e).

9.7.3 The Legacy of Sea-Level Change

Changing sea level controls coastal position, configuration and the marine processes that act on coasts. Evidence of former high sea levels in the Outer Hebrides is provided by emerged rock platforms and raised beach gravels overlain by till on Lewis (von Weymarn 1974, 1979; Fig. 9.5a) and Vatersay (Selby 1987); these indicate that relative sea levels reached up to +10 m OD prior to the last glacial stage. Subsequent glacial and marine erosion have failed to erase these emergent shoreline features, which constitute the only evidence of relative sea level (RSL) being substantially higher than present sea level (Smith et al. 2019). Following deglaciation, the RSL trend over much of the Lateglacial and Holocene has been one of marine transgression submerging the coastal landforms of the Outer Hebrides. In North Uist, the Minch coastline comprises low (<30 m high) cliffs with gradients of 20–50° that continue underwater and lack erosional features such as caves, arches and stacks (Ritchie 1968). The presence nearby of glacially abraded rock surfaces that plunge below sea level with no clear marine erosional signature indicates that these 'pseudo-cliffs' are subaerial slopes drowned by rapid Early Holocene submergence. Jordan et al. (2010) have shown most of the local sea-level rise in south Harris took place before the Middle Holocene.

Holocene sea-level rise resulted in the Outer Hebrides lacking the well-developed suites of emerged postglacial shore platforms and raised beaches that characterize most of the Scottish mainland and Inner Hebrides (Chaps. 4, 10 and 11). Along the Atlantic seaboard of the Uists, Holocene sea-level rise resulted in offshore glacigenic sediment being driven onshore by waves and, in common with much of the Scottish coast, these finite offshore sources became progressively exhausted (Carter 1992; Hansom 2001). In the Outer Hebrides, however, offshore sediment sources were progressively augmented by calcareous shell sand that first began to accumulate in the North Uist and Benbecula machair at ~8.7 ka, in Barra by ~6.8 ka and in Pabbay by ~4.3 ka (Gilbertson et al. 1999). This wide range of dates implies that the onset of shell sand accumulation and the initiation of machair was asynchronous, possibly because sediment within offshore basins was affected by the rising Holocene sea at different times.

Progressive reduction in the rate of sea-level rise after ~6.5 ka marked a transition to a slower rate of transgression that initiated steady shoreface reworking, together with increased shell sand delivery from offshore (Hansom 2003b). These conditions favoured widespread development of beaches and dunes, with strong onshore winds blowing beach sand into dunes and machair plains, a process that continues today.

9.7.4 Present-Day Morphogenetic Controls

The main morphogenetic controls on the Outer Hebrides coasts include wave climate, sea level and sediment supply. Along the Atlantic coast the offshore wave climate is severe, with dominant waves from between 230° and 270° N over an unrestricted fetch (Ramsay and Brampton 2000), whereas the Minch coast is more sheltered. However, inshore along both coasts, wave activity is moderated by nearshore bathymetry and shelter provided by headlands, islands and skerries.

The impact of waves is conditioned by sea level, and estimates by Dawson et al. (2001) placed local relative sea-level rise at ~ 1 mm a^{-1}. However, using tide gauge records for 1992–2013, Hansom et al. (2017) identified a recent rise of ~ 4.3 mm a^{-1} at Stornoway, although it remains to be seen whether this short-term rate persists. Recent sea-level rise has likely contributed to local beach erosion since the 1970s, particularly in the Stornoway area and along the Atlantic coast of the Uists, for example at Baleshare, Askernish and Kilphedar (Hansom et al. 2017). Moreover, if sea-level rise accelerates, a phase of more widespread erosion and marine flooding is anticipated on the beaches and dunes that protect low-lying machair (Angus and Hansom 2004). Apart from the machair-clad west coast of the Uists, much of the Outer Hebrides coast is crenulate and headland-dominated, with limited potential for longshore sediment transport; large self-contained sediment transport cells do not occur in the Outer Hebrides, so a cell-based approach to understanding beach and dune development is inappropriate (Hansom et al. 2004).

9.7.5 Machair

Machair is the name given to the distinctive, floristically diverse, dune-plain system whose global distribution is restricted to the west coasts of Scotland and Ireland. On the western seaboard of the Outer Hebrides, it is most extensively developed along the west coasts of the Uists, where it achieves its highest biodiversity and where traditional and modern agriculture supports scattered rural settlement (Fig. 9.10a). The machair geomorphic system is a flat or gently sloping coastal dune-plain of windblown calcareous shell sand, derived from offshore and nearshore biogenic production. Areas of machair usually incorporate a dune cordon to seaward and species-rich grassland (managed by traditional low-intensity agriculture), wetlands, lochs and blacklands (peat on rock) to landward (Hansom and Angus 2006). However, the seaward dune cordon may now be narrow or missing owing to sea-level rise and frontal wave erosion (Fig. 9.10b, c). Characteristic machair surfaces are lime-rich, subject to strong, moist, oceanic winds and exhibit detectable, long-term human modification from grazing, cultivation, addition of natural fertilizer such as kelp, and sometimes artificial drainage. The term 'machair' has botanical, geomorphological and cultural connotations in the Outer Hebrides.

Hansom and Angus (2006) have summarized the main features of machair evolution as a seaward-migrating dune system developed where a positive sand budget on the fronting beach produced coastal accretion, with some landward movement of windblown beach sand anchored by dune grasses (Fig. 9.10b). However, when sediment supply to the feeder beach diminishes, the seaward-moving sequence is halted and replaced by a landward-moving erosional sequence. This is a common pattern on the Scottish coast where once-abundant but finite offshore sediment sources became progressively exhausted as sea-level rise began to slow in the Middle Holocene, and accreting systems were replaced by erosional ones (Carter 1992; Hansom 2001; Fig. 9.10b). On machair coasts, wave erosion of frontal dunes has favoured landward sand blow and sedimentation onto the machair surface, producing a 'layer-cake' deposition of progressively higher machair surfaces fronted by steep eroded bluffs, with intervening low-lying landward-dipping deflation surfaces. The current machair system is essentially erosion-driven and landward-migrating, with new surfaces developing as old ones are consumed (Dawson et al. 2004). This process is most advanced where high-energy wave conditions are compounded by ongoing relative sea-level rise, resulting in loss of dune cordons and frontal erosion of the machair plain itself (Fig. 9.10b), as at Solas and Udal in North Uist (Fig. 9.10c). At some South Uist sites, breaches in the dune cordon have been artificially filled to prevent marine flooding of the low-lying machair hinterland (Angus and Hansom 2021; Fig. 9.10d).

Archaeological evidence supports the geomorphological reconstruction depicted in Fig. 9.10b, with early sand surpluses followed by declining sand supply, so that sites located originally within accreting dunes are now undergoing frontal wave erosion. Near Solas in North Uist, for example, machair sand layers separate an intact occupational sequence from the Neolithic, Bronze Age, Iron Age, Viking, Medieval, Post-Medieval and later periods. In South Uist, organic palaeosols dating from Bronze Age to Medieval times occur intercalated with machair sand (Gilbertson et al. 1999). Most remarkable of all, at Uig on the west coast of Lewis, the world-famous Lewis Chessmen of Viking age were found buried within the machair.

Recent analysis of the status of the beaches fronting machair systems has identified erosional hotspots of up to 20–30 m of erosion since the 1970s, particularly in the Uists (Hansom et al. 2017). Loss of some dune cordons will result in higher machair surfaces accreting from sand blown inland, but wave erosion and overwash may also replace low-lying coast-parallel lochs and machair with marine lagoons (Angus and Hansom 2004, 2021).

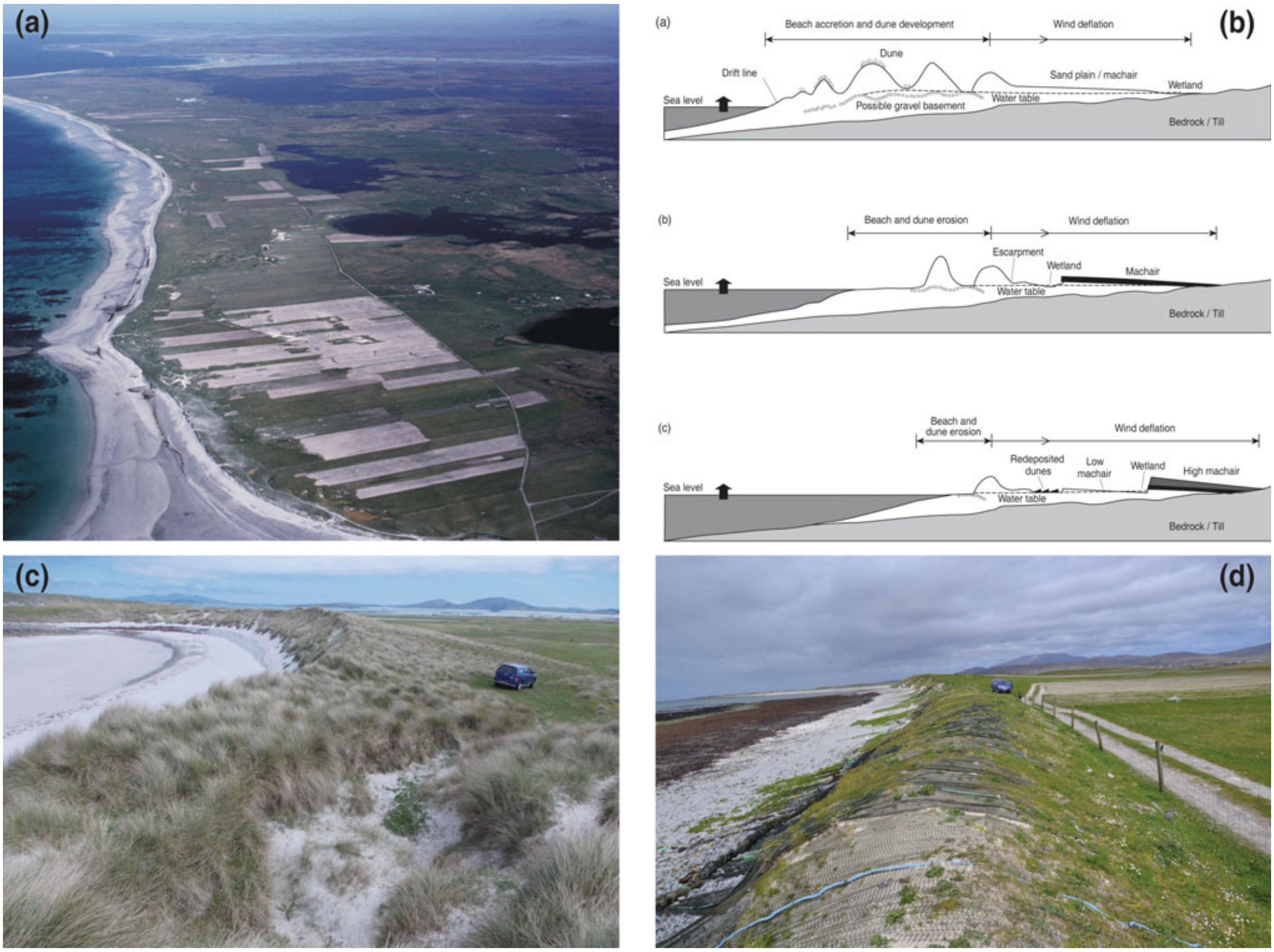


Fig. 9.10 A View north over the west coast machair on South Uist, showing a narrow, eroding frontal dune cordon backed by landward-sloping, low-lying machair grassland utilized for traditional agricultural cropping. **b** Sequence of machair development depicting: (a) slowing of relative sea-level rise before ~6.5 ka and plentiful sediment fuelling seaward accretion; (b) ongoing relative sea-level rise after ~6.5 ka, when reduced sediment supply drives landward migration; (c) present day, when relative sea-level rise and reduced sediment supply are driving frontal erosion, loss of dune cordon and erosion of machair grassland (modified from Hansom 2003b). **c** Frontal wave and wind erosion of the dune cordon at Solas and Udal, North Uist, leads to reduced protection of the low-lying machair behind. **d** Loss of the dune cordon at Kilphedar, South Uist, has led to construction of an embankment to prevent marine flooding of the adjacent low-lying machair. (Images: **a** © P. and A. Macdonald/Aerographica/NatureScot; **c**, **d**: James Hansom)

9.8 The St Kilda Archipelago

9.8.1 Introduction

The four small islands and outlying sea stacks of the St Kilda archipelago (57.817° N, 08.583° W; Fig. 9.2) represent the terrestrial expression of a Palaeogene igneous complex ~14 km in diameter. Much of the complex forms a shallow platform or platforms 40–80 m below sea level rising above a deeper platform at depths of ~120 to 130 m (Sutherland 1984). As the highest point on the archipelago is 426 m above sea level, the igneous complex represents a prominent upland rising to ~550 m above the adjacent shelf. Of the four islands, Hirta is the largest (~6 km^2), and comprises two valleys, Gleann Mòr in the northwest and Village Bay in the southeast (Fig. 9.11a). The rest of the island consists of rolling hills that are flanked on their seaward margins by precipitous cliffs.

In common with the Palaeogene igneous complexes of the Inner Hebrides (Chaps. 2 and 10), that of St Kilda was formed during a period of intense magmatic activity related to lithospheric thinning above the proto-Iceland mantle plume, which ultimately led to opening of the North Atlantic Ocean. The age of the St Kilda complex is constrained by gabbroic rocks dated to ~60 Ma, and the youngest intrusive body, the Conachair Granite, to ~55 Ma. Hirta is composed of a range of intrusive rocks, principally gabbros, dolerites,

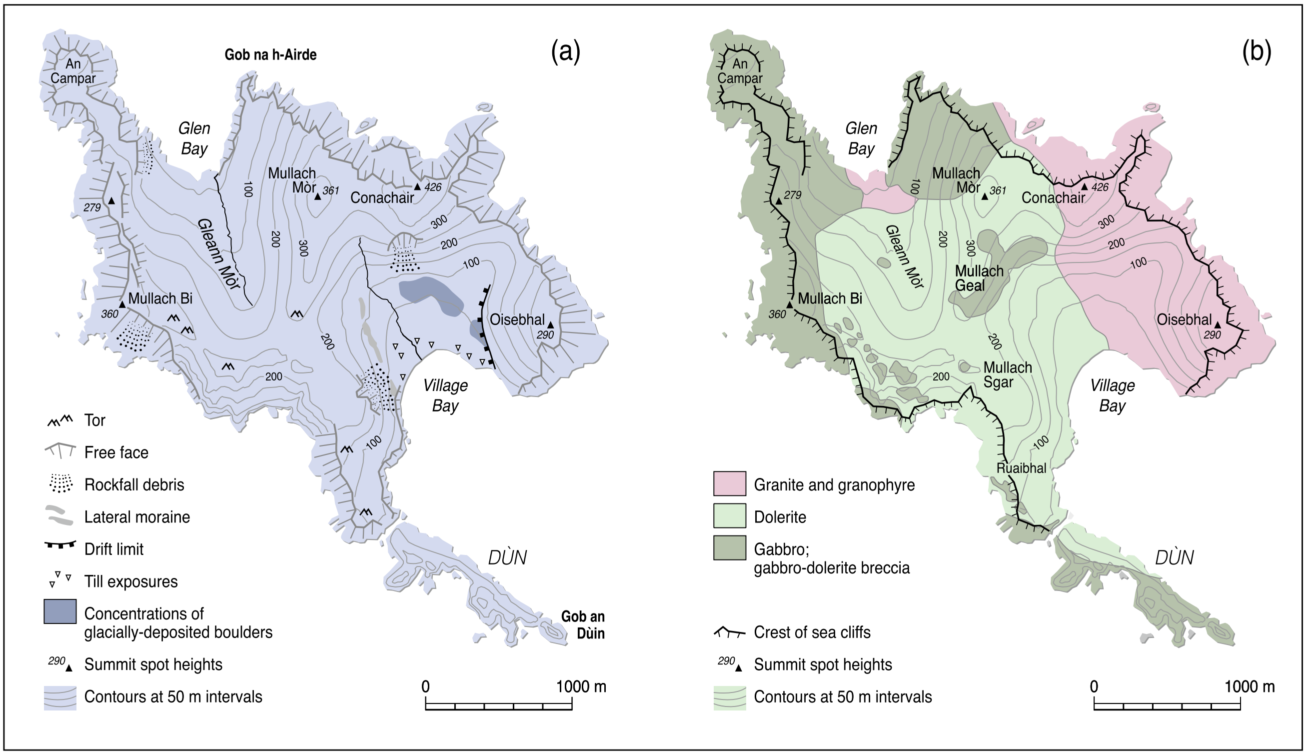


Fig. 9.11 **a** The island of Hirta in the St Kilda archipelago, showing the location of glacial landforms and deposits, tors and rockfall or rockslide deposits. **b** Solid geology of Hirta. (Adapted from Ballantyne et al. (2017) © 2017 Collegium Boreas)

gabbro-dolerite breccias, granites, granite-dolerite hybrids and granophyres, which collectively represent successive intrusion of igneous bodies of widely different composition (Harding et al. 1984; Ritchie and Hitchen 1996; Fig. 9.11b).

9.8.2 Glaciation and Periglaciation

St Kilda is the most westerly landmass on the Hebrides shelf, ~65 km west of the Outer Hebrides and 40–60 km east of the shelf break, and a key site for constraining the maximum westwards extent of the last Outer Hebrides Ice Cap. Sutherland et al. (1984) proposed four glacial or periglacial episodes on Hirta, but reinvestigation of the Quaternary sediments, stratigraphy and landforms of the Village Bay area produced evidence for only two glacial events (Hiemstra et al. 2015). The earlier, recorded by exotic grains in alluvial deposits and fragments of gneiss and sandstone dredged from the adjacent seabed, has been interpreted as indicating that St Kilda was engulfed by an ice mass of probable pre-Devensian age advancing from the Outer Hebrides (Harding et al. 1984). The latter is represented by a lateral moraine enclosing locally derived subglacial till deposits in the Village Bay area. Hiemstra et al. (2015) interpreted this as indicating that a small glacier formed in Village Bay during the LLGM, implying that St Kilda lay beyond the westward reach of the Outer Hebrides Ice Cap. TCN exposure dating supports this conclusion. Exposure ages obtained for glacially deposited boulders in Gleann Mòr indicate that this valley was last occupied by a small local glacier at 30.9 ± 3.2 ka, and five boulders within the limits of the Village Bay glacier yielded consistent exposure ages averaging 19.2 ± 2.3 ka. These results indicate that during the Late Devensian St Kilda supported only two small locally nourished glaciers, and that the Outer Hebrides Ice Cap terminated east of St Kilda, tens of kilometres short of the shelf-edge (Ballantyne et al. 2017).

This conclusion is supported by geomorphological evidence. The easterly summits of Conachair and Oisebhal support granite and granophyre blockfields, but no erratics derived from these blockfields have been found west of the granite-granophyre outcrop, as might be expected if these summits had been over-run by ice from the east. Moreover, high ground along the southeast margin of Hirta is crowned with dolerite tors (Fig. 9.12a), yet none exhibit evidence of modification by glacier ice. Finally, the slopes outside the Village Bay glacier limit are covered by periglacial slope deposits up to 13 m thick overlying fractured and weathered bedrock (Fig. 9.12b). These deposits comprise angular, locally derived dolerite debris in the form of a diamicton that resembles the gelifluctate deposits that form coastal cliffs outside the limit of the last ice sheet in SW England (Ballantyne and Harris 1994) . They imply a very prolonged period of rock weathering and mass movement under severe periglacial conditions, consistent with St Kilda having remained outside the limits of not only the last ice sheet, but possibly earlier episodes of ice-sheet glaciation. St Kilda is therefore unique in Scotland as the only terrestrial location not over-run by the last (and possibly earlier) Pleistocene ice sheet(s).

The most distinctive postglacial landform on Hirta is a massive, arcuate debris ridge at the head of the Village Bay valley, below the Conachair cliff. This ridge is 215 m long, rising to 20 m above the adjacent ground and covered by granophyre boulders derived from the cliff (Fig. 9.12c). Although originally interpreted as a protalus rampart, the ridge resembles others in Scotland formed by cascading rockslide debris impacting a basal break of slope and rebounding to form an arcuate ridge, implying catastrophic failure of over 0.1 Mt of rock from the cliff face.

9.8.3 Offshore and Coastal Landforms

Apart from Village Bay, the coastline of the archipelago is impressively cliffbound, with Conachair (430 m) on Hirta and nearby Stac an Armin (191 m) forming the highest sea cliff and sea stack in Great Britain (Hansom 2003c). The variety of dramatic cliffs and cliff-related forms (including geos, arches, stacks, caves and blowholes) contributed to the archipelago being designated as Scotland's first UNESCO Natural World Heritage Site in 1987.

At the peak of the Last Glacial Maximum, global sea level fell to –120 m or so (Clayton 2003) and evidence of planation at this depth occurs around St Kilda where a near-horizontal bedrock surface occurs at –120 to –130 m depth with only a patchy sediment cover (Sutherland 1984) . A prominent 40 m high submerged cliff separates the –120 m platform from another rock platform or platforms at –40 to –80 m depth. All the islands and sea stacks rise steeply from a depth of –40 m and the valleys on Hirta can be traced down to this level. Despite the extensive occurrence of submerged platforms and cliffs at depth around St Kilda, virtually no shore platform development has occurred

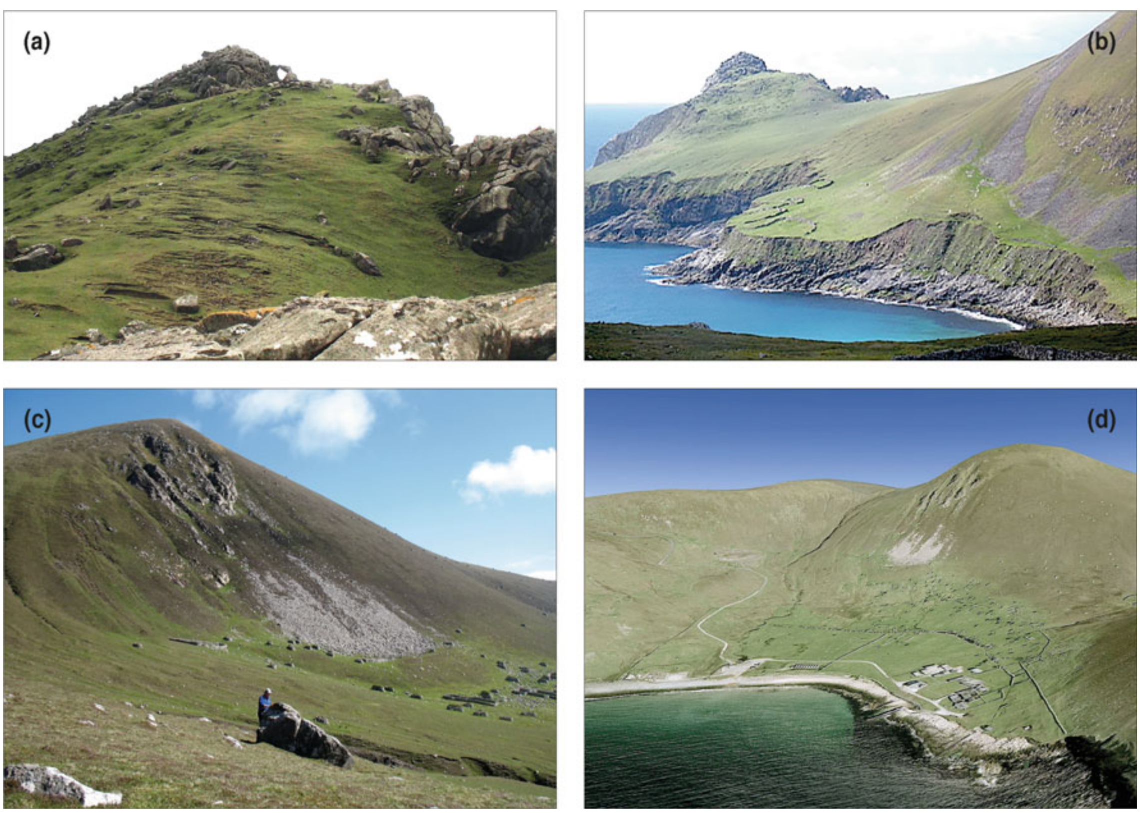


Fig. 9.12 Landforms of Hirta. **a** Dolerite ridge-crest tor. **b** Periglacial slope deposits up to 13 m thick resting on weathered bedrock, Village Bay. **c** Arcuate ridge of rockslide debris at the foot of the Conachair cliffs. **d** Village Bay gravel beach. (Images: **a–c** Colin Ballantyne; **d** Google Earth™)

at present sea level and only a few low skerries exist (Hansom 2003c).

Surrounded by deep water, the St Kilda coastline experiences unattenuated waves from all directions and extreme weather conditions that exceed those anywhere else in the British Isles. The predicted 50-year wave height of 35 m for St Kilda is the highest in the British Isles; the highest individual wave ever recorded reached 29.1 m, 300 km west of St Kilda near Rockall (Holliday et al. 2006). Large, high-energy Atlantic swell and storm waves impact directly on the exposed coastline so that plumes of water and spray overtop the cliffs of Dùn and cascade into Village Bay during storms. The effect is so persistent that halophytic saltmarsh species dominate the steeply sloping cliff vegetation communities, not only on the exposed SW-facing coast of Mullach Bi, but also the 150 m high lee slopes of the Ruaibhal cliffs facing Village Bay (Walker 1984a). Stripping of soil and turf by strong winds occurs at a variety of altitudes, reaching 400 m on Conachair. The 45 m wide neck of An Campar, consisting of boulders within a sand and clay matrix, is being actively eroded by waves and spray on both sides, together with wind-stripping of soil and turf cover. In 1998, the neck was narrowing at 0.26 m a^{-1} on both sides (Hansom 2003c; Fig. 9.13). In Gleann Mòr, coastal gradients are less severe and allow erosional wave run-up in Glen Bay to reach heights of up to 40 m OD.

The only substantial beach occurs at Village Bay (Fig. 9.12d). The beach is characterized mainly by coarse gravels in the west, but these coarsen to large boulders towards the east, where the shoreline is backed by rock gabions to restrict coastal erosion of the land behind. Even in this relatively sheltered location, periodic storms cause substantial erosion, damaging infrastructure and cultural sites close to the shore (Hansom 2003c). Along the west coast of Village Bay, the coastline consists of a low-angled bedrock ramp surmounted by steep bluffs cut in massive diamictic gelifluctate (Fig. 9.12b). Along the exposed outer coasts of the rest of the islands the steep cliff faces plunge directly into deep water and are associated with a wide variety of stacks, caverns, arches, tunnels, blowholes and deep geos at various stages of formation. Particularly impressive natural arches occur on Dùn (Gob an Dùn, 50 m long and 24 m high) and in Glen Bay (Gob na h-Airde, 91 m long and 30 m high). These caves, geos and other coastal features have generally been eroded along dykes or thin, inclined, sheet intrusions, resulting in irregular notched and stepped cliff profiles. Free faces in the high cliffs are locally skirted by talus but the lower cliffs are often vertical. On the north and east coasts, the dolerite outcrops of Mullach Mòr and the dolerite sheets and dykes intruded into the granites and granophyres of Conachair and Oisebhal form irregular,

Fig. 9.13 View south from An Campar, with Glen Bay to the left and the serrated cliffs of Mullach Bi to the right. In the foreground, both sides of the neck of An Campar are rapidly eroding. (Image: © Mick Crawley, licenced for reuse under Creative Commons Licence CC BY-SA 2.0)

stepped cliff profiles interspersed with steep grassed slopes, with many small protuberances and pinnacles.

9.9 Conclusion

The Outer Hebrides Platform is a distinctive, ancient geological terrain with a long history of landform development. The present basement surface above and below sea level is an extensive erosion surface that was subjected to intense weathering and erosion through the Palaeogene and continuing modification in the Neogene and Pleistocene. The partly submerged surface of the platform is an example of a strandflat, shaped first by subaerial weathering and later by glacial, periglacial and marine processes. A succession of independent ice caps developed on the Outer Hebrides during the Pleistocene, but parts of the island chain were over-run at intervals by mainland ice. Glacial erosion roughened the gneiss surface on the platform and carved out the Minch and Sea of the Hebrides troughs. The last ice sheet deposited only a patchy till cover, but more limited mountain glaciation during the Loch Lomond Stade resulted in the formation of end, lateral and recessional moraines. Postglacial marine transgression flooded a glaciated coastline and moved shoreward large volumes of sand that have supplied beach, dune and machair land systems along Atlantic coasts. Wave and wind dynamics and agricultural practices continue to shape the distinctive machair landscape. The exposed outlying islands of the St Kilda archipelago represent the only terrestrial location not over-run by the last Scottish Ice Sheet and are notable for the highest sea cliffs and stacks in Scotland. The Outer Hebrides display an exceptional diversity of distinctive landscapes and landforms that have developed over the Late Cenozoic in response to interactions between tectonics, rock type and structure, and the subaerial processes of weathering, fluvial, glacial and marine erosion and sedimentation set within the context of changing sea levels.

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10 The Islands of the Hebridean Igneous Province: Skye, Mull, Rùm and Arran

Colin K. Ballantyne

Abstract

The islands of the Hebridean Igneous Province are largely composed of Palaeogene igneous rocks relating to extensional tectonics and magmatic activity within the interval ~62–55 Ma. Their geology is dominated by plateau lavas, subvolcanic central igneous complexes, and sills and dyke swarms intruded into earlier sedimentary and metasedimentary rocks. Modification of the igneous terrain by successive Pleistocene ice sheets and mountain glaciers has produced radiating troughs and cirques, pinnacled arêtes and glaciated trap topography. During the local Last Glacial Maximum (~30–26 ka), Skye, Mull and Arran sustained independent ice centres that fed ice streams on the adjacent shelves; following decoupling from these ice streams, locally nourished glaciers readvanced on Skye and Arran. During the Loch Lomond Stade (~12.9–11.7 ka) all four islands hosted mountain glaciers that deposited end, lateral and recessional moraines. Postglacial landforms include periglacial and aeolian landforms on high ground, and spectacular landslides (one of which is the largest in Britain) at sites where failure of sedimentary rocks has caused extensive collapse of the overlying lavas. A wide range of coastal features are represented, including shore platforms of various ages, raised deltas, raised shingle ridges and both Lateglacial and Holocene raised beaches. The islands of the Hebridean Igneous Province are internationally renowned not only for their outstanding geodiversity, but also the haunting beauty of their strikingly dramatic landscapes.

Keywords

Subvolcanic plutons • Trap topography • Sills • Radiating glacial troughs • Cirques • Moraines • Solifluction • Patterned ground • Aeolian deposits • Basalt scarp landslides • Coastal rock platforms • Raised beaches

10.1 Introduction

The islands of Skye, Rùm, Mull and Arran on the western seaboard of Scotland (Fig. 10.1) share a common geological heritage: all are largely composed of igneous rocks that were intruded into, or erupted onto, earlier rocks of Neoproterozoic to Jurassic age during a comparatively brief period of intense Palaeogene magmatic activity. All support central igneous complexes that represent the roots of Palaeogene volcanoes; all are seamed with Palaeogene dyke swarms and intruded by sills; and two (Skye and Mull) are largely composed of stacked mafic lava flows that extend far offshore and onlap NW Rùm. All four islands have experienced modification by successive Pleistocene mountain glaciers and ice sheets; all were completely buried under polythermal glacier ice during the local Last Glacial Maximum (LLGM); all supported glaciers during the Loch Lomond Stade of ~12.9–11.7 ka, and all are scarred by postglacial landslides and surrounded by raised shorelines. Yet despite this common heritage, each island contains landscapes and landforms that are distinct from those on the others and often unique in Scotland. Collectively, the islands of the Hebridean Igneous Province capture a range of geodiversity greater than that of any other Scottish region: in the early twentieth century they formed the topic of some of the most seminal Memoirs of the Geological Survey; they contain 45 Sites of Special Scientific Interest (SSSIs), selected for their international geological or geomorphological significance; and they have proved inspiring training grounds for successive generations

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of Earth science students. The most striking geological and geomorphological features have captured the imagination of writers, photographers, poets, artists, composers and other visitors, stimulating an interest in the origin and evolution of some of the most varied and dramatic landscapes in the British Isles.

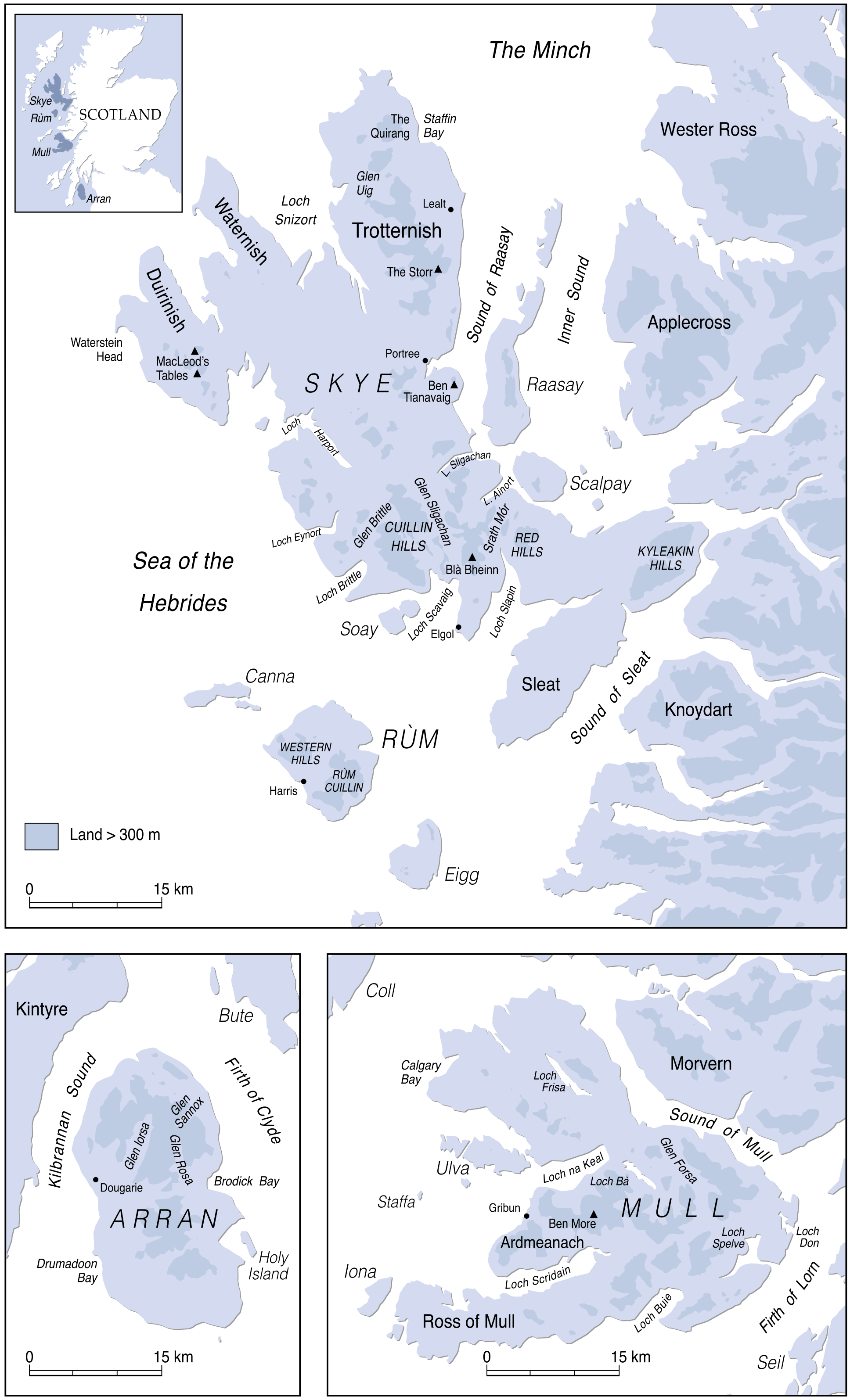

Fig. 10.1 The islands of Skye, Rùm, Mull and Arran: key locations mentioned in the text

10.2 Geology and Geological Evolution

Much of the landscape of the islands of the Hebridean Igneous Province owes its origin to crustal stretching and associated magmatic activity along western Scotland in the interval ~62–55 Ma, immediately prior to the opening of the North Atlantic Ocean. Such magmatic activity was initially associated with fissure eruptions that produced huge volumes of mainly mafic lava, forming the extensive plateaux of stacked, often horizontal lava flows of the Skye and Mull Lava Groups (Fig. 10.2). These underlie most of both islands and extend up to 50 km offshore to form the islands of Canna, Eigg and Muck. Later eruptions were sourced from stratovolcanoes, now eroded to their roots to form central igneous complexes. These had differing individual histories, generally involving intrusion of magma, uplift of surrounding and overlying rocks (possibly to heights of 2000 m or more), eruption of lavas and pyroclastic debris then partial collapse and unroofing by rapid erosion, which exposed their plutonic cores. Thick transgressive sills relating to Palaeogene magmatic activity crop out extensively in northern Skye and southern Arran, and all four islands are seamed by numerous NW–SE aligned dykes of basalt and dolerite, mainly fed from the Skye and Mull central complexes, that indicate intrusion under an early Palaeogene NE–SW extensional stress regime.

The geological complexity of the four islands has engendered a cornucopia of publications. The geological evolution of the entire province is summarized by Bell and Williamson (2002), Emeleus and Bell (2005) and Brown et al. (2009), that of Skye by Bell and Harris (1996), that of Rùm by Emeleus (1997) and Emeleus and Troll (2008, 2011), and that of Arran by MacDonald and Herriot (1983).

10.2.1 Geology

Geologically, Skye consists of three regions (Fig. 10.3). Southeast Skye is underlain by pre-Cenozoic metamorphic and sedimentary rocks. These include thrust sheets of Lewisian gneiss, Moine schists and Torridonian sandstones in Sleat and easternmost Skye, and sedimentary rocks of various ages intruded by Palaeogene sills in the Broadford–Loch Eishort area. The mountains of central Skye are underlain by Palaeogene intrusive rocks emplaced at ~59–56 Ma in the form of four subvolcanic plutons. The earliest, westernmost pluton, the Cuillin complex, comprises concentric masses of layered mafic and ultramafic rocks (mainly gabbros and eucrites) that have been penetrated by thin dolerite cone sheets and numerous discordant basalt dykes. To the northeast, the Cuillin complex is truncated by two further complexes of silicic rocks (mainly epigranites), the Srath na Crèitheach and Western Red Hills complexes. A third epigranite complex (the Eastern Red Hills complex) occupies the ground between Loch Ainort and Loch Slapin. The four complexes represent a gradual eastward shift in the locus of magmatic activity, accompanied by progressively increasing magma acidity, probably due to fractionation and contamination by quartz-rich crustal rocks. Northern Skye is underlain by horizontal or westward-dipping Palaeogene basaltic lava flows, with an aggregate thickness of at least ~1800 m, that were emplaced at ~60 Ma on low-relief terrain underlain by sedimentary rocks. Individual flows are typically 5–15 m thick and frequently separated by tuffs, breccias and laterite horizons (Fig. 10.2a). Mesozoic sediments (primarily shales) and thick transgressive dolerite sills crop out north and east of the lavas on the Trotternish Peninsula.

Rùm comprises the remains of an igneous complex, dated to ~60.5 Ma, that occupies most of the western, central and southern parts of the island, and is intruded into Neoproterozoic sandstones of the Torridon Group, which underlie the north and east. The central complex developed in several phases within a ring fault. The earliest phase involved silicic magmatism, now represented by the felsic rocks of the Western Hills, with uplift of overlying Torridonian rocks and caldera subsidence within the ring fault. The second phase was dominated by intrusion of layered gabbros and ultrabasic rocks of the Rùm Cuillin. Finally, vigorous erosion exposed the plutonic roots of the complex, and basaltic lavas from Skye invaded northern Rùm, and now crop out in parts of the Western Hills.

The geology of Mull is dominated by Palaeogene plateau basalt lavas, which surround the exhumed remains of a central igneous complex that underlies the most mountainous terrain. The main lavas underlie all of northern Mull, most of western Mull and eastern coastal areas, have an aggregate thickness of ~1000 m, and are composed of lava flows typically 10–15 m thick. These are overlain in central Mull by younger lavas that erupted as the central igneous complex collapsed to form a caldera. The central complex itself represents the sequential intrusion of three volcanic centres, and comprises both granitic rocks and layered gabbros, and intrusive rocks in the form of cone sheets and ring dykes. Pre-Cenozoic rocks on Mull include outcrops of Devonian granite and Dalradian metasedimentary rocks on the Ross of Mull, and various Mesozoic sedimentary rocks (sandstones, mudstones and limestones) that underlie the main basalt lavas and crop out in narrow bands along eastern coastal areas and the coast of the Ardmeanach Peninsula.

Arran differs from its Hebridean cousins in lacking extensive Palaeogene lava fields. The pre-Cenozoic geology of

Fig. 10.2 Palaeogene lavas. **a** Stacked basalt lavas of the Trotternish escarpment, northern Skye. **b** Columnar jointing developed in a lava flow on the Isle of Staffa, western Mull. (Images: **a** Colin Ballantyne; **b** John Gordon)

the island is dominated mainly by metasediments of Dalradian age and sedimentary rocks, mainly of Devonian to Triassic age. The Palaeogene igneous components of Arran's geology are fourfold: an unroofed pluton in north Arran; a central igneous complex that represents the remnants of a collapsed volcano; an extensive outcrop of sills in south Arran; and a remarkably dense dyke swarm. The Northern Granite forms a near-circular outcrop, intruded at ~60 Ma, that comprises an outer zone of coarse-grained granite and a slightly younger inner zone of fine-grained granite, both of which have been exhumed from under a thick cover of older rocks. The ring complex to the south is ~5 km in diameter and marks the site of a central volcano that collapsed to form a caldera and contains intrusive rocks of variable composition with localised outcrops of lavas and pyroclastic rocks. The sills that crop out extensively in southern Arran are mainly intruded into Permian sandstones and are of variable composition, though dolerites predominate. The Arran Dyke Swarm is the best example of such a feature in Britain: over 500 individual dykes ranging from 0.3 to 30 m in width have been mapped, most of which are aligned NW–SE or NNW–SSE.

10.2.2 Geology and Landscape

In few parts of Scotland is the relationship between geology and landscape more strikingly illustrated than on the islands of the Hebridean Igneous Province. The plutonic rocks of the subvolcanic central igneous complexes and the Northern Granite of Arran tend to be associated with the highest relief and most prominent mountains, though there are striking contrasts between upland areas. The gabbros and eucrites of the Cuillin Complex on Skye form a serrated ridge of mountains, locally dissected by preferential weathering and erosion of dykes, but the granites of the adjacent Red Hills

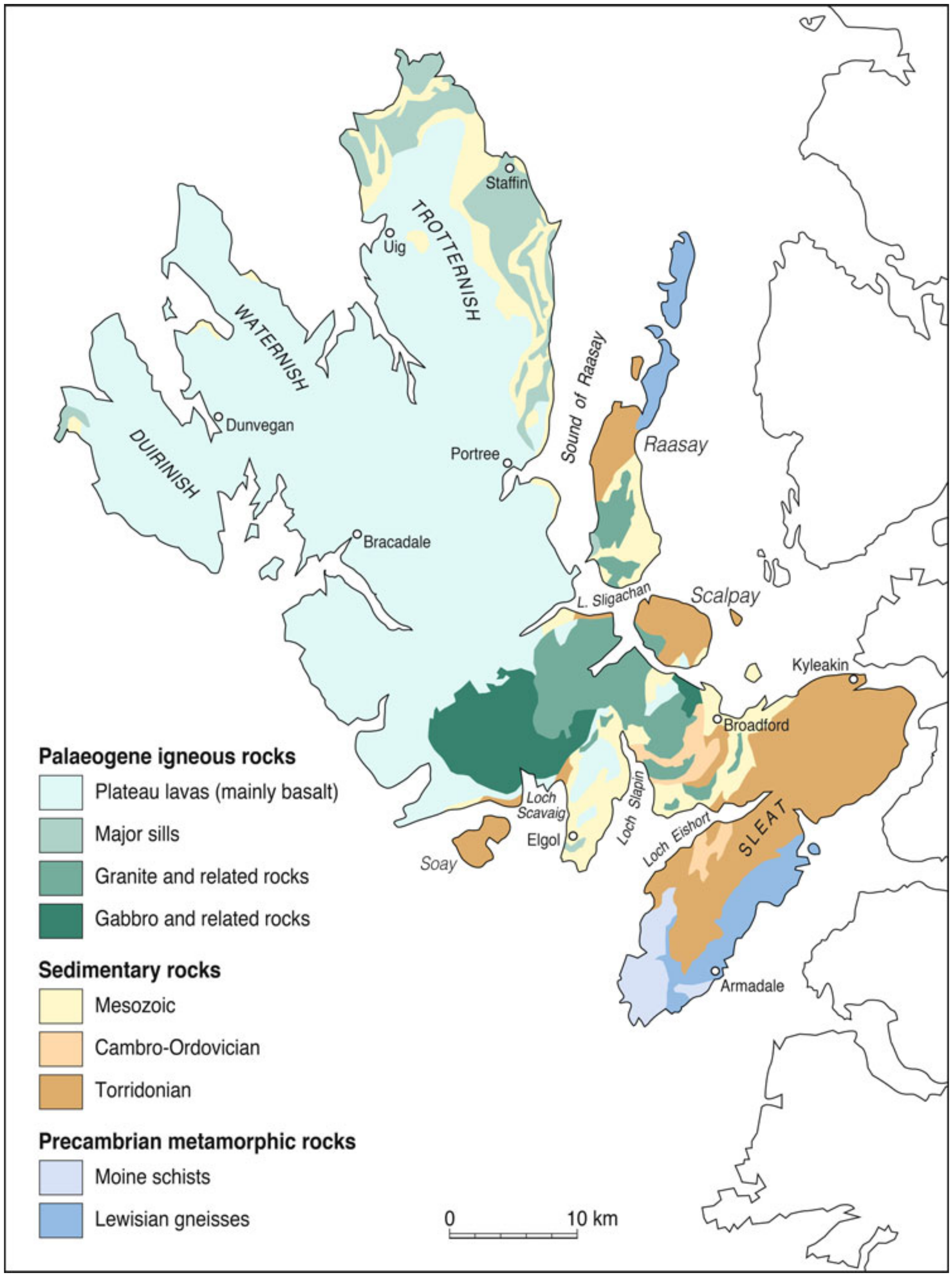

Fig. 10.3 Bedrock geology of Skye

now form domed or conical peaks (Fig. 10.4a, b). The mafic and ultramafic rocks of the Rùm Cuillin form stepped pyramidal peaks, whereas the microgranite that underlies the Western Hills forms a rolling plateau. The fine-grained inner granite of northern Arran forms rolling, domed summits, but the more resistant coarse-grained outer granite forms ridges and peaks that almost rival the Skye Cuillin in topographic ruggedness (Fig. 10.4c). On Mull, the highest ground (Ben More, 966 m) is underlain by the younger lavas that erupted during caldera subsidence.

The extensive basalt lavas of Skye and Mull form the most striking examples of glacially scoured trap topography in Scotland. Over much of the lava terrain, glacial erosion has focused along the tuffs, breccias and laterite horizons that separate near-horizontal lava flows, creating a stepped topography of low hills encircled by lava scarps up to 30 m high, most notably on the Duirinish Peninsula of Skye, where they culminate in the twin ziggurats of MacLeod's Tables (488 m and 468 m). On the Trotternish Peninsula, however, the stacked lavas dip to the west, terminating eastwards in the Trotternish escarpment (Fig. 10.2a), which is crowned by several summits over 500 m and reaches its apogee in The Storr (719 m). The sills and sedimentary rocks that surround the lava fields and central igneous

Fig. 10.4 Mountains of Skye and Arran. **a** The Cuillin Hills on Skye, underlain by gabbro and eucrite. The highest summit shown is Sgùrr na Banachdich, 965 m. **b** The granite Red Hills, Skye. **c** Mountains of the outer (coarse-grained) granite in north Arran. The highest summit is Goat Fell, 874 m. (Images: Colin Ballantyne)

complexes rarely form high ground, except in eastern Skye where sandstones of the Torridon Group rise to the twin summits of the Kyleakin Hills (739 m and 733 m), and southern Arran, where hills up to 500 m are underlain by sills or sandstone.

10.2.3 Columnar Jointing

Few geological formations in the Hebridean Igneous Province have attracted more admiration (and tourists) than the columnar jointing exposed in coastal cliffs, particularly on the tiny island of Staffa. The island is composed of four olivine-tholeiite basalt lava flows overlying a basement of tuff, and superb columnar jointing is present in the lowermost flow, which is overlain by an entablature zone of complex jointing (Fig. 10.2b). The development of columnar jointing on Staffa is attributable to convective cooling of the lava by water (Phillips et al. 2013). Inset into the lava colonnades is the most famous sea cave in Scotland, Fingal's Cave, 20 m high and 75 m deep. The cave and its spectacular columnar jointing have inspired a painting by Turner, an overture (*The Hebrides*) by Mendelssohn, and enthusiastic accounts by several poets (Scott, Keats, Wordsworth and Shelley) and even Queen Victoria (Gordon 2012). Fine columnar jointing also occurs in basalt cliffs elsewhere on the coast of Mull, particularly on Ulva and the Ardmeanach peninsula.

Similar columnar jointing is evident in a sea cliff near Lealt in Skye (Fig. 10.1). Aptly named Kilt Rock on account of its pleated appearance, the vertical columnar jointing in the upper part of the cliff is developed in a dolerite sill that rests on Jurassic sandstones. Here the columnar jointing is thought to have developed through slow cooling of the sill from its contacts with the rocks into which it was intruded. The overlying sandstones have been removed by erosion, but the sill forms a caprock that has protected the underlying sandstones from a similar fate. Columnar jointing in sill rocks is also conspicuous on Arran, most notably at the margins of the quartz-felspar-porphyry sill that underlies The Doon, a prehistoric fort near Drumadoon Bay.

10.3 Glacial History

Although there is evidence for pre-Late Devensian sediments on the Hebrides Shelf (Fyfe et al. 1993; Dove et al. 2015), no terrestrial Pleistocene deposits predating the local Last Glacial Maximum (LLGM) have been found on any of the four islands. During the advance, culmination, retreat and demise of the last Scottish Ice Sheet (~35–14 ka), the islands experienced a similar glacial history. As they all nourished glaciers during the Loch Lomond Stade (~12.9–11.7 ka; Sect. 10.4), it seems inevitable that independent icefields or ice caps formed on each island during the initial phase of ice-sheet expansion (~35–32 ka). On Skye, Mull and Arran, the absence of mainland erratics from the main mountain areas indicates that local ice caps persisted throughout the lifetime of the last ice sheet, forming centres of ice dispersal that were confluent with ice flowing from the Scottish mainland; Rùm, however, was over-run by ice from the western Highlands. During the LLGM, ice from central Skye, Mull and north Arran fed ice into the Hebrides Ice Stream, which terminated at the shelf break northwest of Ireland at ~26.7–25.9 ka (Dove et al. 2015; Callard et al. 2018; Fig. 10.5). The Skye Ice Cap also fed the Minch Ice Stream, which terminated at or near at the shelf edge at

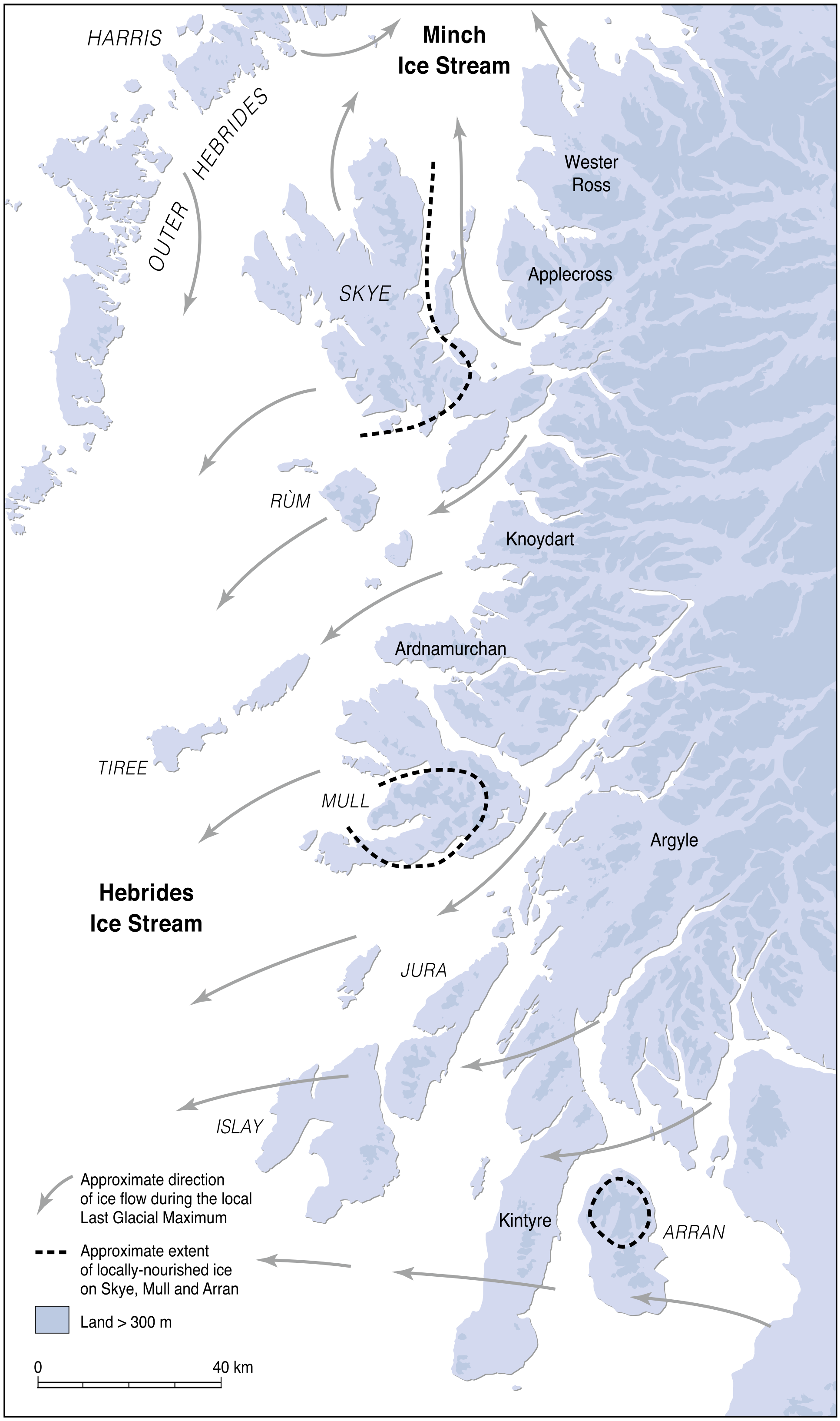

Fig. 10.5 Approximate directions of ice movement in western Scotland and the adjacent shelves during the local Last Glacial Maximum (~30–26 ka)

~30.2–27.5 ka (Bradwell et al. 2019). Retreat of these ice streams resulted in decoupling of local icefields or valley glaciers from mainland ice within the period ~17.0–15.5 ka.

10.3.1 The Last Ice Sheet on Skye and Rùm

On Skye, an independent ice cap covered much the island before being partly encircled by ice advancing from the east (Ballantyne et al. 2016a). Erratics of mainland rocks are restricted to eastern Skye, the Sleat Peninsula and islands of Scalpay and Raasay, but are absent from the mountains of central Skye and the northern peninsulas, suggesting that these areas remained within the domain of the Skye Ice Cap. The distribution of mainland erratics implies that the ice cap diverted mainland ice both northwards into The Minch to feed the Minch Ice Stream and southwestwards to feed the Hebrides Ice Stream (Fig. 10.5).

Terrestrial cosmogenic nuclide (TCN) exposure ages from various sites allow the chronology of the ensuing deglaciation of Skye to be reconstructed. These indicate that the Trotternish hills in northern Skye were deglaciated by ~16.5 ka, Raasay by ~16.2 ka and a site near Strollamus in central Skye at ~16.0 ka (Stone et al. 1998; Small et al. 2012; Bradwell et al. 2019), suggesting that the mainland ice sheet decoupled from the northern margin of the Skye Ice Cap at ~16.0 ka. At several locations there is stratigraphic evidence for expansion of mountain glaciers on Skye following retreat of the mainland ice (Benn 1997). Samples from boulders on readvance moraines in lower Glen Brittle have yielded a mean ^{36}Cl exposure age of ~17.6 ka (Small et al. 2016), but it is likely that the youngest age of 15.5 ± 1.4 ka more accurately reflects of the timing of the readvance. A readvance of Skye ice is also represented by a submarine end moraine that loops around the mouth of Loch Scavaig and onlaps the island of Soay. TCN dating of boulders on the Soay moraine places the culmination of the Loch Scavaig Readvance at 15.2 ± 0.9 ka, approximately coeval with the Wester Ross Readvance on the adjacent mainland (Small et al. 2016; Ballantyne and Small 2019; Chap. 13). The apparent contemporaneity of the Wester Ross Readvance, Loch Scavaig Readvance and (possibly) the readvance represented by the Glen Brittle moraines suggests that late-stage expansion of the residual Skye Ice Cap and mainland ice margin were driven by a pronounced cooling at this time that has been detected in North Atlantic deep-ocean cores (Scourse et al. 2009).

On Rùm, westward-aligned striae and roches moutonnées indicate that the island was over-run by the mainland ice sheet during the LLGM, though basal ice movement was deflected around the mountains of the Rùm Cuillin. Erratics from the mainland are scattered over the north and centre of the island and occur near the summit of Barkeval (591 m) in central Rùm. Ice-moulded bedrock occurs up to 695 m, but the highest summits support shattered rock and periglacial blockfields (Ballantyne and McCarroll 1997), suggesting that they were occupied by a persistent cover of cold-based ice. The timing and pattern of deglaciation on Rùm has not been established, though it is likely that local glaciers persisted on Rùm for some time after retreat of the offshore ice margin.

10.3.2 The Last Ice Sheet on Mull

On Mull, the evidence provided by striae and the distribution of mainland erratics indicates that during the LLGM a persistent ice divide over the central mountains diverted ice from the mainland westward along the Sound of Mull, and southwestward down the Firth of Lorn (Figs. 10.1 and 10.5). Ice from the mainland eventually over-ran all of northern and eastern Mull, as well as the Ross of Mull and Iona in the extreme southwest. Ice-moulded bedrock occurs up to 760 m in the area occupied by the Mull Ice Cap, but the summit of the highest mountain (Ben More, 966 m) hosts a periglacial blockfield of basalt boulders (Ballantyne 1999). The alignment of streamlined submarine bedforms on the shelf adjacent to Mull demonstrates that it occupied the onset zone of the Hebrides Ice Stream, feeding ice to the shelf edge (Dove et al. 2015). The Sea of the Hebrides southwest of Mull was extensively deglaciated by ~20.2 ka (Callard et al. 2018), but for the following ~3 ka the ice margin oscillated in a narrow corridor between Tiree and Mull, backstepping to the Ross of Mull at ~17.5 ka (Small et al. 2016). Over the following millennium the margin of mainland ice gradually retreated, probably leaving residual glaciers occupying the mountains of Mull until the climate warmed at the onset of the Lateglacial Interstade (~14.7 ka).

10.3.3 The Last Ice Sheet on Arran

The history of ice-sheet build-up and changing flow directions on Arran has been reconstructed by Finlayson et al. (2014) from the evidence provided by sequential flowsets derived from streamlined bedforms on Arran and the neighbouring Kintyre Peninsula. During the early stages of ice advance (~35–32 ka), glaciers nourished in the mountains of north Arran merged with southward flow of ice from the SW Highlands down the Firth of Clyde. Subsequent development of an ice divide to the east resulted in westward flow of ice across Arran and Kintyre to feed the Hebrides Ice Stream (Fig. 10.5). Finlayson et al. (2014) suggested that westward ice flow across Arran persisted to ~18.0–17.0 ka, after which southwards flow was re-established and the ice margin advanced into NE Ireland, possibly at ~16.5 ka. The final stage of ice-sheet deglaciation on Arran involved progressive decoupling of glaciers radiating from the mountains of north Arran from southward-flowing ice in the Firth of Clyde. End moraines near valley mouths suggest that glaciers occupying the glens of north Arran readvanced after the retreat of the Clyde ice (Ballantyne 2007). TCN exposure ages obtained for two granite erratics on lateral moraines at Dougarie in west Arran suggest that the readvance of local ice occurred at ~16.2 ka or earlier (Finlayson et al. 2014), but the representativeness of these ages is uncertain (Ballantyne and Small 2019).

Allochthonous erratics occur only along the coastal margins of north Arran, suggesting that during and after the LLGM an ice dome centred over the mountains of northern Arran diverted and fed into the mainland ice sheet. Glacially emplaced perched boulders on several summits suggest that ice over-ran all of the highest ground, but the presence of unmodified tors along ridge crests suggests that the overlying ice was cold-based throughout much or all of the duration of the last ice-sheet glaciation (Finlayson et al. 2014).

10.3.4 The Loch Lomond Readvance

The term 'Loch Lomond Readvance' (LLR) describes the final glaciation in Scotland, which culminated during the Loch Lomond (≈Younger Dryas) Stade of ~12.9–11.7 ka, following the disappearance or near-disappearance of glacier ice under the comparatively temperate conditions of the Lateglacial Interstade (~14.7–12.9 ka). Though it is possible that glaciers persisted on high plateaux and cirques on mainland Scotland during the interstade, it is likely that glaciers disappeared completely from the Hebrides and Arran at this time, as cirque floors are at lower altitudes than those on the mainland, and extensive high plateaux are absent.

The earliest account of a readvance of mountain glaciers in the Hebrides is that of the pioneering Scottish glaciologist James Forbes (1846), who recorded striated and ice-moulded bedrock in the Cuillin Hills of Skye and published a map of an end moraine fronting Coir' a' Ghrunnda in the southern Cuillin. Forbes' account was the forerunner of numerous observations of glacial landforms produced by the readvance of mountain glaciers on Skye, Rùm, Mull and Arran, but until the 1970s no systematic attempt was made to map the landforms associated with these glaciers or reconstruct their former dimensions. The landforms relating to the readvance were later mapped or re-mapped on all four islands (Ballantyne and Wain-Hobson 1980; Ballantyne 1989, 2002, 2007) and employed to delimit its former extent and to reconstruct the three-dimensional form of the readvance glaciers.

The evidence provided by end and lateral moraines, recessional moraines, drift limits and trimlines indicates that during the Loch Lomond Stade substantial icefields developed on the mountainous areas of Skye and Mull and fed outlet glaciers that terminated on land or a short distance offshore (Fig. 10.6). The Cuillin Icefield on Skye occupied an area of ~156 km^2 and the Mull Icefield extended across ~143 km^2. On both islands the central icefield was flanked by cirque glaciers that ranged in size from ~0.4 to ~6.0 km^2. Outlying glaciers also formed on Skye in the form of a small icefield in the Kyleakin Hills, one or two cirque glaciers at the foot of the Trotternish escarpment in northern Skye (Ballantyne 1990) and a small glacier in Duirinish (Ballantyne and Benn 1994). A Loch Lomond Stadial age for the readvance glaciers on Skye and Mull is indicated by stratigraphic evidence: Lateglacial sediments occur in kettle holes and other enclosed basins outside the readvance limits but are absent from similar sites within these limits (Walker and Lowe 1982, 1985; Walker et al. 1988; Benn et al. 1992). Cosmogenic ^{10}Be exposure dating of boulders on readvance moraines at two sites on Skye yielded mean ages of 12.5 ± 0.7 ka and 12.4 ± 0.6 ka (Small et al. 2012; recalibrated in Ballantyne et al. 2016b), suggesting that some readvance glaciers reached their maximum extent in mid-stade. On Mull, a Loch Lomond Stadial age for the readvance is confirmed by calibrated radiocarbon ages of ~13.9–12.6 ka obtained for marine shell and barnacle fragments recovered from till at the western limits of the readvance near Loch Spelve (Bromley et al. 2018).

On Rùm and Arran there is evidence for the formation of only small cirque or valley glaciers during the Loch Lomond Stade and none of these apparently exceeded 2.5 km^2 in extent. There is evidence for eight or nine former small glaciers in the Rùm Cuillin, two below the north-facing cirques of the Western Hills on Rùm, and ten in the mountains of north Arran. The readvance moraines on Rùm and Arran have not been dated, and a Loch Lomond Stadial age for the glaciers that formed them has been inferred mainly from absence of Lateglacial periglacial landforms inside the mapped readvance limits.

10.4 Glacial Landforms

10.4.1 Large-Scale Erosional Landforms

Although the islands contain a wealth of depositional landforms and small-scale erosional features that developed during the last (Late Devensian) glacial stage, large-scale landforms such as glacial troughs and breaches, cirques (corries), rock basins, fjords and ice-scoured terrain have a much longer history of development, potentially extending back to the beginning of the Pleistocene at ~2.6 Ma. During

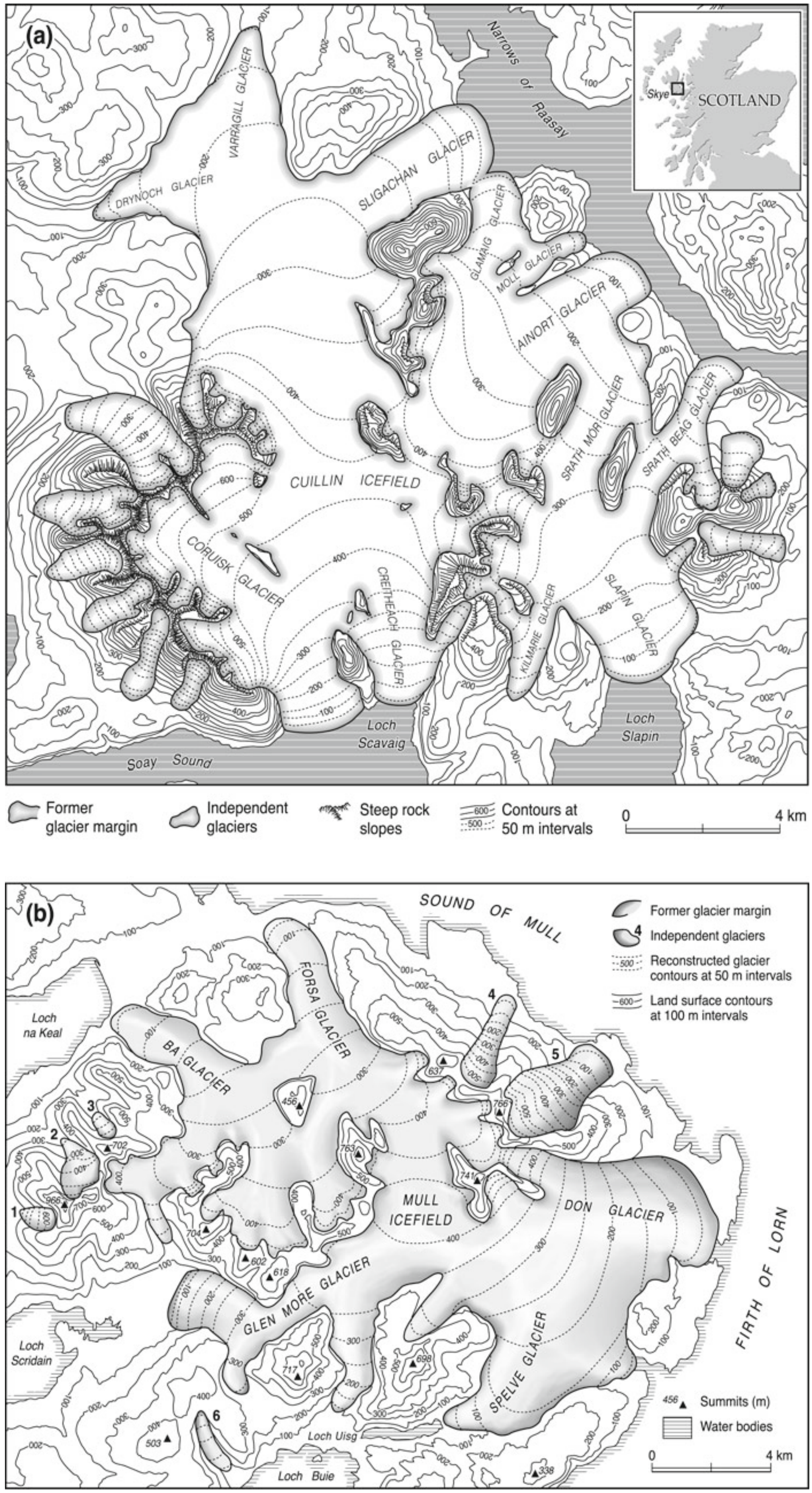

Fig. 10.6 Contoured reconstructions of the maximum extent of Loch Lomond Readvance glaciers in **a** south-central Skye and **b** Mull. (**a** is adapted from Ballantyne (1989) © Longman Group UK Ltd 1989; **b** is reproduced from Ballantyne (2002) © 2002 John Wiley & Sons Ltd)

successive glacial episodes, erosion by mountain glaciers and ice sheets has progressively widened and deepened the pre-glacial valley system, a process aided by rockfall and landslides from the rockwalls of glacial troughs and cirques during intervening interglacials. One consequence of such long-term glacial erosion has been the development of glacial breaches that severed Mull and Skye from the mainland (Sissons 1983).

Glacial troughs on the four islands tend to radiate away from the main mountain masses (Fig. 10.1). This is particularly evident in central Skye, where troughs terminate in fjords that radiate away from the highest ground. A similar radial pattern of troughs is evident in the mountains of Mull and north Arran. Although this pattern is almost certainly inherited from a pre-glacial valley system formed by rivers draining away from the uplifted central igneous complexes, it also suggests that during successive glacial stages radial ice movement predominated. It is notable that glacial breaches on the islands, such as the Srath Mòr and Glen Sligachan–Srath na Creicheach valleys on Skye, are aligned athwart the westward direction of ice flow from the mainland. The most elegant troughs are those developed in granite, such as Srath Mòr on Skye and Glens Sannox and Rosa on Arran (Fig. 10.7a), which in cross-section describe almost perfect parabolas, reflecting the isotropic nature of the rock. Inland rock basins are few and relatively shallow: Loch Coruisk in the Skye Cuillin is no more than 38 m deep (Fig. 10.7b) and Loch Bà on Mull has a maximum depth of 44 m; the largest rock basin is that occupied by Loch Frisa on Mull, 7.2 km long and up to 62 m deep, which is aligned SE–NW along the axis of a dyke swarm and formed a major artery of ice flow across northern Mull.

The numerous cirques within and at the margins of the central igneous complexes and the Northern Granite of Arran also reflect radial outflow of former glaciers, being oriented at all aspects. This radial tendency is strikingly illustrated by the great arc of cirques of the Cuillin Hills on Skye but can also be detected on Mull and Arran (Fig. 10.6). Most cirque floors on Skye, Mull and Rùm are below 400 m, reflecting a westward decline in cirque altitude across Scotland, and are flanked by steep, cliffed headwalls and sidewalls that are skirted by talus accumulations that represent rockwall retreat since the disappearance of the last glaciers (Fig. 10.7c). Three cirques deserve particular mention. Coire Làgan is the archetype of the Cuillin cirques, being surrounded by precipitous rockwalls, flanked by bouldery talus cones, and containing a rock basin; the lip of the cirque supports the finest examples of ice-moulded and striated whalebacks in Scotland (Fig. 10.7d). Coire Cuithir is a broad, low-lying (<200 m elevation) embayment at the foot of the Trotternish escarpment, containing a shallow loch that is dammed by a moraine deposited by a LLR glacier (Fig. 10.8). The floor of the cirque below Sròn an t-Saighdeir in NW Rùm is almost entirely covered by large microgranite boulders that extend in a series of transverse ridges to the terminal LLR moraine, suggesting that the last glacier to occupy the cirque was fed by an extraordinary volume of rockfall debris from the headwall.

The extensive lowland areas of northern Skye and much of Mull that are underlain by stacked lava flows form a distinctive glaciated trap topography of low, stepped, flat-topped hills flanked by lava scarps up to 30 m high. The most impressive landscape of such ice-scoured trap topography is that of the Duirinish peninsula in NW Skye, which is dominated by the flat-topped hills known as MacLeod's Tables, Healabhal Mhòr (468 m) and Healabhal Beag (488 m). By contrast, the low-lying Devonian granite intrusion that forms the westernmost part of the Ross of Mull forms an area of knock-and-lochan topography, where low, ice-scoured hills, whalebacks and roches moutonnées alternate with water- and peat-filled hollows excavated by ice along fracture zones within the granite.

10.4.2 Landforms of Glacial and Glacifluvial Deposition

The most impressive end moraines on the islands are those that formed at the limits of small LLR cirque glaciers. Examples include that below Coir' a' Ghrunnda in the Skye Cuillin, the two end moraines marking the limits of cirque glaciers in the Western Hills of Rùm and the bouldery moraines that record the extent of two small glaciers at the head of North Glen Sannox on Arran (Fig. 10.7e). Elsewhere, limits of cirque glaciers are often marked by arcs of scattered boulders, as at the head of Glen Rosa on Arran (Fig. 10.7f). The development of pronounced end moraines appears to have been conditioned by the amount of sediment carried by the parent glaciers, which in turn appears to reflect the volume of debris deposited on glacier surfaces by rockfalls and avalanches or reworked from talus that accumulated prior to the readvance (Benn 1989). This relationship is strikingly illustrated by the end moraine that forms the limit of a small glacier that occupied Coire Fearchair in the Red Hills of Skye during the Loch Lomond Stade: a massive drift ramp up to 20 m high is located downslope from a rock-slope failure scarp, suggesting that one or more rockslides or major rockfalls supplied the debris that was transported by the glacier to build the ramp (Ballantyne et al. 2016b).

Recessional moraines occur only within 200–300 m of the former termini of most LLR cirque glaciers, suggesting that the termini of these glaciers oscillated close to their limits for a prolonged period before experiencing uninterrupted retreat. Conversely, multiple recessional moraines occur within the limits of the outlet glaciers of the Skye and

Fig. 10.7 Glacial landforms of Skye and Arran. **a** Glacial trough in granite, Glen Sannox, Arran. **b** Lock Coruisk, which occupies a shallow (<40 m deep) rock basin excavated in gabbro, Skye. **c** Coir' a'Ghrunnda, Cuillin Hills, Skye. **d** Striated ice-moulded bedrock whaleback, Coire Làgan, Skye. **e** End moraine of granite boulders at the head of North Glen Sannox, Arran. **f** Arcuate spreads of boulders define the limit of a small glacier that occupied the head of Glen Rosa, Arran, during the Loch Lomond Stade. (Images: Colin Ballantyne)

Mull Icefields (Benn 1992; Ballantyne 2002). In some areas such moraines extend back to glacier source areas, notably on Mull and in the Kyleakin Hills on Skye, indicating that ice-margin retreat was interrupted by minor readvances until the glaciers had shrunk to their sources. In central Skye, however, multiple hummocky recessional moraines occur within a broad zone inside the limits of some of the larger outlet glaciers but are rare or absent farther upvalley. Benn et al. (1992) interpreted this pattern as representing an initial period of slow oscillatory retreat caused by reduction in snowfall, followed by uninterrupted ice recession and localised ice stagnation initiated by rapid warming after ~11.7 ka. However, July temperature reconstructions based on Lateglacial chironomid assemblages at various sites in

Fig. 10.8 Coire Cuithir, a deep embayment in the Trotternish escarpment of northern Skye. The floor of the cirque is only 180 m above sea level. The cirque was occupied by a small ($\sim$1.7 km^2) LLR cirque glacier that deposited the end moraine on the left. (Image: Colin Ballantyne)

Scotland indicate gradual summer warming during the later part of the stade (Brooks et al. 2016), suggesting that initial pulsed retreat may also represent a response to gradual increases in ablation-season temperatures.

Outwash deposits associated with the Loch Lomond Readvance are fragmentary on Skye, Rùm and Arran but well represented on Mull by broad terraces of kettled outwash 10–20 m above present sea level at the outlet of Loch Bà and in lower Glen Forsa. These features indicate that relative sea level was below $\sim$10 m when the glaciers occupying these locations achieved their maximum extent (Gray and Brooks 1972).

10.5 Periglacial and Aeolian Landforms

10.5.1 Relict Periglacial Features

Relict periglacial phenomena include both features of pre-Late Devensian age that escaped glacial erosion during the LLGM because they were covered by cold-based ice (Fabel et al. 2012; Fame et al. 2018) and Lateglacial landforms that developed after ice downwastage exposed mountains to frost action on ground underlain by permafrost. Of the former, the most widely represented are summit blockfields. Some are expanses of openwork, frost-heaved boulders, as on the basalt summit of Ben More on Mull and the microgranite summits of the Western Hills on Rùm. Others are composed of clasts embedded in fine sediment, particularly on the granite summits of northern Arran and Skye. On Arran, granite tors are studded along ridge crests (Fig. 10.4c), notably those at the summit of Caisteal Abhail (859 m), and low tors rise above basalt blockfields along the Trotternish hills on Skye. In some areas a trimline representing the upper limit of glacial erosion marks the lower limit of blockfield debris, notably on the Trotternish hills (Ballantyne 1990). Although such trimlines were originally interpreted as indicating the maximum altitude of the last ice sheet, they are now considered to represent an englacial boundary between warm-based sliding ice on lower ground and cold-based ice on summits, implying that the last ice sheet was polythermal (Fame et al. 2018). Above-trimline bedrock surfaces on the Trotternish hills have yielded apparent ^{36}Cl exposure ages >50 ka, indicating limited erosion by the last ice sheet, whereas bedrock outcrops on cols below the trimline have produced (recalibrated) exposure ages of $\sim$16.5 ka, consistent with the timing of ice-sheet deglaciation (Stone et al. 1998).

During the Lateglacial period, frost debris moved downslope to form bouldery stone-banked lobes. These are most abundant on moderate upper slopes underlain by granite in Skye and Arran, but also occur on high ground underlain by sandstone on the Kyleakin Hills of Skye and on both felsic and mafic igneous rocks on Rùm. On Arran, they terminate at or close to the limits of LLR glaciers (Ballantyne 2007), indicating that movement ceased at the Lateglacial–Holocene transition ($\sim$11.7 ka) as the climate warmed and permafrost degraded.

10.5.2 Active (Holocene) Periglacial Landforms

Active periglacial landforms on the four islands are best represented on high ground underlain by lavas that have weathered to produce silt-rich, frost-susceptible soil. Shallow active solifluction lobes, ploughing boulders and frost-sorted circles and stripes up to 0.6 m wide occur above ~580 m on the basalt lavas of The Storr and Hartaval in northern Skye and Ben More on Mull (Godard 1965). Miniature frost-sorted nets and stripes produced by growth and collapse of needle ice occur on bare ground underlain by various igneous rocks on Skye, mostly above 600 m. Earth hummocks (thúfur) up to 0.3 m high are developed on aeolian deposits near the summit of The Storr in Trotternish, and locally grade into nonsorted relief stripes that run obliquely across gentle slopes. As the parent aeolian deposits are of Holocene age (Sect. 10.5.5), such features may still be actively developing, though whether they represent the products of frost heave or selective erosion by surface wash (or both) is unknown.

10.5.3 Periglacial Landscapes of the Western Hills of Rùm

The rolling plateau and surrounding slopes of the Western Hills of Rùm (Sròn an t-Saighdeir, 522 m and Orval, 571 m) occupy an area of only ~5 km^2 yet contain a greater range of periglacial features than any comparable mountain area in Scotland (Godard 1965; Ryder and McCann 1971; Ballantyne 1993; Fig. 10.9). To a large extent this reflects geological contrasts: Sròn an t-Saighdeir and the surrounding slopes are underlain by densely-jointed microgranite, which has weathered to produce an openwork blockfield of boulders up to ~1.2 m long, whereas much of Orval is underlain by hawaiite lavas that support a regolith of smaller clasts embedded in frost-susceptible soil.

On Sròn an t-Saighdeir the summit blockfield is flanked by cirques to the north and debris-mantled slopes (blockslopes) on other aspects. West of the summit, the latter extend down to 270 m, where they terminate at the crest of sea cliffs. Former downslope movement of debris by

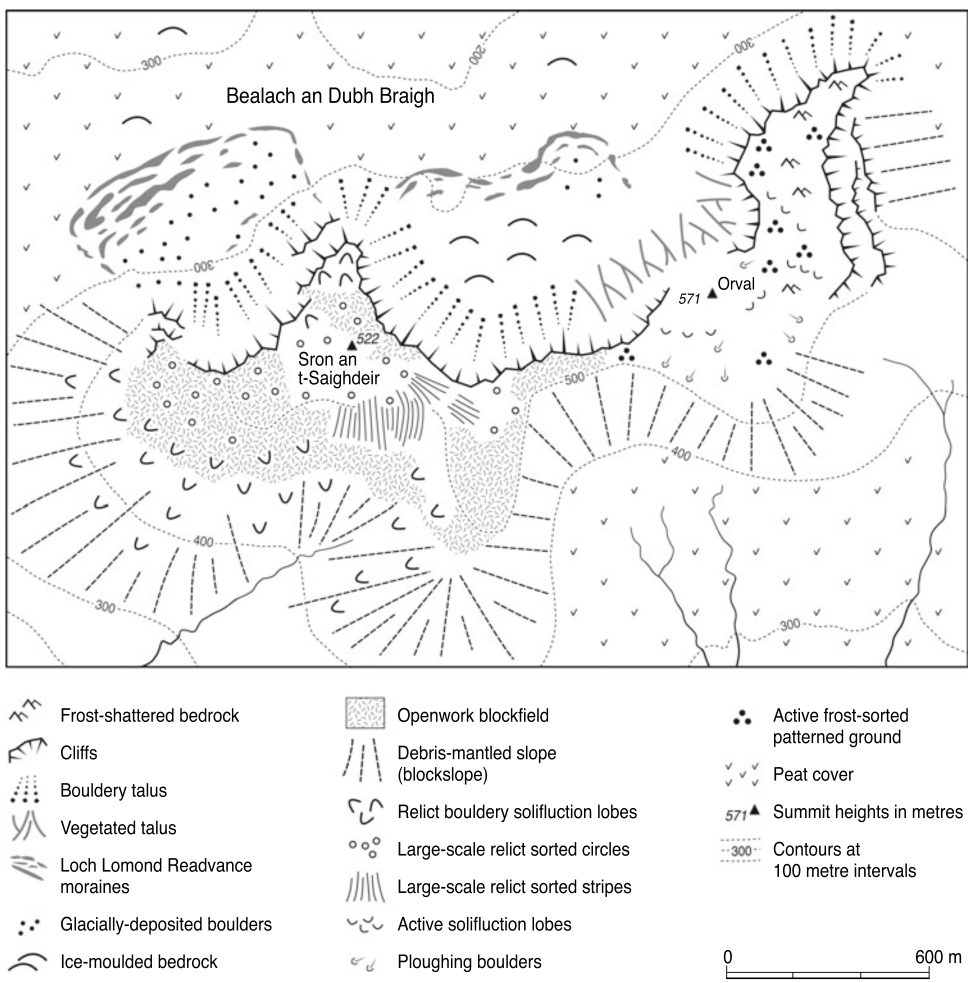

Fig. 10.9 Glacial and periglacial landforms on the Western Hills of Rùm (Adapted from Ballantyne 1984)

solifluction operating over permafrost is evident in the form of relict boulder terraces and lobes that extend down to ~380 m. The plateau and adjacent gentle slopes also support relict frost-sorted circles and nets 2.0–3.0 m in diameter, and sorted stripes of similar width, now represented by vegetation-covered cells or stripes of clasts embedded in fine sediment alternating with borders or stripes of openwork boulders. Such large-scale relict sorted patterns are thought to develop through differential frost heave in the active layer above permafrost (Ballantyne and Harris 1994; Ballantyne 2018) and thus provide evidence for permafrost during the Lateglacial period.

By contrast, the frost-susceptible regolith on the lavas adjacent to Orval supports a range of active periglacial features, including vegetated solifluction lobes with low, steep risers, ploughing boulders and small-scale frost-sorted patterned ground. The sorted patterns mainly occur on deflation scars above 500 m, and include sorted circles, polygons and nets up to 0.5 m in diameter and miniature sorted stripes, 0.2–0.3 m in width, formed by differential growth of needle-ice during periods of superficial ground freezing and enhanced by frost heave of the finer sediment during periods of deeper freezing (Ballantyne 2018).

10.5.4 Aeolian Landforms

The effects of wind erosion are evident on high ground on all four islands. Deflation surfaces (expanses of boulders and lag gravels from which fine soil and vegetation cover have been stripped by wind) occur on the Red Hills and Trotternish hills on Skye, and smaller deflation scars carpeted by lag gravels are present on plateaux, ridge crests and cols above 450–500 m, notably on Ben More on Mull and the Trotternish hills. On some upper slopes, regularly-spaced wind stripes (alternating bands of vegetated and unvegetated ground) have been transformed into flights of turf-banked terraces, where bands of vegetation aligned across-slope have arrested downslope creep of debris to form horizontal step-like terraces with vegetated risers up to a metre high. On Skye, turf-banked terraces are widespread above 550 m on the Red Hills, and north of the summit of Hartaval (668 m) in Trotternish. An uncannily regular suite of turf-banked terraces comprising vegetated risers 0.3–0.6 m high and perfectly horizontal treads covered in small clasts occurs at ~600 m on the Ruinsival ridge in SW Rùm, probably the most perfect example of this type of landform on any Scottish mountain.

Aeolian deposits in the form of vegetation-covered sand sheets up to a metre thick occur on several mountains in the Inner Hebrides, notably on the lavas of Ben More on Mull, the mafic and ultramafic rocks on Rùm, and the granitic rocks that underlie the Red Hills on Skye. Some of those on the Red Hills consist of a lower unit of weathered, windblown sand overlain by a unit of fresh sand. The latter represents recent widespread erosion of soils and sand deposits from upwind deflation surfaces; luminescence dating of these deposits has shown that the onset of upper-unit sand deposition coincided with the exceptionally stormy conditions of the Little Ice Age of the sixteenth to nineteenth centuries (Morrocco et al. 2007).

10.5.5 Aeolian Deposits on the Storr

The most remarkable high-level aeolian deposits in the Hebrides are those that blanket the summit of The Storr (719 m) in northern Skye. Here the highest ground is covered by a cap of vegetation-covered windblown sediment up to 2.9 m thick that covers an area of 33,000 m^2 (Fig. 10.10). Both the thickness and particle size of the deposits decline away from the edge of the adjacent cliff. These trends suggest that grains released from the cliff by weathering have been entrained by strong winds that accelerated upwards on meeting the cliff face, and that deceleration of airflow above the cliff caused the suspended particles to rain down on the summit plateau, where they were trapped by vegetation cover, slowly accumulating to form the summit sand sheet (Ballantyne 1998). The cliff that formed the source of the windblown sediment was exposed by the Storr landslide at 6.1 ± 0.5 ka (Sect. 10.6.2), and radiocarbon dating of organic material from the base of the sand sheet has shown that it began to accumulate sometime between 7.1 ka and 5.7 ka, suggesting that exposure of the cliff initiated sediment accumulation on the summit. Radiocarbon dating of organic material within the windblown sediments has allowed calculation of average rates of sand accumulation, which ranged from about 14 mm per century near the downwind edge of the sand sheet to roughly 60 mm per century in the thickest deposits near the cliff crest.

10.6 Postglacial Landslides, Talus Accumulations and Debris Flows

Large inland postglacial rock-slope failures are rare on most lithologies on the four islands. Two rockslides have been identified on the Torridonian sandstone of the Kyleakin Hills in eastern Skye, and another occupies the northeast face of Glas Bheinn Mhòr, a granite mountain in the Red Hills. Foundering of basalt lavas that overlie weak sedimentary rocks, however, has produced the largest and most dramatic landslides in the British Isles.

Fig. 10.10 Aeolian sand deposits on the summit of The Storr (719 m), northern Skye. The inset shows a pit, 2.9 m deep, excavated in the deposit. At the base of the pit an organic soil dated to ~7.1 ka overlies frost-weathered regolith. (Images: Colin Ballantyne)

10.6.1 The Trotternish Landslides

The Trotternish Peninsula is dominated by the Trotternish escarpment, which extends north–south along much of the peninsula and comprises stacked lava flows overlying Jurassic sedimentary rocks that crop out to the north and east (Fig. 10.2a). Transgressive dolerite sills occur deep within the Jurassic strata and the entire sequence dips gently westwards. Displaced rock masses extend continuously along the foot of the escarpment. The area of landsliding comprises an inner zone of detached blocks, representing slope failure since retreat of the last ice sheet at ~16.5 ka, and an outer region of undulating terrain consisting of older landslide blocks that have been over-ridden by successive ice sheets (Fig. 10.11). Postglacial rock-slope failure along the escarpment takes several forms. Resting against or detached from the scarp are large, intact lava blocks have experienced lateral movement with limited vertical displacement. Elsewhere, detached lava blocks are back-tilted or tilted away from the scarp face. Not all failures have involved block displacement: others take the form of large-scale rockfalls, sometimes with debris runout, as on the northern spur of The Storr, where a tongue of coarse debris 260 m wide and 500 m long descends from the foot of the scarp (Ballantyne 2016).

Interpretation of the structure of the Trotternish landslides has focused on the Storr and Quiraing landslides. Few landforms in Scotland are more impressive than the Storr landslide—'cette topographie anarchique' as Godard (1965) described it. The landslide receives over 200,000 visitors each year and is probably the most photographed landform in Scotland. The entire south face of The Storr (719 m) has collapsed to create a large hollow, Coire Faoin, bounded by lava cliffs 200 m high. The undercliff zone is a labyrinth of lava blocks, narrow clefts and pinnacles of shattered rock, of which the 49 m high Old Man of Storr is the most impressive (Fig. 10.12a). The dip of the lavas in the zone of postglacial landsliding is eastwards, indicating that failure was dominantly translational, along a failure plane within the underlying Jurassic shales. ^{36}Cl exposure dating of rock samples chiseled from the tops of basalt pinnacles indicates that the landslide occurred at 6.1 ± 0.5 ka (Ballantyne et al. 1998, 2014), about 10,000 years after deglaciation of the site.

The Quiraing landslide complex is the largest in Scotland, covering ~8.5 km^2 and extending ~2.2 km from the scarp crest to the coast. It comprises an inner zone of tabular blocks (Fig. 10.12b), narrow corridors and rock pinnacles formed by failure since deglaciation, and an outer zone of more subdued landslide blocks that represent ancient landslides that occurred before the last (and probably earlier) ice sheet(s) crossed the area. The backscarp and some of the slide blocks have rectilinear geometries controlled by pre-existing discontinuities (Fenton et al. 2015). The deep structure of the Quiraing landslide is complex. Anderson and Dunham (1966) inferred that a thickness of ~200 m of sedimentary rocks and tuff had failed under the weight of ~300 m of stacked lava flows as successive rotational slides seated on a westward-dipping dolerite sill. However, several large landslide blocks have moved laterally from the

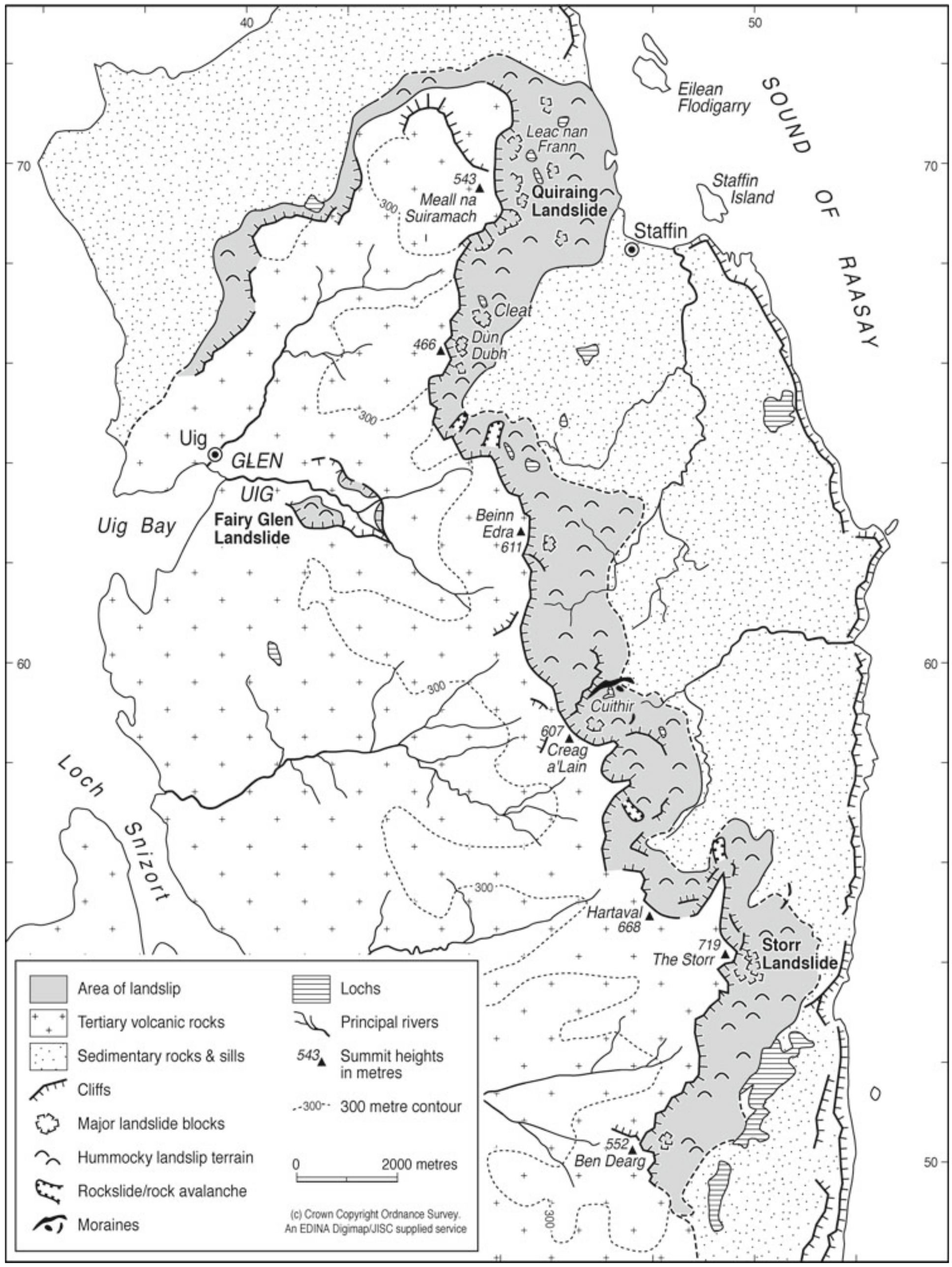

Fig. 10.11 The Trotternish Peninsula, northern Skye, showing the approximate extent of landslide terrain and the location of the Storr, Quiraing and Fairy Glen landslides. (From Ballantyne 2016)

headscarp without back-tilting, leading Martin (2011) and Fenton et al. (2015) to conclude that postglacial landsliding has been dominated by translational sliding of detached blocks over a failure plane within the underlying shales, with toppling of intact rock masses in the gaps created by extensional block detachment and minor rotation of blocks at the margins of the zone of postglacial failure. More speculatively, they suggested that the outer zone of the landslide involved block rotation as a result of reactivation of listric faults in the underlying Jurassic sediments.

The causes of the Storr and Quiraing landslides are incompletely understood. Both may represent a response to paraglacial stress release, but it is possible that the thick sills that are intruded into the underlying shales have acted as aquicludes, allowing high water pressures to build up in the shales, thereby destabilizing the overlying lava pile. At the Quiraing there is evidence that although most of the landslide complex is now stable, occasionally this sleeping giant stirs: there are reports of earth tremors at Flodigarry, on the seaward margin of the complex. Anderson and Dunham (1966, p. 192) considered that 'continuous though not extensive movement' had caused frequent dislocation of the road crossing the distal margin of the complex, and Fenton et al. (2015, p. 182) noted 'areas of ongoing movement,

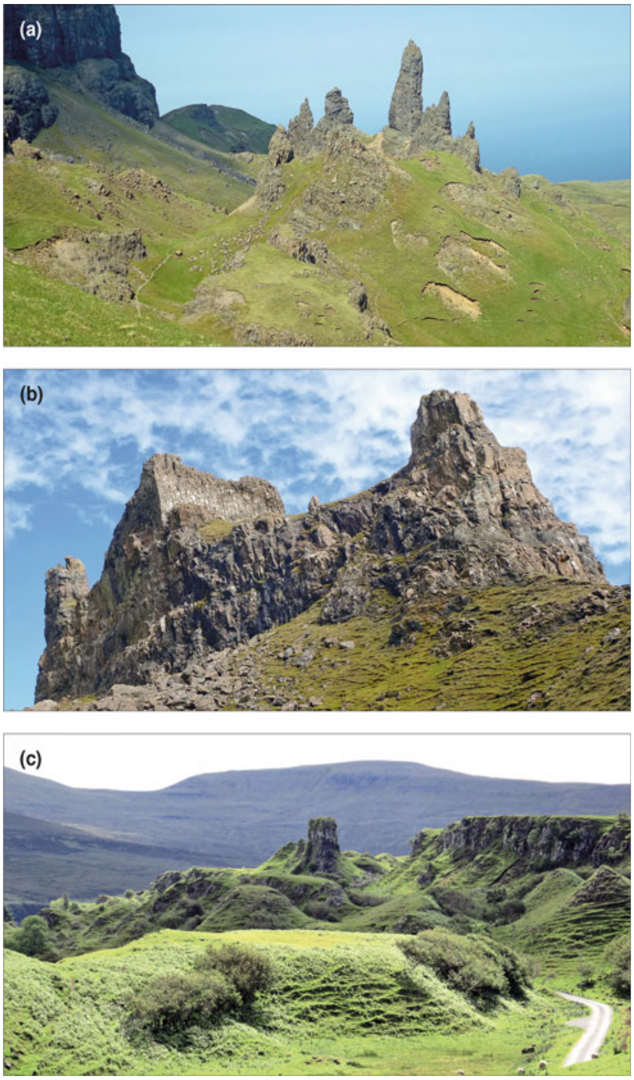

Fig. 10.12 The Trotternish landslides. **a** The Storr landslide. The pinnacled area represents a shattered lava block that slid and tilted eastwards from the cliffs in the background. The highest pinnacle is the 49 m high Old Man of Storr. **b** The Prison, a lava block that forms part of the postglacial failure zone of the Quiraing landslide. **c** Displaced lava blocks of the Fairy Glen landslide. The conspicuous tower is Castle Ewen, which apparently moved laterally from the basalt headscarp on the right with limited vertical displacement. (Images: Colin Ballantyne)

especially where the toe of the slide is expressed in weak, thinly laminated shales'.

The dramatic geomorphology of the Storr and Quiraing landslides has overshadowed that of landslides in Glen Uig, a valley underlain by Jurassic shales in western Trotternish. Here the landslides are located on the dip slope, and the lava headscarps are much lower. The largest (0.5 km^2) of the Glen Uig rock-slope failures is the Fairy Glen landslide on the south side of the glen, which extends up to 370 m northwards from its headscarp. The outer landslide blocks form degraded hummocks that may have been modified by glacial erosion (Fenton et al. 2015) but the inner zone consists of tabular blocks, of which Castle Ewen (Fig. 10.12c) is the most impressive. Murphy (2011) observed that the dip of the lavas in the inner zone is identical to that of the headscarp and concluded that failure had occurred through lateral translational sliding of stacked lavas over the underlying shales, with localized rotation and toppling of segments as gaps opened between detached blocks.

10.6.2 Other Failures of Palaeogene Lavas

The structural configuration of stacked Palaeogene lava flows overlying weaker rocks that has favoured large-scale postglacial rock-slope failure on Trotternish is replicated at several other sites on Skye and on Mull. At Ben Tianavaig, 4 km southeast of Portree on Skye, Jurassic sedimentary rocks ~300 m thick are sandwiched between the overlying lavas and an underlying dolerite sill. The entire seaward face of the mountain has collapsed, forming a landslide complex ~2 km^2 in area that extends over a kilometre from the headscarp to terminate at coastal cliffs. Cheng (2013) has shown that the landslide consists of a series of largely intact, back-rotated lava blocks interrupted by localized toppling failures, and that the planform of the headscarp follows discontinuities (faults and dykes) that formed zones of weakness in the lava pile.

On Mull, the largest rock-slope failure involving Palaeogene lavas is the Coireachan Gorma landslide on Ardmeanach. The failed mass covers ~0.8 km^2 and extends up to 0.5 km from the cliffed headscarp to terminate at coastal cliffs up to 50 m high. The landslide is dominated by four large back-tilted blocks indicative of deep-seated rotational sliding that was probably seated in the underlying schists or sandstones. The interpretation of a large, multi-ridged steep-fronted rampart at the foot of the same lava scarp near Gribun has proved contentious. Dawson et al. (1987) concluded that the rampart is a moraine deposited by a niche glacier during the Loch Lomond Stade, but Ballantyne (2002) argued that the Gribun debris accumulation represents the runout of a postglacial rock-slope failure. Pollen-stratigraphic evidence suggests that the ridge formed prior to ~11.9 ka and possibly prior to ~12.5 ka.

10.6.3 Talus Accumulations and Debris Flows

Talus slopes formed through the incremental accumulation of rockfall debris at the foot of cliffs are common on the islands; most are paraglacial landforms that mainly accumulated soon after deglaciation. The most impressive taluses are those skirting cliffs in the Cuillin Hills, where the coarseness of rockfall debris has inhibited vegetation and soil development. By contrast, the talus slopes that skirt the foot of basalt cliffs, such as those near Gribun on Mull or at the foot of Ben Meabost on Skye, are completely vegetated.

Much of our understanding of the evolution of talus accumulations in Scotland stems from research on the Trotternish escarpment. These taluses support a complete cover of soil and vegetation but are incised by deep gullies, exposing sections that have permitted investigation of their structure, radiocarbon dating of buried organic soil horizons and reconstruction of their evolutionary history (Hinchliffe et al. 1998; Hinchliffe 1999).

The Trotternish taluses form rectilinear slopes resting at 36°–42° with short slope-foot concavities, and ~27–30% of the talus debris comprises fine sediment derived from granular weathering of the source rockwalls. Sections in gully walls have revealed in situ rockfall debris overlain by stacked debris-flow deposits intercalated with slopewash deposits and palaeosols. Radiocarbon dating of these palaeosols has shown that reworking of talus by debris flows has occurred episodically throughout the Holocene. Gully erosion and deposition of debris-flow deposits was apparently initiated at the crest of talus slopes, then extended downslope, ultimately resulting in the deposition of small debris cones at the slope foot. Calculations based on talus volume indicate that 4.3–7.8 m of cliff retreat has occurred since deglaciation (~16.5 ka). It is likely, however, that most of the talus accumulated during the Lateglacial period as a result of joint opening through paraglacial stress release and frost-wedging under severe periglacial conditions prior to ~11.7 ka (Hinchliffe and Ballantyne 1999). During the Holocene, gully erosion has replaced rockfall accumulation as the dominant process operating on the Trotternish taluses.

Recent debris-flow activity on the islands is evident from fresh flow tracks (marginal levées and terminal lobes) on many mountain slopes, particularly on granites that have weathered to produce sandy regolith. By far the highest density of debris-flow tracks (locally more than 30 per kilometre of slope) occurs in the Red Hills of Skye, where tracks originating in bedrock gullies on upper slopes terminate in mid-slope, at slope-foot debris cones or occasionally at valley-floor stream channels (Milne et al. 2015). The east slope of Beinn Dearg Mhòr (731 m) near Loch Ainort is a representative example. At this location at least four rainstorm-triggered debris-flow events occurred in the period 1981–2016, making this and neighbouring areas amongst the most active sites of debris-flow activity in Scotland.

10.7 Coastal Geomorphology

Rock coasts dominate the littoral of Skye, Mull and Rùm, in the form of cliffs, ramps, boulder-strewn intertidal shore platforms and raised rock platforms, sometimes partly buried under Lateglacial or Holocene beach deposits. Sandy beaches and coastal dunes are rare, and machair deposits (windblown sediment composed mainly of shell fragments) are mainly limited to Iona and the Ross of Mull. The coastal geomorphology of Arran is more varied, with fewer coastal cliffs but extensive areas of intertidal or raised shore platforms, and several sandy beaches. Of these, the most interesting are those along the south coast of the island, where numerous Palaeogene basalt dykes have acted as natural groynes to trap small sandy beaches, one of which is sufficiently secluded to be designated one of Scotland's three naturist beaches.

10.7.1 Coastal Landslides

The cliffed coastlines of all four islands are scarred by landslides, including some historic failures attributable to undercutting by wave action, such as that at An Scriodan in northern Arran, which occurred about 300 years ago. The largest coastal landslides are those where lavas or sills overlie weaker rocks, and even some 'inland' rock-slope failures such as the Quiraing, Ben Tianavaig and Coireachan Gorma landslides terminate at the shoreline, where wave erosion is thought to contribute to ongoing local instability (Fenton et al. 2015).

At least 18 major coastal landslides occur along the cliffed coasts of northern Skye. The largest coastal landslide in Scotland is that at Scòre Horan, on the east coast of the Waternish peninsula. The headscarp of this landform coincides with the outcrop of Palaeogene lavas and is over 2 km long and up to 60 m high. The area of failed rock covers ~1.2 km^2 and is mainly underlain by Jurassic siltstone and mudstone, though a dolerite sill underlies the uppermost part of the failed mass and another crops out at its toe. The inner part of the slide consists of back-tilted blocks, but whereas those abutting the headscarp are angular, most are rounded, indicating that they have been modified by glacial abrasion and that postglacial failure has been limited to a narrow zone below the headscarp. A Lateglacial raised shore platform at 12.5–14.0 m OD (Ordnance Datum) cuts across the landslide toe and is overlain by raised beach deposits (Richards 1971). Another huge coastal landslide has occurred at Waterstein Head in NW Skye, where the 50–60 m high headscarp also coincides with the edge of the lava outcrop, and a failed mass of rock ~0.45 km^2 in area is seated on Jurassic sedimentary rocks. Multiple transverse ridges on the landslide suggest that multiple or successive rotational failures occurred within the sedimentary rocks. The toe of landslide terminates in a low cliff that is currently being eroded by wave action.

Not all coastal landslides caused by failure within sedimentary rocks overlain by lavas terminated on land. The

Fig. 10.13 Coastal landforms on Skye and Mull. **a** Coastal landslide at Carn Mòr, on the shores of Loch Scavaig, Skye. Failure seated in shales has caused collapse of the overlying cap of basalt lavas. **b** The Lateglacial rock platform in eastern Mull, which is thought to have formed through a combination of frost-wedging of rock and wave erosion during the Loch Lomond Stade. **c** High rock platform fragment ~19 m above sea level near Staffin Bay, northern Skye. **d** Postglacial raised beach, ~7 m above sea level, at the head of Loch Slapin, Skye. The moraines on the right were deposited by an outlet glacier of the Skye Icefield during the Loch Lomond Stade at ~12.4 ka. (Images: **a, d** Colin Ballantyne; **b** Murray Gray; **c** John Gordon)

runout zone of a coastal landslide on the seaward slopes of Ben Cleat, near Elgol in southern Skye (Fig. 10.13a) extends up to a kilometre offshore in the form of chaotic submarine hummocks and ridges (Small et al. 2016). Several large indentations in basalt or dolerite cliffs along the shores of Skye and Mull probably represent the sites of major coastal landslides where the runout zone is submerged or represented by low skerries.

10.7.2 Rock Platforms

Rock platforms (coastal benches eroded in bedrock, often with a backing cliff) are present on all four islands. Sissons (1983) proposed that these fall into three categories: (i) low-level interglacial platforms; (ii) a glacio-isostatically tilted Lateglacial rock platform; and (iii) high rock platforms at altitudes above ~18 m OD. The first category consists of platform fragments close to the present tidal range that locally support glacially moulded rock outcrops and sometimes pass below till bluffs. Such low-level platforms were considered by Dawson (1980a) to be inherited landforms that developed during interglacials, locally modified by postglacial marine erosion. The Lateglacial rock platform (also referred to as the Main Rock Platform or Main Lateglacial Shoreline) is generally 50–150 m wide with backing cliffs typically 15–20 m high (Fig. 10.13b), and supports relict sea stacks and arches, indicating formation since ice-sheet deglaciation. This shoreline has a maximum altitude of 10–11 m OD in the Loch Linnhe area and eastern Mull, and declines away from the centre of glacio-isostatic uplift in the western Highlands at average gradients of 0.13–0.18 m km^{-1} (Gray 1974; Dawson 1988). It is consequently present as an intertidal feature in southern Skye, eastern Rùm and southern Arran. Because it occurs only outside the limits of the LLR, exhibits no evidence of glacial modification and is well developed in areas of limited fetch, several authors have argued that it formed rapidly under periglacial conditions during the Loch Lomond Stade through a combination of frost-wedging of bedrock and removal of debris by storm waves and sea ice (Sissons 1974, 1983; Dawson 1980b). This platform exhibits abrupt vertical dislocations of 1.0–2.7 m at sites on Mull, providing evidence of postglacial fault movements (Firth and Stewart 2000).

The high rock platform fragments of the Inner Hebrides are amongst the most intriguing landforms in Scotland. They are up to 700 m wide and are most extensively developed on the west coasts of Islay and Jura (Chap. 11); those on the coasts of western Mull and NW Rùm tend to be narrower (~20–200 m) but extend for several kilometres along the coastline. They range in altitude from ~18 m OD to ~51 m OD (~24–30 m on Skye; ~18–33 m on Rùm and ~25–51 m on Mull and adjacent islands; Fig. 10.13c), though some of the higher (>40 m OD) fragments mapped on Mull may be lava benches rather than the products of marine erosion. In some areas, platform fragments at different altitudes occur in proximity, indicating that they are not part of the same former shoreline. Some are reported to be ice-moulded or overlain by till, but others exhibit no evidence for glacial modification. High rock platforms are limited to west of a line that extends through Skye, Mull and Jura to eastern Islay (Fig. 10.14). This distribution prompted Sissons (1982, 1983) to argue that they represent fragments of glacio-isostatically-uplifted intertidal rock platforms that formed through frost action and storm-wave activity during successive glacial periods when ice-sheet margins lay east of the line depicted in Fig. 10.14. More radically, Dawson et al. (2013) have suggested that the high rock platforms are remnants of a formerly more extensive glaciated strandflat surface of Pliocene age. It is also possible that some of the high rock platform fragments of the Inner Hebrides were formed under periglacial conditions during the retreat of the last ice sheet. TCN dating has shown that retreat of the Hebrides Ice Stream was succeeded by a period of ~3000 years (~20–~17 ka) when an oscillating ice margin was located amid the islands of the Inner Hebrides (Small et al. 2017; Fig. 10.14). Palaeotidal modelling indicates that this zone experienced megatidal conditions during this period (Scourse et al. 2018), which would have favoured coastal erosion and removal of debris released by frost wedging of bedrock. Oscillation of the ice margin during this period may explain why some high-level platform fragments exhibit evidence for glacial modification, whereas others do not.

10.7.3 Other Raised Shorelines

Coastal terraces eroded in drift and raised depositional features on the islands fall into two groups: high Lateglacial shorelines and lower shorelines that represent Holocene marine transgressions. Both sets of features decline in elevation away from the centre of glacio-isostatic uplift in the western Highlands, though the gradient of the Holocene features is much gentler than that of the Lateglacial shorelines.

Lateglacial shorelines occur at various altitudes up to 34 m OD on all four islands, mainly at the heads of sheltered bays, and are absent at coastal sites on Skye and Mull where LLR glaciers entered the sea. In some locations, abrupt drops in the Lateglacial marine limit are associated with stillstands or readvances that interrupted the retreat of the last ice sheet. At the mouth of Glen Brittle on Skye, for example, terraces marking the marine limit at 18–20 m OD terminate abruptly upvalley and the marine limit drops to ~7 m OD (Walker et al. 1988). This fall in the marine limit coincides with the maximum extent of a readvance of glacier ice in Glen Brittle, represented by end and lateral moraines that have been TCN-dated to ~17.6 ka (Small et al. 2016), though as noted in Sect. 10.3.1 an age of ~15.5 ka seems more probable. Similarly, on Arran, the Lateglacial marine limit is represented by raised beaches and raised deltas in sheltered bays, such as Brodick Bay and Drumadoon Bay, and rises northwards from ~27 m in the south of the island to ~32 m farther north (Gemmell 1973). In the northernmost part of the island, however, it occurs no higher than ~18 m, suggesting that glacier ice continued to occupy northern Arran after the rest of the coastline was exposed to the high Lateglacial sea.

Some of the most impressive raised shoreline features are of Lateglacial age. On Skye, these include a well-developed erosional platform up to 500 m wide cut across till at ~18 m OD at the head of Staffin Bay, and a raised tombolo, up to 30 m above present sea level, that connects a small offshore island to the coastline at The Braes near the mouth of Loch Sligachan. At Harris on the west coast of Rùm, superb Lateglacial shingle and cobble ridges rest at up to 29 m OD on the high rock platform (McCann and Richards 1969). On Mull, the most striking Lateglacial depositional coastal feature is a raised delta at the head of Loch Don, which was deposited at the retreating ice-sheet margin when relative sea level stood at an altitude of ~34 m OD and was later partly modified by ice advance during the Loch Lomond Stade (Benn and Evans 1993). On Arran, the most impressive Lateglacial coastal feature is a raised delta at Dougarie on the west coast, which occupies an area ~0.3 km^2, slopes seaward from ~32 m to ~30 m OD and formed when glacier ice occupied adjacent Glen Iorsa (Finlayson et al. 2014).

Because of oscillations in relative sea level at a time of continuing glacio-isostatic uplift, Holocene raised shorelines around Scotland are not all of the same age (Smith et al. 2012, 2019). In the Inner Hebrides, the Holocene marine limit is thought to have been reached between 5.8 ka and

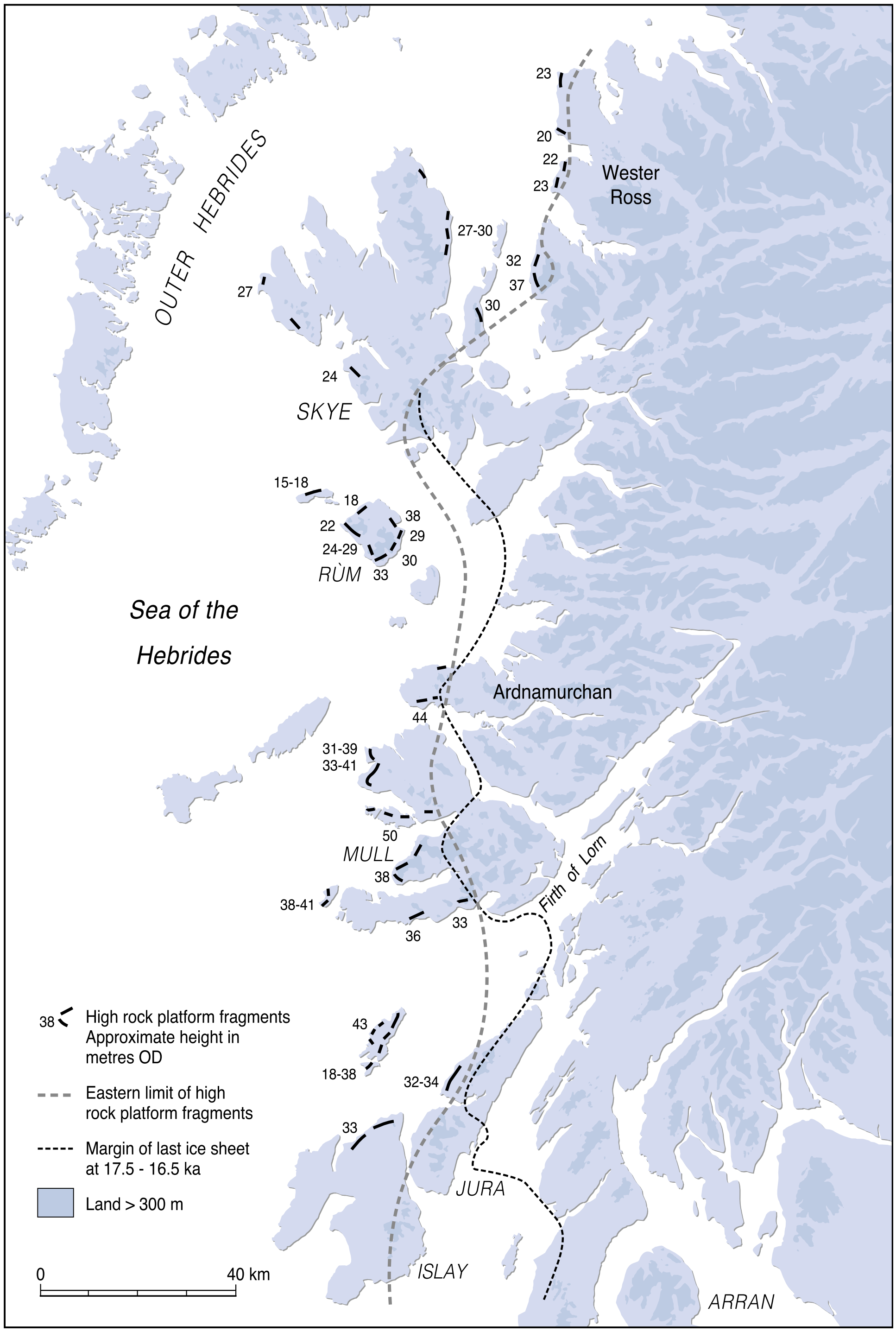

Fig. 10.14 Altitudes of high rock platform fragments in the Inner Hebrides, showing the eastern limit of platform occurrence. (After Sissons 1983) and the position of the margin of the last ice sheet at ~17.5–16.5 ka as depicted by Small et al. (2017)

3.6 ka (Selby and Smith 2007, 2016), whereas on Arran the Holocene marine limit is represented by the Main Postglacial Shoreline, which formed at 7.8–6.2 ka. Around the coasts of Skye and Rùm, Holocene raised beaches up to ~7 m OD occur in sheltered locations, mainly at fjord heads (Fig. 10.13d). The Main Postglacial Shoreline on Arran, however, takes the form of a raised beach that extends around the island at up to ~7–10 m OD (Gemmell 1973), one of the finest examples of a Holocene raised shoreline in Scotland.

10.8 Conclusion

It is difficult to describe the landscapes and landforms of the islands of the Hebridean Igneous Province without resorting to superlatives. They contain the finest igneous geology in Scotland, the most striking alpine scenery, glaciated landscapes dominated by radial outflow of ice, deep, cliffed cirques, numerous end, lateral and recessional moraines deposited by glaciers during the Loch Lomond Stade, an exceptional range of high-level periglacial and aeolian features, the largest and most spectacular landslides in the British Isles, and an unsurpassed range of coastal landforms. They include some of the most striking, internationally important and most-visited geological and geomorphological sites in Scotland and continue to yield new insights into both their long-term geological evolution and Quaternary environmental history. Geomorphologically, the four islands are unique in Scotland in illustrating the effects of successive Pleistocene glaciations on terrain underlain by complex igneous geology, and the aftermath of glaciation in terms of postglacial landscape evolution on such terrain.

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11 The Islands of Islay, Jura, Colonsay, Tiree and Coll

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Abstract

The Hebridean islands of Islay, Jura and Colonsay are mainly composed of Neoproterozoic metasedimentary rocks of the Dalradian supergroup, whereas Tiree and Coll are underlain by Archaean rocks of the Lewisian Gneiss Complex. Most of the terrain is low-lying; only the three peaks of the Paps of Jura exceed 600 m. Successive episodes of ice-sheet glaciation have produced areas of ice-moulded topography, and retreat of the last (Late Devensian) ice sheet was accompanied by the deposition of outwash deposits and moraines, including the longest medial moraine in Scotland. The most impressive and internationally renowned landforms on the islands are those related to coastal erosion and deposition: high-level glacimarine deposits; ancient (possibly Pliocene) glaciated strandflat; extensive areas of high rock platforms and intertidal platforms that demonstrably pre-date the last ice sheet; an isostatically tilted rock platform that formed under Lateglacial periglacial conditions; magnificent flights of Lateglacial shingle ridges extending up to 37 m above present sea level; and lower Holocene raised beaches locally backed by sand dunes and areas of machair (calcareous aeolian shell-sand) deposits. Inland, the quartzite flanks of the Paps of Jura are scarred by paraglacial landslides possibly triggered by seismic activity that accompanied rapid glacio-isostatic uplift.

Keywords

Strandflat • Rock platforms • Glacimarine deposits • Raised beaches • Machair • Medial moraine • Paraglacial rock-slope failures • Glacio-isostatic uplift

11.1 Introduction

The islands of Islay, Jura, Colonsay, Tiree and Coll collectively make up much of the Inner Hebrides outside the main area of the Hebrides Igneous Province (Fig. 11.1). The geology of the islands ranges from the Archaean Lewisian rocks of Tiree and Coll to the Neoproterozoic Dalradian quartzites, metamudstones and limestones of Islay and Jura. The islands also bear witness to Palaeogene igneous activity in the form of swarms of dolerite dykes that occur in huge numbers across SW Jura but more sparsely in neighbouring islands. All of the islands were over-ridden by the last ice sheet, which extended westward towards the continental shelf. None of the islands, however, are of sufficient relief to have supported glaciers during the Loch Lomond Stade. In terms of their geomorphology, the islands are internationally renowned for the various raised shorelines that dominate the coastal landscape. On Tiree and Coll, for example, most of the landscape comprises strandflat. By contrast, the coastlines of Islay, Jura and Colonsay are characterised by spectacular raised shore platforms together with a range of Lateglacial and Holocene raised beach features. Most spectacular of all are the famous spreads of unvegetated raised gravel beach ridges of western Jura, which extend along the coastline for many kilometres. The complex assemblages of raised marine features that occur in western Jura and northern Islay has led to their selection for the Geological Conservation Review (Dawson 1993a, b) and the subsequent designation of the entire coastal area as a Site of Special Scientific Interest (SSSI).

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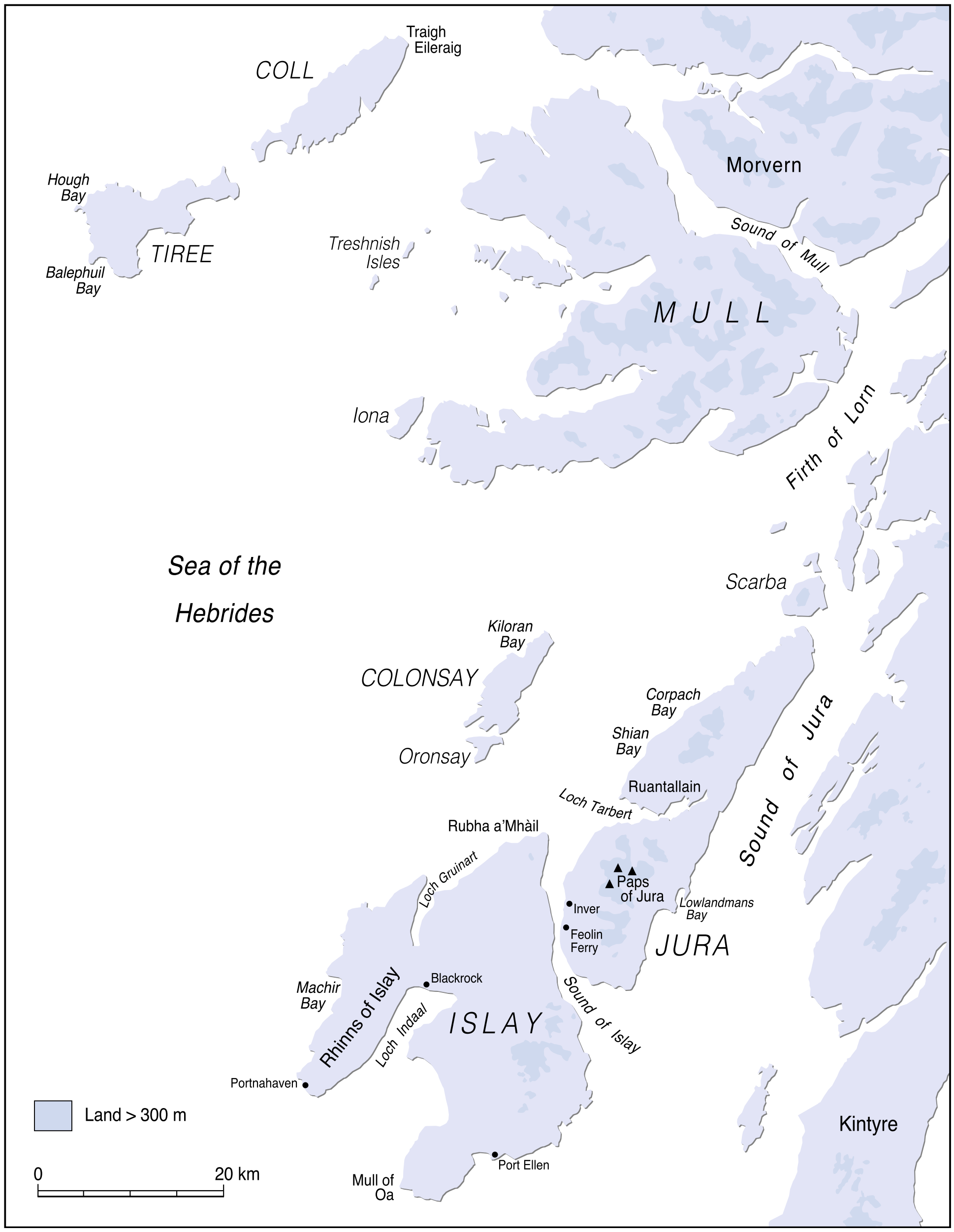

Fig. 11.1 The islands of Islay, Jura, Colonsay, Tiree and Coll, showing key locations mentioned in the text

11.2 Geology and Landscape

The islands of Islay, Jura and Colonsay are largely composed of Neoproterozoic metasedimentary rocks of the Dalradian Supergroup (Stephenson and Gould 1995) and form part of the Grampian Highlands terrane (Chap. 2). Tiree and Coll are underlain by the much older rocks of the Archaean Lewisian Gneiss Complex and represent the southernmost terrestrial extent of the Hebridean terrane (Park et al. 2002).

Jura possesses a relatively simple geological structure, being composed primarily of a uniform series of mid-Dalradian quartzites that underlie all of the highest ground, culminating in the three mountains of the Paps of Jura, all of which exceed 700 m (Fig. 11.1). The Dalradian series is Neoproterozoic in age and forms part of the Caledonian belt of Scotland and Ireland (Chap. 2). The strike of the Jura quartzite runs parallel to the length of the island, and the strata dip ESE between 25° and 40°, giving the landscape a jagged appearance. The quartzite attains an aggregate thickness of over 5 km but thins to the northeast and southwest; it represents sediments deposited as part of a tidal delta system (Levell et al. 2020), then metamorphosed, uplifted and tilted during the Caledonian Orogeny. In eastern Jura the quartzite is overlain by metamudstones and occasional conglomerates that are thought to have been deposited within a subsiding basin; metabasite sills are intruded through the metamudstones and form headlands and prominent inland ridges. In northern Jura, metabasite dykes of pre-Devonian age also form upstanding ridges, but some have been preferentially weathered and eroded to form narrow gorges. The dykes attain widths of up ~6 m, trend north–south and NNW–SSE, and are largely responsible for the crenulate configuration of the coastline of NW Jura. Lamprophyre sills are also locally well developed, forming a prominent element of the coastal cliffs at Ruantallain in western Jura. The most conspicuous igneous rocks, however, are NW–SE trending Palaeogene dolerite dykes that form part of the larger dyke swarm of the Inner Hebrides. These are ubiquitous in southern Jura and NE Islay, where they locally form dense concentrations of coastal stacks, arches and minor headlands as a result of preferential erosion of the adjacent vitreous quartzite.

The orientation of the coastline of western Jura north of Loch Tarbert conforms to a NE–SW trending fault located offshore. The quartzite close to the fault is typically vitreous and forms an elongate shatter belt. The Dalradian rocks that are so vividly illustrated in the Jura landscape continue southwest across Islay and also to the north of Jura on the island of Scarba, where differential erosion of quartzite gives the landscape the appearance of sets of inclined sawtooth ridges and depressions.

Most of eastern and southeastern Islay is also composed of Dalradian quartzite, which underlies low (400–500 m) hills in the southeast, but the central part of the island is composed of Dalradian metamudstones, limestones and well-sorted quartzites that form undulating, peat-covered lowlands. Areas of Palaeoproterozoic orthogneiss also occur in the Rhinns of western Islay. North of Islay, the island of Colonsay and its smaller neighbour, Oronsay, are underlain by Dalradian metamudstones and conglomerates.

The islands of Tiree and Coll are almost entirely composed of Lewisian gneiss. Most of Tiree and southern Coll constitute an essentially flat landscape with only limited areas of ground above ~30 m OD (Ordnance Datum); the highest point on both islands is a low (141 m) hill in southern Tiree. The interior of northern Coll constitutes a classic knock-and-lochan landscape characterised by an apparent chaotic arrangement of ice-moulded gneiss ridges separated by peat-filled hollows.

11.3 Glacial History

Although there is evidence for offshore pre-Late Devensian sediments on the Hebrides Shelf (Fyfe et al. 1993; Dove et al. 2015), no terrestrial deposits predating the Last Glacial Maximum (LGM) have been found on low ground on any of the islands considered in this chapter. During the Late Devensian (~31 to 11.7 ka), all of the islands were completely over-run by ice moving westward or southwestward from the Scottish mainland. On the Paps of Jura (the highest mountains in this island group), ice-moulded bedrock and perched, glacially transported boulders occur up to 660 m, and rhyolitic erratics of mainland origin have been found at ~780 m near the summit of the highest mountain (Ballantyne 1999). The summits of the Paps of Jura nevertheless support thick quartzite blockfields at altitudes over 700 m. Although these blockfields were initially interpreted as evidence for palaeonunataks that remained above the maximum surface level of the last ice sheet, they are now considered to be pre-Late Devensian periglacial regolith covers that escaped erosion by the last ice sheet because the upper parts of the ice sheet were cold-based and frozen to the underlying substrate (Fabel et al. 2012). This reinterpretation implies that the ice crossing Jura was polythermal, with erosive warm-based ice occupying ground below ~660 m but cold-based ice cover persisting over the very highest ground.

During the LGM, the Inner Hebrides and adjacent western seaboard formed the onset zones of the Hebrides Ice Stream, a major artery of the last ice sheet that extended westward to the shelf break, over 220 km WSW of Jura (Ballantyne and Small 2019; Chap. 4). The alignment of offshore submarine bedforms demonstrates southwestwards ice flow across Tiree and Coll and westwards to southwestwards movement of ice across Colonsay, Islay and Jura at this time (Dove et al. 2015). The Hebrides Ice Stream reached its maximum extent at ~26.7 ka and had begun to retreat by ~25.9 ka; by ~20 ka much of the Sea of the Hebrides was deglaciated (Callard et al. 2018). Cosmogenic ^{10}Be exposure ages obtained for samples from glacially deposited boulders indicate that southernmost Tiree became ice-free at 20.6 ± 1.1 ka.

Thereafter ice streaming apparently ceased, and retreat of the ice margin slowed as it became progressively grounded amongst the Inner Hebrides. Cosmogenic ^{10}Be exposure ages obtained for various sites fringing the Sea of the Hebrides imply that by 17.5–16.5 ka Islay, Colonsay, Tiree and Coll were completely deglaciated and the ice margin was located over mainland peninsulas and the easternmost Hebridean islands (Small et al. 2017; Fig. 11.2). This chronology is supported by radiocarbon ages obtained for marine shells recovered from offshore sediment cores; these indicate deglaciation of SW Islay before 15.7–14.9 cal ^{14}C ka, and outer Loch Sunart (on the mainland east of Coll) before 16.8–16.2 cal ^{14}C ka (Peacock 2008; Baltzer et al. 2010). During the period ~20 to 17 ka, regional southwesterly or westerly ice movement was succeeded by topographically confined flow along sounds and fjords (Dove et al. 2015), with the progressive isolation of independent icefields on some Hebridean islands (Chap. 10).

Though there is no evidence for the stranding of remnant icefields on the lower islands in this group, the alignment of striae and moraines indicates that a small residual ice mass became isolated at ~16.6 ka around the Paps of Jura following retreat of the last ice sheet northeastwards across the adjacent shelf (Ballantyne and Dawson 2019; Sect. 11.4.3).

Retreat of the ice margin to the eastern Inner Hebrides and western mainland seaboard by ~17.5 to 16.5 ka was accompanied by marine transgression across coastal areas. The altitude of the Lateglacial marine limit reflects subsequent differential glacio-isostatic rebound: in western Tiree it lies at ~10 m OD, in NE Coll at ~22 m OD, in central Islay at ~23 m OD and in NW Jura it reaches ~38 m OD. This pattern of differential glacio-isostatic uplift is illustrated by shoreline uplift isobases that decline in altitude from northeast to southwest across Jura and Islay and from ESE to

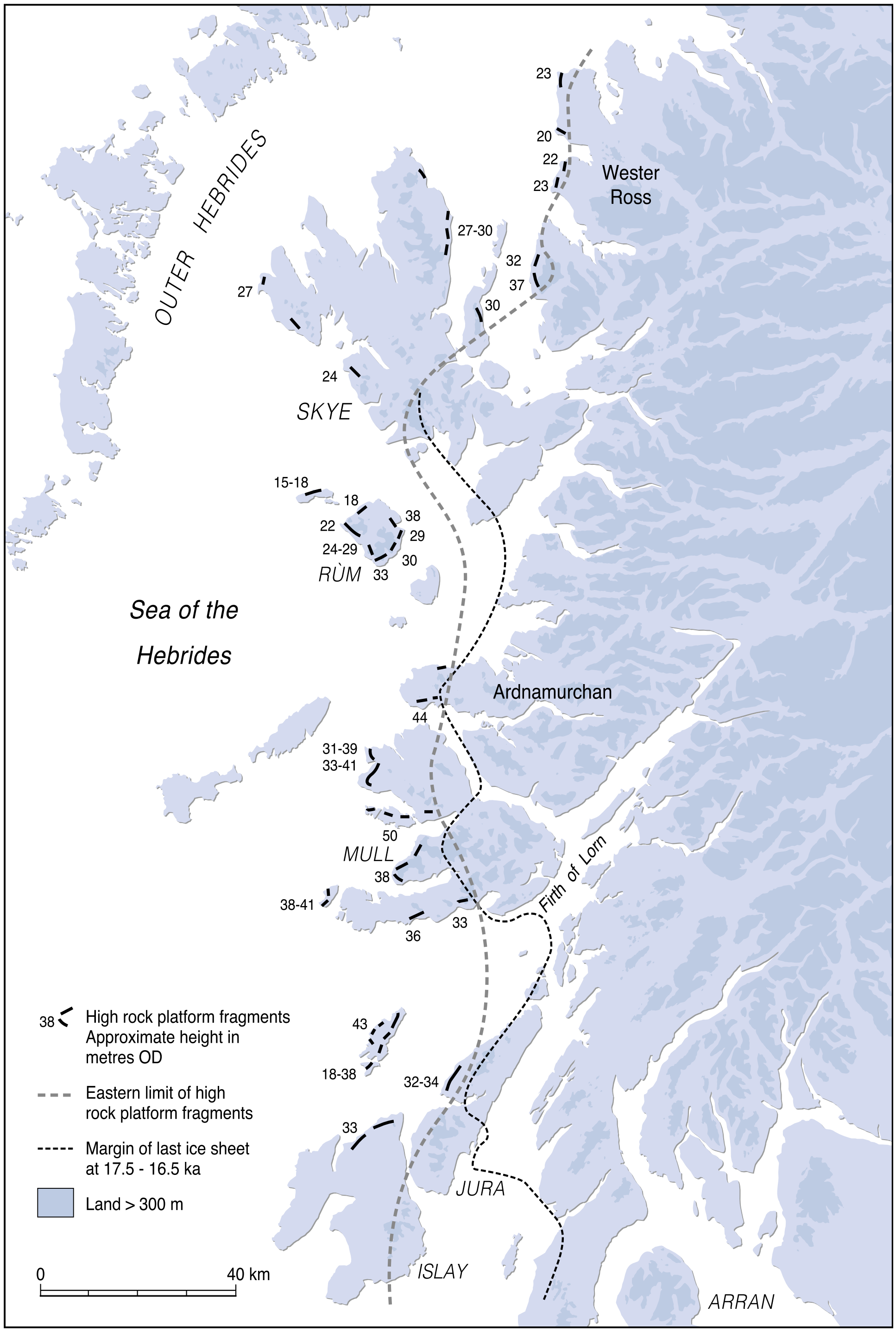

Fig. 11.2 The position of the margin of the last ice sheet at ~17.5 to 16.5 ka as depicted by Small et al. (2017). Also shown are the locations and altitudes of high rock platforms in the Inner Hebrides; the heavy dashed line indicates the eastern limit of high rock platform distribution identified by Sissons (1983)

WNW between northern Jura and Tiree (Smith et al. 2019; Chap. 4). The regional shoreline tilt across these areas is ~0.6 m km^{-1}, indicating the pronounced differential glacio-isostatic rebound that took place across this part of western Scotland in the millennia that followed ice-sheet deglaciation (Dawson 1982).

There is no evidence for the formation of glaciers on any of the islands considered here during the Loch Lomond Stade, when they experienced severe periglacial conditions, with permafrost extending down to low ground (Ballantyne 2019). During this time hillslopes were modified by frost weathering, solifluction, rockfall and talus accumulation, while coastal areas were subject to cold-climate shore erosional processes that led to the development of an extensive shore platform, the Main Rock Platform (Sect. 11.5.3).

11.4 Glacial and Glacimarine Landforms

11.4.1 High-Level Glacimarine Deposits in Western Islay

High-level glacimarine sediment accumulations represent distinctive landscape features in western Islay, across the Rhinns of Islay and the Mull of Oa (Dawson et al. 1997a). These deposits extend up to ~80 m OD and give the impression of a landscape choked with sediment above which lie a series of upstanding bedrock ridges. The sediments are typically ~60 m thick; a borehole south of Machir Bay has revealed a thickness of over 50 m of stoneless pink clay resting on bedrock (Dawson et al. 1997a). The surfaces of the deposits are terrace-like, albeit with surface undulations in places. That the sediments are of glacimarine origin is indicated by their sedimentological characteristics and by their incorporation, for example at Machir Bay, of stenohaline marine foraminifera (Dawson et al. 1997a). Across the Rhinns, the seaward edge of the sediments is defined by scores of shallow landslides that tend to obscure the glacimarine strata. In the western Rhinns the sediments contain numerous flint clasts as well as, unusually, a Cretaceous chalk boulder (Benn and Dawson 1987).

High-level glacimarine deposits also occur in southernmost Kintyre (Sutherland 1981), along the northern coast of Ireland (McCabe et al. 1986) and on the southern margin of Donegal Bay in NW Ireland (McCabe et al. 2007). Marine shells and foraminifera in glacimarine deposits at ~80 m OD at the margin of Donegal Bay have yielded pre-LGM radiocarbon ages, suggesting that the deposits were emplaced on an isostatically depressed coastline prior to extension of the last ice sheet to the edge of the continental shelf, and survived subsequent over-running by the ice sheet, possibly in permafrost. Thermoluminescence dates obtained for the inferred glaciomarine sediments in the Rhinns fall within the range 41–54 ka (Dawson et al. 1997a), suggesting that these were also deposited prior to the LGM.

11.4.2 Ice-Marginal Landforms

An extensive suite of ice-marginal landforms occurs in central Islay north of Loch Indaal (Fig. 11.1), where a widespread area of hummocky drift ridges covers an area ~500 to 700 m wide and ~8 km long. Near the head of Loch Indaal, a broad outwash terrace extends westwards from the former ice margin for over a kilometre. The terrace is graded to a series of raised marine terrace fragments at 21–23 m OD. East of Loch Gorm in western Islay, a complex of raised marine terraces, esker fragments and outwash terraces also indicates outwash deposition when relative sea level was ~21 to 22 m OD. This evidence linking ice-marginal outwash deposition to a specific level of the Lateglacial sea suggests that retreat of the ice margin across Islay was interrupted by a stillstand (Dawson 1982) or possibly minor readvance (Peacock and Merritt 1997).

Similar field relations occur along a stretch of the coastline of NW Islay, where a small arcuate ice-marginal moraine ~1 km long and ~5–8 m high occurs at Coir Odhar. The frontal edge of this moraine forms the backing cliff of two raised shoreline fragments at ~26–27 m OD, and outwash deposits fronting the central area of the moraine are graded to the same elevation; outwash terrace fragments inside the moraine descend seaward from ~42 m OD to merge with the raised shoreline fragments.

11.4.3 The Sgriob Na Caillich Medial Moraine

Medial moraines deposited during the retreat of the last ice sheet are rare in Scotland, but a superb example occurs in western Jura, mid-way between Inver and Loch Tarbert. Known as Sgriob na Caillich (the Witch's Slide), this consists of two to four belts of boulders that extend northwest from an altitude of 450 m on the western footslopes of Beinn an Oir (785 m) to within 300 m of the present coastline (Ballantyne and Dawson 2019; Fig. 11.3). Along most of their length the boulder belts form subdued ridges that rise less than a metre above the adjacent terrain. They are composed of angular quartzite boulders 0.2–1.3 m long, some of which are perched on others indicating deposition from the surface of the last ice sheet as it retreated southeastwards towards the Paps of Jura. In planform the boulder belts are straight throughout their length but fan out near their source towards the southwest flank of Beinn an Oir.

The source of the medial moraine coincides with bedrock gullies and cliffs that Ballantyne et al. (2014b) interpreted as

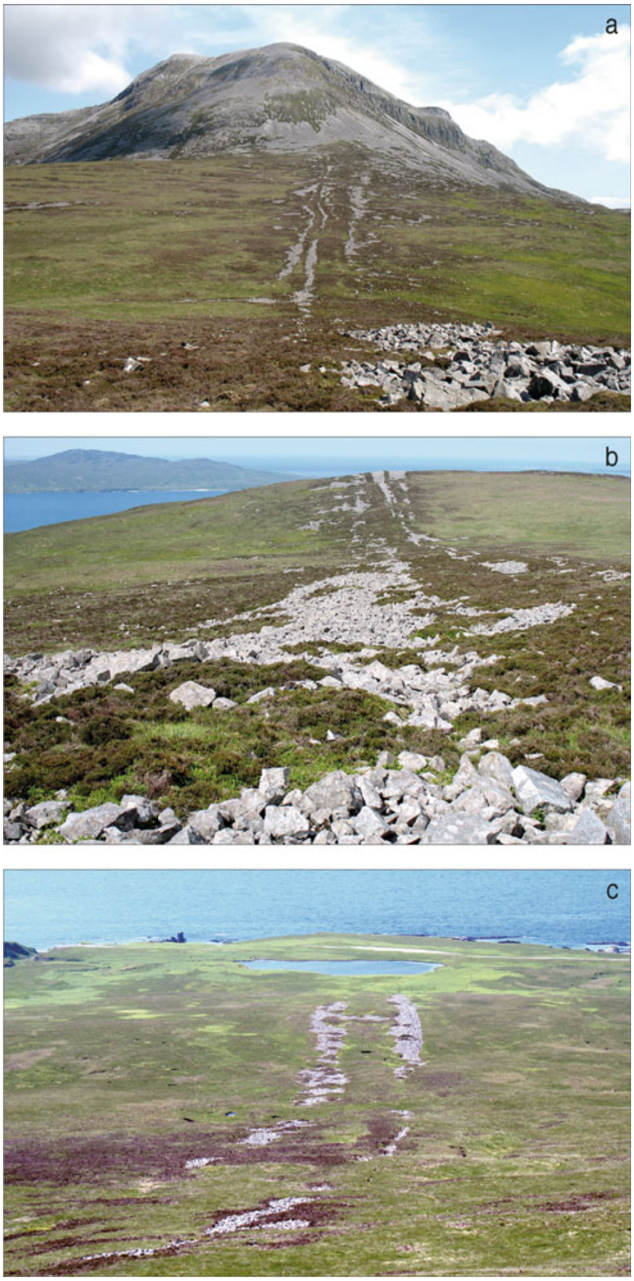

Fig. 11.3 The Sgriob na Caillich medial moraine, SW Jura. **a** The proximal section of the moraine, looking southeast towards its source on Beinn an Oir (785 m). **b** The central section of the moraine, looking northwest towards where it rises to cross a low hill, Cnoc nan Sgrioba (385 m). The coarse, angular nature of the debris indicates deposition from the surface of the retreating and thinning ice sheet. **c** The distal part of the moraine, where it forms two low parallel ridges that descend towards the Lateglacial marine limit at ~27 m OD. (Images: Colin Ballantyne)

debris-free failure scarps, sites of former rock-slope failure where the runout debris has been removed by glacier ice. The most prominent features are two bedrock gullies, ~20–30 m deep and ~70–80 m wide, that converge downslope from near the summit of Beinn an Oir, and an arrested slide block that terminates downslope at a cliff up to 60 m high. The volume of the gullies implies loss of at least 200,000 m^3 (~0.52 Mt) of rock. The coincidence of these debris-free scarps with the upslope limit of the boulder belts suggests that one or more rockslides or a series of rockfalls formed the source of the debris in the medial moraine. It appears that emergence of Beinn an Oir from the thinning ice cover was succeeded by one or more rockslides or rockfalls onto the surface of the ice sheet, and that ice converging around both flanks of the emerging nunatak carried the debris supraglacially to (and possibly beyond) the present coastline; subsequent thinning and retreat of the ice then deposited the debris to form the longest medial moraine in Scotland. Cosmogenic ^{10}Be exposure dating of boulders on the moraine indicates that it was deposited at 16.6 ± 0.8 ka (Small et al. 2017), indicating final deglaciation of SW Jura at this time. The northwestwards alignment of the moraine suggests that a residual ice cap was stranded over southern Jura as the margin of the ice sheet retreated towards the northeast (Ballantyne and Dawson 2019).

11.5 Coastal Rock Platforms

Rock platforms (coastal benches eroded across bedrock, often with a backing cliff) are present on all the islands of the Inner Hebrides. They fall into four categories: (i) extensive areas of strandflat, mainly on Coll and Tiree; (ii) high rock platforms, with an altitudinal range of ~18 to 44 m; (iii) a glacio-isostatically tilted Lateglacial rock platform (the Main Rock Platform), which is thought to be of Loch Lomond Stadial (~12.9 to 11.7 ka) age; and (iv) glaciated low rock platforms that occur within the present intertidal zone (Dawson 1980b; Sissons 1981, 1983).

11.5.1 Strandflat

Well-developed raised rock platforms cutting across gneiss are ubiquitous features on the islands of Tiree and Coll. Almost the entire landscape of Tiree is represented by very wide ice-moulded raised rock platforms that cover an area of ~80 km^2 (Fig. 11.4). Similar rock platforms occur on SW Coll, where they terminate at an ice-moulded cliffline. The rock platform surfaces on these islands typically occur at ~15 to 30 m OD and comprise two discrete surfaces at ~15 to 20 m and 20–30 m OD, though it is difficult to define precisely the elevation of individual rock surfaces since in most areas the gneiss bedrock is extensively ice-moulded.

The Tiree and Coll platforms represent part of a much wider area of planated and glacially moulded rock platform surfaces that attain their finest development across the Outer

Fig. 11.4 Strandflat, Isle of Tiree. Note that the ice-moulded rock surfaces are locally mantled by raised Lateglacial and Holocene marine deposits and vegetated sand dunes. (Image: John Gordon)

Hebridean islands of North Uist, Benbecula and South Uist (Chap. 9). These various planated rock surfaces represent an extensive area of strandflat, similar to those of the coastal areas of northern Norway and Alaska (Dawson 1994). The origin of the strandflat remains controversial, but Dawson et al. (2013) have argued that the Scottish examples may be of Pliocene age since the prolonged periods of relative sea level stability required for strandflat formation did not exist during the Quaternary.

11.5.2 High Rock Platforms

High-level raised shore platforms at altitudes of ∼27–44 m OD are continuously developed along the western coastlines of Jura and northern Islay (Fig. 11.2). On Islay, the High Rock Platform is locally up to 600 m wide and backed by cliffs up to 90 m high (Fig. 11.5a); it is partly covered by till deposits that are locally overlain by Lateglacial raised beach gravels. A similar high platform is present along a 7 km long stretch of coast in western Jura north of Loch Tarbert, where it attains widths of up to 500 m, is backed by degraded drift-mantled quartzite cliffs up to 50 m high, is extensively mantled by Lateglacial beach gravels (Fig. 11.5b) and, at two localities, also supports an overlying cover of till. Farther northwest, however, the Jura platform becomes narrower and fragmented and it is absent northwest of Corpach Bay. At locations in NE Islay and western Jura where the cliff-platform junction is not obscured, it has an altitude of 32.1–34.1 m OD. The raised beach gravels that almost entirely bury the platform on western Jura extend up to ∼36 to 37 m OD. Similar platforms occur elsewhere in the Inner Hebrides: particularly fine examples occur on the west coast of Colonsay, where they are developed across Neoproterozoic sedimentary rocks (Fig. 11.5c).

The age and origin of the high rock platforms in the Inner Hebrides has proved controversial. Early researchers concluded that they are of interglacial or even pre-glacial age. A remarkable feature of their distribution is that they occur only west of a line that extends southwards from Wester Ross in NW Scotland through Skye, Mull and Jura to NE Islay (Fig. 11.2). This restricted distribution prompted Sissons (1981, 1982, 1983) to argue that they represent uplifted intertidal shore platforms that originally developed through a combination of frost wedging of bedrock and storm-wave activity during successive glacial periods when former ice margins lay east of the line depicted in Fig. 11.2. By contrast, Dawson et al. (2013) have suggested that they represent remnants of formerly more extensive strandflat surfaces of Pliocene age, analogous to (and broadly contemporaneous with) the glaciated strandflats of Coll and Tiree but truncated by subsequent glacial and coastal erosion. A final possibility, explored in Chap. 10, is that they formed during the period 20–17 ka, when an oscillating ice margin was grounded amid the islands of the Inner Hebrides (Small et al. 2017).

11.5.3 The Main Rock Platform

A conspicuous lower rock platform, the Main Rock Platform (also known as the Main Lateglacial Shoreline), is developed along much of the coastline of Jura and Islay. This platform

Fig. 11.5 **a** High Rock Platform on the western coast of northern Islay. The seaward edge of this platform represents the backing cliff of the Main Rock Platform in the foreground. This platform occurs in the present intertidal zone at this location and is locally overlain by Holocene beach gravels. **b** High Rock Platform, partly overlain by Lateglacial raised beach deposits (grey), western Jura. **c** High Rock Platform, Kiloran Bay, Isle of Colonsay. (Images: **a**, **c** John Gordon; **b** © Edward Z. Smith)

extends almost continuously along the west coast of Jura, where it is typically 50–150 m wide with a backing cliff ~15 to 25 m high cut in quartzite. The platform is tilted away from the centre of glacio-isostatic uplift: the cliff-platform junction is at ~6 m OD in NW Jura, declining southwestwards at a gradient of 0.16 m km^{-1} so that it occurs at ~1 to 2 m OD on the northwest coast of Islay (Fig. 11.5a) and below present sea level on Colonsay, Tiree, Coll and SW Islay. North of Jura, the platform entirely fringes the island of Scarba but it is a fragmentary feature along the eastern Jura coastline. A distinguishing feature of the platform is that it is often well developed in areas of restricted fetch, as in Lowlandmans Bay in SE Jura, where the fetch is no more than several hundred metres. Between the north of Jura and NE Islay the foot of the backing cliff is indented by over 100 raised sea caves. Both the platform itself and the floors of the sea caves are extensively mantled by Holocene beach gravels.

The absence of till cover and of evidence for glacial modification of the platform demonstrates that it developed after the retreat of the last ice sheet, and early researchers concluded that it was of postglacial (Lateglacial and Holocene) age, and possibly contemporaneous with overlying Holocene beach gravels. It is now believed to have formed rapidly through a combination of frost wedging of bedrock and removal of debris by storm waves and sea ice under the severe periglacial conditions of the Loch Lomond Stade (Sissons 1974, 1983). This interpretation is supported by several lines of evidence: the platform is developed at coastal sites with limited fetch, suggesting that wave action alone cannot account for its formation, and on the Scottish mainland the platform is limited to coasts that escaped glaciation during the Loch Lomond Stade. Moreover, the tilt of the platform due to differential glacio-isostatic rebound is steeper than that of Holocene raised shorelines (implying that the platform pre-dates the Holocene) but consistent with that of shorelines of Loch Lomond Stadial age in eastern Scotland. The Main Rock Platform is now regarded as one of the most outstanding examples of periglacial shore erosion in the world (Dawson 1980a).

11.5.4 The Low Rock Platform

On the coasts of the Inner Hebrides, glaciated intertidal shore platform surfaces are widespread. Those along the coastline of NW Jura are discontinuous, essentially flat-lying, typically 10–50 m wide and exhibit evidence of both ice-moulding and marine abrasion, including innumerable potholes. Intertidal shore platforms frequently occur adjacent to fragments of the Main Rock Platform. Along the coastline of SW Jura, areas of low rock platform can be traced inland beneath both Holocene raised beach gravels and glacial drift. Perhaps the finest development of the platform occurs at Rubha a' Mhail, the northernmost tip of Islay, where it is up to ~100 m wide, and contrasts markedly with the more rugged surface of the adjacent Main Rock Platform. The widest areas of Low Rock Platform occur in western Colonsay, where they are up to 1 km wide and cut across Neoproterozoic sedimentary rocks.

The origin and age of the Low Rock Platform(s) are controversial. Dawson (1980b) suggested that those in the Inner Hebrides represent part of a more extensive shoreline that is represented by intertidal platforms elsewhere in Scotland, such as those of northern Caithness, Orkney and eastern Scotland. Sissons (1982), however, argued that these intertidal platforms represent a series of discrete glacio-isostatically tilted interglacial shorelines that occur within the present intertidal zone. Though the ice-moulded nature of the intertidal platforms in the Inner Hebrides demonstrates that they pre-date at least the last ice-sheet glaciation, their age and significance remain uncertain.

11.6 Coastal Deposits

11.6.1 Lateglacial Raised Shoreline Features

The raised beaches of western Jura represent some of the finest examples of such features in Europe. Long stretches of coastline are characterised by continuous staircases of unvegetated quartzite gravel beach ridges and intervening swales reaching in some locations up to ~37 m OD (Fig. 11.5b). The outstanding development of these coastal landforms has contributed to the entire coastline of western Jura south of Corpach Bay and the neighbouring western coastline of northern Islay being designated as a Site of Special Scientific Interest (SSSI). In southern Jura, the raised beach ridges and terraces can also be traced intermittently along the coast, although along the coasts bordering the more sheltered Sound of Islay the highest level reached by the Lateglacial sea is represented by a single wide, peat-covered raised marine terrace that is a continuous feature as far as Feolin Ferry, where it is at ~23 m OD.

The most impressive sequence of raised beach ridges occurs at Loch an Aoinidh Dhuibh, 2 km southwest of Shian Bay in western Jura. At this site a remarkable suite of 55 raised, arcuate, unvegetated gravel ridges and intervening swales descends seaward over a distance of ~650 m from a high Lateglacial marine terrace fragment at ~35 m OD to terminate at the relict cliffline backing the Main Rock Platform (Dawson et al. 1997b). The highest shingle ridges rest upon a raised rock platform (the High Rock Platform) that, in turn, terminates inland at a drift-mantled cliff ~50 to 70 m high. The quartzite clasts that make up each beach ridge are typically sub-rounded and rarely exceed 25 cm in length. The site is flanked both to the north and south by inclined quartzite rock ridges that protrude above the otherwise relatively even surface of the High Rock Platform. Despite their low (~5 m) relief, these ridges have influenced the curvature and geometry of beach ridge development owing to processes of wave refraction and diffraction. As a result, the highest beach ridges form sweeping arcuate features along their lengths.

Similar flights of raised beach ridges and intervening swales occur across the southern part of Shian Bay. At this site and elsewhere along the coast, exposures of till are locally visible beneath the raised beach deposits, demonstrating that whereas the underlying rock platform pre-dates at least the last glaciation, the overlying raised beach gravels post-date deglaciation. The raised beach ridges in Shian Bay are distinctive in that the upper suite of ridges is separated from the lower ridges by a ~500 m long transverse beach ridge that truncates the adjacent ridges and forms a striking landscape element. The significance of this pronounced beach ridge is uncertain, but it may indicate a brief period of relative sea-level stability that interrupted the otherwise continuous lowering of relative sea level during the Lateglacial (Dawson et al. 1997b).

In SW Jura, from Loch Tarbert to Feolin Ferry, Lateglacial raised marine terraces and beach ridge staircases are also present, but in this zone considerable thicknesses of glacial drift have resulted in a rather different coastal landscape. The lower limit of the Sgriob na Caillich medial moraine (Sect. 11.4.3), for example, is defined by a wide raised marine terrace and till bluff that truncates the moraine at right angles. The terrace is almost 200 m wide and mantled by a series of raised beach ridges, some of which are vegetated. Two prominent sections of the terrace at ~27 m OD define the marine limit at ~16.6 ka, when the local ice margin backstepped southeastwards from the present coastline (Ballantyne and Dawson 2019). An adjacent small lochan (Loch na Sgrioba) is dammed along its seaward margin by several distinctive raised shingle ridges. Along this stretch of coastline the continuity of raised coastal features is interrupted by a swarm of upstanding Palaeogene dolerite dykes.

11.6.2 Holocene Raised Marine Deposits

Along the western seaboard of Scotland, isostatic rebound resulted in a fall in relative sea level between the end of the Loch Lomond Stade and the earliest Holocene, followed by marine transgression as the rate of isostatic uplift slowed and eustatic sea level rose in response the final melting of the last northern hemisphere ice sheets (Shennan et al. 2018; Smith et al. 2019). In the Inner Hebrides, this marine transgression resulted in the development of a younger generation of raised beaches, storm ridges and coastal terraces that typically form the broad swathes of low ground extending inland from just above high tide level to a raised coastal bluff cut in glacial sediments. These low coastal terraces range in width from a few tens of metres to hundreds of metres, and often consist of vegetation-covered shingle banks or gently sloping ramps of cobbles and gravel, and the terrace sediments invariably incorporate marine shell fragments. In exceptional cases, flights of over 40 unvegetated Holocene beach ridges and intervening swales extend continuously inland, as along the coast north of Inver, in SW Jura (Fig. 11.6).

Collectively, the highest of these low marine terrace features and banks of vegetated gravel represent the Holocene marine limit, the highest level reached by the sea over the past ~11.7 ka, now elevated above present sea level by glacio-isostatic uplift. Although traditionally referred to as the Main Postglacial Raised Beach and assumed to reflect marine transgression between ~7.8 ka and ~6.2 ka, it has now been established that only the Holocene raised beaches nearer the centre of isostatic uplift reflect marine transgression at that time. Farther from the centre of uplift, the Holocene marine limit (for example on Coll, Tiree, Colonsay and western Islay) may have been reached later, at ~5.8–3.4 ka (Smith et al. 2012, 2019). At coastal sites proximal to the axis of glacio-isostatic rebound, the Holocene marine limit is much lower than the Lateglacial marine limit. On the coast of Loch Tarbert in western Jura, for example, the highest Lateglacial marine terraces are at ~30 m OD and the highest Holocene marine gravel terraces occur at ~12 m OD. Farther from the axis of uplift, however, the Lateglacial and Holocene marine limits converge because of the steeper gradient of the former. On Tiree, the highest Holocene marine gravel terraces occur at a similar altitude (~9 to 11 m OD) to the highest Lateglacial raised beach deposits, making it almost impossible to distinguish between the two.

Only limited stratigraphic evidence is available to establish the nature and timing of Early Holocene relative sea-level fall and the subsequent marine transgression in the islands considered here. In a coastal inlet (Tràigh Eileraig) on northern Coll, intertidal peat deposits overlain by beach sands have yielded a radiocarbon age of 9.1–8.7 cal ^{14}C ka, implying that relative sea level was similar to that of the present at that time but subsequently rose to bury the peat under beach sands (Dawson et al. 2001). This pattern is broadly consistent with the results of stratigraphic investigations at Loch Gruinart in northern Islay, which indicate that relative sea level appears to have risen from ~0 m OD at the Lateglacial–Holocene transition to ~5 m OD during the Middle Holocene, before falling thereafter to its present level (Dawson et al. 1998). It is notable, however, that the majority of Holocene raised marine gravels extend up to ~10 m OD or slightly higher, highlighting the important role that storm sedimentation has played in their formation.

11.6.3 Machair and Coastal Sand Dunes

The Gaelic term *machair* is used to describe the low-lying dune grassland plains that occur along Hebridean coasts, particularly those facing the Atlantic Ocean (Chap. 9). The machair deposits are predominantly composed of sand intermixed with abundant shell fragments and support a floristically rich vegetation cover that owes its diversity to the calcareous nature of the sediments. Machair deposits are aeolian in origin, having accumulated throughout the Holocene as a result of offshore winds blowing sand and comminuted shell fragments (originally derived from offshore) from the beaches onto the adjacent plains, where they

Fig. 11.6 Staircase of 40 unvegetated gravel beach ridges of Holocene age, Inver, SW Jura. (Image: John Gordon)

are anchored by the vegetation cover (Ritchie 1976). Many machair areas are designated SSSIs on account of their rich floral diversity. Areas of machair are often separated from the present coastal margin by sand dunes, many of which accumulated as a result of windstorm activity during the Little Ice Age of the sixteenth to nineteenth centuries. Similar areas of dunes also occur on lowland areas at the rear of some modern beaches in the absence of machair.

The Isle of Tiree is fringed by numerous sandy beaches, most of which are backed by vegetated and unvegetated dunes and some of the most extensive machair areas in the Inner Hebrides. The most spectacular site exhibiting this assemblage is Hough Bay on the westernmost tip of Tiree, where the entire coastal landscape is blanketed by machair interspersed by vegetated coastal dunes.

On the neighbouring Isle of Coll, machair and coastal dune areas are mainly restricted to the south of the island, as most northern coasts consist of low rock cliffs or skerries. Similarly, on Jura, machair is restricted to a small area near Knockrome on the east coast. Around the coasts of Islay both machair and coastal sand dunes are widespread, notably near the mouth of Loch Gruinart in the north of the island. Some coastal areas support multiple generations of sand dunes representing successive episodes of Holocene windstorms (Dawson et al. 2003). Striking examples occur at Kiloran Bay in NW Colonsay, Balephuil Bay on Tiree and Machir Bay in westernmost Islay. At Loch Gruinart, the geomorphological characteristics of loch-head and loch-side saltmarshes are particularly well displayed (Hansom 2003).

11.7 Rock-Slope Failures in the Paps of Jura

The most spectacular postglacial landforms inland from the coast are catastrophic rock-slope failures (rock avalanches and fragmented rockslides) and rock-slope deformations that have occurred since deglaciation on the flanks of the Paps of Jura (Beinn a' Chaolais, 733 m; Beinn an Oir, 785 m; and Beinn Shiantaidh, 757 m; Fig. 11.7). Of these, the most impressive is a rock avalanche on the eastern flank of Beinn Shiantaidh, where catastrophic collapse of the entire 300 m high slope has occurred (Fig. 11.7a). The runout debris of quartzite boulders extends for 380 m along the slope foot and 180 m outwards over the adjacent level ground and terminates at a massive arcuate distal ridge that encloses a boulder-filled depression up to 6 m deep (Fig. 11.7b). Dawson (1977) calculated that the deposit has a volume of at least 185,000 m^3, implying failure of over 0.37 Mt of rock.

The distal ridge has been interpreted as having formed through the impact of avalanching debris at the basal break of slope, causing the debris to rebound and accumulate as a crescentic ridge around the impact zone (Ballantyne et al. 2014b). On the slopes of all three mountains there is evidence for rock-slope deformation and arrested landsliding in the form of displaced rock masses, bulging slopes, antiscarps and rock benches, as well as more localised catastrophic failure in the form of runout lobes of coarse debris (Fig. 11.7c, d).

The timing of five of the catastrophic rock-slope failures has been determined using cosmogenic ^{10}Be exposure dating (Table 11.1). All appear to have occurred during the Lateglacial period between ~15.4 and ~13.7 ka, and thus within three millennia of deglaciation as determined by the age (~16.6 ka) of the nearby Sgriob na Caillich medial moraine. The primary cause of failure has been attributed to fracture propagation due to deglacial unloading and associated paraglacial stress release, but the timing of the dated failures coincides with the period of most rapid glacio-isostatic uplift, suggesting that some or all of the Jura rock-slope failures were triggered by large-magnitude earthquakes due to fault reactivation as the crust responded to deglacial unloading (Ballantyne et al. 2014a, b).

11.8 Conclusion

The islands of the Inner Hebrides described here present striking contrasts to the much more mountainous islands of the Hebrides Igneous Province (Chap. 10), being composed of much older Lewisian and Dalradian rocks and being mainly low-lying. All were over-ridden by successive Pleistocene ice sheets that moulded and scoured bedrock

Fig. 11.7 Rock-slope failures on the Paps of Jura. **a** The Beinn Shiantaidh rock avalanche, where rebound of cascading debris has produced an arcuate ridge around the runout zone. **b** The terminal ridge of the Beinn Shiantaidh rock avalanche. **c** Trench and antiscarp formed by large-scale rock-slope deformation on Beinn an Oir. **d** Lobe of runout debris at the foot of one of the rock-slope failures on Beinn a'Chaolais. (Images: Colin Ballantyne)

Table 11.1 Cosmogenic ^{10}Be exposure ages of catastrophic rock-slope failures on Jura

Beinn a'Chaolais West rock-slope failure	15.4 ± 0.9 ka
Beinn Shiantaidh rock avalanche	15.1 ± 0.8 ka
Beinn a'Chaolais South rock-slope failure	14.7 ± 0.7 ka
Beinn an Oir East rock-slope failure	14.4 ± 0.9 ka
Beinn a'Chaolais East rock-slope failure	13.7 ± 0.7 ka
Timing of deglaciation (Sgriob na Caillich moraine)	16.6 ± 0.8 ka

Data from Ballantyne et al. (2014a, b)

surfaces, and the most recent of these deposited ice-marginal moraines and outwash deposits on Islay and the longest medial moraine in Scotland on Jura.

The most spectacular and scientifically significant landforms on the islands are those produced by coastal erosion and deposition. Some of these, such as the high and intertidal rock platforms (and probably the high-level glacimarine deposits on Islay) pre-date the passage of the last ice sheet; others such as the glaciated strandflat of Coll and Tiree may even owe their origins to pre-Quaternary marine erosion. Retreat of the last ice sheet was followed by the formation of spectacular flights of Lateglacial gravel ridges on Jura and Islay, now uplifted by glacio-isostatic rebound to form the most outstanding raised beach features in the British Isles. Cutting across many coastlines is the isostatically tilted Main Rock Platform, which formed rapidly during the Loch Lomond Stade through periglacial shore erosion. The final episode of coastal change on the islands is represented by Holocene raised marine terraces and shingle ridges that represent the products of Holocene marine transgression, now elevated up to $\sim$12 m above present sea level by isostatic uplift, and associated areas of sand dunes and machair. No other part of Scotland contains such a complete assemblage of well-preserved relict coastal landforms, covering an age range that extends from the Middle Holocene possibly as far back as the Pliocene.

Inland, the flanks of the Paps of Jura provide classic examples of paraglacial rock-slope failures on deglaciated quartzite mountains. These occurred within a few millennia following deglaciation and may have been triggered by uplift-generated earthquakes of a magnitude no longer experienced in the British Isles. Collectively, the islands of Islay, Jura, Colonsay, Tiree and Coll contain a range of landforms distinctive from those elsewhere in Scotland, and in some cases the most outstanding examples of their kind.

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12 The Far Northwest: Sutherland, Assynt and Coigach

Tom Bradwell and Colin K. Ballantyne

Abstract

The far northwest of mainland Scotland is renowned for its scenery, structural complexity and geodiversity, and is designated as a UNESCO Global Geopark. The region is bisected by the Moine Thrust Zone (MTZ) , west of which a foreland of undeformed Archaean gneiss supports inselbergs of Neoproterozoic and Palaeozoic rocks, and east of which are thrust-stacked, deformed metasedimentary rocks of the Neoproterozoic Moine Supergroup. The MTZ forms a north–south belt within which rocks were extensively thrust and folded during the Caledonian Orogeny. Successive Pleistocene glaciations have resulted in an array of erosional landforms: troughs, rock basins, cirques, glacially steepened inselbergs, extensive areas of knock-and-lochan terrain and clusters of glacial megagrooves. During the last and earlier ice-sheet glaciations, the region sourced northwestward-flowing ice feeding the Minch Ice Stream, which extended far across the adjacent shelf, but by $\sim$15 ka the last ice sheet had retreated to its mountain heartland. The Loch Lomond Stade ($\sim$12.9 to 11.7 ka) witnessed reoccupation of the main mountain axis by a substantial ($\sim$350 km^2) icefield, and cirque glaciers formed on peripheral mountains; the extent of the former is mainly delimited by multiple recessional moraines, the latter by end-moraine belts. Lateglacial and Holocene landforms include outwash or delta terraces at fjord heads, sea stacks, beaches backed by sand dunes, rock-slope failures, relict talus accumulations, and active periglacial and aeolian features on high ground. Karst terrain developed on dolostones comprises sinkholes, resurgences and extensive cave networks formed by water-table lowering due to Middle and Late Pleistocene valley deepening.

Keywords

Stable foreland • Moine Thrust Zone • Inselbergs • Glacial troughs • Cirques • Knock-and-lochan topography • Glacial megagrooves • Glacial landsystems • Hummocky recessional moraines • Coastal landforms • Landslides • Periglacial features • Karst • Caves

12.1 Introduction

The region of West Sutherland, Assynt and Coigach considered in this chapter is bounded by Loch Kanaird and Loch Broom to the south, Cape Wrath in the north, and extends eastwards beyond the present north–south watershed to Loch Eriboll and the Loch Shin basin (Figs. 12.1 and 12.2). It incorporates the entire area of the North West Highlands Geopark, accredited as a UNESCO Global Geopark on account of its outstanding geodiversity. To the west lies a rugged coastline of headlands and fjords that borders The Minch, the wide strait that separates mainland Scotland from the Outer Hebrides. The region is one of the great geological complexities and contains a wide range of landscape types, including knock-and-lochan terrain, glaciated inselbergs and mountains, till-covered lowlands, areas of megagrooves and the most outstanding area of karst landforms in Scotland (Fig. 12.2). This chapter presents and discusses the key terrestrial landforms in the region, focusing in turn on the bedrock geology and outstanding structural landforms, glacial history and geomorphology, coastal, paraglacial, postglacial and coastal landforms, and karst landscapes. The offshore submarine geomorphology of The Minch and adjacent areas is considered in detail in a series of published accounts (Bradwell and Stoker 2015a, b, 2016; Bradwell et al. 2019; Chaps. 4 and 6).

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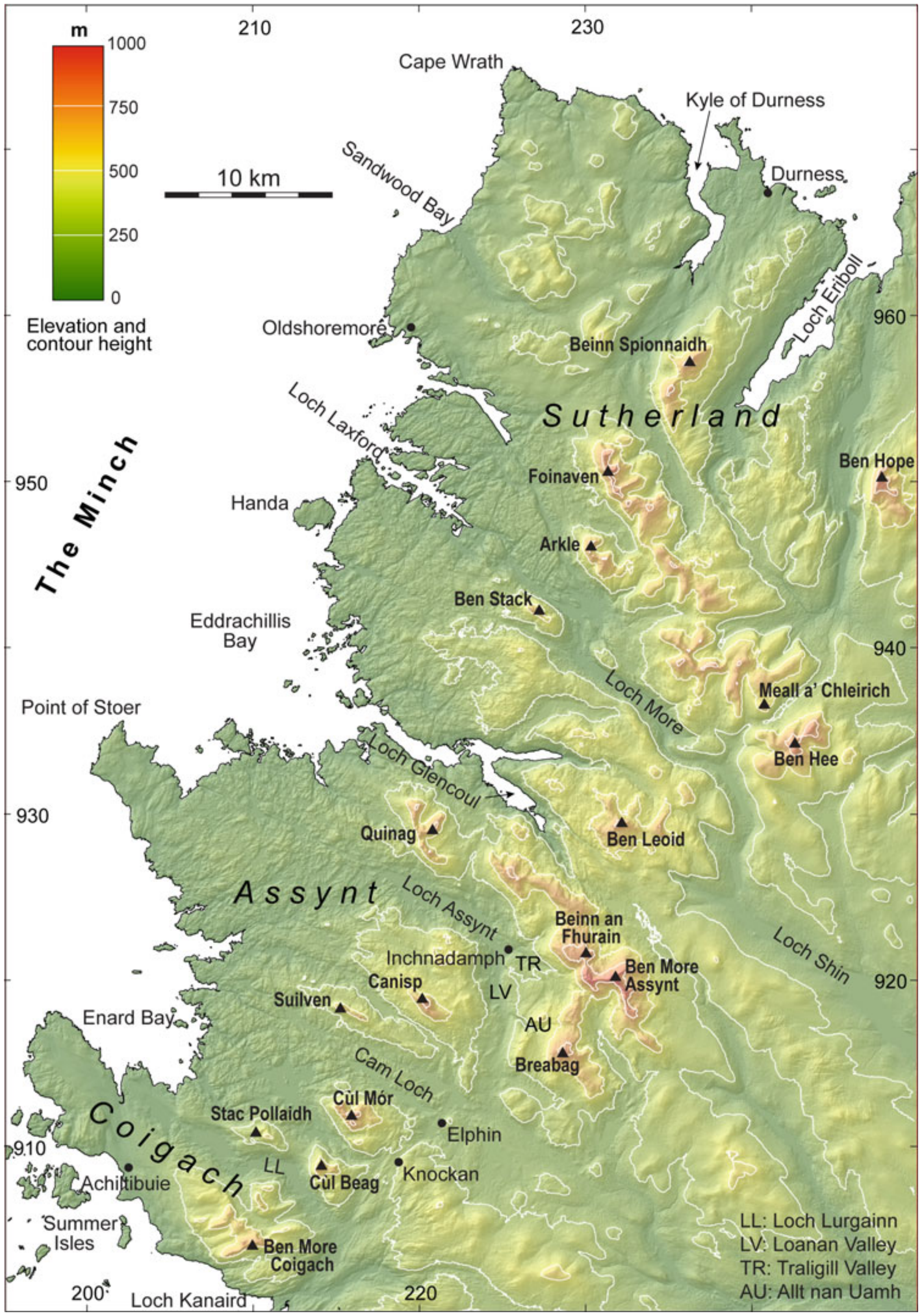

Fig. 12.1 Far northwest of mainland Scotland, showing key locations mentioned in the text. Elevation base map derived from NEXTMap data (Intermap Techologies)

12.2 Bedrock Geology, Relief and Pre-Quaternary Landscape Evolution

The region is bisected by the north–south trending Moine Thrust Zone (MTZ), which developed during the later stages (~430 to 410 Ma) of the Caledonian Orogeny (Chap. 2). West of the MTZ is a basement foreland of Archaean Lewisian gneiss, unconformably overlain by Proterozoic Torridonian sandstones and Cambro-Ordovician sedimentary rocks. East of the MTZ is the Moine Nappe, which is largely composed of metasandstones (metapsammites) of the Northern Highlands terrane (Park et al. 2002; Strachan et al. 2002; Mendum et al. 2009; Fig. 12.2a). The MTZ is bounded by two major low-angle thrust planes, the Moine Thrust and the Sole Thrust, and within this belt, older rocks are extensively thrust and folded (Fig. 12.3a). In Assynt, the

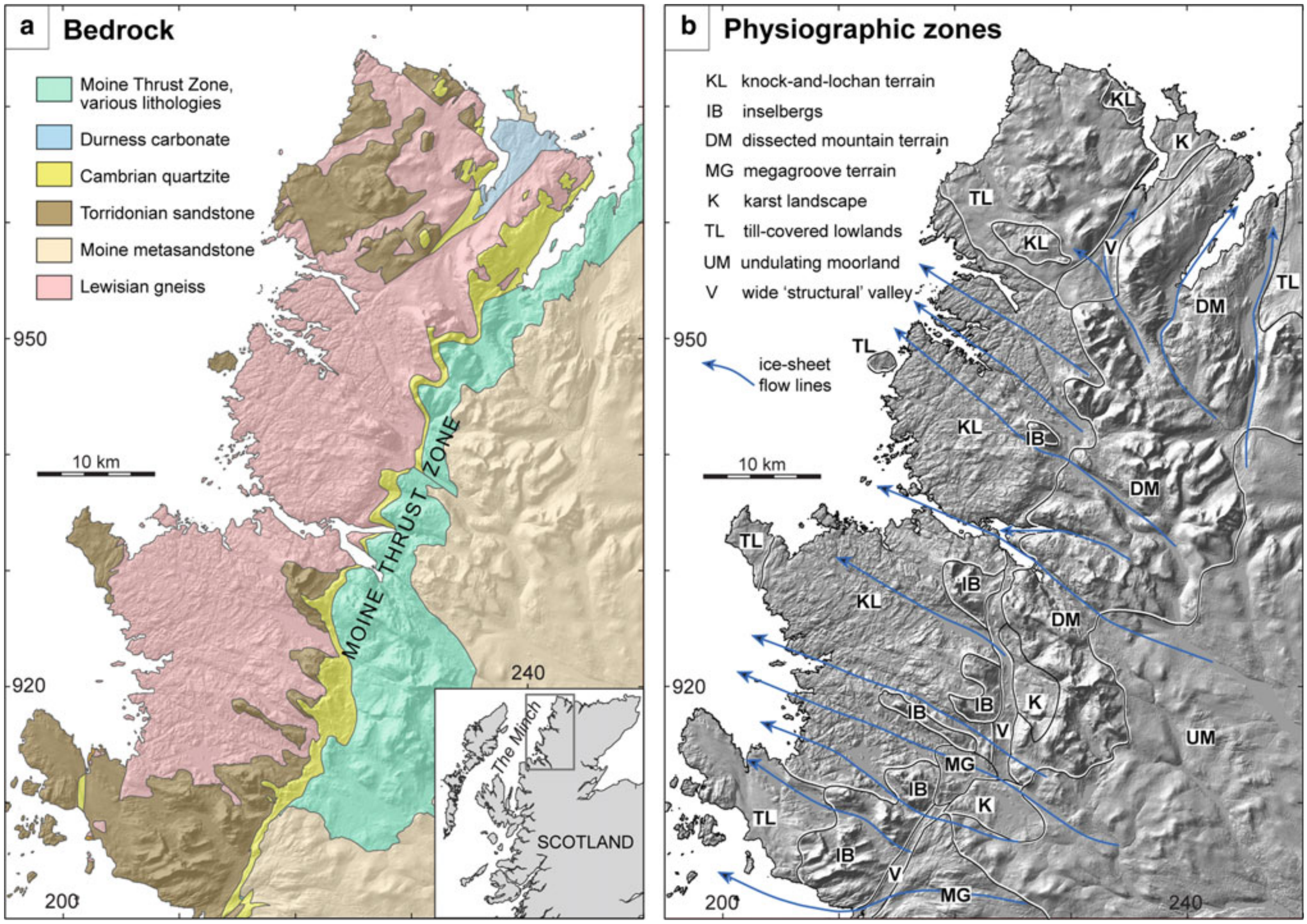

Fig. 12.2 Far northwest of mainland Scotland. **a** Generalised bedrock geology. **b** Physiographic regions, showing generalised ice-sheet flow lines (modified from Krabbendam and Bradwell 2014). Hill-shaded relief derived from NEXTMap elevation data (Intermap Technologies)

MTZ widens considerably to form the Assynt Culmination, a complex stack of thrust sheets, best exposed around Inchnadamph. In Coigach, however, the Moine Nappe directly overlies foreland rocks, a succession highlighted at the Geopark visitor centre at Knockan.

The Archaean Lewisian Gneiss Complex comprises banded felsic to intermediate orthogneisses, with pods and lenses of mafic gneiss, and is exposed across much of the ground from Cape Wrath to Enard Bay (Fig. 12.2a). The gneiss is cross-cut by WNW–ESE dolerite dykes of the Palaeoproterozoic Scourie Dyke Swarm and crossed by several shear zones up to a kilometre wide that impart a prominent WNW–ESE structural grain to the rocks and the landscape (Fig. 12.3b). Thrust slices of Lewisian gneiss in the MTZ also form the highest summit in the region (Ben More Assynt, 998 m).

A major unconformity separates the Lewisian gneiss Complex from the overlying, gently dipping rocks of the Stoer and Torridon Groups, here informally referred to as Torridonian sandstones. The rocks of the Stoer Group are limited to the peninsulas of Stoer and Coigach, and comprise sandstone and mudstone, with minor components of limestone and volcanoclastic rocks. Those of the much more extensive Torridon Group form a series of spectacular isolated inselbergs rising from the gneiss basement terrain, such as Stac Pollaidh (613 m; Fig. 12.3c), Suilven (731 m; Fig. 12.3d), Canisp (846 m) and Quinag (808 m), but also underlie low and hilly ground around Cape Wrath and in Coigach. The rocks of the Torridon Group constitute a thick sequence of strikingly uniform, coarse, thick-bedded, gently dipping arkosic red sandstones with occasional conglomerate beds and thin siltstone layers.

Both the Lewisian gneiss and Torridonian sandstones are unconformably overlain by the Cambro-Ordovician sequence, which dips 10–15° to the east along a narrow corridor west of the MTZ (Fig. 12.2a). In terms of outcrop, the dominant Cambro-Ordovician rocks are Cambrian quartzites (quartz-arenite) and Durness Group dolostones (dolomitic limestones). The Cambrian quartzite forms a dipping slab on the foreland, caps several Torridonian sandstone summits and has been thickened by thrust stacking in the MTZ to form a north–south aligned chain of

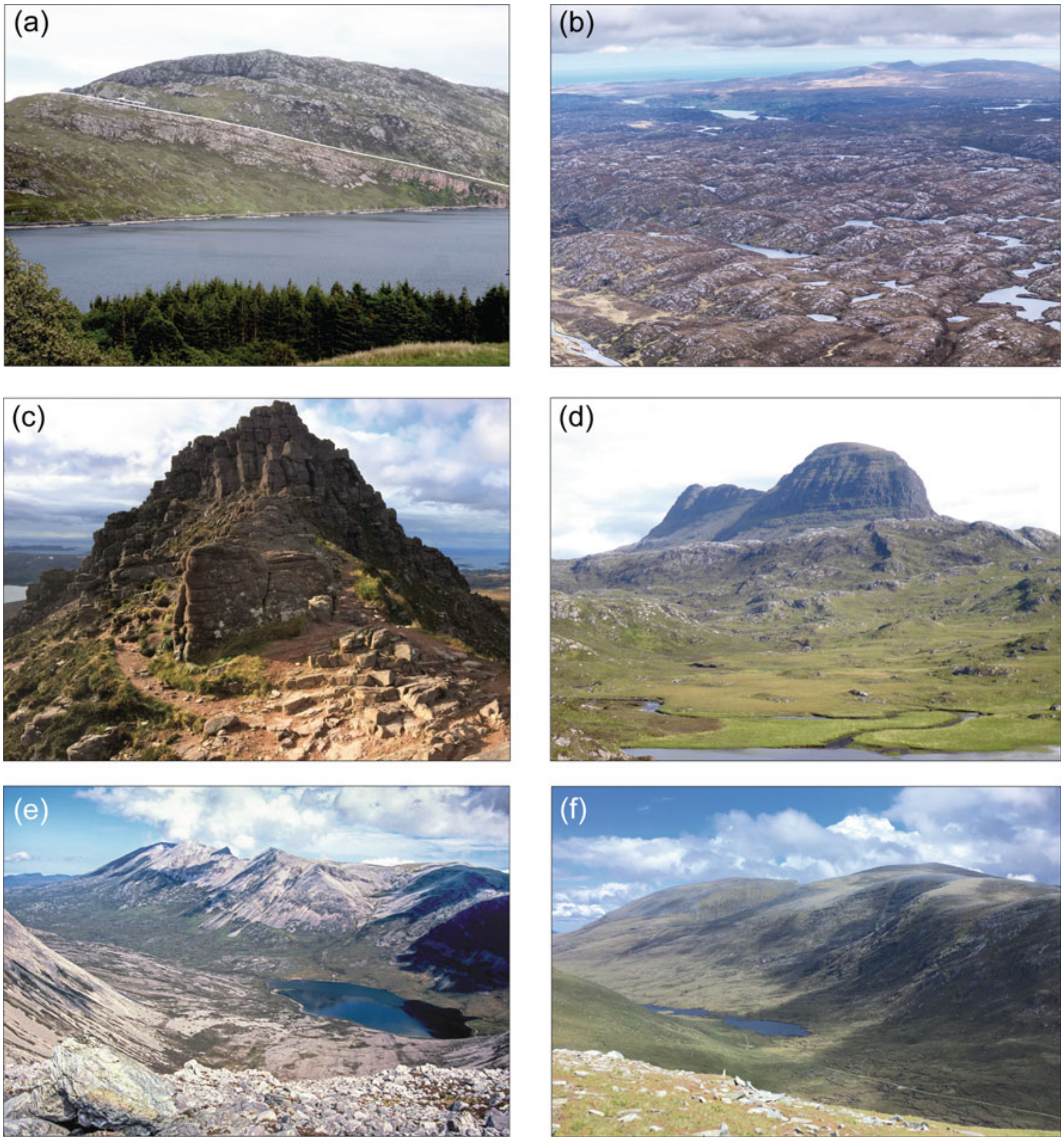

Fig. 12.3 Landscapes and rocks of the far northwest of Scotland. **a** The Ben More Thrust, part of the MTZ, where Lewisian gneiss has been thrust over a band of Cambrian quartzite that rests unconformably on Lewisian gneiss of the stable foreland. **b** Knock-and-lochan terrain on Lewisian gneiss, looking north from Ben Stack; shear zones impart a WNW–ESE structural grain. **c** Summit of Stac Pollaidh (613 m) in Coigach, a Torridonian sandstone inselberg modified by ice-sheet erosion and subsequent frost action. **d** Suilven (731 m), a Torridonian sandstone inselberg rising above a knock-and-lochan landscape of Lewisian gneiss in Assynt. **e** Thrust-stacked eastward-dipping Cambrian quartzite unconformably overlying Lewisian gneiss, Foinaven (911 m), Sutherland. **f** Ben Hee (873 m) lies east of the MTZ and is composed of Moine metasedimentary rocks. Blockfields mantle the summit plateau, and the glacial trough beyond the loch is occupied by Loch Lomond Readvance recessional moraines. (Images: **a**, **d**, **f** Colin Ballantyne; **b**, **e** John Gordon; **c** Tom Bradwell)

quartzite mountains, from Ben Spionnaidh (772 m) in the north to Breabag (815 m) in the south (Fig. 12.3e). Dolostone crops out around Durness on the north coast and within the Assynt Culmination, where it has been thrust-stacked over an area of $\sim$34 km^2 to form a glaciated karst landscape (Sect. 12.7). The Assynt Culmination also contains two large syenite plutons, the Loch Borrolan and Loch Ailsh Plutons, both intruded during the Caledonian Orogeny, and dykes and sills of microdiorite, vogesite, rhyolite and trachyte.

East of the Moine Thrust are the metasedimentary rocks of the Morar Group, the lowest part of the Neoproterozoic

Moine Supergroup. These comprise mainly arkosic to sub-arkosic metamorphosed sandstones (metapsammites) that have been interpreted as the deformed and metamorphosed equivalent of the Torridon Group rocks (Krabbendam et al. 2008). These rocks form an axis of high ground that stretches from Ben Hope (927 m) in the north to Ben Leoid (792 m) in the south (Fig. 12.3f) as well as intervening troughs and lower ground farther south and east.

The bedrock geology of the region has profoundly influenced its physiography (Krabbendam and Bradwell 2010; Fig. 12.2b). Much of the Lewisian gneiss comprises ice-scoured knock-and-lochan terrain, locally punctuated by inselbergs of Torridonian sandstone, Cambrian quartzite or, in the case of Ben Stack (721 m), Lewisian gneiss. Till-covered undulating lowland occupies the Torridonian sandstones of NW Sutherland, Stoer and Coigach. The north–south axis of mountainous terrain, comprising glacial troughs, rock basins, arêtes and cirques, forms the present drainage divide and coincides approximately with the MTZ; the mountains themselves comprise a range of lithologies (Lewisian gneiss, Cambrian quartzite, syenite and, east of the MTZ, metasandstones). East of the mountains lies an area of undulating lowland of broad hills and wide valleys, underlain by Moine rocks and extensively mantled by till and peat. A zone dominated by glacial megagrooves occurs in the extreme south. Outcrops of Cambro-Ordovician dolostone form areas of glaciated karst around Durness, Inchnadamph and Knockan. The final topographic element is the long, north–south trending composite valley system (labelled 'V' in Fig. 12.2b) that coincides with the lowest thrusts of the MTZ and with the outcrop of weak dolomitic mudstone and silty dolomite (Fucoid Bed Member) that overlies the resistant quartzite in the Cambro-Ordovician sequence.

Although the present relief of the far northwest of Scotland is thought to represent Cenozoic uplift of a relatively low-relief land surface and its subsequent modification by fluvial and glacial dissection, some elements have much more ancient origins (Chap. 3). The sub-Torridonian/sub-Cambrian platform of Lewisian gneiss (Fig. 12.3b) represents the remnants of an exhumed primeval landscape, albeit modified by subsequent uplift, weathering and erosion. Similarly, the absence of intervening Torridonian rocks between the gneiss and overlying basal Cambrian quartzite in some areas (Fig. 12.3e) indicates that dissection of the Torridonian sandstones, which ultimately led to isolation of sandstone inselbergs, was initiated prior to the deposition of the Cambrian quartzite. The main topographic elements, however, are thought to reflect: (i) major uplift in the Palaeocene (~60 to 55 Ma) in response to magmatic activity at that time, followed by episodes of renewed uplift during the Miocene and Pliocene; (ii) development of palaeosurfaces graded to successive base levels (high eustatic sea levels) during periods of relative Cenozoic tectonic quiescence; (iii) deep chemical weathering and formation of saprolite cover that was subsequently stripped by glacial erosion to reveal underlying etch surfaces (Krabbendam and Bradwell 2014); and (iv) modification of pre-glacial relief by selective glacial erosion, particularly along troughs (Hall 1991; Hall and Bishop 2002; Chap. 3). In NW Scotland, however, this sequence is complicated by adaptation of the evolving Cenozoic landscape to litho-structural controls, particularly along the MTZ where lithologies of contrasting resistance are juxtaposed.

Godard (1965) recognised evidence for four main Cenozoic palaeosurfaces of declining altitude and age in the region, each (apparently) graded to successively lower base levels as a result of episodic uplift: at ~700 to 900 m (the *surface supériere*, represented by the dissected remnants of undulating plateaux that form broadly accordant summits), ~400 to 500 m, ~180 to 300 m (the *surface écossaise*) and ~90 to 180 m (the *niveau pliocène*, which coincides with much of the area of the basement Lewisian gneiss). The validity of this scheme remains debatable because of the possibility of landscape inheritance (particularly on Lewisian gneiss), dominance of litho-structural controls of relief along and west of the MTZ, and the possibility of Cenozoic downwarping, tilting or differential uplift defined by faults.

12.3 Pleistocene Glacial History

12.3.1 Pre-Late Devensian Glaciations

Although the major relief elements of NW Scotland were established prior to the beginning of the Quaternary Period at ~2.6 Ma, the end-Neogene landscape has been extensively modified by glacial erosion during successive Pleistocene glaciations, creating some of the most spectacular glaciated scenery in the British Isles: over-deepened offshore basins, fjords, glacial troughs, breached watersheds, truncated mountain spurs, arêtes, cirques and, on the Lewisian gneiss, some of the most impressive ice-worn knock-and-lochan terrain in the British Isles (Figs. 12.2 and 12.3b, d). During the Early Pleistocene, it is likely that periods of limited (icefield or ice-cap) glaciation were predominant (Chap. 4), but after the Mid-Pleistocene Transition (~1.25 to 0.7 Ma) several successive Middle Pleistocene ice sheets covered the area and extended far out over the adjacent shelves, in some cases probably surpassing the dimensions of the last (Late Devensian) ice sheet (Stoker et al. 1993; Chap. 4).

Although the extent and timing of the Early and Middle Pleistocene glaciations of NW Scotland are uncertain, information is available for Late Pleistocene events in the region from $^{230}Th/^{234}U$ disequilibrium ages for speleothem formation in dolostone caves at the foot of mountains in

Assynt (Lawson and Atkinson 1995; Lawson 2010). These indicate temperate ice-free conditions during Marine Isotope Stage (MIS) 5e (the last interglacial), and a cluster of ages for the period ~95 to 75 ka (MIS5c–5a) implies mainly glacier- and permafrost-free conditions during this interval. Conversely, MIS 4 (~71 to 57 ka) is represented by only two ages, both with very large uncertainties, suggesting that Assynt was ice-covered during all or part of this period, consistent with wider evidence for ice expansion at this time (Merritt et al. 2019). The extent of ice cover during the Middle Devensian (~MIS 3; ~57 to 32 ka) is uncertain, but evidence from offshore cores suggests the intermittent presence of marine-terminating glaciers in Scotland during this period (Hibbert et al. 2010).

12.3.2 The Last (Late Devensian) Ice-Sheet Glaciation

The last ice sheet to cover Scotland, during the Late Devensian (≈Late Weichselian) Glacial Substage (~32 to 11.7 ka), provides an analogue for the dimensions, evolution and flow patterns of earlier ice sheets in NW Scotland. The Late Devensian ice sheet expanded from mountain source areas after ~35 to 32 ka, and for much of its existence was drained by several large ice streams (Clark et al. 2012; Hughes et al. 2014; Ballantyne and Small 2019). One of these, the Minch Ice Stream, dominated the ice-sheet flow configuration in NW Scotland (Bradwell et al. 2007; Hubbard et al. 2009; Bradwell and Stoker 2015a). The trunk of this ice stream occupied The Minch and continued northwestwards across the continental shelf within a well-defined cross-shelf trough, probably following the route of previous ice streams that developed during earlier periods of ice-sheet glaciation. As outlined below (Sect. 12.4.1), many erosional landscape features in the region are believed to have formed during similar episodes of ice streaming when ice flow was relatively rapid, warm-based (wet-based) and spatially focused.

Evidence provided by sequential flowsets (Hughes et al. 2014) and the distribution of erratic boulders (Lawson 1990) indicates that the ice divide across the far northwest of Scotland migrated during the build-up and retreat of the last ice sheet. The orientations of striae, roches moutonnées and whalebacks demonstrate that ice flow across most of the region was westwards to northwestwards, towards The Minch, probably from an ice divide located east of the present north–south watershed (Lawson 1996). However, in areas east of the watershed, there is evidence that the final directions of ice flow were southeastwards along major troughs (Finlayson and Bradwell 2008).

The retreat chronology of the Minch Ice Stream is based mainly on terrestrial cosmogenic nuclide (TCN) exposure ages and inferred connections with a sequence of 17 seabed grounding-zone wedges in The Minch and on the continental shelf (Bradwell et al. 2019). These ages suggest that the ice stream reached its maximum extent at ~30.2 ka and had begun to retreat before ~27.5 ka, and by ~23.3 ka the ice margin lay at the northern mouth of The Minch. After ~18.5 ka, retreat was apparently accelerated by the opening of a calving bay east of the Outer Hebrides, with the ice margin retreating to the northwest mainland by 16.2–15.4 ka. TCN exposure ages obtained by Bradwell et al. (2008a) for boulders on a moraine in the Loanan valley at the foot of the highest mountains in Assynt indicate deglaciation at 15.5 ± 0.8 ka (Ballantyne and Small 2019), implying that the last ice sheet in this region had retreated to its mountain source areas prior to rapid warming at the onset of the Lateglacial Interstade (~14.7 to 12.9 ka).

12.3.3 The Loch Lomond Readvance

It is not known whether glacier ice persisted in the far northwest of Scotland under the cool temperate conditions of the Lateglacial Interstade, but during this period any surviving ice masses were probably confined to high plateaux or cirques (Finlayson et al. 2011; Chaps. 4 and 13). There is abundant geomorphological evidence, however, for the renewed expansion of glaciers in the region during the Loch Lomond (≈Younger Dryas) Stade (LLS) of ~12.9 to 11.7 ka, primarily in the form of end, lateral and recessional moraines, drift limits and trimlines. Collectively, these features constitute distinctive glacial landsystems that contrast with that associated with retreat of the last ice sheet (Lukas 2006; Sect. 12.4.3) and have enabled detailed reconstruction of the dimensions of the associated Loch Lomond Readvance (LLR) glaciers.

Early reconstructions of the extent of the LLR identified only small cirque and valley glaciers in this region (Sissons 1977; Lawson 1986), but detailed remapping has demonstrated that the mountains of Assynt and west Sutherland hosted a substantial icefield, ~350 km^2 in extent, that took the form of a transection complex feeding a radial pattern of outlet glaciers that descended to low ground (Bradwell 2006; Lukas 2006; Lukas and Benn 2006; Lukas and Bradwell 2010; Fig. 12.4). The LLS age of the icefield has been confirmed by TCN exposure ages obtained for samples from boulders on the crest of an end moraine that marks the limit of one of the former outlet glaciers (Bradwell 2006; Lukas and Bradwell 2010). In addition, Sissons (1977) mapped geomorphological evidence for the development of twelve small cirque glaciers on isolated mountains, such as Ben Hope and Cul Mòr, peripheral to the icefield. Three of these palaeoglaciers, on Ben Mòr Coigach, were reconstructed by Chandler and Lukas (2017), who argued that these formed during the LLS on the basis of morphostratigraphic evidence and contrasting landsystems inside and outside the associated end moraines.

Contouring of the reconstructed Assynt-Sutherland icefield (Fig. 12.4) permitted Lukas and Bradwell (2010) to estimate the former equilibrium-line altitudes (ELAs) for individual palaeoglaciers (268–429 m) and for the entire icefield (334 m). The area-weighted ELA calculated for the three cirque glaciers on Ben Mòr Coigach (328 ± 16 m) is almost identical. In both cases, however, it is possible that some glaciers were fed by thin, cold-based ice on high ground or by snow blown from upwind plateaux, so the true ‘regional’ ELA may have been slightly higher. The reconstructed ELAs are nonetheless reasonably consistent with a

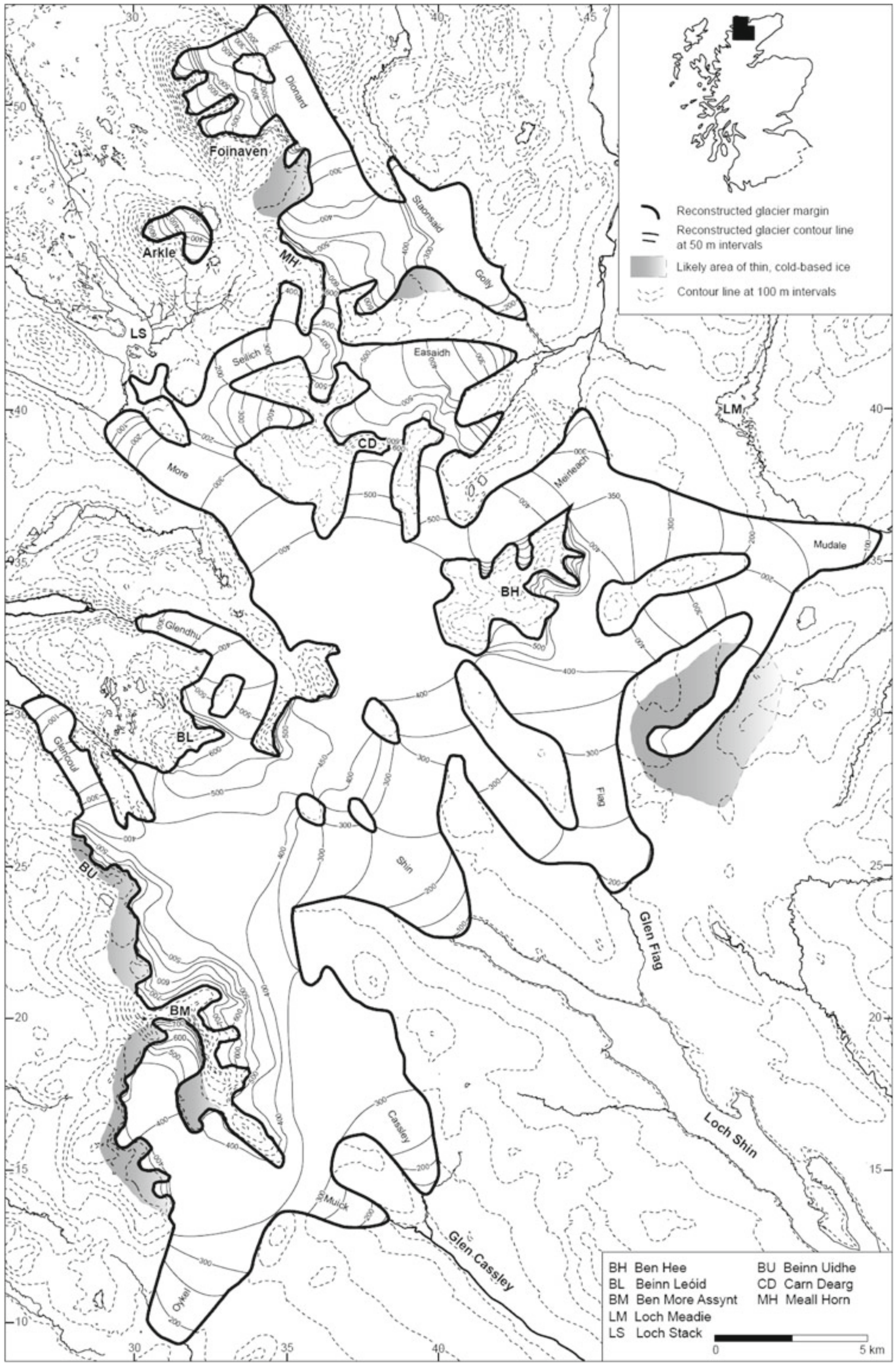

Fig. 12.4 Reconstruction of the icefield that developed on the mountains of Assynt and Sutherland during the Loch Lomond Stade. (From Lukas and Bradwell (2010) J Quat Sci 25:567–580 © 2010 John Wiley & Sons Ltd)

general pattern of westward and northward decline in LLS ELAs across the Scottish Highlands (Chandler et al. 2019), a pattern attributed to the dominance of moisture-bearing westerly airmasses during the stage. Estimation of associated mean annual precipitation is fraught with uncertainties, particularly relating to seasonality (Golledge et al. 2010). By assuming a seasonally neutral estimate, however, and July sea-level temperature of 8.5 ± 0.3 °C, Chandler et al. (2019) estimated that stadial sea-level precipitation in the region was of the order of 1800–1900 mm a^{-1}, implying annual precipitation of ~2100 to 2300 mm at the icefield ELA and suggesting that the LLS glaciers in this area had a high mass turnover, flowed rapidly and probably responded nimbly to short-term climatic forcing (Benn and Lukas 2006; Lukas and Benn 2006).

12.4 Glacial Landforms

As is the case for much of the Western Highlands and Hebrides, the glacial landforms of the far northwest fall into three categories: erosional landforms, including both large-scale features that developed over successive glacial cycles and small-scale features that achieved their current form during the last ice-sheet glaciation; depositional landforms associated with the retreat of the last ice sheet across the area; and the distinctive assemblage of landforms associated with mountain glaciation during the LLS.

12.4.1 Erosional Landforms

The upland spine of the region (Fig. 12.2b) is a zone of classic mountain glaciation, characterised by deep glacial troughs with a pronounced NW–SE alignment. Within this zone, the northwesterly flow of successive glaciers has excavated deep rock basins, some to well below present-day sea level; Loch More and Loch Assynt occupy basins more than 80 m deep. Almost all the mountains in this zone, irrespective of lithology, are indented with deep cirques with predominant northerly to easterly orientations; these exhibit a gradual eastward rise in cirque-floor altitude that is thought to reflect the influence of westerly airmasses in feeding snowfall to highly dynamic cirque glaciers along the coastal margin during successive periods of limited glaciation (Barr et al. 2017).

West of the main mountain axis lies a unique zone of glaciated inselbergs, such as Stac Pollaidh and Suilven, which rise above an irregular platform of Lewisian gneiss (Fig. 12.3c, d). Although these inselbergs represent pre-Pleistocene residuals of the Torridonian cover rocks, isolated by prolonged slope retreat, their present cliffed slopes are the product of steepening by ice flowing northwestwards to feed the Minch Ice Stream during successive periods of ice-sheet glaciation.

Equally striking are the extensive areas of low-lying, ice-scoured terrain on the Lewisian gneiss that forms most of the zone along the western seaboard north of Coigach (Fig. 12.2). Here, glacial erosion has gouged out small rock basins and fracture belts to create a landscape dominated by rocky knolls separated by lochs and peat-filled hollows (Fig. 12.3b, d). Taking their name from the Gaelic words *cnoc* (low hill) and *lochain* (small loch), the knock-and-lochan landscapes of Assynt and Sutherland represent the finest example of ice-scoured basement topography in the British Isles, rivalled only by similar terrain in the Outer Hebrides (Chap. 9). Krabbendam and Bradwell (2014) have argued that the evolution of glaciated gneiss terrain in NW Scotland was a multi-stage process, involving: (i) long-term pre-glacial chemical weathering, forming a deep saprolite cover overlying an irregular weathering front; (ii) stripping of the saprolite cover by glacial erosion during the initial periods of glacier expansion across the area to form a rugged, uneven etch surface; and (iii) subsequent modification of the exposed resistant bedrock by subglacial abrasion and plucking during later periods of glaciation. Glacial erosion of the etch surface has focused on belts of weathered or shattered rock, leaving rounded bosses of more resilient rock upstanding as whalebacks, roches moutonnées or low hills.

The knock-and-lochan landscape of NW Scotland has traditionally been regarded as an area of monotonously uniform 'areal scouring' by glacier ice (Fig. 12.3b). Detailed mapping of the distribution of a range of small- to medium-scale erosional features (whalebacks, roches moutonnées, P-forms and striae) over a ~200 km^2 area inland from Loch Laxford in Sutherland, however, has demonstrated that such features exhibit a distinct zonation that is unrelated to lithology (Bradwell 2013). This research identified a NW-aligned corridor, dominated by wholly abraded bedrock features and P-forms, flanked by transitional zones where stoss-and-lee forms (roches moutonnées) become increasingly common, and outer zones where stoss-and-lee forms predominate. As Loch Laxford and its hinterland were occupied by ice flowing NW–NNW to feed the Minch Ice Stream, Bradwell (2013) interpreted this pattern in terms of palaeoglaciological contrasts between a fast-flowing ice-stream onset zone along the central low-lying corridor, flanked on both sides by transitional flow zones, with outer zones dominated by slower-flowing ice. He also noted that the degree of glacial modification of bedrock surfaces declines with altitude on mountains in the area (Ben Stack and Arkle), and that the summits exhibit no modification by glacier ice and are mantled by periglacial regolith (see below). Bradwell related this trend to contrasts in thermal regime and rheology of the ice within the last ice sheet, with 'hard' cold-based ice occupying the highest ground, 'soft' warm-based ice occupying the intervening low ground and a gradual altitudinal transition between the two. Lithological contrasts also influence the nature of landforms in areas of 'areal scouring' by successive ice sheets. Erosional bedforms on Torridonian sandstone include roches moutonnées with smooth surfaces and concave lee sides, whalebacks and elongate P-forms, indicating a predominance of glacial abrasion. Bedforms on resistant but densely jointed Cambrian quartzite, however, are typically more angular with plucked lee faces, suggesting that plucking rather than abrasion dominated their evolution (Krabbendam and Glasser 2011).

Amongst the most striking landforms of the far northwest of Scotland are megagrooves, large linear erosional features incised into resistant bedrock by the passage of successive ice sheets. First recognised by Bradwell (2005) in Assynt, bedrock megagrooves have now been identified quite extensively in NW Scotland (Bradwell et al. 2008b) and farther afield (Newton et al. 2018). Individual megagrooves range from 100 m to 10 km in length and from 3 m to 30 m in depth, and typically occur in groups with a strong common orientation (Fig. 12.5a). In Assynt, megagrooves are cut into very hard bedrock lithologies, including Cambrian quartzite, Moine psammite, Torridonian sandstone and Lewisian gneiss, sometimes crossing lithological boundaries without a change in form, suggesting that they developed as a consequence of powerful focused erosion under successive thick ice sheets. Megagrooves occur over a range of elevations but typically below 300 m. They are absent on the permeable Durness Limestone, possibly suggesting a genetic link between their excavation and subglacial hydrology (Bradwell 2005).

At their type site in Assynt, on the quartzite dip slope west of Elphin and on the low ground around Cam Loch, the megagrooves are gorge-like dry channels 5–20 m deep with width:depth ratios of <3:1. Most trend ESE–WNW along the former direction of ice-sheet flow as inferred from erratic carry and other independent evidence (Lawson 1990; Bradwell et al. 2008b). Rock outcrops between and within the grooves host an abundance of smaller erosional forms including striae, chattermarks, P-forms and plucked surfaces, indicating ice-bed contact and possibly the presence of high-pressure water in subglacial cavities or channels (Bradwell 2005). These large-scale megagrooves and the intervening ridges represent a type of mega-scale glacial lineation, a feature widely considered to be an erosional hallmark of former ice streams or their tributaries (Clark 1993; Bradwell et al. 2008b; Newton et al. 2018).

Small-scale features of glacial erosion are well represented on the hard bedrock lithologies of Assynt, notably Cambrian quartzite, Torridon Group sandstone and quartz veins in Lewisian rocks. Scalloped (plastically sculpted) P-forms and striae > 5 mm deep are particularly well-preserved on coastal rock outcrops at Loch Kanaird in Coigach, where overlying sediment has been removed by wave action to reveal pristine glacially abraded and ice-moulded rock surfaces (Fig. 12.5b). Excellent examples of ice-abraded and plucked rock slabs also occur on the lower parts of the quartzite dip slope of Canisp, where well-preserved striae, friction cracks, percussive marks and other stress-induced fractures record predominantly WNW-directed former ice flow (Fig. 12.5c). These subtle small-scale landforms attest to a thick ice sheet with abundant basal debris flowing upslope against the local gradient, polishing and comminuting bedrock with a high compressive strength. In some localities, evidence for more than one flowset is preserved, indicating switches in flow direction during the lifetime of the last ice sheet (Lawson 1996). The pristine condition of these striae, and others in NW Scotland, indicates that these ice-abraded surfaces have been subject to little or no surface loss since deglaciation.

Although erosional bedforms dominate much of the low ground in the region, evidence for glacial erosion is absent or very limited on the summits of the highest mountains. These are mantled by periglacial blockfields (Fig. 12.3f), mostly less than 2 m deep and consisting of openwork boulders or clasts embedded in fine sediment. A trimline representing the boundary between glacially modified outcrops and in situ blockfields has been mapped by McCarroll et al. (1995). This trimline descends gradually northwestwards across the region, from 810 to 850 m on Ben More Assynt to $\sim$600 m on Ben Spionnaidh near the north coast. It was initially interpreted as representing the maximum altitude of the last ice sheet, but this interpretation is inconsistent with extension of the ice sheet far offshore across the adjacent shelf. On some summits, moreover, areas of ice-scoured bedrock occur, and erratics rest on blockfield debris (Mathers et al. 2010). Erratics resting on blockfields in Wester Ross (Chap. 13) have yielded TCN exposure ages averaging 16.1 $\pm$ 1.0 ka, demonstrating that they were emplaced by the last ice sheet (Fabel et al. 2012). This finding indicates that the last ice sheet over-rode all mountains in the region, but that the blockfields escaped erosion. It is now generally considered that the trimline represents a former englacial thermal boundary within the last ice sheet, separating the upper limit of erosion by wet-based ice at lower altitudes from a zone of cold-based ice on summits (Bradwell 2013; Hopkinson and Ballantyne 2014). It also indicates that the last, and probably earlier, ice sheet(s) in this region were polythermal and thermally zoned (Mathers et al. 2010; Fame et al. 2018).

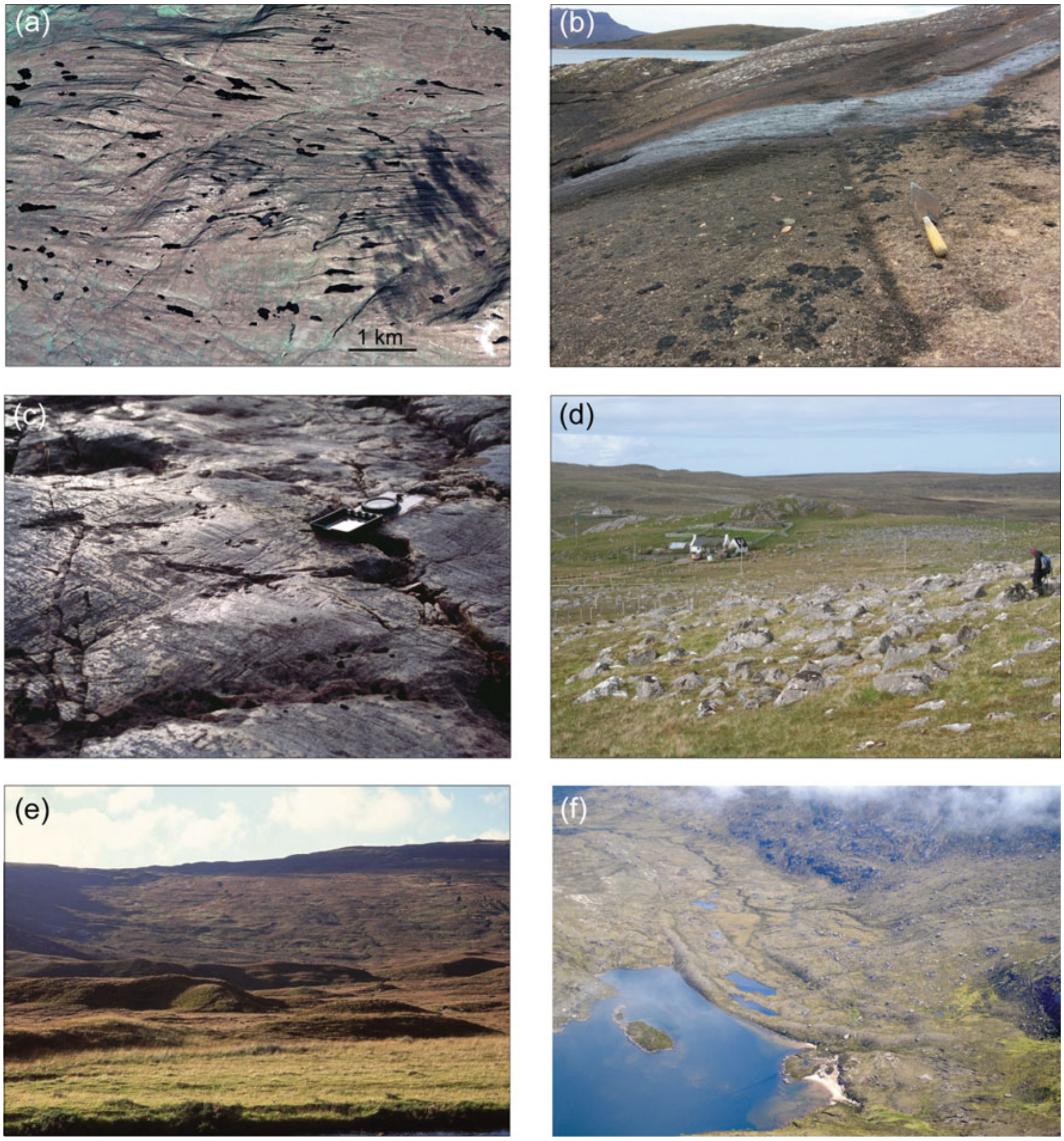

Fig. 12.5 Glacial landforms in NW Scotland. **a** Glacial megagrooves on Moine rocks north of Ullapool. **b** Glacially abraded and ice-moulded sandstone outcrops with large striae and small grooves, Loch Kanaird, Coigach. **c** Shallow, cross-cutting striae and friction cracks on glacially polished Cambrian quartzite, Assynt. **d** The Oldshoremore boulder belt moraine, Sutherland. **e** Loanan valley moraines, Assynt. **f** End moraine-belt representing the limit of a cirque glacier of probable Loch Lomond Stadial age on Ben Mòr Coigach. (Images: **a** Google Earth™; **b–e** Tom Bradwell; **f** Ben Chandler)

12.4.2 Depositional Landforms Associated with the Last Ice Sheet

Although moraine sequences representing readvances of the last ice sheet have been detected widely offshore of NW Scotland (Bradwell and Stoker 2015b), onshore landforms and deposits relating to the retreat of the last ice sheet in Sutherland and Assynt are much rarer (Bradwell 2003; Bradwell et al. 2008a). Following retreat of the Minch Ice Stream from the North Minch at ~22 to 20 ka (Bradwell et al. 2019), ice-sheet moraines were deposited on projecting headlands as the ice-sheet margin stabilised at the marine-terrestrial transition. On the Stoer headland in Assynt, for example, ice-marginal moraines form an assemblage of ridges and mounds up to 5 m high surrounded by gently undulating, boulder-strewn, till-covered ground extending over an area of ~4 km^2. The ridges are composed of poorly sorted, sandy, matrix- and clast-supported diamicton of Torridonian sandstone debris with occasional exotic Lewisian gneiss and Cambrian quartzite clasts. TCN exposure-age dating of perched glacially transported boulders closely associated with the morainic ridges indicates deposition at ~18 to 16 ka (Bradwell et al. 2019).

A linear boulder accumulation 40–150 m wide related to ice-sheet retreat occurs NW of Oldshoremore on the west Sutherland coast, where it is traceable for 6–8 km in a NW–SE direction (Mathers 2014). This 'Oldshoremore boulder belt moraine' takes the form of a relatively well-sorted and evenly distributed spread of boulders, rarely more than one boulder deep, comprising clasts of local gneiss bedrock with occasional far-travelled erratics (Fig. 12.5d). Samples from boulders within the moraine belt and from erratic boulders perched on surrounding bedrock outcrops have yielded a spread of apparent TCN exposure ages (Mathers 2014), leaving the timing of moraine formation inconclusive. Its relation to dated offshore moraines around The Minch (Bradwell et al. 2019), however, indicates that the boulder belt pre-dates the age (~18 to 16 ka) of the Stoer moraines described above, consistent with a general southwards retreat of the ice-sheet margin.

Farther inland, a chaotic assemblage of large mounds and ridges up to 10 m high and ~50 m wide occurs in the Loanan valley, near the foot of the highest mountains in Assynt (Fig. 12.5e). Although originally attributed to deposition by a small glacier during the LLS (Lawson 1986), remapping by Bradwell (2006) related the landform assemblage at this site to deposition at the margin of a residual ice-cap or outlet glacier complex centred on the Assynt mountains or farther south. Recalibration of TCN exposure ages obtained for samples from boulders on moraine crests at this site (Bradwell et al. 2008a) has yielded a recalculated mean age of 15.5 ± 0.8 ka (Ballantyne and Small 2019), confirming deposition of the moraines at the margin of a residual ice mass as the last ice sheet retreated. The age of the Loanan valley moraines is consistent with that of the Wester Ross Readvance (15.3 ± 0.7 ka) farther south, suggesting that they represent a climatically driven readvance of the ice margin at that time (Chap. 13). Boulders on lateral moraines near Achiltibuie in Coigach, independently dated by Bradwell et al. (2008a) and Ballantyne et al. (2009), have yielded (recalibrated) mean TCN exposure ages of 15.2 ± 0.9 ka and 15.1 ± 0.8 ka, respectively. These moraines represent the northern margin of a marine-terminating outlet glacier that occupied outer Loch Broom at the time of the Wester Ross Readvance. The exposure ages for the Loanan valley and Achiltibuie moraines confirm that the remnant ice sheet was largely land-based by ~15 ka, but with tidewater glaciers still existing in the fjords south of Coigach.

12.4.3 Landsystems of the Loch Lomond Readvance

The mountains of the far northwest of Scotland provide outstanding examples of two of the characteristic LLR landsystems identified by Bickerdike et al. (2018): the cirque glacier landsystem and the icefield landsystem. The landforms associated with the three small (0.3–0.6 km^2) cirque glaciers of inferred LLS age that formed below the cliffs of Ben Mòr Coigach are archetypes of the former: the terminal zone of all three palaeoglaciers is defined by arcuate end moraines (Fig. 12.5f), inside of which are belts of nested moraine ridges, hummocks and glacially deposited boulders (Chandler and Lukas 2017). Overlapping of moraine ridges and the restriction of depositional landforms to the moraine belt suggest that these glaciers, and most other cirque glaciers in the region (Sissons 1977), oscillated close to their terminal limits before experiencing uninterrupted retreat.

The Assynt-Sutherland icefield took the form of a transection glacier complex, overlooked by nunataks and fed by ice nourished in cirques, from which twelve main outlet glaciers radiated to terminate on low ground (Fig. 12.4). The limits of the icefield are constrained by trimlines, drift limits, lateral moraines and in some valleys, such as Glen Oykel, by large end moraines. The dominant components of this landsystem, however, are suites of nested hummocky recessional moraines that mark the oscillatory retreat of outlet glaciers towards their source areas (Fig. 12.6). The recessional moraines are up to 15 m high, usually tens of metres apart, often with steep (~30 to 35°) ice-proximal slopes and gentler distal slopes, and have undulating crestlines, giving them a hummocky appearance. They typically trend obliquely across hillsides, forming chevron or arcuate patterns. Detailed study of the sediments and structures within these moraines by Lukas (2005) and Benn and Lukas (2006) has demonstrated that they represent ice-contact fans composed of stacked glacigenic debris-flow deposits and/or glacifluvial sediments formed at successive ice-margin positions during overall glacier retreat, but often modified by subglacial deformation during subsequent ice-margin advance. Analysis of the spatial patterns of such recessional moraines in the valleys occupied by the Assynt-Sutherland icefield indicates moraine formation every 3–23 years, depending in part on topography and glacier size and hypsometry. Such frequency implies that the retreating glaciers had short (ice-margin) response times, indicative of dynamic behaviour associated with a maritime glacial environment (Lukas and Benn 2006).

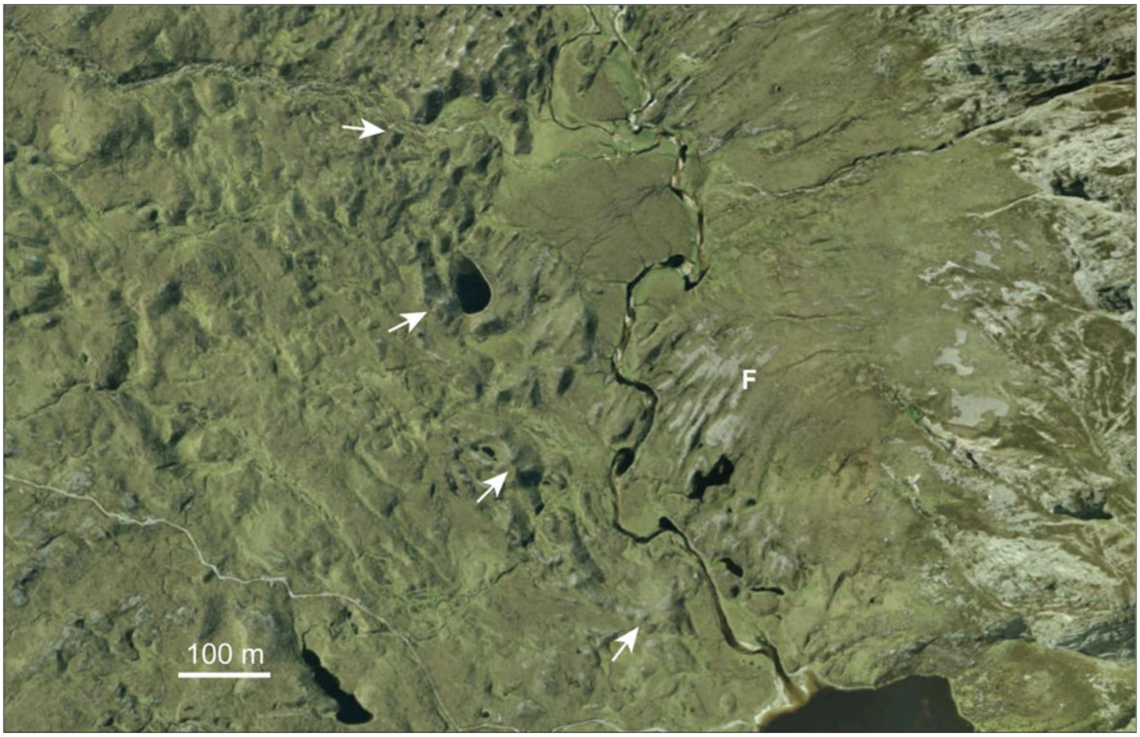

Fig. 12.6 Hummocky recessional moraines deposited in Strath Dionard (at the foot of Foinaven) during retreat of a LLR outlet glacier of the Assynt-Sutherland icefield. The white arrows indicate prominent moraine fragments. F: fluted moraines. (Google Earth™ image)

12.5 Coastal Geomorphology

Changes in relative sea level (RSL) during the Lateglacial and Holocene are poorly constrained in the region, but both model projections and the available field-based data indicate high RSLs during ice-sheet deglaciation, succeeded by rapid RSL fall and marine regression due to rapid Lateglacial glacio-isostatic uplift, with RSLs probably falling below present sea level. Following an Early Holocene lowstand, RSLs rose as eustatic sea-level rise outstripped rates of glacio-isostatic uplift, leading to a mid-Holocene sea-level highstand, before RSLs fell gradually to present levels (Shennan et al. 2000, 2018; Hamilton et al. 2015; Long et al. 2016).

The altitude of the Lateglacial marine limit has not been systematically determined. Gravel terraces at altitudes of 14–30 m OD (Ordnance Datum) occur at the heads of several sea lochs (e.g. Lochs Eriboll, Laxford, Glencoul, Kanaird and the Kyle of Durness) and are interpreted as thick outwash and glacideltaic sediments deposited during ice-sheet retreat (Bradwell 2010). Exposures in the terraces at Loch Kanaird (Fig. 12.7a), for example, exhibit mass-flow units, syn-sedimentary slump bedding, low-angle sand and gravel foresets and climbing ripples, suggesting that the lowest terrace sediments represent a prograding delta front deposited by glacial meltwater when RSL was ~10 to 15 m higher than present. Other accounts, however, place the Lateglacial marine limit in Assynt no higher than +6 to 8 m OD (Hamilton et al. 2015; Long et al. 2016). The Holocene RSL highstand in Coigach did not exceed +2.6 m OD and on the north coast was no higher than +1.0 m, and RSL had begun to fall by ~6.2 to 5.9 ka (Shennan et al. 2000; Long et al. 2016). Studies of salt-marsh stratigraphy at two sites in Sutherland (Barlow et al. 2014) indicate that the last 2000 years have been characterised by a very gradual fall in RSL to present levels.

Rock coasts dominate the present littoral of Sutherland, Assynt and Coigach, in the form of cliffs, bedrock ramps and boulder-strewn intertidal shore platforms, with numerous low, ice-scoured islets and skerries in the nearshore zone. The highest cliffs are those of Lewisian gneiss and Torridonian sandstone on the exposed coastline south and east of Cape Wrath, where they achieve heights of up to 280 m. The most iconic coastal landforms are two sea stacks that have been severed from the adjacent Torridonian sandstone cliffs by marine erosion: the 65-m-high stack of Am Buachaille (the Herdsman) near Sandwood Bay in Sutherland, and the 60-m-high Old Man of Stoer, near Point of Stoer in Assynt (Fig. 12.7b). Smoo cave, on the north coast, is a large coastal cave formed partly by solutional processes and partly by marine erosion of Durness limestone. Extensive sandy bay-head beaches are limited to two locations on the west coast of

Fig. 12.7 **a** Outwash terrace terminating at 10–15 m OD above the modern beach at the head of Loch Kanaird, Coigach. **b** Am Buachaille sandstone sea stack, Sandwood Bay, Sutherland. **c** Sandwood Bay beach and eroding dunes, Sutherland. **d** Failure scar and hummocky fan of rock avalanche debris, Cùl Beag, Coigach. **e** Rock-slope failure on Lewisian gneiss, Ben Stack, Sutherland. **f** Arrested rockslide on Cambrian quartzite, Breabag, Assynt. (Images **a**, **f** Colin Ballantyne; **b**, **c** John Gordon; **d**, **e** Google Earth^{TM})

Sutherland, at Oldshoremore and Sandwood Bay; the latter has been designated a Site of Special Scientific Interest on account of its shifting dunes, back-barrier lagoon and *machair*, a floristically diverse, windblown, calcareous shell-sand plain (Fig. 12.7c; Chap. 9). In northern Sutherland, a sand beach at Balnakeil Bay (near the mouth of the Kyle of Durness) is backed by a dune belt incised by several large, steep-sided blowthrough corridors. These funnel windblown transgressive sand waves onto higher ground, where sand is stabilised by machair vegetation, or over east-facing cliffs, where it is lost to the beach-dune system (Hansom 2003).

12.6 Postglacial Landforms

12.6.1 Landslides and Talus Accumulations

Postglacial rock-slope failures (RSFs) are less common in this region than elsewhere in the NW Highlands but include several interesting examples. Of 34 postglacial RSFs >0.01 km^2 in area, 18 are located on Moine metasedimentary rocks and the remainder on Cambrian quartzite (7), Torridonian sandstone (6) and Lewisian gneiss (3).

Translational rockslides predominate, most of which are arrested slides where the slipped mass has retained some integrity, but at a few sites the mobile rock mass has disintegrated to form a fragmented slide; rock-slope deformations are limited to four or five sites, mostly on Moine rocks. The region also contains a similar number of debris-free failure scars, sites of rock-slope failure where displaced blocks or runout debris have been entirely removed by glacier ice. The persistence of such failure scars outside the limit of the LLR indicates limited erosion of upper slopes by the last ice sheet (Cave and Ballantyne 2016; Chap. 14).

Evidence for possible rock avalanching is limited to Cùl Beag (769 m), a Torridonian sandstone mountain in Coigach, where a rupture scar on the uppermost slopes overlooks an extensive fan of hummocks with a partial veneer of boulders (Fig. 12.7d). The scar is 0.04 km^2 in area, and the failure plane descends from ~740 m to ~625 m at a gradient of ~35° before terminating downslope at a line of cliffs 80 m high. The fan of hummocks descends from ~300 m to ~60 m, covering >0.7 km^2. The hummocks are rounded, typically 20–40 m wide and up to 5 m high. The unusual form and exceptionally long runout of this RSF suggest that the rock released from the rupture scar accelerated and disintegrated as it passed over the subjacent cliff, acquiring sufficient kinetic energy to splay out across the footslope as the hummocky runout fan.

The largest RSF on Lewisian orthogneiss is a failure on Ben Stack (721 m) in Sutherland. A 480-m-long scarp along the summit ridge marks the downslope displacement of the SW flank of the mountain by up to 15 m, with partial collapse and runout of debris at its NE shoulder (Fig. 12.7e). On Cambrian quartzite, the largest RSF is an arrested slide below Breabag in Assynt (Fig. 12.7f). The main slide block occupies an area of 0.04 km^2 and exhibits limited displacement, being separated from the plateau rim by a trench 20–30 m wide. Most of the detached block is intact, but the lower part of the block is partly separated from the upper by a 40 to 50 m wide gash that suggests that the block rotated during failure. Secondary rock toppling is evident along the gash headscarp and several oblique rock ribs immediately downslope. The northernmost part of the landslide has collapsed completely, with the associated runout debris forming a 210 m long cone of bouldery debris.

A cluster of RSFs on Moine metasedimentary rocks occurs on and near Ben Hee (873 m) in Sutherland. The largest (0.43 km^2) and most enigmatic of these is located in a cirque east of Ben Hee summit and consists of a slipped mass ~500 m long that terminates in broad transverse ridges. Jarman and Lukas (2007) concluded that failure involved translational sliding along a failure plane inclined at 18°, but as the residual friction angle of psammite is typically much higher (>30°), this seems unlikely; an alternative explanation is that this RSF represents quasi-rotational slumping (Chap. 14). On nearby Meall a' Chleirich (628 m), the central part of an arrested slide disintegrated to form a runout tongue of large boulders at ~7.9 ka, ~3 ka after deglaciation of the site at the end of the LLS (Ballantyne et al. 2014).

Outside the limits of LLS glaciation, cliffs on Torridonian sandstone and Cambrian quartzite mountains are skirted by thick talus accumulations. These are extensively vegetated, gullied and scarred by the tracks of debris flows, indicating that erosion now predominates over rockfall deposition (Fig. 12.8a). Like the vegetated taluses on Skye (Chap. 10), these are believed to be paraglacial landforms that accumulated as a result of deglacial stress release and frost wedging promoting rockfall from jointed rockwalls during the Lateglacial period. Gully-wall exposures in the talus accumulations below sandstone cliffs show that in situ rockfall debris is overlain by debris-flow and slopewash deposits, locally intercalated with organic soil horizons (Hinchliffe and Ballantyne 2009). Radiocarbon dating of the buried soils suggests that reworking of the talus by debris flows commenced on upper slopes in the Middle to Late Holocene and was succeeded by progressive gully incision and extension, with intermittent debris flows feeding the growth of slope-foot debris cones (Chap. 13).

12.6.2 Holocene Periglacial and Aeolian Landforms on High Ground

Evidence for active solifluction in the region takes the form of vegetated solifluction terraces and lobes with steep risers up to about a metre high but is limited to frost-susceptible soils above ~580 m on metasedimentary rocks (Ballantyne and Harris 1994). Radiocarbon dates obtained for peat buried by the downslope advance of a solifluction lobe on Arkle (757 m) range from ~5.9 to ~4.3 cal ^{14}C ka (Mottershead 1978), implying that solifluction has been at least intermittently active since the Middle Holocene. Active frost-sorted patterned ground is mainly limited to pockets of unvegetated ground on quartzite regolith, notably on Beinn an Fhurain (804 m) in Assynt, where active sorted nets 15–60 cm wide co-exist with much larger sorted patterns of Lateglacial age (Godard 1965; Ballantyne 1995).

Holocene aeolian landforms, such as deflation surfaces and aeolian sand sheets, are mainly restricted to the highest ground on some Torridonian sandstone mountains, though aeolian sands also occur on some quartzite summits (Pye and Paine 1983). High-level aeolian features are best represented on the Torridonian sandstones of Ben Mòr Coigach (743 m), where areas of deflation surface and vegetation-covered sand sheets occupy most ground above 550 m (Fig. 12.8b). The deflation surfaces are carpeted by lag gravels and locally

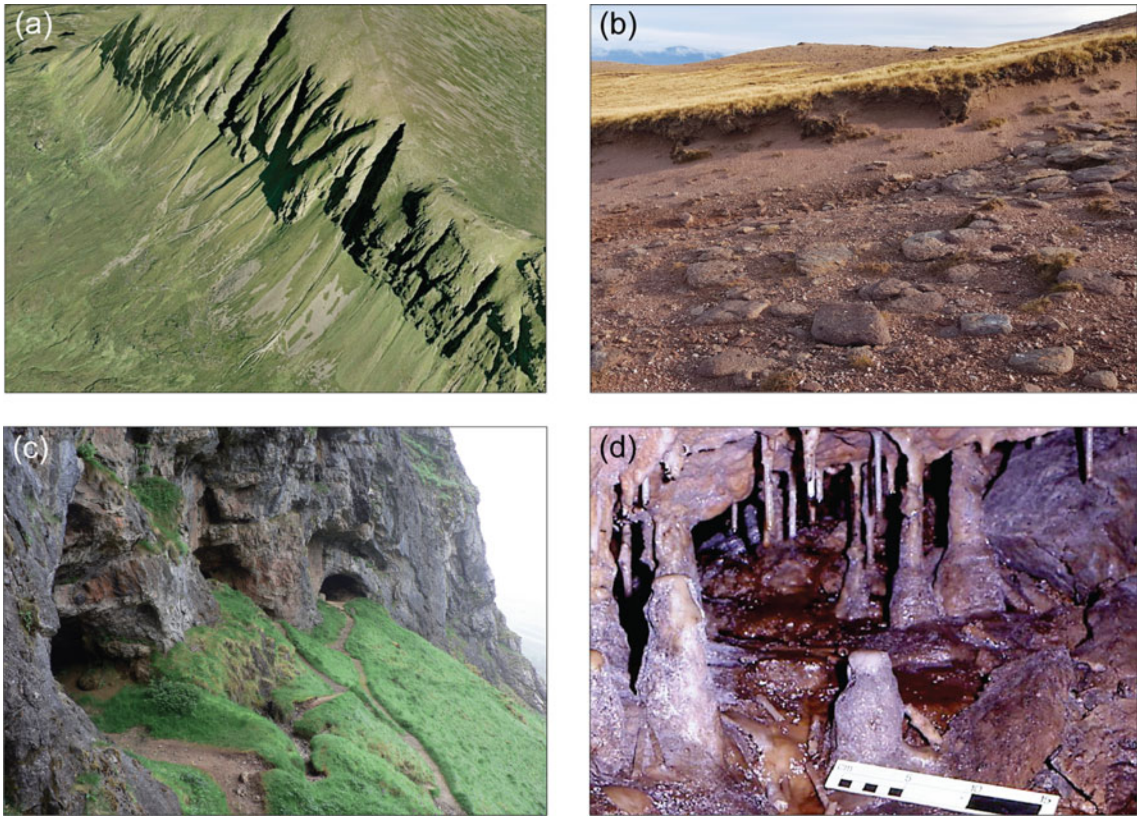

Fig. 12.8 **a** Relict talus accumulations of Torridonian sandstone rockfall debris, Quinag, Assynt. **b** Deflation surface and vegetation-covered niveo-aeolian sand sheets on Ben Mòr Coigach. **c** The Bone Caves of the Allt nan Uamh valley in Assynt, where faunal remains have been found in cave sediments. **d** Stalactites and stalagmites in the Traligill caves. (Images: **a** Google Earth™; **b** Jim Hansom; **c** Wojsyl, Creative Commons; **d** Tim Lawson)

form stone pavements; the sand sheets are bounded by eroding scarps, with isolated outliers that represent remnants of formerly more extensive sand cover. The sands are predominantly niveo-aeolian, having been deposited on snow-cover and then lowered during snowmelt onto the underlying vegetation cover. The deposits are poorly sorted and dominated by medium-grained sand with minor components of silt and fine gravel, characteristics attributable to gusty winds, short transport paths and admixture of different grades during snowpack ablation (Ballantyne and Whittington 1987). As with the similar deposits on An Teallach in Wester Ross (Chap. 13), those on lee slopes exhibit two units: a lower unit of weathered sand that progressively accumulated throughout most of the Holocene, and an upper unit of fresh sand. The onset of upper sand unit deposition has been dated on other Scottish mountains to AD 1550–1700, suggesting that it represents widespread stripping of aeolian deposits and aeolisols from exposed plateaux under the stormy conditions of the Little Ice Age, and redeposition on sheltered lee slopes (Morrocco et al. 2007).

12.7 Karst Landforms

Karst landforms are rare in Scotland, as carbonate strata are generally thin and interbedded with clastic sedimentary rocks. The largest area of karst development is the ~34 km^2 outcrop of Cambro-Ordovician dolostone (dolomitic limestone) in central Assynt, where stacking of successive layers of dolostone between two major thrust faults (the Sole and Ben More Thrusts) has created a thick, imbricate sequence of carbonate rocks. Surface karst features in this area are limited to peat-covered dolines, sinkholes, resurgences and areas of bedrock exhibiting small-scale solutional features (Atkinson et al. 1995), but the area also contains extensive cave networks that have formed the focus of detailed research (Lawson 1988; Fig. 12.8c); one of these, the Uamh an Claonite cave system, is the longest in Scotland, with over 3 km of passages. The most interesting karst features are contained in two valleys, the Traligill valley and the Allt nan Uamh valley, which drain westwards from high ground, mainly of Cambrian quartzite. In both valleys, surface drainage sinks underground in the upper catchment, flowing through a complex of cave passageways to emerge as resurgences farther downvalley, with dry stream beds marking the routes of surface flow when floods exceed the capacity of the underground drainage. Large abandoned phreatic tubes in the upper parts of the cave complexes are succeeded downwards by smaller vadose passages, implying sequential lowering of the former water table due to progressive valley deepening. The caves contain occasional speleothems (stalactites, stalagmites and flowstone floors; Fig. 12.8d) and fluvial sediments attributed to both slow-flowing subglacial meltwater sedimentation and high-velocity flow under vadose conditions (Lawson 1995a). From the uranium disequilibrium-series ages of the oldest (~200 ka) speleothems in the uppermost passages, Hebdon et al. (1997) have inferred subsequent deepening of the Allt nan Uamh valley by 47–68 m, implying an average erosion rate of ~2 mm a^{-1} over each glacial-interglacial cycle.

Dating of speleothems in the Traligill and Allt nan Uamh caves also provides evidence of periods of ice-free conditions during the Devensian (Lawson 2010). As outlined above (Sect. 12.3.1), a cluster of ages for the period ~95 to 75 ka (MIS 5c–5a) implies mainly glacier- and permafrost-free conditions during this interval, but two dates fall within the interval ~25 to 20 ka when Assynt was completely ice-covered, suggesting that these are erroneous or, less plausibly, that speleothem formation continued under a cover of wet-based glacier ice. The large phreatic 'bone caves' of Creag nan Uamh in the Allt nan Uamh valley (Fig. 12.8c) are famous for containing a wide range of faunal remains, including those of species now extinct from Scotland, such as brown bear, wolf and northern lynx, and others indicative of cold conditions, such as reindeer, arctic fox, arctic lemming and possibly polar bear (Lawson 1995b; Kitchener and Bonsall 1997). Some of these remains have yielded radiocarbon ages that fall within Marine Isotope Stage 3 (~57 to 31 ka), prior to the extension of the last ice sheet; most others fall within the Lateglacial or Holocene, but a few anomalous ages straddle the earlier part of the Last Glacial Maximum (Lawson 2010).

12.8 Conclusion

The remarkable geological and geomorphological diversity of the far northwest of Scotland has led to the region being designated a UNESCO Global Geopark. The region contains landscapes unlike any others in Scotland, including steep-sided glaciated inselbergs rising above an ice-scoured basement of knock-and-lochan topography, areas of spectacular bedrock megagrooves and the finest karst landscapes in the country. Much of this diversity results from its location astride the Moine Thrust Zone, which separates the Archaean Lewisian basement gneisses and overlying Neoproterozoic Torridonian sandstones and Cambro-Ordovician rocks from the Neoproterozoic metasedimentary rocks of the Moine Supergroup. The Thrust Zone itself is famously an area of great geological complexity, where movement of rocks along a series of low-angle thrust planes has resulted in stratigraphic inversion, overturning of strata and stacking of rock sequences of similar age.

During the last (Late Devensian) glaciation, ice flow across much of the region was northwestwards, feeding the Minch Ice Stream, but by ~15 ka the last ice sheet had shrunk to its mountain source areas in Assynt and west Sutherland. During the subsequent Loch Lomond Stade (LLS) an extensive icefield reoccupied the mountainous spine of Sutherland and Assynt, and a dozen or so cirque glaciers formed on outlying mountains. Successive periods of glacial erosion have produced cirques, glacial troughs, rock basins and steep-sided inselbergs, and the most spectacular area of knock-and-lochan terrain on the Scottish mainland. Glacial megagrooves and areas of wholly abraded bedrock represent the onset zones of tributaries of the Minch Ice Stream during successive periods of ice-sheet glaciation. Rare but important depositional landforms record the retreat of the last ice sheet, including moraines in Assynt and Coigach that formed at the time of the Wester Ross Readvance. Both cirque and icefield landsystems developed during the LLS, the former represented mainly by arcuate end-moraine belts and the latter being dominated by hummocky recessional moraines.

Sand and gravel terraces at the heads of sea lochs represent outwash and glacideltaic deposits graded to high relative sea level during ice-sheet deglaciation; relative sea levels then fell before reaching a mid-Holocene highstand, no more than ~2.6 m above present, before gradually falling. Rock coasts dominate the present littoral, with spectacular sea stacks up to 65 m high. Inland, large-scale rock-slope failures occur on all lithologies, and relict rockfall talus accumulations skirt steep mountain slopes. Active solifluction and patterned ground features occupy the high ground on some mountains, and deflation surfaces and windblown sand deposits occur on sandstone and quartzite plateaux. The karst terrain of Assynt is represented by peat-infilled dolines, sinkholes and resurgences, and by the most extensive cave networks in Scotland. These caves record sequential lowering of the water table associated with Middle to Late Pleistocene valley deepening and contain the remains of several animals now extinct in Scotland.

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13 Wester Ross

Colin K. Ballantyne and Tom Bradwell

Abstract

Wester Ross is an area of striking geodiversity, containing some of the most impressive glacial and postglacial landforms in Scotland. The region is bisected by the Moine Thrust Zone, west of which a platform of Archaean gneiss supports Torridonian sandstone mountains, and east of which the geology is dominated by metasedimentary rocks of the Moine Supergroup. Successive episodes of Pleistocene glaciation have resulted in the formation of glacial troughs that continue westward as fjords, and some of the most dramatic cirques in Scotland. The last ice sheet flowed across the area in a northwesterly direction, feeding a large ice stream in the trough between mainland Scotland and the Outer Hebrides. During its retreat, the ice front deposited a nested sequence of submarine recessional moraines in fjords and a chain of terrestrial moraines representing a late-stage ($\sim$15.3 ka) readvance of the retreating ice-sheet margin. During the Loch Lomond ($\approx$Younger Dryas) Stade of $\sim$12.9 to 11.7 ka, Wester Ross was partly reoccupied by an icefield centred near the present drainage divide and supported numerous independent cirque and valley glaciers that deposited prominent terminal, lateral and recessional moraines. After deglaciation, the landscape was modified by rock-slope failures, frost action, solifluction and aeolian activity, whilst glacio-isostatic uplift resulted in the formation of raised marine shorelines of both Lateglacial and Holocene age.

Keywords

Glacial troughs • Cirques • Fjord landsystems • Seabed moraines • Wester Ross Readvance • Blockfields • Solifluction lobes • Aeolian sand deposits • Landslides • Talus • Debris flows

13.1 Introduction

'Wester Ross' is the informal name given to that part of the Northern Highlands bounded by Loch Broom to the north, Loch Carron and Strathcarron to the south, and the main north-south drainage divide to the east (Fig. 13.1). It is bisected by the Moine Thrust Zone (MTZ) , which separates the Hebridean terrane from the Northern Highlands terrane (Fig. 13.2; Chap. 2). West of the MTZ, the geology is dominated by Lewisian gneisses of Archaean age that are unconformably overlain by Neoproterozoic Torridonian sandstones, which are in turn unconformably overlain by Cambro-Ordovician sedimentary rocks (Park et al. 2002). East of the MTZ, the terrain is dominated by metapsammites and metapelites of the Neoproterozoic Moine Supergroup (Strachan et al. 2002). The area contains some of the most impressive mountains in Scotland, several of which exceed 1000 m in elevation. The character of the mountain scenery, however, changes at the MTZ: many of the sandstone mountains to the west form massive, steep-sided, often isolated mountains (Fig. 13.3), whereas the schist mountains east of the MTZ are less dissected, often with gentler slopes and broad summit ridges. The record of glaciation includes not only large-scale erosional features (glacial troughs, rock basins, fjords and cirques) that developed during successive Pleistocene glacial stages, but also moraines deposited during a late readvance (the Wester Ross Readvance) that interrupted retreat of the last ice sheet. More limited glaciation during the Loch Lomond ($\approx$Younger Dryas) Stade of $\sim$12.9–11.7 ka resulted in the deposition of

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conspicuous moraines. During the Holocene, Wester Ross was the site of the largest rock avalanche in Britain. The mountains in the area host spectacular assemblages of Holocene hillslope, periglacial and aeolian landforms.

13.2 Geology, Structure and Cenozoic Landscape Evolution

West of the MTZ, Archaean gneisses of the basement Lewisian complex crop out extensively east and south of Gruinard Bay and in smaller pockets farther south (Fig. 13.2). The Lewisian rocks are typically composed of grey banded orthogneiss with layers and pods of basic and ultrabasic meta-igneous rocks. They form both high and low ground and represent a primeval landscape that has been partly exhumed from under the Torridonian cover rocks. In places, the Lewisian relief can be seen under the Torridonian rocks, as at the foot of Slioch, where the view from Loch Maree shows Lewisian hills and valleys underlying the Torridonian strata. The Lewisian landscape reaches its highest point on A'Maighdean (967 m).

In Wester Ross, the overlying strata are dominated by the rocks of the Torridon Group (hereafter referred to as 'Torridonian sandstones'), which began to accumulate around 1000–970 Ma, eventually achieving an aggregate thickness of 5–6 km. The rocks of the Torridon Group consist of gently-dipping arkosic red sandstones with occasional conglomerate beds and were originally deposited as alluvial fans and plains under semi-arid conditions. These rocks form both low and high ground, including several mountains over 900 m, such as Slioch, Beinn Alligin and Liathach (Fig. 13.1). Many of the highest sandstone mountains along the eastern margin of the Hebridean terrane are capped by Cambrian quartzite (quartz-arenite) that was deposited unconformably over the Torridonian rocks. The three eastern summits of An

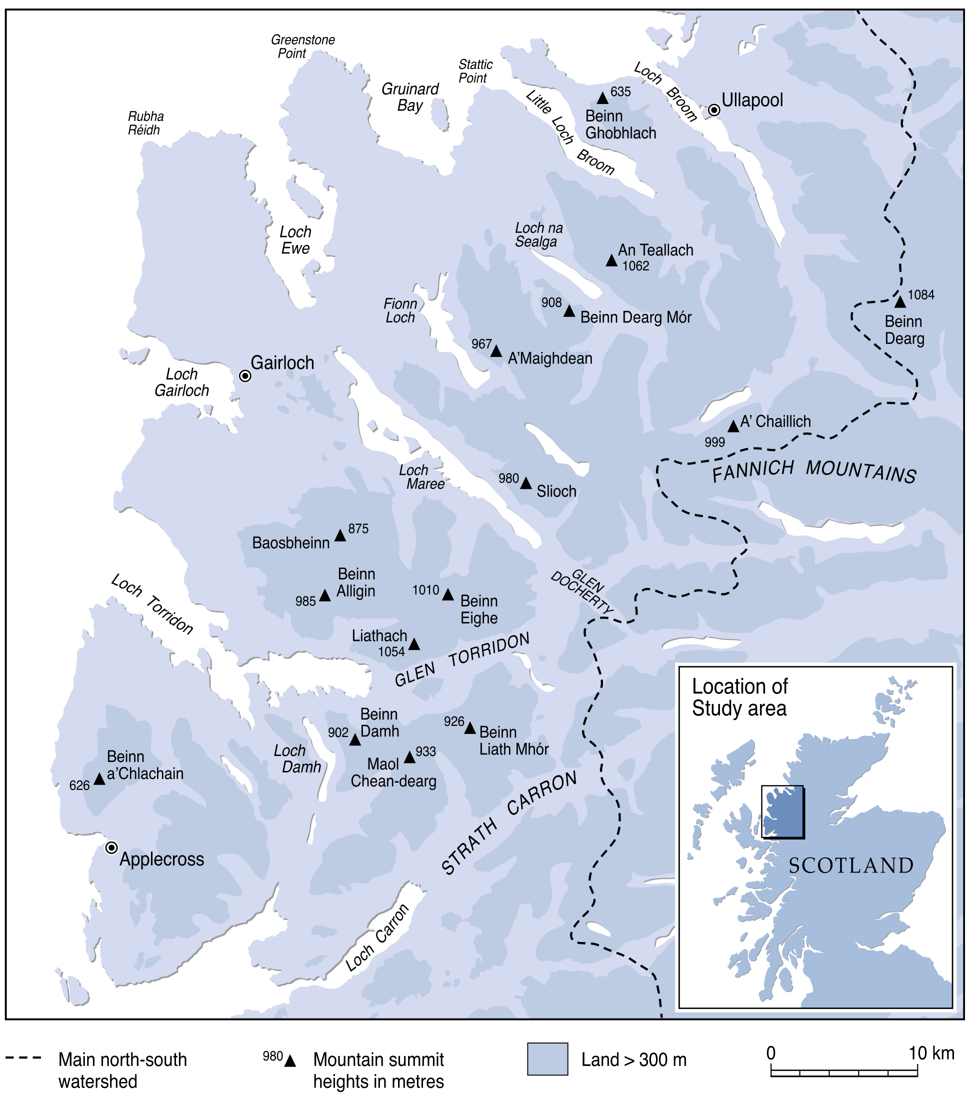

Fig. 13.1 Wester Ross, showing the location of sites mentioned in the text

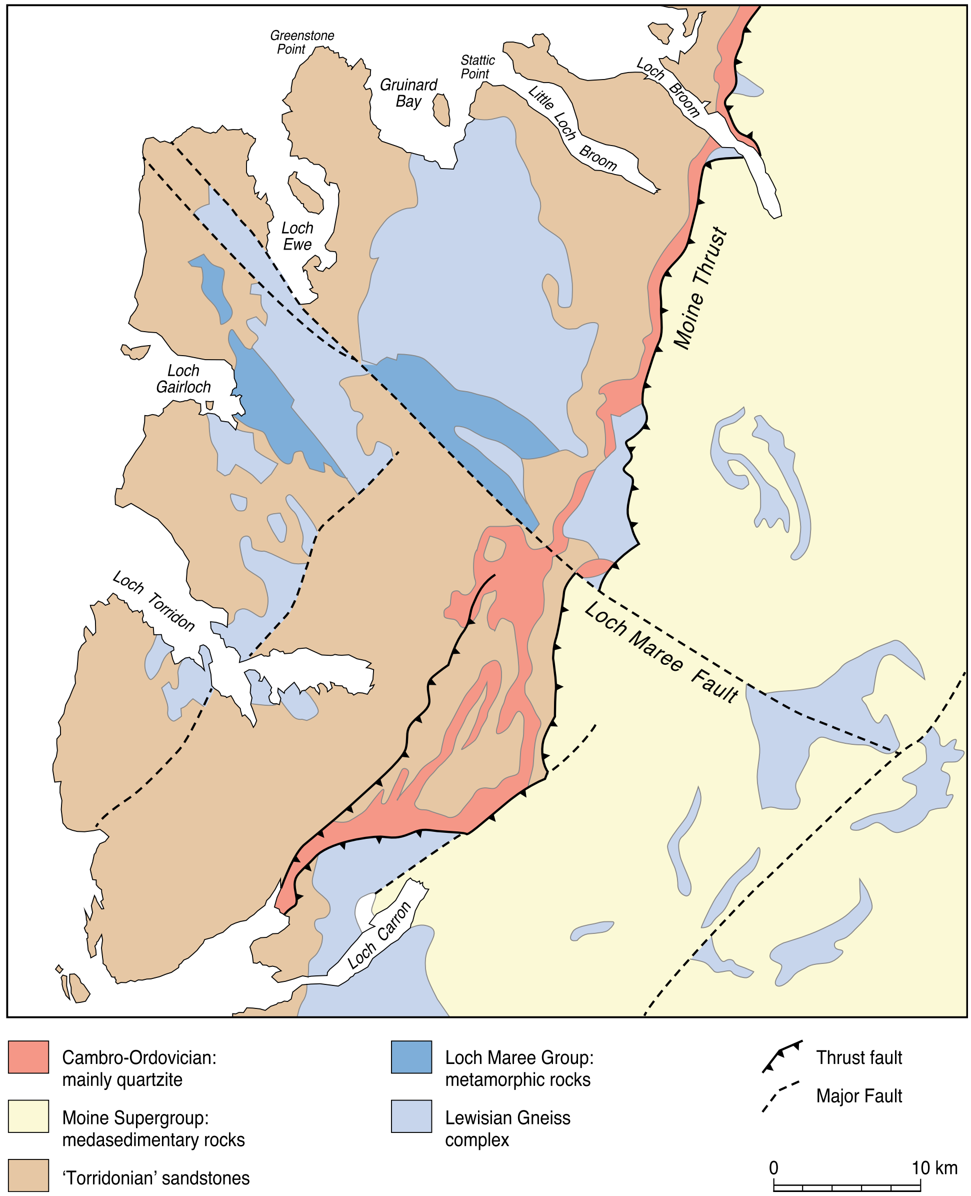

Fig. 13.2 Generalised geology of Wester Ross. The Moine Thrust Zone separates the Hebridean terrane to the west from the Northern Highlands terrane to the east

Teallach, for example, are crowned by quartzite outliers that represent the upslope continuation of a quartzite escarpment east of the mountain, and on Liathach the quartzite cap thickens eastwards, following the trend of the underlying unconformity (Fig. 13.3).

The Hebridean terrane is bounded to the east by the MTZ, along which rocks of the Moine Supergroup were thrust (by at least 70 km) over the rocks of the Hebridean terrane during the Scandian event (~435 to 425 Ma) of the Caledonian Orogeny (Chap. 2). The structure of the thrust zone is often complex, involving superimposition of successive low-angle thrusts that sometimes incorporate bodies of Lewisian, Torridonian and Cambro-Ordovician rocks, so that the orderly sequence of rocks is dislocated: rafts of Lewisian rocks have been thrust over Torridonian rocks, thrust Torridonian rocks locally overlie Cambrian quartzites, or there are multiple sequences of the same lithology.

East of the MTZ, the mountains forming the main north-south watershed are composed of metapelites and metapsammites (metasandstones) of the Moine Supergroup, interleaved with occasional 'Lewisianoid' inliers of gneiss (Fig. 13.2). The Moine rocks were originally deposited as a thick sequence of sandy, silty and muddy sediments on the margin of Laurentia within the period ~980 to 870 Ma and have been interpreted as the deformed and metamorphosed equivalent of the Torridon Group rocks (Krabbendam et al. 2008). These sediments were initially deformed during tectonic events centred around ~800 and ~740 Ma before being thrust westwards against and over the stable foreland of the Hebridean terrane during the Caledonian Orogeny. The

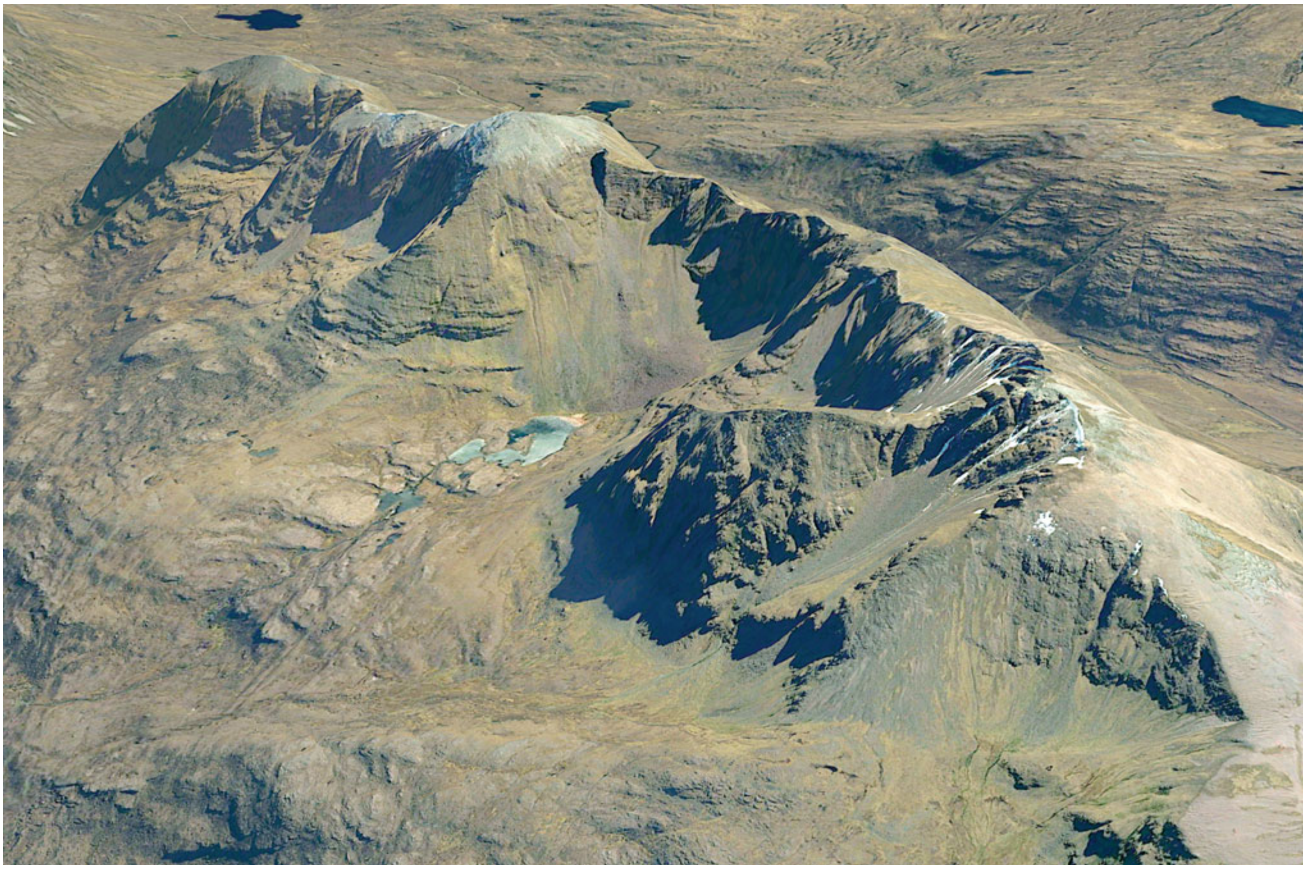

Fig. 13.3 Oblique satellite image of Liathach (1054 m), a Torridonian sandstone mountain, viewed from the northwest. The highest summit and eastern part of the ridge support a cap of Cambrian quartzite, and the northern flank of the mountain is indented by six cirques with intervening rock buttresses up to 500 m high. The runout debris of a postglacial rock avalanche can be seen on the floor of the large central cirque. (Image: Google Earth™)

Moine schist terrain is dominated by mountains that culminate in the Beinn Dearg massif and Fannich Mountains.

The present relief of Wester Ross reflects Cenozoic uplift and subsequent dissection of a land surface originally of relatively low relief. Major uplift occurred in the mid- to late-Palaeocene (~60 to 55 Ma), probably in response to magmatic activity along the western seaboard of Scotland, and resulted in vigorous erosion that removed almost all Mesozoic cover rocks (Hall and Bishop 2002). Renewed uplift occurred during the Miocene and Pliocene. The initial uplift is thought to have created a steep west-facing scarp, now strongly dissected, that extends southwestwards from Beinn Ghobhlach (635 m) in the north to Beinn a'Chlachain (626 m) on the Applecross peninsula (Hall 1991; Fig. 13.1). Eastward tilting of the Northern Highlands during the same episode is thought to have positioned the pre-glacial watershed a few tens of kilometres inland from the west coast, so that most major rivers in the Northern Highlands flow east to the Moray Firth.

Despite extensive dissection by rivers following Palaeogene uplift and even more vigorous erosion by Pleistocene glaciers, the higher mountains of Wester Ross display accordant summits within the altitudinal range ~900 to 1100 m, and some preserve fragments of rolling plateaux at lower altitudes. Areas of accordant summits and plateaux in Northern Scotland were interpreted by Godard (1965) as palaeosurfaces (erosion surfaces) formed during pulsed Cenozoic uplift, though the possibility that some areas were downwarped or tilted makes correlation of palaeosurface fragments uncertain. During much of the Cenozoic, Scotland experienced a warm to subtropical humid climate, and there is evidence that the pre-glacial landscape evolved through differential deep chemical weathering, with subsequent erosion of saprolite covers to reveal underlying etch surfaces (Hall 1991). Occasional outcrops of chemically weathered rocks occur on Torridonian sandstones and Moine schists. Saprolite remnants also occur in fissures within areas of glaciated Lewisian rocks, suggesting that the Lewisian knock-and-lochan terrain represents etch surfaces modified by varying degrees of glacial erosion (Krabbendam and Bradwell 2014).

13.3 Pleistocene Glacial Erosion

The landscape legacy of repeated episodes of Pleistocene glaciation partly reflects the geology and pre-glacial relief. On low-lying Lewisian gneiss terrain, successive ice sheets have excavated areas of weathered or shattered rock, forming knock-and-lochan landscapes of rounded, ice-moulded bosses or small hills separated by shallow linear valleys and lake-filled depressions. The overall relief is usually less than 100 m, but locally larger hills rise above a chaotic pattern of small lochs and boggy hollows. On the Lewisian gneiss at the mouth of Upper Loch Torridon and west of Loch Maree, for example, rugged hills rise up to 200 m above adjacent lochans (Fig. 13.4a). Ice-abraded knock-and-lochan terrain also characterises much of the low ground underlain by Torridonian sandstones, mainly in the form of low ridges and bedrock scarps etched out by selective glacial abrasion along bedding planes, though drift cover is more extensive than on the Lewisian rocks.

Glacial erosion of the Torridonian sandstone mountains has produced a transection complex of deep glacial troughs that isolate spectacular, steep-sided mountains such as An Teallach, Beinn Alligin, Liathach and Beinn Eighe (Figs. 13.3 and 13.4b). The dominant orientation of these troughs reflects westwards or northwestwards ice movement, locally along fault lines, as is the case with the Loch Maree trough. Some troughs extend westwards into glacially deepened fjords, such as Loch Broom, Little Loch Broom, Loch Torridon and Loch Carron (Fig. 13.1 and Sect. 13.4.2 below). West of the MTZ, subsidiary troughs aligned across the direction of regional ice-sheet movement link the major trunk valleys to create a network of deeply dissected terrain. East of the MTZ, troughs developed on Moine rocks form glacial breaches that cut through the main north-south watershed, as in upper Strathcarron and the broad breach that separates the Beinn Dearg massif from the Fannich Mountains (Fig. 13.1).

The higher (>750 m) mountains throughout the region are indented with cirques (corries) that preferentially occupy north- and east-facing slopes. Those on Torridonian sandstone mountains are amongst the most impressive in Scotland. The north face of Liathach, for example, is scalloped by six cirques with headwalls, sidewalls and intervening buttresses up to 500 m high (Fig. 13.3). Most cirques are skirted with thick talus accumulations, suggesting that cirque widening by rockfall has occurred during successive interglacial periods (Ballantyne 2013a, 2019a). As elsewhere in Scotland, there is a general eastward rise in cirque-floor altitudes, from ~300 m near the west coast to ~500 m around the main north-south watershed, a trend that has been attributed to an eastward rise in glacier equilibrium-line altitudes during periods of restricted glaciation (Barr et al. 2017).

13.4 The Last Ice-Sheet Glaciation

Although the large-scale glacial erosional landforms described above were formed over multiple glacial cycles throughout much of the Pleistocene, in Wester Ross there is no terrestrial record of glacial events prior to the advance and retreat of the last (Late Devensian) ice sheet. The presence of ice-rafted detritus of Scottish provenance in deep-water marine cores in the North Atlantic suggests that marine-terminating glaciers were at least intermittently present in the western Highlands after ~43 ka (Hibbert et al. 2010) and consistently present after ~38.5 ka (Peters et al. 2008). Radiocarbon ages obtained from organic material

Fig. 13.4 Glacial landforms of Wester Ross. **a** Ice-moulded hills of Lewisian gneiss near the mouth of Loch Torridon. **b** Glacial trough excavated in Torridonian sandstone north of Liathach; Beinn Dearg (914 m) is the mountain on the right. (Images: Colin Ballantyne)

buried by till in the Midland Valley, Outer Hebrides and North Sea Basin place the expansion of the nascent ice sheet across low ground and the adjacent continental shelf at ~35–32 ka (Chap. 4). Various lines of evidence suggest that the northwest sector of the ice sheet reached its maximum extent on or at the margin of the Atlantic shelf within the period ~30 to 27 ka (Bradwell et al. 2008a, 2019; Clark et al. 2012; Bradwell and Stoker 2015a; Ballantyne and Small 2019).

The orientation of striae, subglacial bedforms and ice-moulded bedrock suggests that the dominant direction of ice movement across Wester Ross was northwesterly, although basal ice flow was diverted around the western mountains such as An Teallach and Liathach. Analysis of ice-sheet flowsets suggests that northwesterly ice flow persisted throughout the growth, culmination and shrinkage of the last ice sheet. The regional ice divide, however, migrated eastwards from the present north-south watershed as the ice sheet thickened, before returning to occupy the high ground near the present watershed as the ice sheet thinned and retreated (Hughes et al. 2014). During at least part of the lifetime of the last ice sheet, ice from Wester Ross fed the Minch Ice Stream, which drained an area of 10,000–15,000 km^2 of the northwest mainland, Skye and eastern Lewis (Bradwell et al. 2007, 2019; Bradwell and Stoker 2015b).

13.4.1 Offshore Evidence of Ice-Sheet Retreat

The landscape of Wester Ross and the adjacent seabed has been shaped by successive glaciations since ~2.59 Ma (Stoker et al. 1993; Thierens et al. 2012). The result is strongly dissected mountainous terrain, glacial troughs, over-deepened offshore rock basins and a deeply indented fjord coastline. The first-order submarine topography of the seabed around Wester Ross, as elsewhere in NW Scotland, closely reflects the offshore bedrock geology, with bedrock highs composed of resistant, largely Precambrian rocks, and basins cut into weaker Mesozoic rocks or Quaternary sediments. The deepest waters, exceeding 200 m in places, are close to shore in the fjords (Fig. 13.5a). Superimposed on the submarine landscape is a wealth of well-preserved geomorphological evidence relating to the last ice sheet. These Late Pleistocene submarine landforms occur at a range of scales and include sedimentary (constructional) forms as well as erosional features.

Wester Ross and the adjacent seabed consitute one of the few places in Scotland where evidence of ice streaming has been clearly identified in the landscape record. The bed signature of fast-flowing ice, or its onset, is represented onshore in the megagrooved bedrock terrain north of Ullapool (Bradwell 2005; Stoker and Bradwell 2005; Chap. 12) and offshore in the streamlined seabed landscape around the Summer Isles (Bradwell et al. 2008b). Large-scale glacial lineations, broad till ridges and other streamlined bedforms on the Applecross peninsula and elsewhere probably formed at the convergence of ice-stream tributaries (Bradwell et al. 2007; Davies et al. 2019), both onshore and offshore. Such evidence has enabled detailed reconstruction of the geomorphological record of the former Minch Ice Stream and its tributaries (Bradwell et al. 2008b, 2019; Stoker et al. 2009; Bradwell 2013; Bradwell and Stoker 2015b).

On the seafloor around northern Wester Ross, a number of prominent seabed moraines record stillstands or readvances of the diminishing ice sheet at the marine to terrestrial transition (Bradwell and Stoker 2015b, 2016a, 2016b; Fig. 13.5). These ice-marginal features are particularly well-preserved and are an integral part of the interest of the Wester Ross Marine Protected Area both for their geomorphology and for the habitats and species they support; they probably formed soon after the demise of the Minch Ice Stream and prior to the Wester Ross Readvance of ~15.3 ka (Sect. 13.4.3). The most conspicuous moraine is a large constructional ridge starting at the mouth of Loch Ewe, in water depths of 30–70 m, and extending as a multi-ridge complex ~1 km wide for over 30 km to the northeast beyond the Rubha Coigach headland (Bradwell and Stoker 2016a; Bradwell et al. 2019). The moraine forms a prominent arcuate ridge in Loch Ewe, 30 m high and ~1000 m wide, before bending sharply to the north near Greenstone Point, beyond which it continues unbroken for over 25 km to the west of the Summer Isles. This large moraine complex represents a significant regional-scale interruption or readvance of the ice-sheet margin as it underwent overall recession following the demise of the Minch Ice Stream (Bradwell and Stoker 2015b). Superimposed on and inshore of these large offshore moraines are discontinuous but smaller-scale ridges with quasi-regular spacing, interpreted as De Geer moraines that formed at the margin of a large tidewater outlet glacier (Stoker et al. 2006; Bradwell and Stoker 2015a, 2016a; Fig. 13.5b). The delicate, well-preserved nature and superimposition of the small De Geer moraines on the larger moraines indicate that they are the most recent moraines to have been deposited on the seafloor around Wester Ross. Seabed photographs show their surfaces to be comprised of densely packed (mostly Torridonian sandstone) boulders (Bradwell and Stoker, 2016b), strongly resembling some of the terrestrial Wester Ross Readvance moraines (Fig. 13.7a).

13.4.2 The Little Loch Broom Fjord Landsystem

The seaboard of Wester Ross is a classic fjord landscape, carved and strongly modified by successive Pleistocene ice

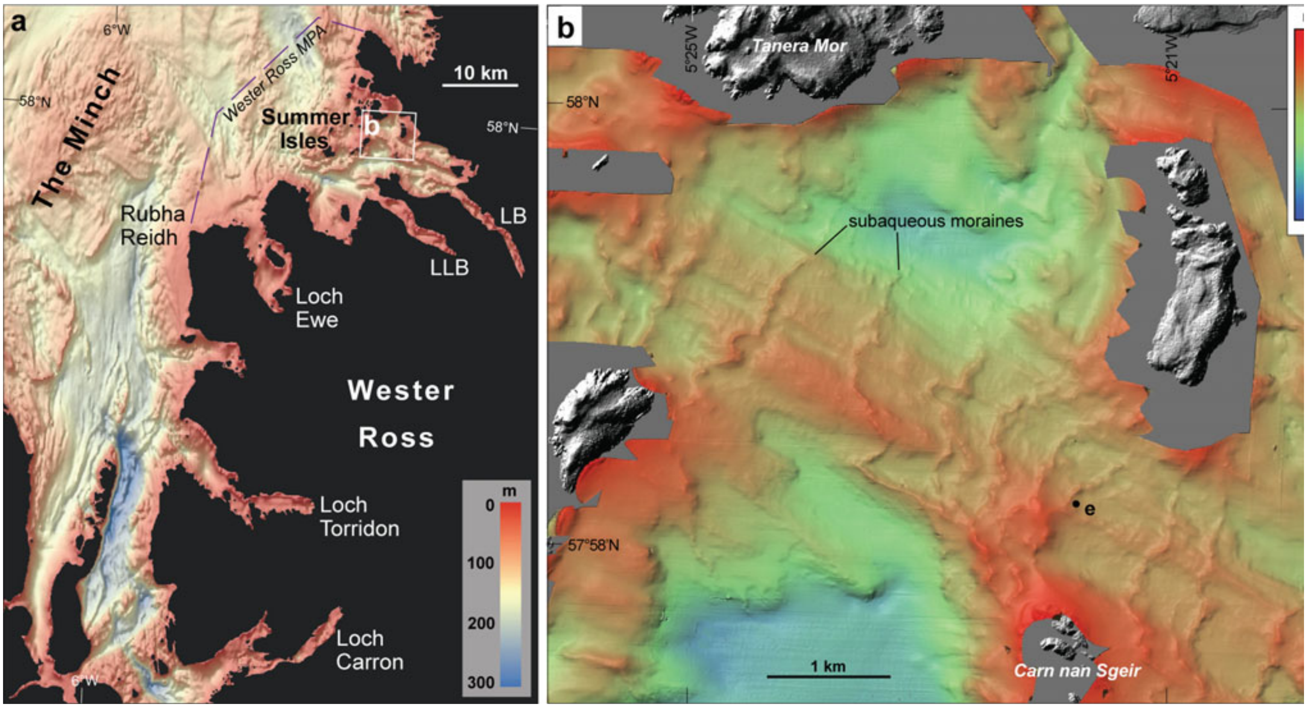

Fig. 13.5 Submarine glacial landforms on the seabed around northern Wester Ross. **a** General bathymetry offshore Wester Ross. Purple dashed line marks the outer extent of the Wester Ross Marine Protected Area (MPA). Colour ramp shows water depth. LB: Loch Broom; LLB: Little Loch Broom. **b** Multibeam bathymetry echosounder data showing a suite of recessional moraines on the seabed in the Summer Isles region, northwest of Loch Broom. Colour ramp shows water depth. (From Bradwell and Stoker (2015a) Earth Env Sci Trans R Soc Edinb 105:297–322 © 2015 Cambridge University Press)

sheets. This focused linear erosion is exemplified in the deep fjords (sea lochs) of Wester Ross, namely Loch Carron, Loch Torridon, Loch Broom and Little Loch Broom. In these locations, steep mountain flanks and over-deepened bathymetry result in enhanced vertical relief that locally exceeds 1 km in less than 2 km horizontal distance. Although of lower relief than the fjords in West Greenland or Norway, their formation and age are directly comparable.

Little Loch Broom, between the An Teallach massif to the south and Beinn Gobhlach to the north, is one of the most thoroughly investigated of all the Scottish fjords. The fjord is 12 km long and ~2 km wide at the present-day coastline. It is divided into two deep basins by a mid-loch bedrock high and has a shallow sill at its mouth. Hydro-acoustic (boomer) reflection profiles show that both basins are typically floored by 50–100 m of poorly consolidated Quaternary sediments, which locally exceed 150 m in depth. This sediment infill varies in character but, where undisturbed, typically forms horizontally bedded packages. Short cores from the basin floors have proved ~2–3 m of soft, dark grey-brown, clay-rich sediment with isolated dropstone pebbles, typical of glaciomarine sediments deposited during final (tidewater to terrestrial) deglaciation of the last ice sheet (Stoker et al. 2009, 2010).

As in other (de-)glaciated settings with steep slopes, sediment recycling by mass movement is common in the fjords of Wester Ross. Four main areas of submarine mass movement, manifest as geomorphologically and geologically disturbed seabed, have been identified in Little Loch Broom. These rotational slumps and planar-slide failure complexes range in size from 0.1 to 1.5 km^2 and include 'slide-mass' sediment packages 10–20 m thick. Although undated, these mass-movement events probably occurred during or soon after deglaciation in response to changing stress conditions caused by the removal of ice-mass support or longer-term, perhaps seismically accompanied, glacio-isostatic adjustments (Stoker et al. 2009, 2010; Bradwell and Stoker 2016a). The size and abundance of mass-movement features in Little Loch Broom, both above and below the marine limit, indicate that the fjords of Wester Ross may still be undergoing paraglacial readjustment to Late Pleistocene ice-sheet deglaciation.

13.4.3 The Wester Ross Readvance

A chain of ice-marginal moraine ridges that cross the peninsulas of Wester Ross from the Applecross peninsula in the south to Achiltibuie in the north marks the limits of an extensive readvance, the Wester Ross Readvance, that interrupted retreat of the last ice sheet in NW Scotland (Figs. 13.6 and 13.7a). Cosmogenic ^{10}Be exposure dating of Torridonian sandstone boulders on six moraines initially produced 22 consistent ages averaging ~13.5 ka (Bradwell et al. 2008c; Ballantyne et al. 2009), apparently implying that extensive ice cover persisted across low ground in Wester Ross during the Lateglacial Interstade of ~14.7 to 12.9 ka. Subsequent recalibration of these exposure ages using locally derived ^{10}Be production rates, however, showed that the reported ages were too young (Ballantyne and Stone 2012). The most recent recalibration suggests that the Wester Ross Readvance culminated at 15.3 ± 0.7 ka (Ballantyne and Small 2019) and thus several centuries before the rapid warming that marked the onset of the Lateglacial Interstade. Though it is possible that the readvance represents a response to internal reorganisation within the retreating ice sheet that reinvigorated ice flow across Wester Ross, it is notable that it occurred at approximately the same time as readvance of an independent ice cap on the adjacent Isle of Skye (Small et al. 2016; Chap. 10). This coincidence in timing suggests that both readvances may have been climatically driven by a pronounced cooling at ~15.5–15.3 ka that has been detected in North Atlantic deep-ocean cores (Scourse et al. 2009).

13.4.4 Periglacial Blockfields and High-Level Erratics

Several of the mountains in Wester Ross support outstanding examples of periglacial blockfields. These take the form of a mantle of debris, typically up to about a metre thick, that has formed through frost weathering of the underlying bedrock (Ballantyne 1998; Hopkinson and Ballantyne 2014). Those derived from quartzite and metasandstones are usually openwork blockfields that lack fine sediment at the surface but contain an infill of subsurface fines, whereas those derived from sandstones and pelitic schists are typically diamictic blockfields, with clasts of varying sizes embedded in a matrix of fine sediment. Erratic boulders are scattered over the blockfield debris on several mountains, in some cases hundreds of metres above the parent outcrops. Boulders of Torridonian sandstone, for example, occur amid quartzite blockfield debris on An Teallach and have been elevated at least 450 m from their source 3–6 km away. Similarly, erratics of Lewisian gneiss near the summit of Slioch imply uphill transport of at least 300 m over a distance of less than 4 km, and quartzite erratics on the same mountain indicate an even steeper uphill trajectory. Striking evidence of upwards transport of erratics occurs on the northern plateau of An Teallach, where a belt of quartzite erratics is scattered across the underlying Torridonian sandstone blockfield (Fig. 13.7b). The source of these boulders is an escarpment east of the mountain, implying that they were entrained by the last ice sheet at an altitude of 200–250 m, transported 4 km uphill and spread across the plateau at 700–710 m altitude.

The periglacial blockfields on the mountains of Wester Ross were once considered to have formed on nunataks that remained above the level of the last ice sheet (Ballantyne et al. 1997). Cosmogenic ^{10}Be exposure dating of erratics on several summits, however, has shown that these were emplaced at 16.1 ± 1.0 ka (Fabel et al. 2012), demonstrating that the erratics were deposited by the last ice sheet. This finding confirms that the last ice sheet buried all summits during the Last Glacial Maximum and indicates that most mountain summits in Wester Ross emerged from the thinning ice sheet several centuries before the Wester Ross Readvance. A further implication is that the blockfields pre-date burial by the last ice sheet but escaped glacial erosion, suggesting that the ice that overrode the highest ground in Wester Ross remained predominantly cold-based and frozen to the underlying permafrozen substrate, indicating that the last ice sheet in this area was polythermal (Fame et al. 2018).

13.5 The Loch Lomond Readvance

Following the Wester Ross Readvance, the margin of the shrinking ice sheet withdrew eastwards to mountain source areas. It is not known whether residual bodies of glacier ice survived the Lateglacial Interstade (~14.7 to 12.9 ka). Finlayson et al. (2011) have argued that a carapace of glacier ice may have persisted over the eastern parts of the Beinn Dearg massif. Conversely, the exposure ages obtained for high-level erratics on other plateaux in Wester Ross show that these remained ice-free throughout the Lateglacial Interstade (Fabel et al. 2012).

Subsequent expansion of glacier ice known as the Loch Lomond Readvance (LLR) commenced in response to cooling during the final centuries of the Lateglacial Interstade and culminated during the Loch Lomond Stade (~12.9 to 11.7 ka). At this time, Wester Ross was partly reoccupied by an icefield centred near the present drainage divide and supported numerous independent cirque and valley glaciers (Sissons 1977; Ballantyne 1986a; Bennett and Boulton 1993a; Finlayson et al. 2011; Fig. 13.6). The timing of the LLR maximum in Wester Ross is uncertain, though cosmogenic ^{10}Be exposure ages for samples from bedrock surfaces inside the readvance limits suggest that some smaller glaciers in Wester Ross had reached their maximum extent and were retreating prior to ~12.3 ka (McCormack et al. 2011; Ballantyne 2012).

The extent of many LLR glaciers is marked by end and lateral moraines, notably twin end-moraine ridges at the

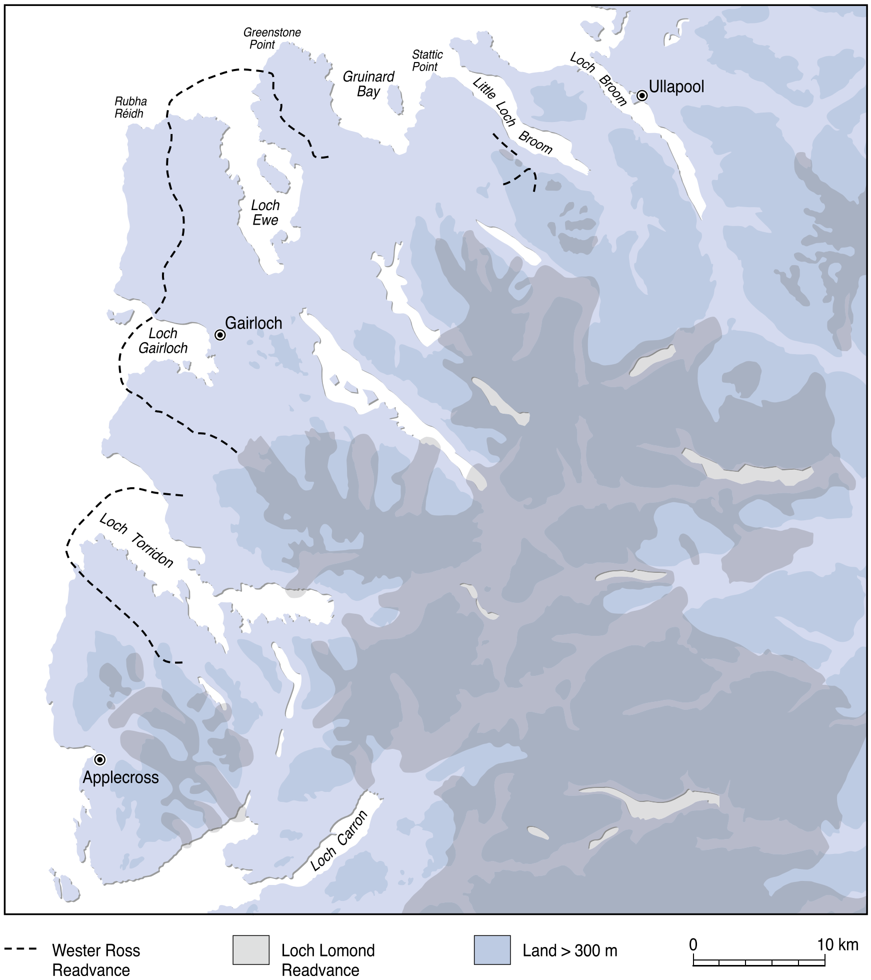

Fig. 13.6 Approximate extent of the Wester Ross Readvance (dashed line) and Loch Lomond Readvance in Wester Ross. Numerous summits protruded above the main icefield complex of the Loch Lomond Readvance as nunataks but are not shown as mapping of the interior of the complex is incomplete

Fig. 13.7 **a** The Gairloch Moraine, part of the Wester Ross Readvance limit. **b** Belt of quartzite erratics crossing a Torridonian sandstone blockfield at 700–710 m on the northern plateau of An Teallach. (Images: Colin Ballantyne)

Fig. 13.8 End moraine and paired lateral moraines that define the limits of a Loch Lomond Readvance cirque glacier on Beinn Dearg Mòr (908 m), south of An Teallach. (Image: Martin Kirkbride)

mouth of Glen Torridon, a massive drift ridge northwest of Loch Glascarnoch and a spectacular arcuate moraine north of Beinn Dearg Mòr (Fig. 13.8). Near Baosbheinn, bouldery lateral moraines deposited by a LLR outlet glacier even truncate those deposited some 3000 years earlier during the Wester Ross Readvance (Ballantyne 1986a). Inside these glacial limits, many valleys contain suites of hummocky recessional moraines that trend obliquely across slopes towards valley axes, forming nested chevron-shaped or arcuate patterns on valley floors (Fig. 13.9). These moraines were produced by ice push and subglacial deformation of sediment at oscillating glacier margins during overall glacier retreat (Bennett and Boulton 1993b; Benn and Lukas 2006); they imply gradual, pulsed retreat of the last glaciers under conditions of slowly warming climate and/or reduced snowfall. The most conspicuous hummocky moraines in Wester Ross are those of Coire a' Cheud-cnoic ('Valley of a Hundred Hills') south of Glen Torridon. At ground level, these appear to be a chaotic assemblage of mounds and conical hummocks (Fig. 13.10). Viewed from above, however, these can be seen to comprise longitudinal ridges superimposed on broad recessional moraines. This pattern has been interpreted by Wilson and Evans (2000) as representing deposition of fluted moraines during the LLR across nested recessional moraines that were deposited during the retreat of the last ice sheet.

13.6 Raised Marine Shorelines

Isolated fragments of Lateglacial and Holocene raised marine shorelines occur in Wester Ross, particularly along the margins of bays and fjords. Survey of these features by Sissons and Dawson (1981) has shown that the altitude of

Fig. 13.9 Hummocky recessional moraines marking successive positions of a Loch Lomond Readvance glacier margin as it underwent pulsed retreat north of Liathach. (Image: Colin Ballantyne)

the Lateglacial marine limit declines approximately south-eastwards across the region, from ~18–24 m OD (Ordnance Datum) on exposed western headlands to ~15.5 m OD at the head of Little Loch Broom and ~13.5 m OD at the head of Loch Ewe. They interpreted this trend as representing a gradual drop in relative sea level as the margin of the last ice sheet retreated inland. Holocene raised shorelines range in altitude from 4.8 to 7.8 m OD and storm-deposited Holocene shingle ridges occur at altitudes of 6.7–9.0 m. The wide altitudinal range of Holocene shoreline fragments suggests that they are also asynchronous (Smith et al. 2019). The most striking example of raised shorelines in Wester Ross occurs where the Little Gruinard River enters the head of Gruinard Bay. At this site, a raised beach at ~18.7 m OD merges with a gravel outwash fan that appears to have been deposited at the time of the Wester Ross Readvance and is itself truncated by a Holocene raised beach at ~5.2 m OD. On exposed headlands, the present coastline consists of low cliffs, rock ramps and boulder-strewn rock platforms, but sandy beaches and coastal dunes are present in sheltered inlets around Gruinard Bay, Loch Ewe and Loch Gairloch.

13.7 Postglacial Landforms

13.7.1 Landslides

Although large rock-slope failures are less common in Wester Ross than elsewhere in the Scottish Highlands (Chap. 14), the area hosts three remarkable examples. South of the Torridonian sandstone summit of Beinn Alligin (985 m), a deep notch defined by upslope-converging scarps marks the release zone of the largest rock avalanche in Scotland (Fig. 13.11a). The runout zone is covered by a tongue of large boulders that laps onto the opposing slope and descends downvalley over a distance of 1.25 km (Ballantyne and Stone 2004). This is the longest runout of any postglacial rock avalanche in Scotland and reflects the huge kinetic energy of the failure, which involved detachment of ~9 Mt of rock from a rockwall over 500 m high. Excess runout of this order has been explained by crushing of grains under large dynamic stresses, causing fluid-like behaviour that reduces frictional resistance within a flowing mass of

Fig. 13.10 Hummocky moraines in Coire a' Cheud-cnoic, Glen Torridon. Image: Colin Ballantyne

debris (McSaveney and Davies 2007). Cosmogenic ^{10}Be exposure dating of landslide boulders indicates that failure occurred at 4.4 ± 0.3 ka (Ballantyne et al. 2014). The cause of failure is uncertain, but it is possible that movement along the fault scarp that marks the eastern margin of the rupture zone may have triggered the landslide (Fenton 1991).

Another spectacular postglacial failure of Torridonian sandstone cliffs occurred from the northwestern extremity of Baosbheinn (875 m), where ~0.6 Mt of rock has collapsed between upslope-converging sidescarps. On impacting the break of slope at the foot of the cliffs, the landslide debris rebounded to form an arcuate boulder ridge, 485 m long and up to 31 m high, that is draped over a Wester Ross Readvance moraine. Cosmogenic ^{10}Be exposure dating of samples from boulders on the ridge indicates that the landslide occurred at 15.0 ± 0.7 ka. It is likely that failure was preconditioned by paraglacial stress release, and the timing of failure suggests that it may have been triggered by an earthquake caused by fault movements due to rapid glacio-isostatic uplift at this time (Ballantyne and Stone 2009).

On the eastern flank of Maol Chean-dearg (933 m), failure of a quartzite cliff has spread quartzite boulders across ~0.20 km^2 of the adjacent gently sloping Torridonian sandstone bedrock. The quartzite debris cover terminates abruptly downslope, locally forming a terminal ridge (Fig. 13.11b). The unusual nature of landslide runout suggests that the landslide may have occurred onto a shrinking glacier during the Loch Lomond Stade (Ballantyne and Stone 2013), an interpretation consistent with a cosmogenic ^{10}Be exposure age of 12.4 ± 0.6 ka obtained for samples from the runout boulders.

Several landslide sites in Wester Ross are represented by debris-free scarps where rockslide runout debris has been removed by glacier ice during the Loch Lomond Stade or earlier (Cave and Ballantyne 2016). One conspicuous example is a large landslide scar east of the summit of Beinn Damh (Fig. 13.11c), which may represent the source of a rock avalanche similar to that at Beinn Alligin (Ballantyne 2013a). Rockslides have also played a major role in the formation of arêtes on mountains such as Liathach and An Teallach: the rock slabs flanking such arêtes represent the failure planes of rockslides, but the associated runout debris has been removed by glaciers (Fig. 13.11d).

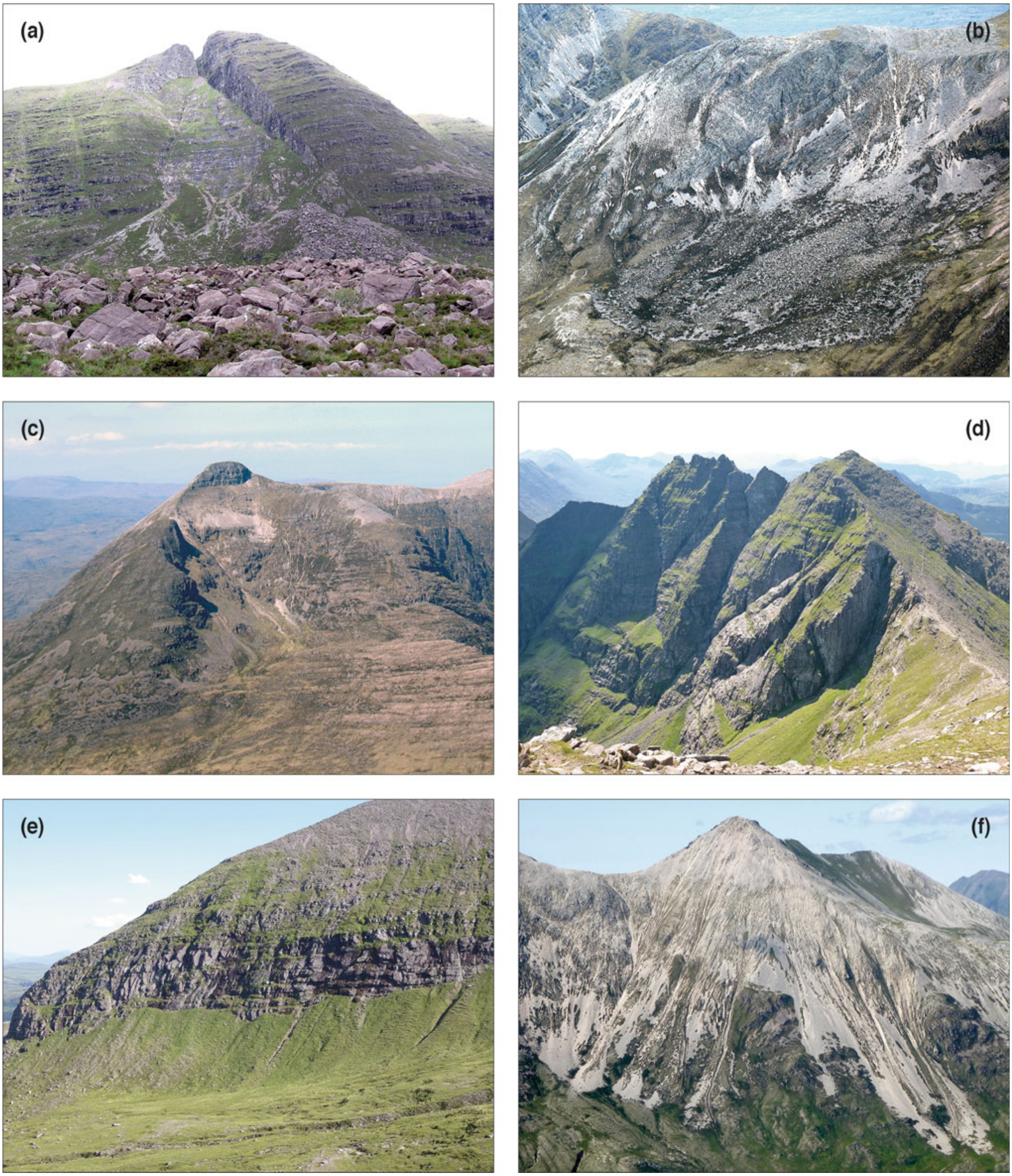

Fig. 13.11 Landslides, talus and debris flows in Wester Ross. **a** The Beinn Alligin rock avalanche. **b** The Maol Chean-dearg rock avalanche, where quartzite runout debris forms a broad lobe that buries the underlying sandstone bedrock. **c** A hollow on Beinn Damh (902 m), thought to represent the scar of a rock-slope failure where the runout debris has been removed by glacier ice. **d** The southeast arête of An Teallach: the bedrock slabs represent failure planes of rockslides, but the runout debris has been removed by glacier ice. **e** Relict, vegetated talus slopes on An Teallach. **f** Debris-flow tracks on Beinn Eighe, Torridon; flows originating in upper-slope gullies have formed tongues of quartzite debris crossing Torridonian sandstone. (Images: Colin Ballantyne)

13.7.2 Talus Slopes and Debris Flows

On mountains throughout Wester Ross, rockfall debris has accumulated at the foot of rockwalls to form talus accumulations, characterised by an upper straight slope at ~34°–38° and short basal concavity. Most talus slopes, however, are partly or entirely vegetation-covered and incised by gullies, suggesting that rockfall input was much greater in the past and is now very limited (Ballantyne and Eckford 1984; Ballantyne 2019b; Fig. 13.11e). Many are scarred by the tracks of debris flows, usually in the form of parallel levées of debris that originate in gullies upslope and terminate downslope in one or more bouldery lobes (Fig. 13.11f). Successive debris flows from the same parent gully have locally built small slope-foot debris cones, typically at gradients of 12–25° and decorated with the tracks of several generations of debris-flow deposits.

Exposures in talus typically reveal rockfall debris overlain by stacked debris-flow deposits intercalated with palaeosols, indicating that rockfall accumulation has been replaced by reworking of sediment by intermittent debris flows as the dominant mode of geomorphic activity; most talus accumulations are essentially relict paraglacial landforms that accumulated within a few millennia following deglaciation (Hinchliffe and Ballantyne 2009; Fig. 13.12). Radiocarbon ages obtained for palaeosols buried by debris-flow deposits within the talus deposits flanking Baosbheinn suggest that the transition from net rockfall accumulation to net erosion by debris flows occurred before ~6.7 ka and thus within a few millennia following deglaciation at the end of the Loch Lomond Stade. In Glen Docherty, gullies up to 10 m deep, flow tracks and slope-foot debris cones provide evidence for recurrent reworking of talus and drift deposits by debris flows. Radiocarbon dating of buried palaeosols exposed in gullies at this location indicates intermittent sediment reworking by successive debris flows since at least ~6.5 ka. As the associated pollen record contains no evidence for significant vegetation changes, burning or woodland clearance, Curry (2000) attributed the triggering of the debris flows to exceptional rainstorms occurring at random intervals throughout the Middle and Late Holocene. From the volume of the Glen Docherty debris cones, Strachan (2015) calculated that approximately 140,000 m^3 of sediment has been deposited by debris flows at the slope foot. The morphology of the source gullies suggests that these have extended and deepened through a combination of translational landslides, removal of sediment by debris flows and erosion of debris-flow tracks by mountain torrents.

Fresh debris-flow tracks mark the sites of recent activity. At least five debris-flow events occurred in Glen Docherty between 1957 and 2014, and two of these blocked the road through the glen. Strachan (2017) has also documented a recent small peaty debris flow in the glen. This involved rupture and downslope movement of a mixture of minerogenic sediment and peat, first as a translational slide, then a viscous debris flow and finally as low viscosity slurry, a pattern of failure that is fairly common (though rarely reported) in the Scottish Highlands.

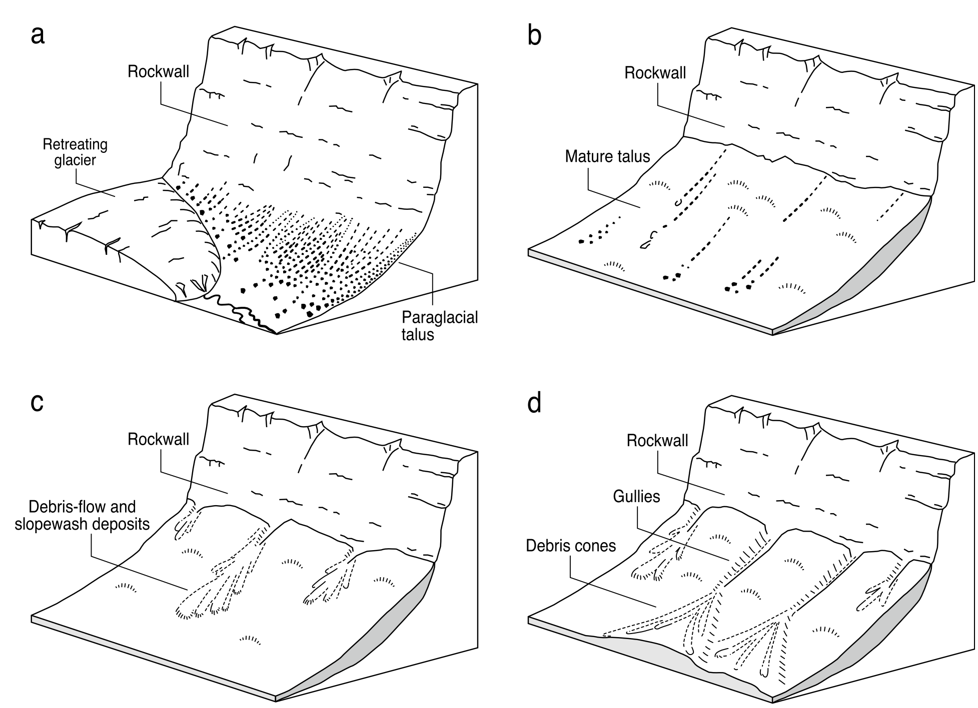

Fig. 13.12 Typical talus slope evolution in Wester Ross (schematic). **a** Rapid paraglacial talus accumulation after glacier retreat. **b** After a few millennia, rockfall has diminished; soil and vegetation cover the talus slope. **c** Gully erosion at the talus crest and deposition of debris-flow deposits downslope. **d** Downslope extension of gullies and accumulation of slope-foot debris cones. (Adapted from Hinchliffe and Ballantyne (2009) The Holocene 19:477–486 © SAGE publications)

13.7.3 Active Solifluction Landforms on the Fannich Mountains

The Fannich Mountains support some of the finest active solifluction landforms in Scotland. Metapelites of the Moine Supergroup have weathered to form a shallow, frost-susceptible regolith that has moved gradually downslope to form conspicuous solifluction terraces and lobes. Above 950 m on the southeast flank of Sgùrr Mòr (1110 m), solifluction terraces form a staircase of broad steps with widths of 6–13 m and lengths (along-slope) of 33–86 m on an average gradient of ~13° (Fig. 13.13a). The front of each terrace comprises a steep, vegetated riser 0.5–1.2 m high; some treads are almost bare of vegetation, whilst others support complete vegetation cover.

Similar terraces occupy gentle to moderate upper slopes on pelitic schists elsewhere in the Fannich Mountains, notably above 800 m on Sgùrr na Clach Geala (1093 m) and on the western Fannichs between A'Chailleach (999 m) and Sgùrr Breac (1000 m). As the gradient steepens downslope, the terraces become increasingly crenulate in planform, forming solifluction lobes (Fig. 13.13b). Recent movement has been measured using segmented columns inserted in three solifluction lobes then exposed by excavation after 35 years. The displaced segments (Fig. 13.13c) showed that recent lobe movement has been very gradual, with average surface velocities of 7.8–10.6 mm a^{-1} declining to zero at depths of ~0.4 m (Ballantyne 2013b). Trenching of lobes has shown that downslope movement has buried organic soil horizons, typically at 0.4–0.5 m depth. Five samples of organic soil recovered from the top of an illuvial (B) horizon of a podzolic soil buried by downslope advance of one lobe yielded radiocarbon ages that are statistically indistinguishable from the radiocarbon age of the same soil horizon immediately downslope (Ballantyne 1986b). This pattern suggests that slow, steady-state advance and thickening of solifluction lobes is sometimes interrupted by rupture of lobe fronts and rapid burial of soil downslope by liquified soil, followed by stabilization and renewed thickening of slowly advancing lobes (Kinnard and Lewkowicz 2006).

Intimately associated with active solifluction terraces and lobes on the Fannichs are numerous ploughing boulders that have moved downslope faster than the surrounding soil, leaving a vegetated furrow upslope and sometimes pushing up a ridge of soil downslope (Fig. 13.13d). Recent activity is indicated by the presence of deep niches that extend to the base of boulders at their upslope ends. Ploughing boulders on the Fannichs are confined to frost-susceptible soils above about 750 m on vegetation-covered slopes of 10–34° and 'ploughing' behaviour is limited to boulders with a mass greater than ~50 kg. Recent rates of downslope movement of seven boulders measured over 20 years averaged 3.6–30.3 mm a^{-1} and are strongly related to slope gradient. Excavation of one boulder in winter revealed thick sub-boulder lenses of segregated ice, leading Ballantyne (2001) to propose that 'ploughing' behaviour is primarily due to the much greater thermal diffusivity of boulders compared with that of the soil in which they are embedded. As the freezing plane penetrates boulders more rapidly than the soil, ice lenses form under boulders. During the ensuing thaw, the thaw plane penetrates to the base of boulders whilst the surrounding soil is still frozen. Water released by thaw of sub-boulder ice lenses cannot escape, generating high pore-water pressures in the soil under boulders. This causes them to lurch downslope over a bed of softened or liquified soil when the surrounding soil eventually thaws. Boulder movement allows trapped water to escape via the upslope niche, and stability is regained until the next period of deep freezing and thawing of the ground.

13.7.4 Aeolian Landforms on An Teallach

The mountains of Scotland constitute a topographic barrier to cyclonic systems tracking eastwards across the North Atlantic. As a result, gale-force winds exceeding 80 km h^{-1} are common on high ground, and gusts can exceed 150 km h^{-1}. Such powerful winds have produced a range of aeolian landforms on and around many high-level plateaux and cols; these are particularly well developed on the Torridonian sandstones of the northern plateau of An Teallach at 700–850 m altitude. Almost all of the plateau is bare ground: soil particles up to 4–6 mm in size have been winnowed away by strong winds to produce a deflation surface that is armoured by rounded boulders and carpeted with a gravel lag deposit. In a few places, isolated islands of vegetation-covered sand with eroded margins protrude above the deflation surface, suggesting that vegetation and sand cover were once much more extensive.

The history of wind erosion of the plateau is captured in aeolian sand deposits on lee slopes, where extensive vegetation-covered sand sheets up to 4.0 m deep decline in thickness downwind from the plateau margin (Fig. 13.13e). The sand sheets contain a lower unit of weathered sand up to 2.4 m thick and an upper unit of fresh unweathered sand up to 2.2 m thick. Both units comprise poorly sorted sand, reflecting their predominantly niveo-aeolian origin: most sand deposition occurs on the winter snowpack, and when the snow melts both fine and coarse grains are trapped by the underlying vegetation cover, which grows upwards through the accumulating sand. Radiocarbon dating of organic material in the lower unit shows that sand blown from the plateau started to accumulate on lee slopes in the Early Holocene, but that accumulation diminished as vegetation

Fig. 13.13 Periglacial and aeolian landforms in Wester Ross. **a** Solifluction terraces at 950–1050 m altitude on Sgùrr Mòr, Fannich Mountains. **b** Solifluction lobe on the Fannich Mountains. **c** Downslope displacement of segmented columns in a solifluction lobe on the Fannich Mountains over 35 years. **d** Ploughing boulder on the Fannich Mountains. **e** Aeolian sand deposits up to 4 m thick at the eastern margin of the An Teallach plateau. **f** Turf-banked terraces on a lee slope near the margin of the An Teallach plateau. (Images: Colin Ballantyne)

cover extended over plateau source areas (Ballantyne and Whittington 1987). The abrupt change to accumulation of the fresh upper-unit sand implies catastrophic wind erosion of sand deposits and sandy soils on the plateau upwind. Luminescence dating of samples from the base of the upper unit places the onset of this event at AD 1550–1700 (Ballantyne and Morrocco 2006), and similar stripping of vegetation and soil cover by wind erosion has been shown to have occurred at roughly the same time on other Scottish mountains (Morrocco et al. 2007). This timing coincides with the Little Ice Age of the sixteenth to nineteenth centuries, a period of long, cold winters and violent storms, suggesting that prolonged snow-lie may have caused vegetation degradation on plateaux, leading to the formation of blow-outs and ultimately removal of almost all vegetation, soil and sand cover from plateaux by strong winds.

Lee slopes on and at the margins of the An Teallach plateau are occupied by flights of turf-banked terraces, characterised by steep vegetated risers and bare stony treads up to 4.0 m wide and 25 m long (Fig. 13.13f). These landforms represent damming of downslope-creeping debris by bands of vegetation aligned across the slope. It is possible that the vegetated risers developed first as wind stripes (alternating bands of vegetated and unvegetated ground aligned normal to the dominant wind direction), that evolved into terraces as they blocked the downslope migration of unvegetated debris by frost creep (Ballantyne and Harris 1994), though the origin of these enigmatic landforms remains unresolved.

13.8 Conclusions

Wester Ross is an area of striking geomorphological diversity. This reflects not only lithological contrasts within and between the terranes on either side of the Moine Thrust Zone, but also a long history of glacial erosion that has produced a wide range of classic glacial landforms, including glacial troughs, fjords and cirques. High-level erratics scattered across intact plateau blockfields provide key evidence that the last ice sheet in this area was polythermal. The Lateglacial history of Wester Ross is well-documented both offshore and onshore: the area contains not only the finest terrestrial evidence in Scotland for a major readvance of the last ice sheet, but also an outstanding geomorphological record of renewed glaciation during the Loch Lomond Stade, captured by remarkable examples of terminal, lateral and recessional moraines. Postglacial landforms include Lateglacial and Holocene raised marine shorelines, the largest rock avalanche in Great Britain, classic debris-flow-modified paraglacial talus accumulations, outstanding examples of active solifluction landforms and, on An Teallach, a suite of high-level aeolian landforms unsurpassed anywhere in the British Isles. Few regions of Scotland (or elsewhere) contain such a magnificent and diverse range of upland landforms within a comparatively small area, an attribute that has contributed to the designation of Wester Ross as a National Scenic Area.

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14 Rock-Slope Failures in the North West Highlands

Colin K. Ballantyne

Abstract

Over 400 postglacial rock-slope failures >0.1 km^2 in extent have been identified in the NW Highlands. Most are rock-slope deformations (40%) or arrested rockslides (38%), with smaller numbers of fragmented slides, rock avalanches and rockfalls or topples; 78% are located on schists. All can be considered paraglacial features, preconditioned by rock damage due to ice loading and unloading, glacial erosion, thermomechanical effects and hydromechanical fatigue during successive glacial episodes. Progressive failure (brittle shearing of internal rock bridges and reduction of friction angles to residual) has probably played a major role in reducing rock-mass strength to critical levels, and there is evidence that seismic activity caused by reactivation of faults during glacio-isostatic uplift may have triggered rock release in some (possibly many) cases. Numerous debris-free failure scars represent rupture surfaces of rock-slope failures where displaced rock or runout debris has been removed by glaciers during the Loch Lomond Stade or earlier. Rock-slope failures operating throughout much of the Quaternary have played a significant role in shaping the present form of mountain summits and ridges, glacial troughs and many cirques.

Keywords

Rock avalanches • Arrested rockslides • Fragmented rockslides • Rockfall • Toppling failures • Rock-slope deformations • Debris-free failure scars • Paraglacial • Progressive failure • Earthquakes

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14.1 Introduction

The NW Highlands constitute mainland Scotland north of the Great Glen. West of the Moine Thrust Zone (Fig. 14.1), the NW Highlands comprise an undulating basement of Archaean Lewisian gneiss above which rise mountains of Neoproterozoic sandstones (here referred to as Torridonian sandstones) and Cambrian quartzite (Chap. 2). East of the Moine Thrust Zone, the terrain is dominated by the thrust and folded schists (metapsammites, metapelites and semipelites) of the Neoproterozoic Moine Supergroup, locally interleaved with inliers of 'Lewisianoid' gneisses or intruded by granitic rocks. In Caithness and around the Moray Firth, the Moine rocks are overlain by younger sedimentary rocks, mainly Devonian sandstones, and parts of Ardnamurchan and Morvern are composed of Palaeogene volcanic rocks. Most rock-slope failures in the NW Highlands occur in steep, mountainous terrain, though some (such as those bordering the Great Glen trough) occur on ground below 400 m.

The term *rock-slope failure* (RSF) describes the downslope movement of large rock masses, irrespective of the mode or amount of displacement or the rate at which it occurred. The term incorporates both rapid rock-slope failures (major rockfalls and toppling failures, rockslides and rock avalanches) and rock-slope deformations caused by the gradual downslope movement of mountain slopes, but excludes small-scale rockfalls and mass movements in unlithified sediments or soil. According to Jarman and Harrison (2019), over 660 and possibly as many as 920 postglacial RSFs are present in the Scottish Highlands. Moreover, the widespread occurrence of RSF cavities (hollows) or rupture surfaces that lack runout debris or displaced rock masses represent the sites of numerous RSFs where mobilised rock has been removed by glacier ice; these sites are referred to here as debris-free failure scars (DFFSs).

The classification of postglacial RSFs employed here follows a simplified version of the revised Varnes

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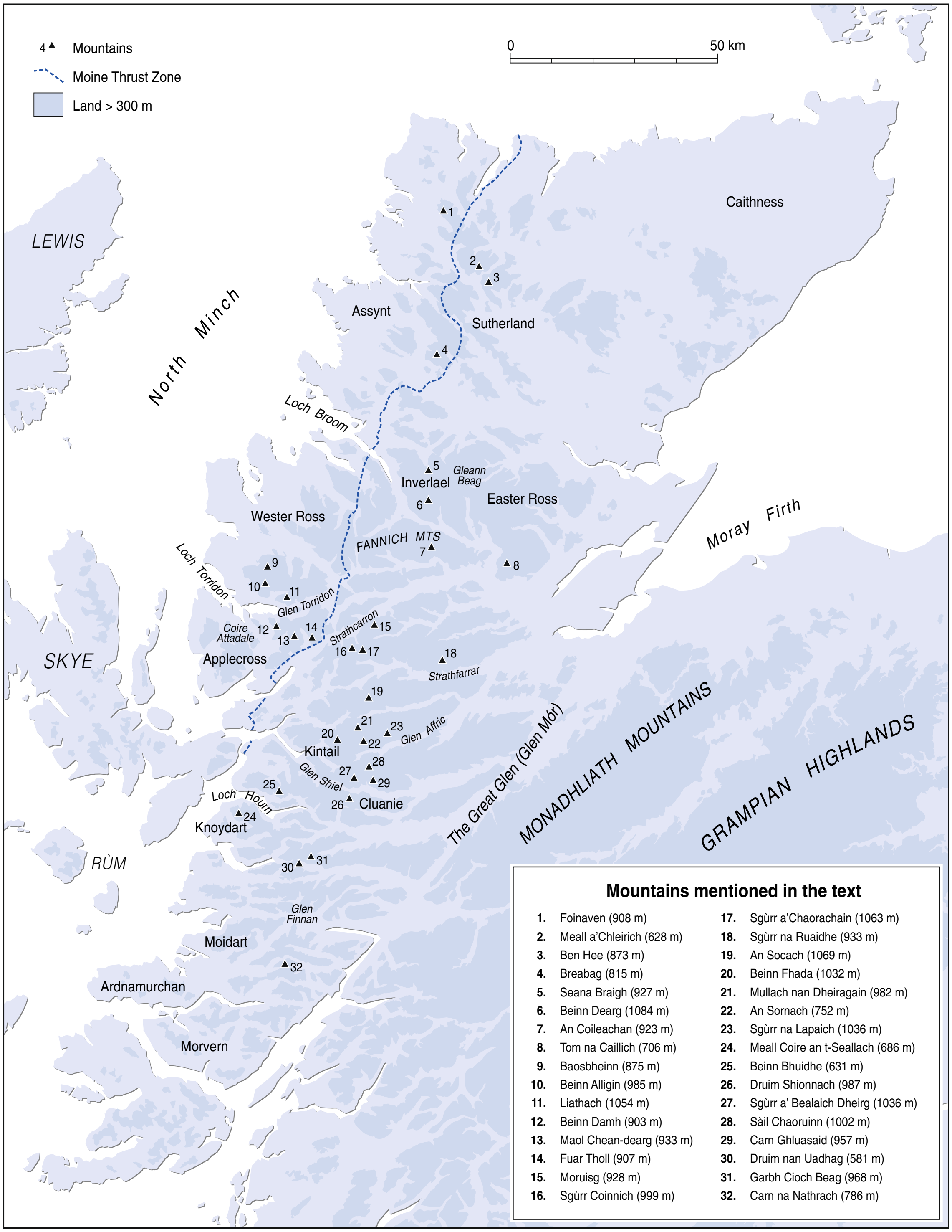

Fig. 14.1 The NW Highlands, showing the Moine Thrust Zone and sites mentioned in the text

classification (Hungr et al. 2014; Fig. 14.2). Category RF (rockfall) includes discrete major rockfalls and topples from cliffs but excludes intermittent small-scale rockfalls that have accumulated as talus. Category AS (arrested slides) incorporates sliding failures where the detached rock mass has largely retained integrity, irrespective of travel distance. Most arrested slides in the NW Highlands are planar, wedge or complex slides; rotational slides are rare. Category FS (fragmented slides and rock avalanches) comprises catastrophic landslides where the mobile rock has largely or completely disintegrated, forming a tongue of runout boulders (Hermanns and Longva 2012); the term 'rock avalanche' is restricted to fragmented rockslides that exhibit extended debris runout. Category RSD (rock-slope deformation) identifies areas of slow gravitational deformation of mountain slopes, typically manifest in arrays of scarps and/or antiscarps (upslope-facing scarps), rock benches, tension cracks, trenches and slope bulging; most RSDs lack a fully-defined rupture surface or clearly defined lateral margins.

The timeline for postglacial RSFs begins with local deglaciation. At ~17 ka, most of the region was still occupied by the shrinking remnants of the last ice sheet, but by ~16 ka nunataks had begun to emerge through the ice cover, and by ~15 ka glacier ice was largely confined to inland valleys and cirques (Ballantyne and Small 2019; Chap. 4). Under the cool temperate conditions of the Lateglacial Interstade (~14.7 to 12.9 ka), glaciers completely disappeared from low ground but may have persisted on some high plateaux. During the Loch Lomond Stade (~12.9 to 11.7 ka), a large icefield reoccupied mountainous terrain across much of the region, with smaller icefields in Assynt and Sutherland and on the Beinn Dearg massif, and numerous independent cirque and valley glaciers on outlying mountains (Bickerdike et al. 2018; Chap. 4). Postglacial RSFs within the limits of the Loch Lomond Readvance (LLR) glaciers must have occurred after these retreated (~12.5 to 11.5 ka), whereas RSFs outside or above these limits may have occurred at any time since ice-sheet retreat (~17.0 to 14.0 ka).

Early research on the RSFs of the NW Highlands was largely limited to pioneering theses by Watters (1972), Holmes (1984) and Fenton (1991), a few accounts of individual RSF sites (e.g. De Frietas and Watters 1973; Holmes and Jarvis 1985; Sellier and Lawson 1998), and a review by Ballantyne (1986). During the present century, the literature on the RSFs in Scotland has expanded markedly, focusing on the development and interpretation of RSF inventories (Jarman 2006, 2007a; Cave and Ballantyne 2016; Jarman and Harrison 2019), terrestrial cosmogenic nuclide (TCN) exposure dating of catastrophic RSFs (Ballantyne and Stone 2013; Ballantyne et al. 2014) and the significance of RSFs in the long-term evolution of Scotland's mountain landscapes over multiple glacial-interglacial cycles (Jarman 2009; Ballantyne 2013, 2019).

14.2 Characteristics and Distribution

An inventory of postglacial RSFs exceeding 0.01 km^2 in area in the NW Highlands has been compiled by the author using satellite imagery (Google Earth^{TM}) and extensive field verification (Fig. 14.3). Excluding coastal RSFs, a total of 403 RSFs were identified (Table 14.1), together with a further 105 possible RSF sites; the latter are not included in the analyses below.

As with all attempts to classify RSFs, the fourfold subdivision outlined above (RF, AS, FS, RSD) raises occasional problems of application, mainly in the form of hybrid landforms. Some RSDs exhibit zones of partial collapse, and over 25% of RSFs classified as arrested slides are partly fragmented. In some cases, differentiation of RSDs from arrested slides involving limited rock displacement is difficult, as both types may share common diagnostic features. At a few sites, delimitation of RSD extent proved problematic, as lateral margins are poorly defined, though the presence of a springline where groundwater percolating through fractured rock emerges at the surface aided identification of the margins of both RSDs and some arrested slides. A further problem arose in deciding whether some laterally-contiguous RSFs should be recorded as a single slope failure or two or more discrete RSFs.

Of the 403 recorded postglacial RSFs, 161 (40%) were identified as RSDs, 151 (37.5%) as arrested slides, 59 (14.6%) as fragmented slides or rock avalanches and 32 (7.9%) as rockfalls or toppling failures (Table 14.1). By comparison, an inventory of 358 RSFs compiled by Jarman and Harrison (2019) for the mountains of the NW Highlands (but apparently excluding RSFs on lower ground) identified 139 (38.8%) RSDs, 149 (41.6%) arrested slides and 70 (19.6%) 'rock avalanches', the latter being broadly equivalent to categories FS plus RF.

Figure 14.4 summarises size distributions of the different RSF categories. All distributions differ significantly ($p < 0.01$), except for categories AS and FS. Category RF failures are small (<0.1 km^2) and largely limited to cliffed scarps (Fig. 14.2f) and cirques. By contrast, 66% of RSDs exceed 0.1 km^2 in area, 32% exceed 0.25 km^2, and most occur on open valley-side slopes, typically on gradients of ~32 to 39° (Fig. 14.2e). Arrested and fragmented slides occupy a range of topographic settings and 33% of these failures exceed 0.1 km^2 in size.

Most RSFs (~78%) are seated on Moine metasedimentary rocks, mainly on dominantly psammitic schists (~54%) and on metapelites and semipelites (~24%). This predominance partly reflects the extensive outcrop of these lithologies east of the Moine Thrust Zone. However, a study by Cave and Ballantyne (2016) of the spatial density of RSFs and DFFSs on different lithologies normalised for slopes >25° in areas bordering the Moine Thrust Zone showed that the density of sites on Moine schists exceeds that on all other lithologies by a factor of two or more. This conclusion supports the long-established view that Scottish mountain slopes underlain by schists have been especially prone to failure. In the NW Highlands, this is particularly true of RSDs, 87% of which are seated on schists. The susceptibility

Fig. 14.2 Examples of different categories of rock-slope failures in the NW Highlands. **a** Arrested slide on Cambrian quartzite, Breabag, Assynt. **b** Large arrested slide, Sàil Chaoruinn, Cluanie. **c** Fragmented slide, Druim nan Uadhag, Dessarry; block toppling has occurred at the slope crest. **d** Rock avalanche in Garbh Choire Mór, An Coileachan, Fannich Mountains. **e** Part of the Sgùrr a'Bealaich Dheirg rock-slope deformation in Glen Shiel, showing slope bulging and well-developed antiscarps. **f** Gleann Beag, Easter Ross: on the left are two rockfalls; on the right, rockwall collapse has resulted in the formation of a hummocky runout lobe. (Images: **a**, **b**, **c**, **e** Colin Ballantyne; **d**, **f** Google Earth™)

of schistose rocks to failure and slope deformation was attributed by Watters (1972) and Holmes (1984) to their relatively low peak and residual friction angles and pronounced mechanical anisotropy, though failure on schistose rocks has occurred on anaclinal and orthoclinal as well as cataclinal slopes, and on slopes underlain by near-vertical foliation. Jarman and Harrison (2019) claimed that thin pelites interleaved within psammitic rocks 'lubricate' mass movement but provided no evidence; moreover, 'lubrication' is not a component of rock-slope mobilisation.

The overall distribution of RSFs in the NW Highlands is uneven (Fig. 14.3). About 75% lie within 20 km of the main

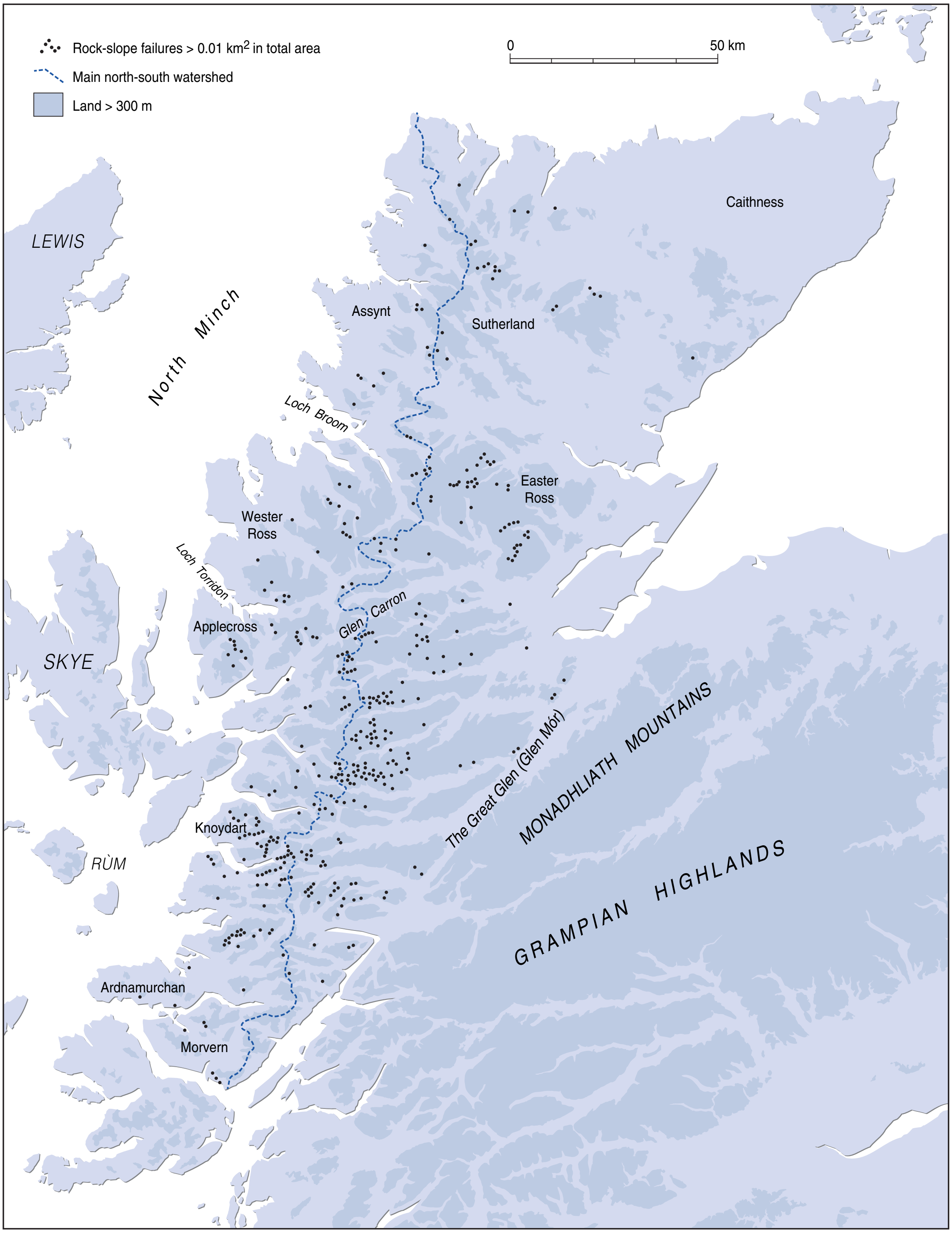

Fig. 14.3 Distribution of rock-slope failures > 0.01 km^2 in the NW Highlands

north-south drainage divide, with clusters in the Kintail-Cluanie-Glen Affric area, particularly east of the main drainage divide, and the Knoydart-Glen Finnan area, with several smaller outlying clusters, such as those in Easter Ross and Moidart (Figs. 14.1 and 14.3). Jarman and Harrison (2019) explained such clustering in terms of zones of 'concentrated erosion of bedrock' operating through 'catastrophic' rates of glacial erosion in 'late-developing' glacial breaches and trough heads during 'recent glacial cycles'. Their argument is circular: 'association' of RSFs

Table 14.1 Characteristics of rock-slope failures (RSFs) in the NW Highlands

			Lithology (%)[b]						Location (%)[c]			
Category[a]	n	%	MPs	MPe	TS	LG	Q	Other	1	2	3	Crest
RSD	161	40.0	64.6	22.4	7.4	0.6	1.9	3.1	73.3	22.4	4.3	69.6
AS	151	37.5	47.7	26.5	5.3	9.9	6.0	4.6	61.6	20.5	17.9	78.1
FS	59	14.6	45.8	25.4	20.3	5.1	1.7	1.7	49.2	22.0	28.8	89.8
RF	32	7.9	43.7	18.7	18.7	6.3	6.3	6.3	53.1	9.4	37.5	78.1
All RSFs	403	100	53.8	24.1	9.4	5.2	3.7	3.7	63.8	20.6	15.6	76.4

[a]RSD: rock-slope deformations. AS: arrested slides. FS: fragmented slides and rock avalanches. RF: rockfalls and toppling failures
[b]MPs: Moine metasediments, dominantly psammitic. MPe: Moine metasediments, dominantly pelitic or semipelitic. TS: Torridonian sandstones. LG: Lewisian gneiss. Q: Cambrian quartzite and Moine quartzite. Other: granite, diorite, Devonian sedimentary rocks and Palaeogene volcanic rocks
[c]1: open valley-side slopes. 2: spurs. 3: cirque headwalls, sidewalls and adjacent slopes. Crest: RSFs that originate at the slope crest

with selected breaches and trough heads is inferred to indicate 'intense erosion in recent [glacial] cycles' and hence 'late-development' of these locations, which in turn is assumed to explain the 'association' of RSFs with such 'late-developing' glacial breaches and trough heads. As they acknowledged, many breaches and trough heads contain few or no RSFs. They argued that slopes at such sites have become 'stress-hardened' by repeated episodes of glacial erosion, but this concept has no physical meaning; it has been demonstrated that cyclic loading, erosion and unloading of slopes during successive glaciations results in rock-slope weakening, not slope 'hardening' (e.g. Grämiger et al. 2017, 2018, 2020; Riva et al. 2018). An additional implication of their argument is that major watershed breaches with numerous 'associated' RSFs, such as Glen Shiel, developed only during 'recent cycles', and were unaffected by throughflow of glacier ice during earlier periods of ice-sheet glaciation, which is implausible. The author has plotted the distribution of RSFs relative to all possible glacial breaches in the NW Highlands (main watershed breaches, secondary breaches and col breaches). Only 71 RSFs (17.6%) occur within 3 km of the site of watershed breaching, and only 28 (6.9%) occur within 3 km of the main watershed breaches. Even a generous interpretation of what constitutes a glacial breach therefore fails to substantiate a causal connection between breaching and the distribution of RSFs.

A more meaningful control is slope height and gradient: RSFs in the NW Highlands tend to cluster in the highest, steepest and most dissected terrain, primarily on valley-side slopes (Table 14.1). Analysis of a sample of such slopes in areas underlain by Moine rocks where there are dense concentrations of RSDs and arrested slides indicates significantly steeper modal slope values in comparison with similar slopes lacking RSFs. This contrast suggests that gradient has played an important role in determining limit-equilibrium conditions for failure (cf. Cossart et al. 2014) and partly explains the variable density of RSF occurrence on Moine schists, and in particular the clustering of postglacial RSFs on the exceptionally steep slopes of the Kintail-Cluanie-Glen Affric and Knoydart-Glen Finnan areas.

14.3 Causes of Postglacial RSFs

As in other deglaciated mountain areas, rock-slope failure in the Scottish Highlands has been conditioned by multiple episodes of glaciation and deglaciation, and postglacial RSFs are essentially paraglacial landforms (Ballantyne 2002). Paraglacial slope failure in deglaciated terrain reflects preconditioning, preparatory and trigger factors (McColl 2012). Preconditioning factors include lithology, structure, stress history and rock-mass strength. Because recurrent cycles of glaciation impose mechanical stress on rock slopes as glaciers advance, erode and retreat, all slopes must have been weakened by incremental rock damage before the onset of the most recent glacial cycle, some to potentially critical levels. Preparatory factors associated with the most recent cycle of glaciation and deglaciation include glacial erosion, differential glacial loading and unloading of high-relief terrain (which induces rock-mass dilation ('rebound'), stress redistribution and fracture extension and criticality), seismic fatigue (rock damage caused by earthquakes), thaw of permafrost ice in joints, rock damage due to thermomechanical effects, particularly thermoelastic strain induced by exposure to seasonal temperature changes during deglaciation, and hydromechanical fatigue caused by long-term changes in groundwater pressures as glaciers advance and retreat (Leith et al. 2014; Gischig et al. 2015; Grämiger et al. 2017, 2018, 2020). Potential triggers include debuttressing (reduction of lateral support provided by glacier ice to a critically-stressed slope), excess joint-water pressures, thaw of permafrost, earthquakes, and progressive failure in the form of rock-mass degradation over long timescales caused by

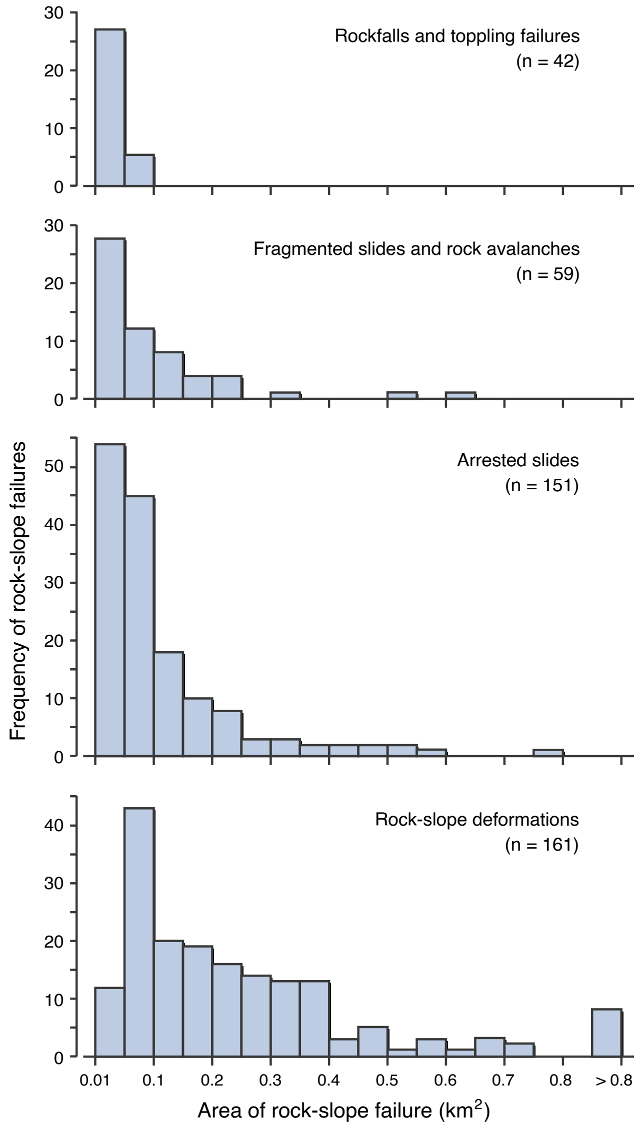

Fig. 14.4 Size distributions of rockfall/toppling failures (category RF), fragmented slides and rock avalanches (FS), arrested slides (AS) and rock-slope deformations (RSD). The measured areas extend from the failure crest to the toe of RSFs, including any debris runout zone

internal deformation and brittle failure of internal rock bridges (Eberhardt et al. 2004; Brideau et al. 2009).

Figure 14.5 illustrates possible trajectories of rock-mass weakening leading to slope failure during and after the last glacial cycle. During glacial advance, resisting forces increase through ice loading and joint closure, but during glacial retreat (ice thinning) accumulated rock damage due to glacial loading and unloading, erosion of rock and a combination of thermomechanical effects and hydromechanical fatigue lowers resisting forces to below previous levels. Depending on the initial disparity between resisting and driving forces, increased rock damage may lead to failure during or soon after deglaciation. Most slopes, though weakened, remain intact, though time-dependent reduction in resisting forces (progressive failure) may lead to rock-slope deformation that eventually stabilises or accelerates to catastrophic collapse. Rapid failure may also be triggered by debuttressing, earthquakes (which cause a transient increase in driving forces) and excess joint-water pressures (which cause a transient decrease in resisting forces). Pathways to failure are therefore both spatially and temporally non-uniform.

The limit-equilibrium analyses of Highland RSFs carried out by Holmes (1984, p 113–116) showed that all of the sliding failures he examined could not have failed under self-weight stresses alone, implying that glacial steepening of rock slopes has been ‘relatively unimportant as a direct cause of slope failure’. At most of the sites he investigated, failure surfaces lie well below peak friction angles but close to residual friction angles, suggesting that progressive failure could have triggered rapid sliding failure. Others lie below the residual friction angle, implying that some other factor triggered failure.

Some insight into the causes of rock-slope failure has been provided by TCN exposure dating of 31 fragmented slides and rock avalanches on mountain slopes in the Scottish Highlands and NW Ireland (Ballantyne et al. 2014). Most of these lagged deglaciation by centuries or millennia, implying that only a small number of the sampled RSFs could have been triggered by debuttressing or elevated joint-water pressures during deglaciation. Similarly, the timing of dated RSFs exhibits no relationship to periods of rapid warming, implying that thaw of permafrost was not a widespread cause of failure. Ballantyne et al. (2014) concluded that the time lag between deglaciation and most catastrophic RSFs indicates that many were the outcome of progressive failure through time-dependent strength degradation within fractured rock, leading to either spontaneous failure or failure triggered by an extrinsic agent. They also noted a temporal correspondence between the timing of RSF activity and rates of glacio-isostatic crustal rebound (Fig. 14.6), suggesting a causal connection via fault (re)activation and triggering of failure of critically-stressed slopes by earthquakes.

The NW Highlands form part of a tectonically-stable intraplate region, and recent seismic events with $M_L > 4$ are rare. However, numerical modelling has demonstrated that faults become unstable during the crustal unloading that accompanies ice-sheet deglaciation (e.g. Hampel et al. 2010), resulting in a pulse of medium- to high-magnitude earthquakes, though shallow fault activation and associated earthquake activity may lag deglaciation by centuries or millennia because of long crustal response times. There is evidence that stable intraplate areas, including Scotland, experienced uplift-driven faulting and seismic activity

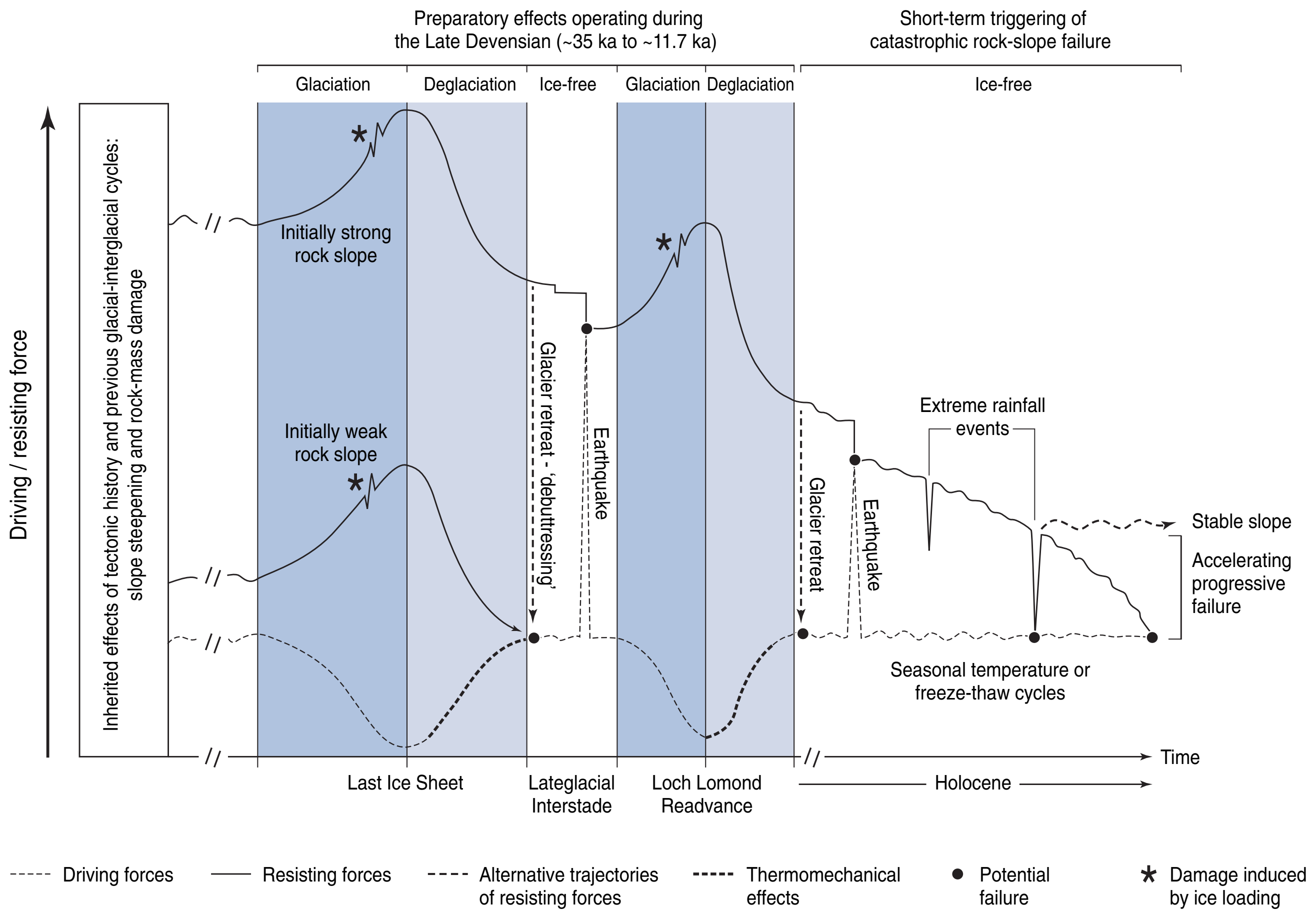

Fig. 14.5 Trajectory of paraglacial preparatory and trigger factors leading to failure (schematic), showing the changes in driving and resisting forces in hypothetical rock slopes after ~35 ka. Rock damage induced by ice loading/unloading and thermomechanical/hydromechanical effects during deglaciation reduces resisting forces during each glacial-deglacial cycle. Failure may be triggered by 'debuttressing' during glacier retreat, a transient increase in driving forces during earthquakes, extreme rainstorm events that cause excess joint-water pressures, or progressive failure due to time-dependent brittle failure of internal rock bridges. Earthquakes also act as preparatory factors causing subcritical rock damage and reduction in resisting forces (seismic fatigue). Rock-slope deformations represent a subcritical response to reduction in resisting forces, ultimately evolving into catastrophic failures or stable slopes. (Adapted from Grämiger et al. 2018)

following ice-sheet deglaciation (e.g. Firth and Stewart 2000; Stewart et al. 2001). Because of topographic amplification of seismic waves (Meunier et al. 2008), coseismic RSFs tend to originate at or near slope crests (Densmore and Hovius 2000), as is the case with 76% of postglacial RSFs in the NW Highlands (Table 14.1). The effect of topographically-amplified ground shaking in Scotland may also be evident in collapses of low anaclinal rock scarps along mountain ridges (Ballantyne 2019). It is also notable that ~75% of RSFs occur within 20 km of the main drainage divide (Fig. 14.3), close to the uplift axis, where rates of glacio-isostatic rebound were greatest. Collectively, both the timing of dated RSFs (Fig. 14.6) and the above evidence suggest that many catastrophic RSFs (and probably arrested slides) in the NW Highlands were triggered by uplift-induced earthquakes, or occurred on slopes weakened by seismic fatigue, leading to later failure (Gischig et al. 2015). Dynamic loading during earthquakes has also been implicated in the initiation and acceleration of RSDs (Moro et al. 2007, 2012; Agliardi et al. 2012; Pedrazzini et al. 2016), so it is possible that some of those in the NW Highlands were activated or influenced by seismic activity.

14.4 Catastrophic RSFs

Only seven or eight RSFs are classified here as rock avalanches, four of which have been dated using cosmogenic ^{10}Be. All exposure ages cited below are from Ballantyne et al. (2014).

The largest rock avalanche in Scotland resulted from failure of 3.3–3.8 × 10^6 m^3 (~9 Mt) of rock from Beinn

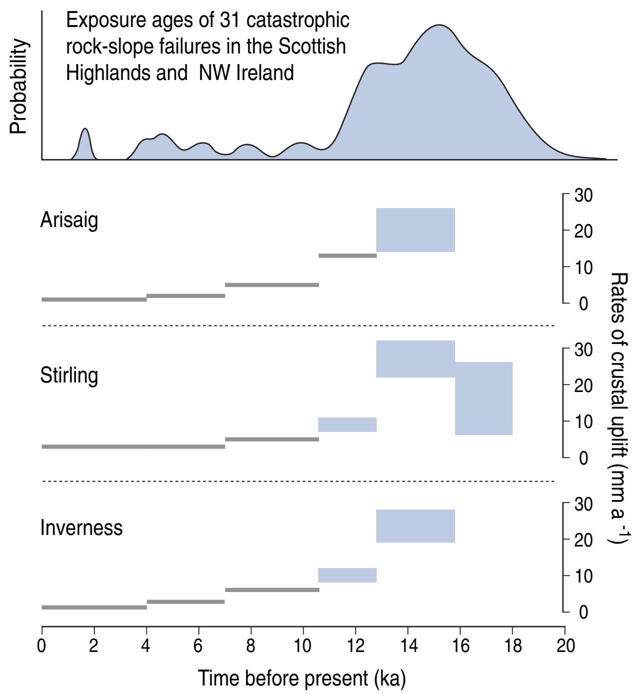

Fig. 14.6 Summed normal kernel density estimates generated by the ages and associated uncertainties of 31 fragmented slides and rock avalanches in the Scottish Highlands and NW Ireland, plotted against average rates of glacio-isostatic crustal uplift for three locations in Scotland. (From Ballantyne et al. (2014) Quat Sci Rev 86:144–157 © 2014 Elsevier Ltd; reproduced under Creative Commons CC-BY Licence)

Alligin (985 m), a Torridonian sandstone mountain in Wester Ross. The source of the rock avalanche is a steep (42°) rupture surface flanked by upslope-converging scarps that meet near the mountain summit; the eastern scarp is a fault scarp, and the almost vertical western scarp is the RSF headwall (Fenton 1991). The adjacent cirque floor is occupied by a tongue of runout debris comprising sandstone boulders up to and sometimes exceeding 5 m in length (Fig. 14.7a) that exhibits crude inverse sorting. This deposit is at least 15 m thick along the runout axis, occupies an area of 0.38 km^2 and extends continuously along the cirque floor for 1.25 km, descending at an average gradient of 8° from near the base of the cirque headwall. The margins of the runout tongue are sharp. The regularity of the rupture surface suggests that failure took the form of a planar slide that completely disintegrated during movement.

The Beinn Alligin rock avalanche occurred at 4.4 ± 0.3 ka, over 7000 years after deglaciation of the site. This long delay suggests that failure was due to progressive failure (reduction of cohesive strength through shearing of rock bridges) initiated by deglacial unloading, though seismic activity caused by reactivation of the fault scarp that borders the rupture zone may have triggered release of the intervening rock mass (Fenton 1991). This rock avalanche represents the most striking example in Scotland of the phenomenon of excess runout that greatly exceeds that which would be expected of a frictional flow of dry, angular debris. Calculations by Ballantyne and Stone (2004) suggest that in this case over 680 m of excess runout occurred. The causes of this phenomenon are still debated, but appear to be related to the high potential energy of the released rock mass (Dade and Huppert 1998), and may be due to the development of a mobile basal layer (Kilburn and Sørensen 1998) or crushing of grains under large dynamic stresses, causing fluid-like behaviour that reduces friction within a flowing mass of debris (McSaveney and Davies 2007).

Two further rock avalanches involved failures on Torridonian sandstones. Partial collapse of a cirque headwall on Liathach (1054 m) in Wester Ross released >0.5 Mt of rock debris onto the cirque floor, and failure of ~0.6 Mt of rock from the NW shoulder of Baosbheinn (875 m) in the same area resulted in the accumulation of a massive arcuate boulder ridge that is draped over a lateral moraine (Ballantyne and Stone 2009). The moraine was deposited by a readvance of the last ice sheet at ~15.3 ka (Chap. 13) and the rock avalanche occurred at 15.0 ± 0.7 ka, suggesting that it may have been triggered by seismic activity during the period of most rapid glacio-isostatic uplift (Fig. 14.6). A further rock avalanche in Wester Ross involved collapse of a quartzite rockwall on the eastern flank of Maol Chean-dearg (933 m). At this site a broad runout tongue of quartzite boulders covers ~0.20 km^2 of the underlying gently-sloping Torridonian sandstone bedrock (Fig. 14.7b) and locally terminates at a low boulder ridge, suggesting that it may represent failure onto the surface of a shrinking LLR glacier (Ballantyne and Stone 2013). The TCN exposure age of 12.4 ± 0.6 ka obtained for this RSF is consistent with this interpretation.

Rock avalanches *sensu stricto* are rare on Moine rocks, though collapse of a cirque headwall in the eastern Fannich Mountains has resulted in deposition of a 370 m long tongue of boulders that extends across the cirque floor (Fig. 14.2d). Holmes (1984) showed that rock-mass release at this site took the form of planar sliding along a major joint set inclined at 35°, 6° below the friction angle of the rock, implying that excess joint-water pressures or possibly seismic acceleration were responsible for triggering failure. A similar rockwall failure from the SW spur of Carn Ghluasaid (957 m) in the Cluanie mountains spread a 450 m long lobe of runout boulders across the subjacent slope (Fig. 14.7c) at 12.7 ± 0.7 ka.

Over 50 fragmented slides have been recorded in the NW Highlands. Most (~70%) occur on Moine schists but few have been documented in detail. That at Druim nan Uadhag (upper Glen Dessarry) occurred at 9.9 ± 0.5 ka and involved disintegration of a detached mass of rock, accompanied by block toppling at the slope crest and release of a

Fig. 14.7 Catastrophic rock-slope failures. **a** The Beinn Alligin rock avalanche, Torridon (4.4 ± 0.3 ka). **b** The Maol Chean-dearg rock avalanche, Torridon (12.4 ± 0.6 ka). **c** The runout lobe of the Carn Ghluasaid rock avalanche, Cluanie (12.7 ± 0.7 ka). **d** The Druim nan Uadhag fragmented rockslide, Dessarry (9.9 ± 0.5 ka). **e** The Carn nan Gillean fragmented rockslide, Sutherland (7.9 ± 0.4 ka). **f** Fragmented rockslide and debris flow, Meall Coire an t-Seallach, Knoydart. (Images: **a**–**d** Colin Ballantyne; **e**, **f** Google Earth™)

spread of large boulders that extends 550 m downslope to the valley floor (Figs. 14.2c and 14.7d). A fragmented slide at Carn nan Gillean (Meall a'Chleirich) in Sutherland (Chap. 12) forms part of a larger RSF, exhibiting tension cracks and displaced blocks at the slope crest and a thick runout tongue of large boulders that extends downslope for 370 m but fails to reach the valley floor (Fig. 14.7e); failure occurred at 7.9 ± 0.4 ka. Both of these examples occur on the flanks of troughs formerly occupied by LLR glaciers.

At a few sites, catastrophic rock-slope failure has been succeeded by downslope flow of debris, either in the form of a broad lobe (Fig. 14.2f) or a narrow tongue bounded by bouldery levées. An example of the latter is a fragmented slide on Meall Coire an t-Seallach (686 m) in Knoydart, where part of the runout debris forms a narrow tongue bounded by bouldery levées up to 10 m high (Fig. 14.7f). Extended downslope flow of debris at this site was attributed by Ballantyne (1992) to momentum transfer through inertial collisions within the mobile mass of boulders.

14.5 Arrested Slides

Most (~66%) arrested slides in the NW Highlands are small failures (<0.1 km^2; Figs. 14.4 and 14.8a), but several exceed 0.25 km^2 and represent major ruptures on steep (>25°) valley-side slopes, spurs and cirque sidewalls, though they are rare on steep rockwalls where catastrophic failures and rockfalls predominate. Rupture surfaces are typically planar, indicating translational sliding controlled by joint alignment or foliation. In most cases, the thickness of the displaced rock mass ranges from 10 m to ~50 m. At some sites, very limited downslope shift of failed rock has occurred (Figs. 14.2a and 14.8d), but many terminate on middle or lower slopes and some reach the valley floor (Fig. 14.8e). Most (~84%) are seated on schist or gneiss, though some outstanding examples occur on other rocks, such as the Breabag RSF on Cambrian quartzite (Fig. 14.2a) and those on Torridonian sandstones in Coire Attadale, Applecross (Fig. 14.8c).

Accounts of individual arrested slides in the NW Highlands tend to be descriptive and provide no information regarding the cause of failure, but study of a RSF on Sgùrr na Lapaich (Affric) by Holmes (1984) is instructive. The rupture surface is inclined at ~26°, and failure apparently took place along the principal joint set. Because the inclination of the failure surface is below the residual friction angle for psammite (~32°), rock-mass release cannot have resulted from progressive failure, even assuming zero cohesion. Holmes concluded that only an exceptionally unfavourable joint-water pressure distribution could have caused kinematic release, though it is feasible that failure was triggered by seismic shaking.

A particularly interesting large arrested slide is located on Moine psammites in a cirque below the summit of Ben Hee (873 m) in Sutherland (Fig. 14.8b). The main body of the slide consists of an upper zone of slipped rock, a middle zone of partly disaggregated rock that extends downslope for ~500 m, and (unusually) a terminal zone of broad transverse ridges. Jarman and Lukas (2007, p 102) described the ridged area as 'tension gashes possibly exploited by fluvial erosion', though the location of the ridges at the slide toe suggests that they may be of compressional origin. Jarman and Lukas concluded that the Ben Hee RSF involved translational sliding along a failure plane inclined at 18°, much lower than the residual friction angle of psammite (~32°), and indeed those of the failure surfaces of other translational slides in the NW Highlands. The cirque was occupied by LLR glacier ice, and Haynes (1976, p 337) considered that the landslide occurred 'while glaciation was in progress', but if so there must have been only a tiny residual glacier in the cirque or RSF debris would have been carried farther. An alternative possible interpretation is that the failure surface is curved, and that the Ben Hee RSF is an atypical case of quasi-rotational sliding in schists, thus accounting for the development of compressional ridges at the slide toe.

14.6 Rock-Slope Deformations

Rock-slope deformations (RSDs), often termed deep-seated gravitational slope deformations (DSGSDs), are widespread in deglaciated mountain environments, but until the present century received remarkably little attention from geomorphologists. The burgeoning recent literature on these massive mountain landforms has been summarised by Agliardi et al. (2012) and Pánek and Klimes (2016), but there is still uncertainty regarding the causes and mechanisms of slow (10^{-1}–10^{2} mm a^{-1}) deep-seated rock deformation. The onset of deformation has been widely attributed to deglacial unloading or debuttressing of slopes leading to immediate or delayed 'creep' of steep mountain slopes, particularly on anisotropic fractured schistose rocks. The nature of movement has been variously attributed to 'rock-mass creep', elasto-plastic deformation, shear along deep-seated discontinuities or shear zones, and/or progressive failure due to sequential shearing of rock bridges and reduction of frictional strength along joints. Seismic activity and high-joint water pressures are known to activate or episodically accelerate RSDs (Moro et al. 2007, 2012), particularly in areas with high uplift gradients (Pedrazzini et al. 2016). Over long timescales, RSDs eventually stabilise or may fail completely as rockslides or rock avalanches.

The most characteristic features of RSDs in the NW Highlands are arrays of antiscarps up to ~10 m high produced by shallow faulting or block toppling within the mobile slope (Fig. 14.9a–d) and slope bulging (Figs. 14.2e and 14.9a, c), though the latter is not always evident. Lateral margins are often indistinct (Fig. 14.9b, c), headscarps are typically subdued, and some RSDs are represented only by a single scarp along the slope crest indicating subsidence of the adjacent slope. Most (87%) are located on schists (Table 14.1) though they also occur on other rocks; on Torridonian sandstones, RSDs are sometimes represented by antiscarps (Fig. 14.9d) and elsewhere by downslope 'sagging' of sandstone strata.

The most extensive area of rock-slope deformation in Scotland occurs on the southern flank of Beinn Fhada (1032 m) in Kintail and covers an area of ~3.1 km^2

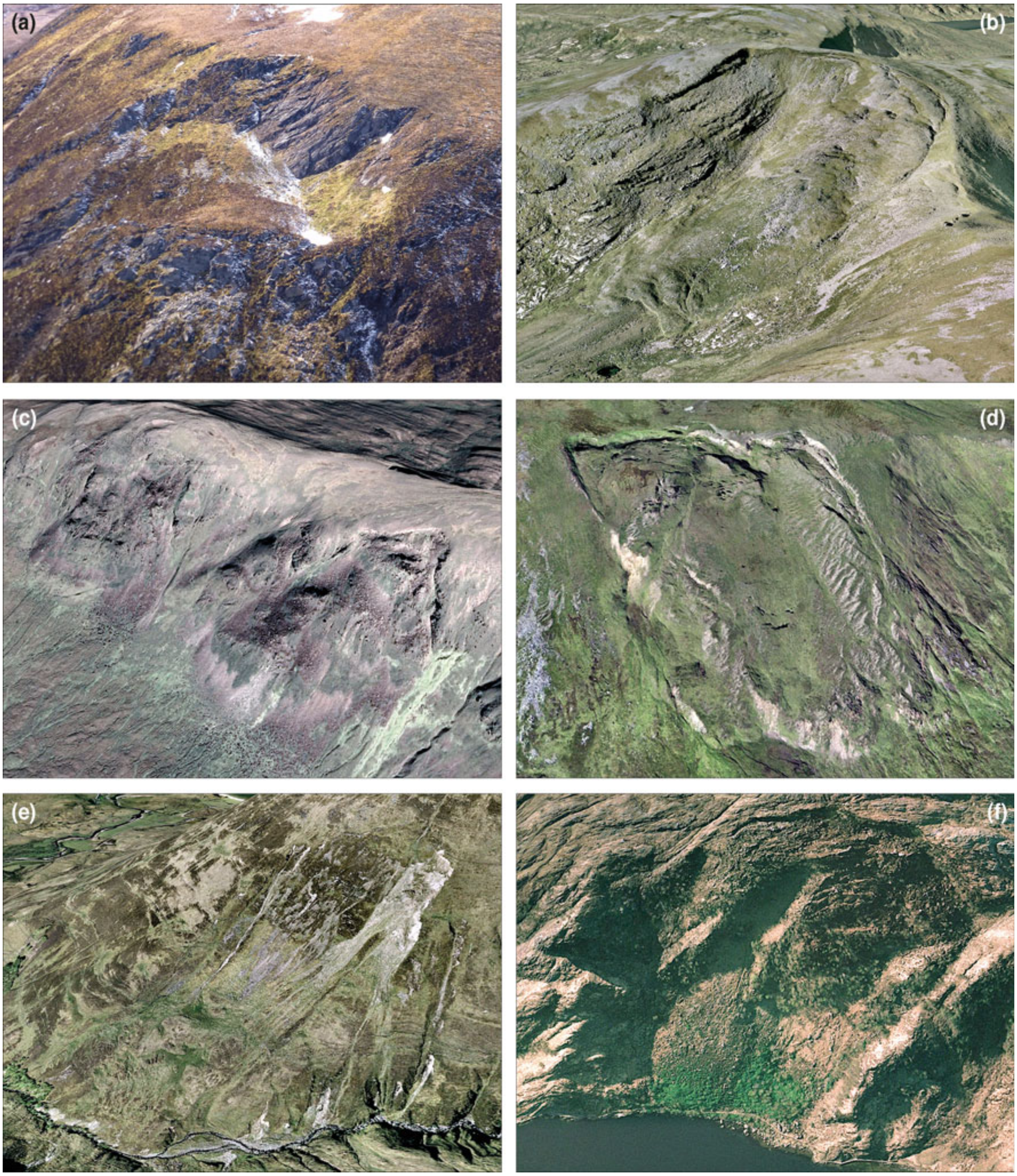

Fig. 14.8 Arrested slides; all except (**c**) are on Moine schists. **a** Small (<0.02 km^2) failure at the slope crest, Tom na Caillich, Easter Ross. **b** The Ben Hee RSF, Sutherland; the main body of the slide covers ~0.40 km^2. **c** Two partly fragmented arrested slides (0.14 and 0.19 km^2) on Torridonian sandstone, Coire Attadale, Applecross. **d** Arrested slide (0.14 km^2) with limited downslope displacement, Sgùrr na Ruaidhe, Strathfarrar. **e** Full-slope rockslide on Beinn Bhuidhe, Arnisdale (0.11 km^2); the slide mass has translated to the slope foot. **f** Arrested slide (0.13 km^2) at the head of Loch Hourn. The headscarp is at 400 m, and the central part of the displaced rock mass has collapsed to form a debris tongue that extends to the shore of the fjord. (Images: **a** Colin Ballantyne; **b**–**f** Google Earth^{TM})

(Ballantyne and Jarman 2007; Fig. 14.9a). The deformed slope rises 790–900 m above the adjacent trough floor at average gradients of 30–35°. Although the valley was occupied by glacier ice ~400 m thick during the Loch Lomond Stade, the toe of the RSD exhibits no evidence of glacial modification, suggesting that movement continued into the Holocene, though deformation of the upper slope may have been initiated much earlier. This huge RSD lacks distinct lateral margins but is defined by a remarkable array of antiscarps that extend from near the valley floor to the

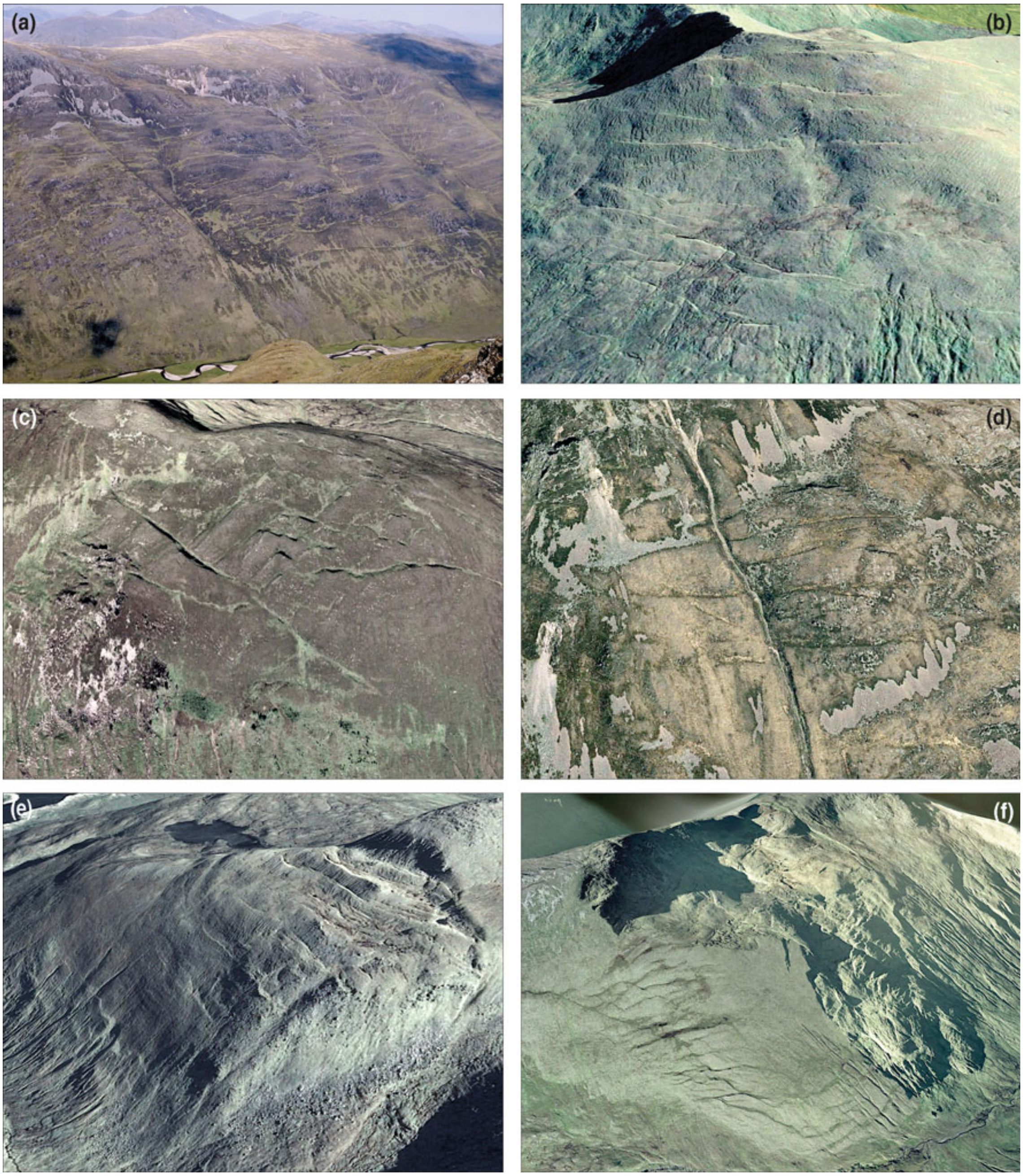

Fig. 14.9 Rock-slope deformations and related landforms. All are on schists except (**d**). **a** The Beinn Fhada RSD in Kintail, the largest in Scotland. **b** Antiscarp arrays of the An Socach RSD, Mullardoch. **c** The An Sornach RSD, Affric, showing partial collapse of the margin. **d** Antiscarps of a small (0.15 km^2) mid-slope RSD on Torridonian sandstone, Liathach, Torridon. **e** The Druim Shionnach RSF, Glen Shiel, a hybrid RSD/arrested slide. **f** RSF on Mullach nan Dheiragain, Affric, where two-thirds of a large ($\sim$1.3 km^2) RSD have apparently collapsed to form an arrested slide. (Images: **a** Colin Ballantyne; **b**–**f** Google Earth^{TM})

slope crest: those on the upper slope are up to 10 m high and 800 m long. Most are aligned across the slope, though some dip gently and a few converge. The main area of deformation forms three prominent bulges below a degraded, discontinuous ($\sim$35°) scarp, but antiscarps link the west and central bulges and extend onto the plateau upslope, indicating that deformation was not confined to the bulges. As with other RSDs in Scotland, the timing, cause and mechanics of movement remain conjectural: Watters (1972) inferred that the Beinn Fhada RSD represents a deep-seated translational slide; Fenton (1991) proposed seismically-triggered sliding failure; and Holmes and Jarvis (1985) postulated deep internal rock deformation, with formation of antiscarps by joint-guided block-flexural toppling.

A large (~ 0.50 km^2) RSD on the southern slopes of An Sornach (Glen Affric; Fig. 14.9c) is defined by low antiscarps, with a headscarp up to 40 m high inclined at $\sim 40°$ and a sidescarp at its western end, below which part of the RSD exhibits toppling collapse. Deformation has been attributed to translational sliding, possibly along slope-parallel stress-release joints, but if the headscarp represents an exposed failure surface this interpretation implies that downslope movement was accommodated entirely by slope bulging. The conspicuous antiscarp or trench that descends eastwards across the feature may represent a neotectonic fault scarp.

Another large (~ 0.56 km^2) RSD on Druim Shionnach (Glen Shiel) illustrates the hybrid nature of some slope deformations. The upper part of this landform consists of a headscarp up to 25 m high and a broad trench (tensional rift) that is bounded on its downslope side by a steep antiscarp (Jarman 2007b). Farther downslope the surface of the displaced block forms a broad bulge, interrupted by minor antiscarps and benches, that descends at an average gradient of 22° before terminating at a 150 m high slope inclined at up to 45° (Fig. 14.9e). Jarman (2007b) depicted movement of the displaced mass over a fracture zone inclined at 16°, ~ 100 to 150 m below the surface. This reconstruction implies (for a 100 m thickness of rock) a normal stress of ~ 2.5 MPa and a shear stress of ~ 0.7 MPa along the fracture zone. Shear under such conditions is improbable, and a more complex structural configuration is likely, but difficult to reconcile with the evidence for extensional movement downslope from the headscarp. It is possible that 'hybrid' RSDs such as that on Druim Shionnach belong to a category of large, slow-moving landslides as opposed to rock-slope deformations *sensu stricto* (Agliardi et al. 2012; Pánek and Klimes 2016), though differentiation of the two from morphology alone is difficult.

Some RSDs in the NW Highlands exhibit zones of catastrophic failure (topples or fragmented slides) at their margins (Fig. 14.9c). These have occurred where slow downslope movement has reduced rock-mass strength to criticality through brittle shear of rock bridges and reduction of frictional strength to residual values. The most striking example of a partly-failed RSD is that on Mullach nan Dheiragain (Affric), where a bulging RSD adjoins a large arrested slide (Fig. 14.9f), suggesting that both initially formed part of a large (~ 1.3 km^2) RSD, two-thirds of which subsequently collapsed. Some (possibly many) catastrophic failures and arrested slides in the NW Highlands almost certainly represent RSDs that accelerated towards kinematic release even as others gradually stabilised.

14.7 Debris-Free Failure Scars and the Role of RSFs in Long-Term Landscape Evolution

TCN dating of catastrophic RSFs has demonstrated that almost all of those located outside the limits of the LLR failed within the Lateglacial interval between ice-sheet deglaciation and the end of the Loch Lomond Stade (Ballantyne et al. 2014). An important implication is that numerous RSFs must have occurred inside the limits of the LLR within the same period, but that the resultant failed rock or runout debris was removed by glacier ice so that failure sites are now represented by debris-free failure scars (DFFSs). Such landslide scars were first recorded in Geological Survey memoirs, and Holmes (1984) recorded 68 large DFFS sites in the Scottish Highlands. Common attributes include a planar, stepped or irregular failure plane or rupture surface, a straight, arcuate, irregular or wedge-shaped headscarp (which may be absent where failure removed a ridge crest), sidescarps and tension cracks (Ballantyne 2013; Fig. 14.10). The author has identified over 250 possible DFFSs > 0.01 km^2 in the NW Highlands. Most occur on schists though they are present on all lithologies (Cave and Ballantyne 2016). Large cavities or hollows representing DFFSs occur, for example, in Torridonian sandstone below the summits of Beinn Damh and Fuar Tholl in the Torridon-Strathcarron area, and spectacular debris-free failure surfaces are present on the Cambrian quartzite slopes of Foinaven in Sutherland. Most DFFSs are small (< 0.1 km^2), but some represent the sites of major RSFs: that on Garbh Cioch Beag (Fig. 14.10b), for example, represents translational sliding of rock over an area of ~ 0.5 km^2.

The fate of RSF debris removed by LLR glaciers is poorly understood. At a few sites, landslides appear to have deposited debris directly on to glaciers, which then transported it farther downvalley (Ballantyne and Stone 2013). Benn (1989) has shown that large lateral moraines occur downvalley from cliffed cirque sidewalls, suggesting that the moraines contain a substantial component of reworked RSF and rockfall debris. It is likely that Lateglacial RSFs that occurred before or during the LLR also contributed to the huge volume of sediment in the multiple recessional moraines that formed during oscillatory retreat of LLR glaciers. At a few sites, rock avalanche debris has survived over-riding by LLR glaciers, forming a lobe or tongue of streamlined hummocks (Ballantyne 2018).

DFFSs also occur outside the limits of the LLR, implying that such sites represent RSFs that occurred prior to the expansion of the last ice sheet. Moreover, some DFFSs inside the limits of the LLR exhibit ice moulding of

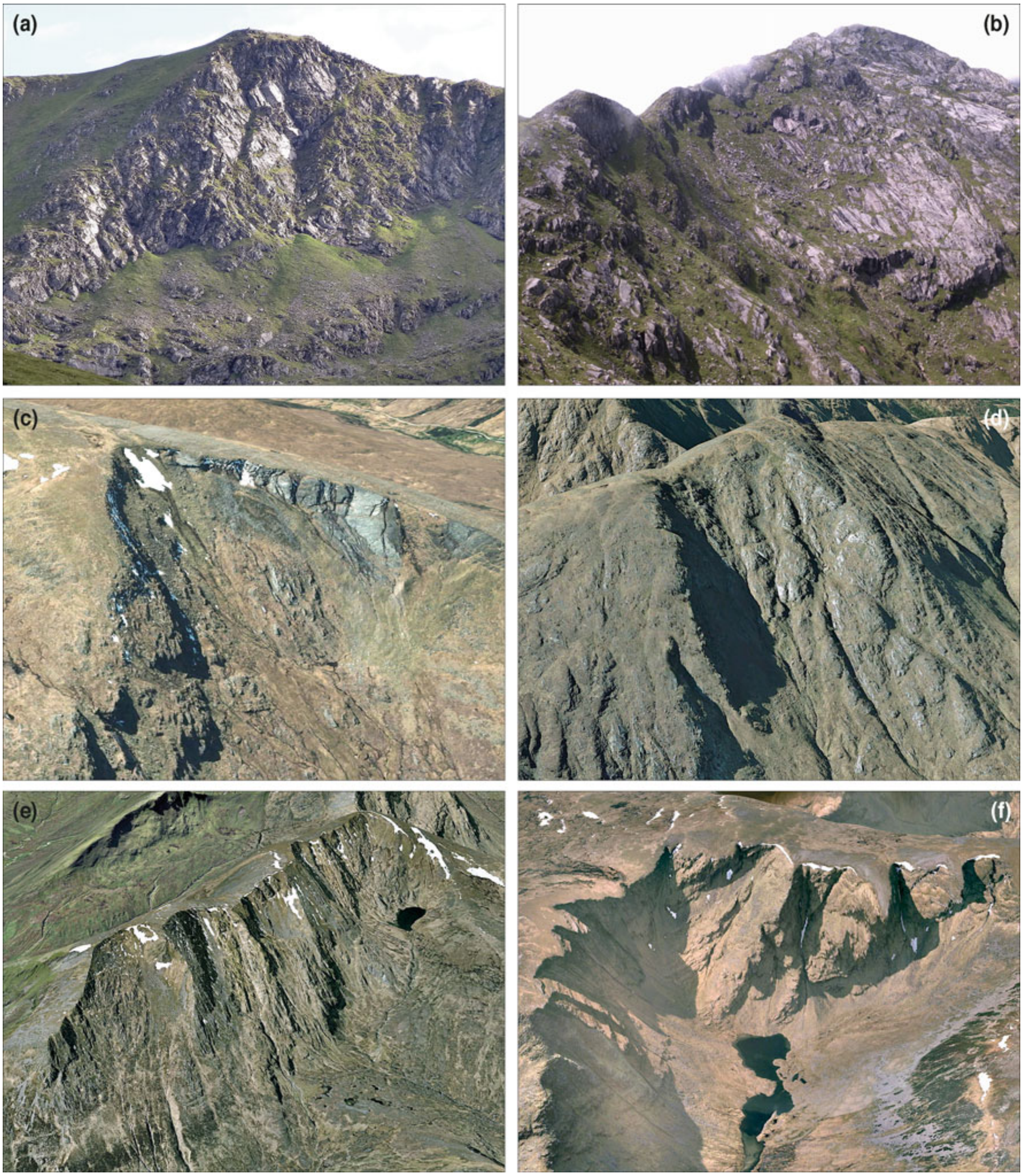

Fig. 14.10 Debris-free failure scars (DFFSs) on schists. **a** Small DFFS on Sgùrr Coinnich, Monar. **b** Stepped failure plane of a large ($\sim$0.5 km^2) DFFS, Garbh Cioch Beag, Dessarry. **c** Headscarp and failure surface of an arcuate DFFS (0.15 km^2), Moruisg, Strathcarron. **d** Cavity marking a large (0.23 km^2) DFFS on Carn na Nathrach, Ardgour. **e** Multiple DFFSs along the east ridge of Sgùrr a'Chaorachain, Monar. **f** Multiple DFFSs along a cirque headwall, Seana Braigh, Inverlael. All of these sites were partly occupied by glacier ice during the Loch Lomond Stade, implying that runout debris may have been removed by glaciers during the Loch Lomond Readvance, but ice moulding of the buttresses flanking (**d**) suggests that failure here may have pre-dated the last ice sheet. (Images: **a**, **b** Colin Ballantyne; **c**–**f** Google Earth^{TM})

sidescarps and intervening buttresses (Fig. 14.10d), suggesting that these also represent the sites of ancient RSFs that occurred prior to ice-sheet expansion. Survival of pre-Last Glacial Maximum DFFSs demonstrates that the cavities formed by ancient rock-slope failures are persistent landscape features and indicates limited erosion of upper slopes by the last ice sheet.

The abundance of DFFSs and postglacial RSFs in the NW Highlands suggests that rock-slope failure at a range of scales has played an important role during successive interglacial periods in widening glacial troughs and tributary valleys, and probably a dominant role in the formation of arêtes (Ballantyne 2013, 2019; Chap. 13; Fig. 14.10e). Additionally, many cirques in the region are skirted by talus slopes that demonstrate postglacial retreat of headwalls and sidewalls by small-scale rockfalls, but others are scalloped by failure scars that demonstrate rockwall retreat and cirque widening by larger RSFs (Fig. 14.10f). As Figs. 14.2 and 14.10 illustrate, the large-scale morphology of many mountain summits and ridges in the NW Highlands has been strongly influenced by interglacial, interstadial and postglacial rock-slope failure, and is not simply an outcome of successive episodes of glacial erosion.

Jarman (2009) proposed a model of trough widening by rock-slope failure in Scotland. This postulates that valley modification by paraglacial RSFs was pronounced during early stages of trough development, then slowed as slopes became 'stress hardened', before being rejuvenated by 'glacial reincision' during the 'late Quaternary'. As noted in Sect. 14.3, however, *all* episodes of glaciation and deglaciation result in rock-slope weakening through ice loading and unloading, glacial erosion and both thermomechanical and hydromechanical rock damage. It follows that all glacial-deglacial cycles (and particularly those involving ice-sheet growth and decay) have induced widespread paraglacial rock-slope failure, with removal of displaced rock masses and runout debris during subsequent glaciations. The present form of the mountains, ridges, troughs, cirques and arêtes of the NW Highlands has been dictated by a synergic relationship between successive glaciations and ensuing episodes of paraglacial rock-slope failure that has operated since the earliest Pleistocene.

14.8 Conclusions

The North West Highlands contain an exceptional range of paraglacial rock-slope failures. Over 400 have been identified, mainly on schistose rocks, though they are represented on all lithologies. Most take the form of rock-slope deformations or arrested slides; fragmented slides are rarer, and rock avalanches *sensu stricto* are restricted to a few localities where rockwall collapse has fed runout tongues of coarse debris. Postglacial rockfalls and rock topples are small-scale failures mainly restricted to steep scarps and cirque headwalls. Understanding of the causes and nature of failure is limited by a paucity of rigorous geotechnical surveys and dating evidence. The available data suggest that many postglacial RSFs could not have occurred by self-weight sliding or toppling and that most catastrophic failures lagged deglaciation by centuries or millennia. This delay and the abundance of RSDs suggests that many rapid RSFs resulted from progressive failure of slopes following deglaciation. The role of uplift-induced seismic activity in weakening rock slopes and triggering failure remains conjectural but may offer a solution to the problem of why many technically stable slopes have failed. Perhaps the most significant conclusion of recent research is that paraglacial rock-slope failure has operated at the end of multiple glacial cycles, playing an important role in trough and cirque widening and the formation of arêtes. The large-scale form of the mountain landscape of the NW Highlands has been dictated not only by successive episodes of glacial erosion acting on a dissected pre-glacial land surface, but also by recurrent failure of steep rock slopes during and after each period of deglaciation.

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The Glacial Geomorphology Around Inverness and the Great Glen

15

Jon W. Merritt and Clive A. Auton

Abstract

Superimposed on ancient landscape elements, the Inverness region includes a palimpsest of subglacial landforms formed during successive Late Devensian ice movements. It contains a particularly rich and diverse set of sediments and landforms created close to retreating glacier margins, together with the legacy of a major oscillating tidewater outlet glacier. At least five phases of glaciation have been recognised, although they remain poorly constrained temporally. The region includes well-documented buried glacial rafts of arctic shelly clay, two internationally important localities where organic deposits pre-date the last ice sheet, and a good geomorphological record of Late Devensian and Holocene relative sea-level change.

Keywords

Glacial geomorphology • Eskers • Kame terraces • Transverse ridges • Raised beaches • Megagrooves

15.1 Introduction

Situated at the mouth of the Great Glen, the Inverness area is unrivalled in the British Isles for the range of Quaternary landforms and glacigenic deposits it contains. It includes many textbook examples of glacial features, good evidence of Late Devensian and Holocene sea-level change, and the legacy of the punctuated retreat and oscillation of a major outlet glacier that terminated in the Inner Moray Firth, then a glacimarine fjord (Merritt et al. 1995). The Great Glen is a remarkably straight, fault-controlled, steep-sided and glacially overdeepened trench that is clearly visible from space and separates the Northern Highlands and Grampian Highlands geological terranes (Chap. 2). Reaching depths of over 225 m, Loch Ness, the largest freshwater body in Scotland, is located at the north-eastern end of the glen, bounded by remote, mountainous ground mainly between 450 and 900 m above Ordnance Datum (OD). The NE–SW structural trend of the bedrock imparted during the Caledonian Orogeny has a controlling influence on the large-scale topography of the region, particularly the orientation of the Great Glen and Strathglass, and the principal river valleys, those of the Rivers Nairn, Findhorn and Spey (Fig. 15.1a). The uplands to the south of the Inner Moray Firth have been deeply dissected by the River Findhorn, which enters a sinuous, 200–300 m deep open gorge (The Streens) downstream of Tomatin (Fig. 15.2a). The A9 trunk road and railway utilise a glacial breach in the regional Spey/Dulnain-Findhorn divide at Slochd Mòr (401 m). Strathdearn, otherwise known as the Moy Gap (Young 1980), is a major glacial breach in the Findhorn-Nairn divide.

Detailed accounts of the Quaternary landforms and deposits in the Monadhliath, Great Glen and to the east of Inverness are given in Boston et al. (2013) and Merritt et al. (2017a).

15.2 Geology and Landscape Evolution

The rugged mountains of the Northern Highlands and Monadhliath bordering the Great Glen are mostly formed of polydeformed and metamorphosed Neoproterozoic metasedimentary rocks (psammite, semipelite and pelite) intruded locally by granitic rocks, some of which have provided useful indicator erratics (Fig. 15.1a). The Northern Highlands and Grampian Highlands terranes docked left-laterally along the

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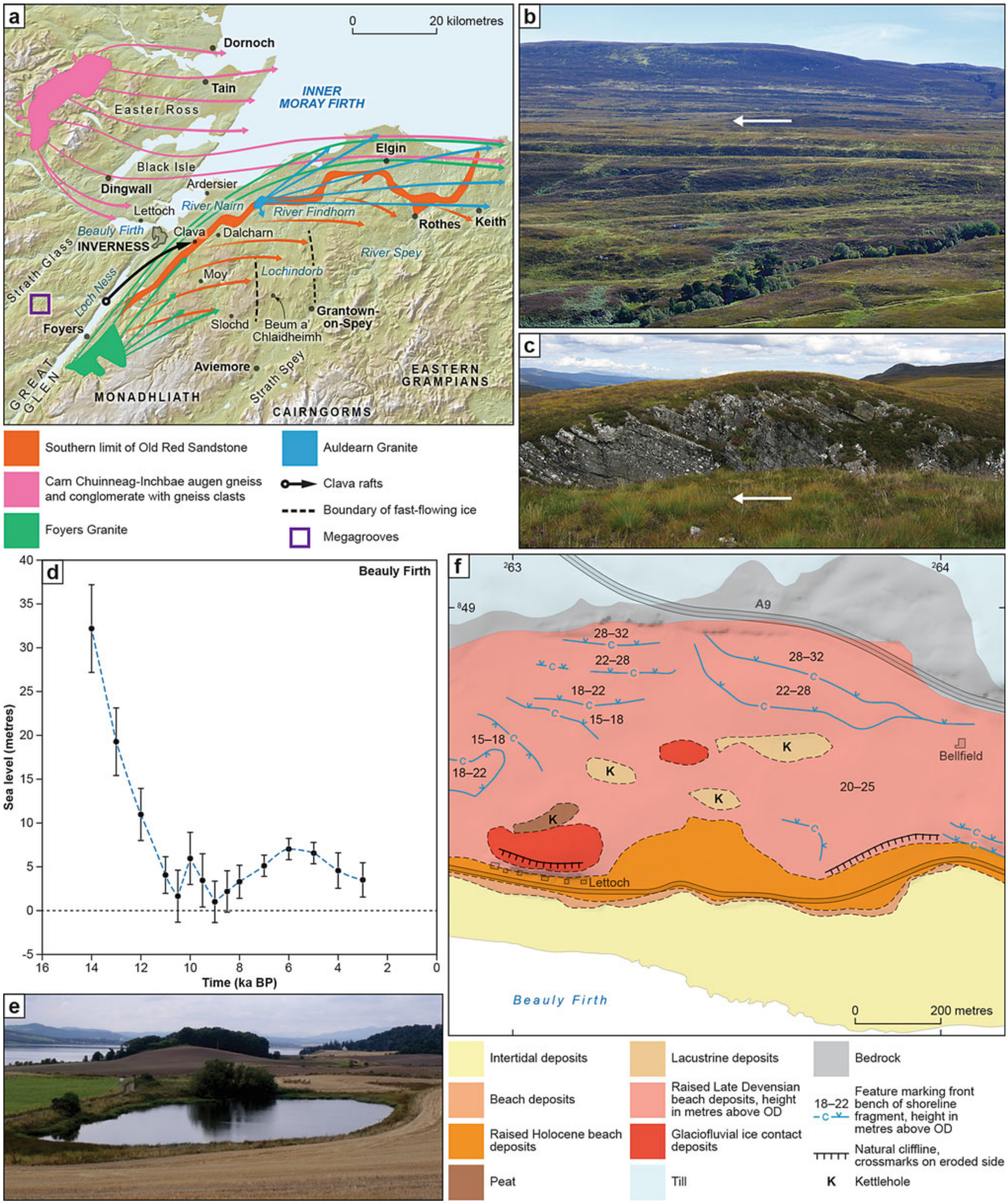

Fig. 15.1 **a** Location map and transport paths of some indicator glacial erratics (after Merritt et al. 2019). **b** Megagrooves on the western flank of Suidhe Ghuirmain in the Endrick valley, Strath Glass, looking SE. **c** Rock drumlin associated with the megagrooves. Ice flowed NE towards the Moray Firth. **d** Relative sea-level curve for the Beauly Firth (after Firth 1989, modified by Lambeck 1995). **e** Kettle hole in Late Devensian raised shorelines near Lettoch, on the northern shore of the Beauly Firth. Ice-contact sands and gravels form the wooded ridge in the distance. **f** Map of the kettled Late Devensian shoreline fragments between Lettoch and Bellfield, on the northern shore of the Beauly Firth. (Images: **b** BGS registered photo P698979; **c** BGS registered photo P698980; **e** Clive Auton)

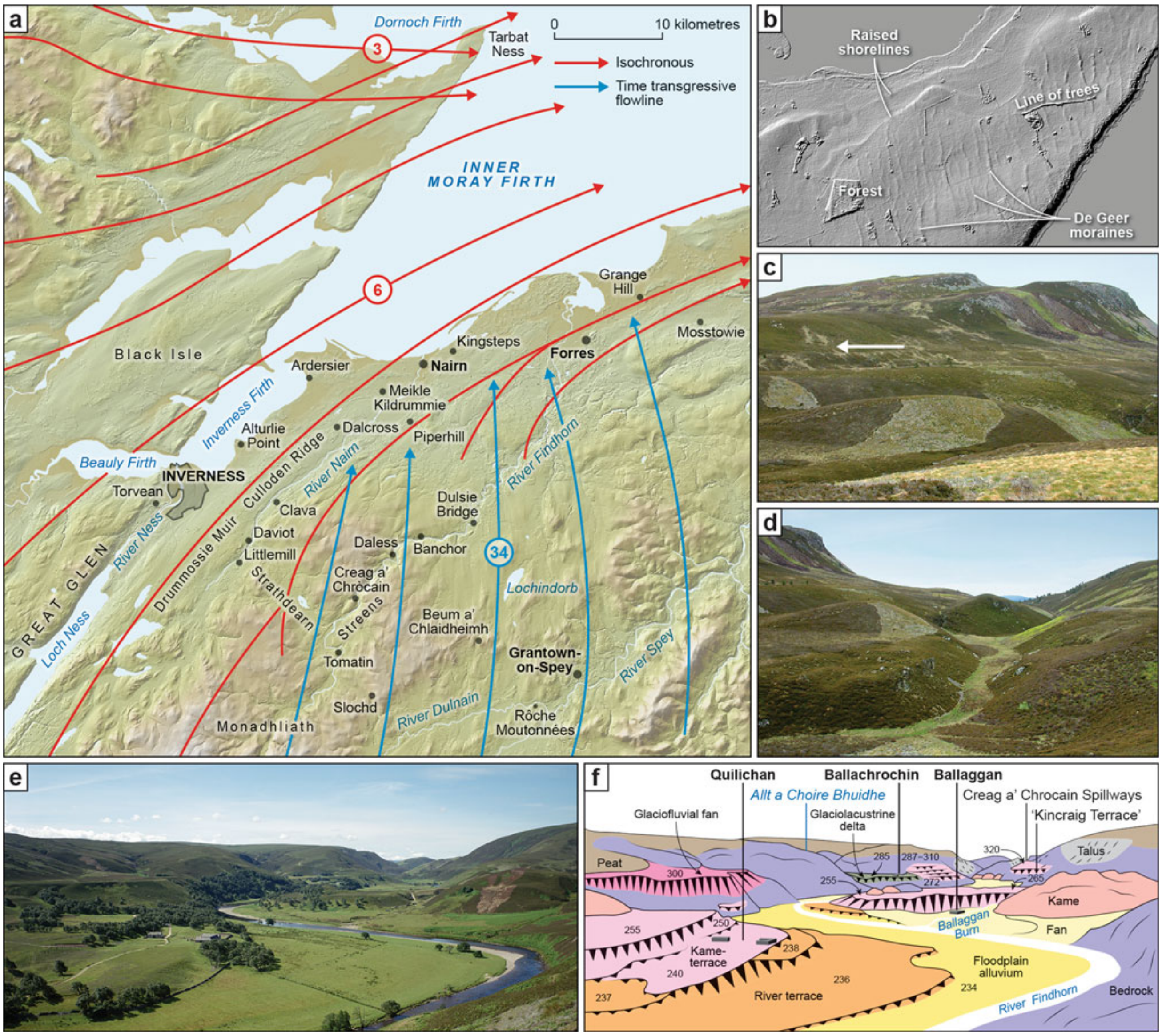

Fig. 15.2 **a** Topography with numbered flowsets identified by Hughes et al. (2014). **b** Transverse till ridges and raised shorelines on Tarbat Ness. **c** Crag-and-tail features at Beum a' Chlaidheimh, looking SE. **d** Glacial spillway at Beum a' Chlaidheimh, looking S towards the breach. **e** The middle Findhorn valley looking SW from Daless. **f** Geomorphological interpretation with elevations in metres above OD (after Fletcher et al. 1996). (Images: **b** Hill-shaded surface model built from NEXTMap™ Britain elevation data © Intermap Technologies with illumination from NW; **c**, **d** Jon Merritt; **e** BGS registered photo P008519)

Great Glen Fault Zone towards the end of the Caledonian Orogeny (430–400 Ma), with lateral displacement of at least 200–300 km (Strachan et al. 2002). This 3 km wide fault zone, formed of relatively weak, intensively sheared and fractured rock, subsequently, experienced 25–30 km of right-lateral strike-slip movement from the Late Palaeozoic, followed by downthrow to the south-east. The ground flanking the mouth of the Great Glen, including the Black Isle and around the Inner Moray Firth, is mostly underlain by sedimentary rocks of the Devonian Old Red Sandstone Supergroup (conglomerate, sandstone and siltstone), which also occur within a narrow graben, bounded on its southern side by the Sron Lairg Fault (British Geological Survey 2012a).

Although deeply ice-scoured during successive Pleistocene glaciations, the Great Glen first emerged as a major landscape element in the Devonian, during the degradation of the mountains formed during the Caledonian Orogeny, but before much of the sinistral transcurrent faulting. Most other major landscape elements in the region had become established by the end of the Mesozoic and continued to develop in response to tectonic uplift and differential chemical weathering under warm, humid climates through the Palaeogene and Neogene (Hall 1991; Chap. 3).

15.3 Key Quaternary Sites

This region includes one of the few sites in Scotland where organic material lying beneath glacial till has been assigned to the Last (Ipswichian) Interglacial (Marine Isotope Stage 5e; Walker et al. 1992; Merritt et al. 2019). At Dalcharn, 15 km east of Inverness (Fig. 15.1a), the Dalcharn Palaeosol contains compressed and disseminated organic material that includes pollen of full interglacial affinity. The pollen record indicates that closed pine forest with birch, alder and holly was succeeded by pine-heathland, which was replaced initially by birch and later by heath and open grassland as the climate deteriorated. As the Dalcharn site lies near to the present northern limit of holly in Britain, the relative abundance of *Ilex* pollen in the profile almost certainly reflects a climate somewhat warmer than that of today. A compressed bed of fibrous peat found beneath till at the nearby Allt Odhar site, near Moy, contains pollen, plant macrofossils and coleopteran remains that reveal climatic deterioration towards the end of a cool interstadial period as reflected in the replacement of birch woodland and willow scrub with grassland and heath, and then by open communities of grass and sedges. An Early Devensian (MIS 5a or 5c) age is most likely on current evidence.

At Clava, in the Nairn valley, there is an important, well-documented site long known for discoveries of enigmatic shelly clay and shelly diamicton (Merritt et al. 2017a, 2019). The Clava Shelly Clay contains well-preserved, high-boreal to low-arctic, shallow-water marine shells, whereas an associated folded body of pebbly diamicton (Clava Shelly Till) contains fragments of *Portlandia arctica*, an indicator of shallow-marine, fully arctic conditions. At elevations of 150 m OD, these sediment bodies are now generally accepted to be glacial rafts transported eastwards from the Loch Ness Basin (Fig. 15.1a), which was probably a marine fjord during the Middle Devensian (Merritt et al. 2019).

15.4 Late Devensian Glacial History

Although the area was glaciated repeatedly during the Pleistocene (Hall et al. 2019), the last (Late Devensian) Scottish Ice Sheet expanded at ~32 ka BP and probably reached its maximum extent before the global Last Glacial Maximum (LGM) at ~24 ka (Clark et al. 2012; Ballantyne and Small 2019; Chap. 4). Several distinct flowsets of subglacially streamlined landforms have been recognised (Fig. 15.2a) and cross-cutting relationships indicate that major switches in flow occurred during the last glaciation (Hughes et al. 2010, 2014). Evidence suggests that there were at least five distinct phases during the last glaciation (Merritt et al. 2017a, 2019), as described below.

15.4.1 Phase 1: Early Eastward Ice Flow

Few, if any, landforms may be assigned with confidence to this phase, but the distribution of indicator erratics and the composition and lithostratigraphy of tills indicate that ice flowed eastwards from the Great Glen across the middle to lower reaches of the Nairn, Findhorn and Spey valleys (Fletcher et al. 1996; Fig. 15.1a). The timing of this early phase of ice-sheet glaciation is uncertain but was influenced by a relatively thick build-up of ice to the north-west. Dated organic beds and ice-rafted shelly material found locally within glacigenic sequences indicate that this flow mainly occurred after MIS 5a and possibly during MIS 3 (Merritt et al. 2017b, 2019).

15.4.2 Phase 2: Last Glacial Maximum (LGM)

The distributions of crag-and-tail features, roches moutonnées and glacial striations indicate that the ice sheet subsequently buried the entire region (Ballantyne and Small 2019). However, a narrow swathe of ground around Lochindorb (Fig. 15.1a) containing streamlined subglacial landforms suggests that a corridor of relatively fast-flowing ice (an ice stream) flowed northwards from the Cairngorms and upper Strath Spey across the lower Findhorn valley and into the Inner Moray Firth Basin (Hughes et al. 2010, 2014; Merritt et al. 2017a). Flow was concentrated through the Beum a' Chlaidheimh breach (flowset 34 in Fig. 15.2a). The presence of interlocking spurs and decomposed regolith within The Streens Gorge upstream of Daless suggests that the middle Findhorn valley has experienced relatively little cumulative glacial erosion, like most of the north-eastern Grampians (Hall et al. 2019). By contrast, the mountains to the north of the Great Glen are heavily ice-scoured. For example, a swathe of splendid glacially streamlined mega-grooves (Newton et al. 2018), whalebacks and rock drumlins occurs in the upper reaches of Strath Glass (British Geological Survey 2012b; Boston et al. 2013; Fig. 15.1b, c). These features probably are the legacy of an ice stream within the last ice sheet that flowed directly through Strath Glass towards the Moray Firth.

15.4.3 Phase 3: Early Upland Deglaciation

Following the LGM, the ice sheet eventually became too thin for ice to be driven northwards over the Spey-Findhorn divide, and a major glacial reorganisation ensued after 21.3 ka (Merritt et al. 2019). Ice flowing from the Cairngorms and from upper Strath Spey gradually parted from ice that flowed north-eastwards into the Moray Firth from the Monadhliath and Great Glen. A substantial outlet glacier became established within Strath Spey upstream of Grantown-on-Spey between 18 and 15 ka (Hall et al. 2016), and an ice-free enclave opened up that included the middle Findhorn valley, within which there was widespread deltaic and glacilacustrine sedimentation (Merritt et al. 2017a; Fig. 15.2e, f).

15.4.4 Phase 4: Establishment of the Moray Firth Ice Stream (MFIS)

The presence of widespread, well-preserved and relatively elongate subglacial landforms across the Black Isle indicates that a vast ice stream drew down ice from the west towards the Moray Firth Basin (Hughes et al. 2010, 2014; flowset 6 in Fig. 15.2a). The MFIS possibly became established during the Elgin Oscillation at ~15 ka when ice re-advanced into lower Strath Spey from offshore (Peacock et al. 1968; Clark et al. 2012; Merritt et al. 2017b). The subglacial landforms of flowset 6 clearly truncate, and are younger than those of flowset 34, a temporal relationship that is corroborated by the lithostratigraphy (Merritt et al. 2017a, 2019). The MFIS was pinned against opposing slopes to the northeast of Inverness, where it also blocked the lower reaches of the Findhorn valley causing ponding upstream.

15.4.5 Phase 5: Punctuated Retreat of the MFIS

Ice-marginal drainage was concentrated at progressively lower elevations along Strath Nairn during deglaciation. Staircases of glacifluvial (kame) terraces and fans formed sequentially, whilst the lateral margin of the MFIS withdrew towards the Inverness Firth. Numerous sets of low transverse till ridges were formed at its westward-retreating frontal margin on the coastal lowland as far east as Elgin. The Inverness Firth was occupied by a receding tidewater outlet glacier that progressively uncoupled from adjacent, stagnating land-based ice and ice-cored glacifluvial deposits (Alturlie Gravels; Merritt et al. 1995). A significant glacial readvance (the Ardersier Readvance) subsequently deformed glacimarine deposits (Ardersier Silts) within the firth to create a push moraine stretching between Castle Stuart and the Ardersier peninsula (Fig. 15.3a). This undated event followed an oscillation of the ice front at Grange Hill (Fig. 15.2a) and preceded a minor readvance from the Beauly Firth, during which a large fan-delta formed at Alturlie Point (Bothyhill Gravels; Merritt et al. 1995, 2017a). The Great Glen glacier had a lightly grounded, or potentially floating ice shelf at the mouth of the glen, but subsequently retreated to a more grounded position at Foyers, where a major end moraine beneath the loch was created during a stillstand or glacial readvance (Turner et al. 2012; Boston et al. 2013). The entire region became deglaciated during the Lateglacial Interstade, a period of ameliorated climate between ~ 14.7 and 12.9 ka (Chap.4).

15.4.6 The Loch Lomond Stade

Although glaciers developed in the Monadhliath, Cairngorms and Western Highlands during the Loch Lomond (≈Younger Dryas) Stade of ~12.9 to 11.7 ka, none extended onto the lower ground of the region (Boston et al. 2013). However, cobble gravels underlying an extensive, low-lying raised marine terrace adjacent to the Kessock Bridge (Fig. 15.3a) represent the top of a marine delta that formed during successive flood events of the River Ness (Merritt et al. 1995; British Geological Survey 1997; Boston et al. 2013). Sissons (1981) proposed that the delta resulted from an enormous glacial outburst flood when the 260 m ice-dammed lake in Glen Roy and Glen Spean drained catastrophically towards the end of the Loch Lomond Stade (Chap. 16).

15.5 Sea-Level Change and Raised Shorelines

A flight of up to ten raised shorelines of Late Devensian age has been identified around the Inner Moray Firth, indicating there was a progressive fall in relative sea level (RSL) from at least 35 m OD whilst ice retreated westwards into the Beauly Firth and south-westwards into the Great Glen (Firth 1989; Fig. 15.1d). Each shoreline is tilted north-eastwards at progressively gentler gradients down the sequence; this occurred as a result of diminishing glacio-isostatic rebound whilst the glaciers retreated. RSL in the Beauly Firth fell to a minimum of +2 m OD early in the Loch Lomond Stade, before rising to create a prominent abandoned cliffline around the Beauly and Inverness Firths later in the Stade. The period of marine erosion was replaced by deposition and falling relative sea level during the Early Holocene when

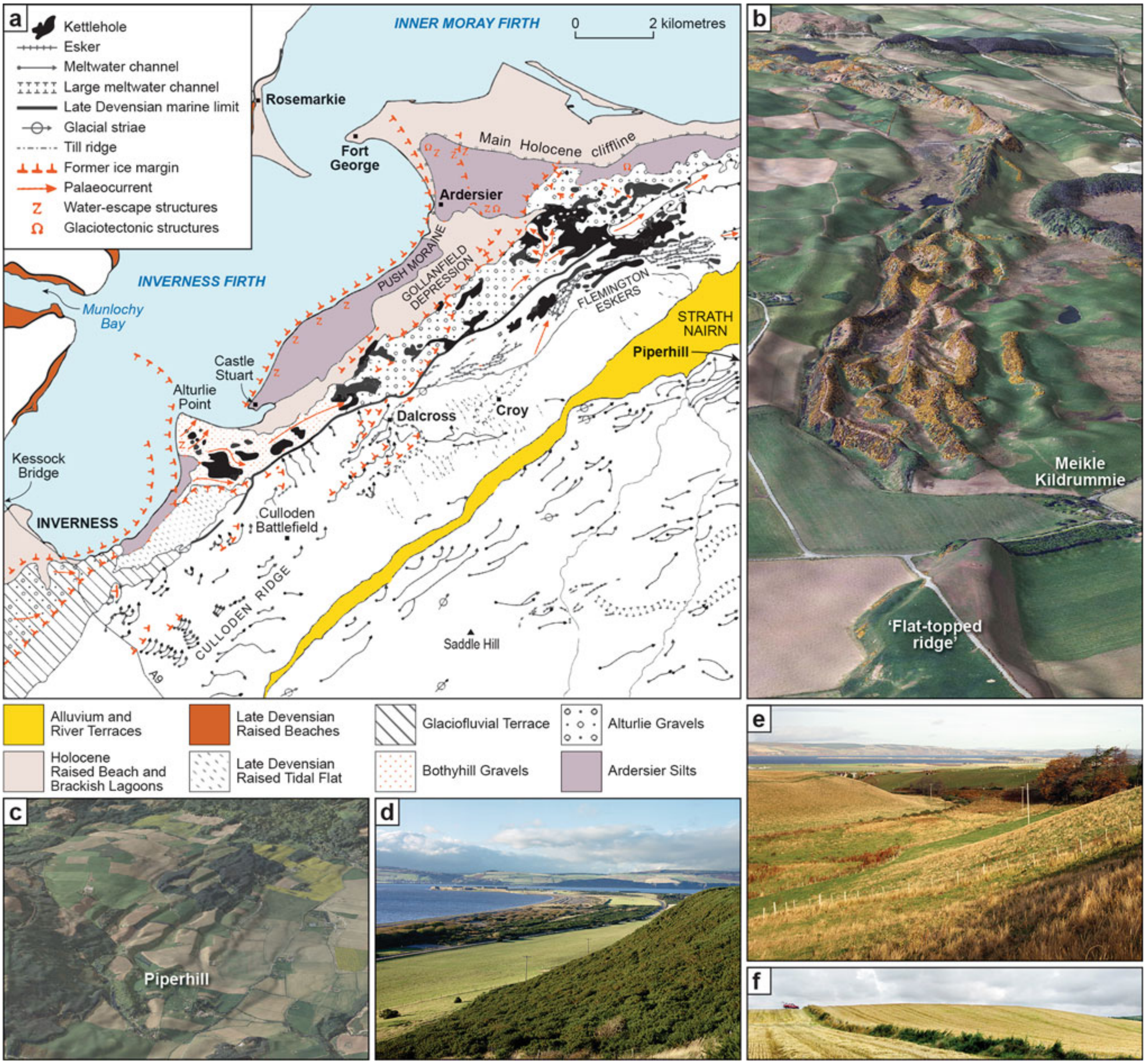

Fig. 15.3 **a** Geomorphological features and deposits associated with deglaciation of the Inverness Firth (after Merritt et al. 1995). **b** The Flemington Esker system looking WSW; ×4 vertical exaggeration. **c** Ice-marginal benches, lateral moraines, glacifluvial terraces and glacial drainage channels in lower Strath Nairn, looking SW towards Piperhill; ×3 vertical exaggeration. **d** View looking NW towards Fort George on the Ardersier peninsula. Mid-Holocene raised beach abutting a slope formed principally during the Ardersier Readvance. **e** Subglacial meltwater channel near Dalcross, looking NNW. **f** Transverse till ridge near Mosstowie, looking E. (Images: **b**, **c** NEXTMap™ Britain elevation data © Intermap Technologies. Aerial photography © UKP/Getmapping Licence No. UKP2006/0.1; **d** BGS registered photo P002879; **e** BGS registered photo P008543; **f** Clive Auton)

estuarine features, now buried, developed at the head of the Beauly Firth (Firth and Haggart 1989). Subsequently, RSL rose and extensive estuarine mudflats were deposited against the abandoned cliffline. RSL peaked locally during the Middle Holocene, resulting in an extensive raised shoreline that tilts gently north-eastwards, backed by the Main Holocene Cliffline at an altitude of 9 m OD at Inverness (Smith et al. 2019; Fig. 15.3a).

Late Devensian raised beaches at up to ~29 m OD occur around Munlochy Bay, whereas others pass laterally at similar elevations into extensive glacifluvial outwash fan-deltas and raised tidal flats beneath eastern Inverness (Fig. 15.3a). The elevations and juxtaposition of shorelines and raised fan-deltas indicate that a relative fall in sea level occurred from 28 to ~22 m OD whilst the outlet glacier retreated westwards through the Beauly Firth (Firth 1989).

A notable feature of the coastal lowlands, particularly between Inverness and Forres, is that raised shingle beaches commonly lie at higher elevations than adjoining large kettle holes within outwash deposits immediately inland. This indicates that the outwash was locally ice-cored and that ice remained buried during the fall in sea level. Kettle holes also occur within raised beaches, notably around the Beauly Firth (Fig. 15.1e, f), indicating that sea-level changes were occurring whilst the outlet glacier was retreating.

15.6 Glacial and Glacifluvial Landforms

15.6.1 The Beum a' Chlaidheimh Breach

The Beum a' Chlaidheimh ('cleft of the sword') at ~360 m OD is the lowest of several breaches in the Dulnain/Spey-Findhorn divide where large crag-and-tail features provide evidence for strong northward flow of ice during the local LGM (Hughes et al. 2010, 2014; Merritt et al. 2017a; Fig. 15.2c). The breaches lead towards a relatively featureless plain underlain by till around Lochindorb. During deglaciation, meltwaters flowed northwards through the Beum a' Chlaidheimh and higher breaches to the west, cutting deep, winding, lee-side drainage channels and spillways (Fig. 15.2d). These features start at the watershed and descend with interlocking spurs, suggesting that they were functioning when ground to the north of the watershed was largely ice free. The main channel links with an esker on the southern side of the breach amidst a suite of ice-marginal channels, benches and block-strewn lateral moraines that arc around the mountain rim, formed at the northern margin of a huge outlet glacier that occupied Strath Spey (Young 1977; British Geological Survey 2013; Merritt et al. 2017a).

15.6.2 The Middle Findhorn Valley

The middle Findhorn, downstream of Tomatin, contains a particularly fine assemblage of glacial features and deposits that can be viewed from Daless (Fletcher et al. 1996; Merritt et al. 2017a; Fig. 15.2e, f). Here, the NE–SW trend of the middle Findhorn is influenced by the NNE–SSW alignment of a pair of faults that form a graben within which Middle Old Red Sandstone strata occur. The higher of two splendid glacial spillways cutting Creag a' Chròcain follows a zone of shattering along the northern of the graben-margin faults (Fletcher et al. 1996).

Topographical benches may be observed at up to twelve levels from Daless. Glacifluvial terrace fragments and benches rise above the floodplain and river terraces. Sections in a bench at 285 m OD on the eastern side of the valley reveal that the feature comprises gravel fining downwards into interlaminated sandy silt and clay with dropstone cobbles, indicating a glacilacustrine origin. The succeeding benches at between 287 and 310 m OD are underlain by gravel, with one including a kettle hole 5 m deep. The sequence terminates with a small outwash fan at 365 m OD. The Kincraig Terrace lies at the foot of the Creag a' Chrocain spillways and is underlain by about 25 m stratified sand and gravel. Many of the benches are probably kame terraces laid down in contact with downwasting ice within the valley, whereas the large sloping terraced feature at 300 m OD south of Quillichan is a glacifluvial fan.

A river cliff section 4.5 km downstream, at Banchor, displays particularly good evidence of glacilacustrine sedimentation (laminated sand and silt with dropstones). This supports traditional views that an ice-dam formed by Moray Firth ice blocked drainage lower in the Findhorn valley, near Dulsie Bridge (Fig. 15.2a), leading to the ponding of Glacial Lake Findhorn (Merritt et al. 2017a).

15.6.3 Piperhill and Lower Strath Nairn

Flights of ice-marginal features extend along the south-eastern flank of Strath Nairn, downstream from Daviot (Fletcher et al. 1996; British Geological Survey 1997; Fig. 15.2a). The anastomosing channels and terraces of this suite formed when the middle Findhorn valley was choked with ice, causing meltwater from the upper Findhorn valley to flow through Strathdearn and into Strath Nairn near Daviot. The features are well displayed around Piperhill (Fig. 15.3a, c), where glacifluvial terraces and moraines either form contiguous staircases, or are bound on their uphill sides by arcuate, steep-sided, glacial meltwater channels up to 15 m deep (Merritt et al. 2017a). Some terraces pass upstream into channels eroded into till and separated by lateral moraine ridges with steep scarps (ice-contact slopes) facing the main valley. An age of 14.5 ± 1.4 ka BP has been determined from the mean of four cosmogenic ^{10}Be exposure ages derived from large glacially transported boulders embedded in the moraine ridges. These hitherto unpublished ages were calculated using CRONUScalc v.2.0 (Marrero et al. 2016).

15.6.4 The Flemington Eskers

The Flemington Eskers form one of the best-preserved braided esker systems in Britain (Merritt et al. 1995, 2017a; Fig. 15.3b). The eskers are mainly aligned with the final north-eastward direction of ice flow (flowset 6) and were created whilst the ice front retreated south-westwards towards the Great Glen. The main system includes at least eight braided ridges, 5–10 m high, with intervening kettle holes infilled by peat. The braided forms occur in three distinct groups, linked together by a single discontinuous ridge (Fig. 15.3b). At the south-western end of the system, a single ridge curves around into the bottom of a deep subglacial drainage channel at Dalcross (Fig. 15.3a). Debate continues as to whether the eskers were deposited within subglacial or englacial conduits within the ice, or within open channels cut into the ice and ice-cored outwash at the front of the receding ice margin (Auton 1992; Merritt et al. 2017a).

The main esker system terminates north-eastwards in a broad, flat-topped and steep-sided ridge near Meikle Kildrummie (Fig. 15.3b). This feature slopes gently towards the east from ~38 m OD and is pitted by small kettle holes. It has been interpreted either as a glacimarine delta laid down at the mouth of a subglacial conduit at the retreating ice front, or as a glacifluvial deposit formed within a large open crevasse, or cleft, at the ice margin, unmodified by marine processes (Merritt et al. 1995). The ridge was initially contiguous with the esker system but has been separated from it by subsequent glacifluvial erosion.

The Flemington Eskers form part of an extensive belt of glacifluvial deposits, including eskers, kames, plateaux and terraces, that lie between Daviot (Fig. 15.2a), in the upper Nairn valley, and Elgin (Merritt et al. 2017a). The features formed sequentially, probably along a suture or shear margin that divided stagnating Monadhliath ice within Strath Nairn from more active ice that issued from the Great Glen during deglaciation. A splendid assemblage of glacifluvial features straddles the valley of the River Ness around Torvean (Fig. 15.2a), on the western fringe of Inverness, where ice retreating back into the Great Glen formed a 68 m high esker, one of the tallest in Great Britain (Gordon 1993).

15.6.5 The Drummossie Muir-Culloden Ridge

A broad, elongate, gently convex and drumlinoid ridge lies between Strath Nairn and Inverness (Fig. 15.2a). The feature was sculptured beneath ice that flowed north-eastwards (flowset 6) from the Loch Ness Basin (Hughes et al. 2010, 2014). The ridge extends past the Culloden Battlefield visitor centre towards Croy (Fig. 15.3a); it has been modified by glacial meltwaters that carved channels of contrasting origin and is surmounted by two distinct suites of low ridges (Fletcher et al. 1996).

Several 15 m deep channels occur around Dalcross and Croy (British Geological Survey 1997; Fig. 15.3a, e). The features were probably incised subglacially as they start and terminate abruptly and are generally aligned with the former north-eastward direction of ice flow. Shallower channels arc around towards Strath Nairn, probably cut when the ice had thinned and the routing of subglacial meltwater was controlled more by the underlying topography being exhumed during deglaciation. A third set of channels occurs only along the north-west-facing slope of the ridge, and cross-cutting relationships indicate that they were cut at progressively lower elevations. They formed at, or close to, the margin of the outlet glacier that retreated into the Great Glen, but some locally have humped longitudinal profiles indicating a subglacial origin. Several flights of 'in–out' channels (Benn and Evans 2010) south-east of Inverness (Fig. 15.3a) were probably cut by meltwater that mostly flowed in conduits within the glacier.

A suite of ridges orientated parallel to the axis of the Drummossie-Culloden Ridge are ubiquitous at the south-western end of the feature, where the land surface has a corrugated form with 2–4 m deep furrows (possibly megagrooves) spaced 10–20 m apart (British Geological Survey 1997). The intervening ridges are formed of sandstone and siltstone rubble derived from the underlying Lower Devonian bedrock. This sediment-landform assemblage was interpreted as fluted moraine by Fletcher et al. (1996), but some of the furrows curve and locally bifurcate suggesting that they were eroded by subglacial meltwater. A contrasting suite of transverse till ridges lies across the distal end of the ridge, where they are typically 2–5 m high, spaced 75–150 m apart, locally coalesce, are generally lobate in plan and have intervening re-entrants (Fletcher et al. 1996; British Geological Survey 1997; Fig. 15.3a). These could have formed at the retreating ice front as winter push ridges, but generally they are not asymmetrical in profile like most such features (Benn and Evans 2010).

15.6.6 Transverse Ridges Between Nairn and Elgin, and on Tarbat Ness Peninsula

Similarly, orientated transverse ridges occur on the coastal lowland as far east as Elgin (Merritt et al. 2017a, b). The features tend to become longer and higher eastwards and are well-developed around Mosstowie (Finlayson et al. 2007; Figs. 15.2a and 15.3f). They are generally 3–7 m high and 50–90 m wide, traversing topographic undulations of up to 20 m amplitude. The major ridges are till-cored, have an average spacing of 190 m and are closely associated with ice-marginal glacial drainage channels that cut across higher ground to the south. Small-scale flutings occur locally between the major ridges. More pristine transverse ridges occur on Tarbat Ness (Fig. 15.2b), beside the Dornoch Firth, where they are up to 6 m high, 40–90 m wide and spaced 80–280 m apart (Finlayson et al. 2007).

Transverse ridges may result from different glacial processes (Benn and Evans 2010). Networks of narrow, locally anastomosing transverse ridges occur in the forelands of modern surging glaciers where they form by water-saturated sediment squeezing up into basal crevasses in highly fractured ice during surging. The sediment subsequently freezes and becomes part of the glacier. However, the straightness, length, height and spacing of the ridges at both Mosstowie and Tarbat Ness are also characteristic of De Geer moraines, especially as the ridges tend to have steeper east-facing ('down-ice') slopes (Finlayson et al. 2007). De Geer moraines are created at, or near, ice margins retreating in water. The features possibly formed during the retreat of the marine-terminating MFIS, but the altitudes of both sets of the ridges (up to 55 m OD) are much higher than predicted by current glacio-isostatic adjustment sea-level modelling for the region (Smith et al. 2019).

15.6.7 Ardersier

The Ardersier peninsula is bordered by a prominent raised beach at ∼11 m OD that underlies Fort George (Fig. 15.3a, d). This shingle beach is of Middle Holocene age (Firth and Haggart 1989), but the abandoned cliffline that borders it represents a former ice-contact slope created earlier during a significant, but so far undated, late-stage glacial readvance within the Inverness Firth (Merritt et al. 1995). This so-called Ardersier Readvance formed a push moraine composed of glacitectonized glacimarine sediments (Ardersier Silts) that arcs around from Castle Stuart towards Rosemarkie, on the Black Isle (Merritt et al. 1995, 2017a; Fig. 15.3a). The Ardersier peninsula is also trimmed by fragmentary Late Devensian shorelines at altitudes of 28.5, 26.6, 21.0–21.6 and 18.5 m OD (Firth 1989).

15.7 Conclusions

A key feature of the Inverness region is that it provides a palimpsest of successive Late Devensian ice movements that are clearly recorded in the landscape. It contains a particularly rich and diverse set of sediment and landform assemblages that were mostly created at glacial margins towards the end of the last glaciation. It includes the legacy of a major oscillating tidewater outlet glacier together with Late Devensian and Holocene sea-level changes. At least five phases of glaciation have been recognised, although they remain poorly constrained temporally.

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Glen Roy and Glen Spean

16

Adrian P. Palmer

Abstract

The landscape around Glen Roy and Glen Spean is dominated by the effects of glacial processes that operated during the Quaternary. The area has been widely studied over the last two hundred years. It was one of the first locations in Britain where the role of glaciation was recognised and remains an important focus of research today. Glen Roy and Glen Spean lie adjacent to the main ice accumulation areas in the Western Grampian Highlands, and geomorphic evidence for processes that operated during the Loch Lomond Stade (~12.9–11.7 ka) is especially well preserved. Particular highlights are the shorelines ('Parallel Roads') and associated landforms of the famous glacilacustrine systems that existed for 518 years between 12,135 and 11,618 cal a BP. The remarkable assemblage of geomorphic features includes terminal and lateral moraine complexes, kames, eskers, kame terraces, kettle holes, shorelines of glacial lakes, subaerial and subaqueous fans, deltas and a cluster of rock-slope failures that were activated before, during and after the glacilacustrine systems. After the demise of the ice-dammed lakes there is evidence for the evolution to fluvial drainage systems of the present day. The geomorphological importance of the area is highlighted by its recognition as a flagship locality within Lochaber Geopark.

Keywords

Ice-dammed lakes • Lake shorelines • Lake deltas • Jökulhlaup • Varve chronology • Alluvial fans • Rock-slope failures • Palaeoseismicity

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16.1 Introduction

Glen Roy and Glen Spean, in Lochaber, provide geomorphological evidence of international significance for Late Quaternary glaciation in the British Isles. This global recognition primarily stems from the distinctive glacial lake shorelines preserved on the hillsides of both valleys and neighbouring Glen Gloy. These shorelines were formed by a glacilacustrine system that developed between ~12.1 ka and ~11.6 ka during the Loch Lomond Stade (equivalent to the Younger Dryas of ~12.9–11.7 ka). In Glen Roy, the shorelines are referred to as 'Parallel Roads' and represent former lake levels at altitudes of 260 m, 325 m and 350 m OD (Ordnance Datum; Fig. 16.1), whereas only fragments of a single shoreline at 260 m are present in Glen Spean and only a single shoreline at 355 m occurs in Glen Gloy. The Parallel Roads represent the focal point of interest, but the shorelines are linked to fine examples of glacial, fluvial and mass-movement landforms that provide important evidence for the Lateglacial evolution of the landscape in the three valleys. These include ice contact subaqueous fans, subaqueous fans, deltas, moraines, kame and esker systems, fluvial terraces and rock slope failures.

The landforms and landscape of Glen Roy and Glen Spean have stimulated debate from the time of the emergence of modern geological principles in the nineteenth century to the present day. After 1840, when the true significance of the lake shorelines was first appreciated, the area was fundamental in establishing the role of glaciers in shaping the Scottish Highlands, and research on its landforms and sediments over the past 50 years has proved critical in determining the extent and chronology of the Loch Lomond Readvance (Younger Dryas) icefield in the Western Highlands. Moreover, Glen Roy and Glen Spean remain at the forefront of research in the UK, through the analysis of annually resolved lake sediment (varve) records to understand glacier dynamics at decadal scales using high-precision chronologies. The discovery within the varve

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Fig. 16.1 **a** ‘Parallel Roads’ of Glen Roy, looking north from the Viewpoint with the ‘Massive Drift Accumulation’ in the foreground and showing the three shorelines at 260 m, 325 m and 350 m OD. **b** Detail of the three distinctive shoreline benches eroded into the valley side. **c** Upper Glen Roy, with the 350 m and 325 m shorelines in the foreground and the Burn of Agie fan surface in the background. (Images: **a**, **b** Adrian Palmer; **c** Colin Ballantyne)

sediments of the Vedde Ash, a key Icelandic cryptotephra horizon of Loch Lomond Stadial age, highlights the area's continuing significant role in refining our understanding of Lateglacial environmental and climatic change.

16.2 Topography

Glen Roy and Glen Spean are located in the Western Grampian Highlands, ~20 km northeast of Fort William and ~15 km east of the Great Glen (Fig. 16.2). The area is dominated by mountainous terrain rising to over 900 m OD, which includes some of the highest peaks in the Scottish Highlands. The main valley is the Spean, which trends east–west. Drainage through the Spean valley is fed by the River Pattack, which rises in the east and enters Loch Laggan, which is drained westwards by the River Spean, which in turn flows into the River Lochy in the Great Glen. The River Roy (~25 km long) is a north-bank tributary of the Spean that flows initially westwards from the Roy–Spey watershed at 350 m, then southwards to join the River Spean at Roybridge (Fig. 16.2a).

A number of tributary valleys to Glen Roy and Glen Spean supplied high sediment loads during the Lateglacial period and are thus important to our understanding on the evolution of the landscape (Fig. 16.2). The Burn of Agie enters the River Roy in upper Glen Roy as a south-bank tributary, and Glen Turret is a major valley entering upper Glen Roy from the northwest; the valley of the Allt a' Chòmhlain, a tributary of the River Turret, connects upper Glen Roy with the head of Glen Gloy via a col at 355 m (Fig. 16.3a). A series of small streams through upper and middle Glen Roy have steep catchments (those of the Canal Burn, West and East Allt Dearg, Allt Na Reinich, Allt Brunachain and Allt Bhreac Achaidh; Fig. 16.3a, b). To the west of lower Glen Roy, beyond Bohuntine Hill, lies the valley of Caol Lairig; to the east of lower Glen Roy, Gleann Glas Dhoire connects to the Spean valley via a col at 325 m. The lower Spean valley is flanked by the Nevis range to the south, forms a wide valley west of Roybridge but narrows eastwards. Two deep and narrow valleys enter Glen Spean from the south (the Laire and Treig valleys) and the River Ossian flows into Loch Laggan (Fig. 16.2a).

The cols between and at the heads of valleys played an important role in the evolution of the Lateglacial lake sequence in Glen Roy and Glen Spean (Fig. 16.2b). At the eastern end of the Spean valley a long, low-lying area at 260 m between the River Pattack (Spean catchment) and River Mashie (Spey catchment) forms the lowest col between the Spean and Spey catchments. The 325 m col at the head of Gleann Glas Dhoire connects the Roy and Spean catchments via a valley that joins the Spean valley at Roughburn. On the western side of middle Glen Roy a col at 300 m connects Caol Lairig and Glen Roy, and a col at 350 m at the head of upper Glen Roy connects the Roy catchment with the Spey catchment. Finally, as noted above, Glen Gloy and Glen Turret are connected by a col at 355 m.

16.3 Geology

The bedrock geology of the area is dominated by Dalradian psammites, pelites and mica-schists that strike towards the northeast (Key et al. 1997). The area on the western flanks of Glen Roy and into Glen Turret is composed of psammites, semipelites and quartzites of the Grampian Group, but lower Glen Roy, the western flanks of middle Glen Roy and upper Glen Roy are underlain by pelites and calc-silicate rocks of the Leven Schist Formation (Appin Group). The head of Glen Roy is underlain by granodiorite of the Corrieyairick Pluton. Middle Glen Spean is underlain by psammites and feldspathic quartzites of the Grampian Group and, farther east, by the Strath Ossian granodiorite intrusion. Surficial deposits from the last glaciation are present across much of the landscape and are recorded as sandy tills or gravelly diamictons that mantle the lower slopes of the area. Large faults are also recorded within both Glen Roy and Glen Spean.

16.4 Historical Context and the Formation and Drainage of the Glacial Lakes

Glen Roy has been a focus of geological research since the emergence of modern geological and glaciological thinking in the nineteenth century. The enigmatic shorelines on the valley sides and their altitudinal relationship to the overspill cols led to the suggestion that they represented former lake systems, although the lack of obvious impounding barriers or dams caused some to doubt this theory. For example, Darwin (1839) argued that the features were marine shorelines of an inland sea, comparable to features he had observed in southern South America. Agassiz (1840) was the first to propose ice as a competent barrier to river drainage and so enabled the lakes to form. Indeed, the landforms around the mouth of the Treig valley, in combination with the shorelines, were a crucial line of evidence used by Agassiz to suggest glacial ice was prominent in the development of the landscape. Jamieson (1863, 1892) developed this idea using Ordnance Survey data to develop a model of lake evolution that remained established for nearly a century. The different nineteenth-century interpretations of the formation of the Parallel Roads are summarised in Rudwick (2017).

A series of papers by Sissons (1978, 1979a, b, c) and Sissons and Cornish (1982a, b, 1983) presented geomorphic

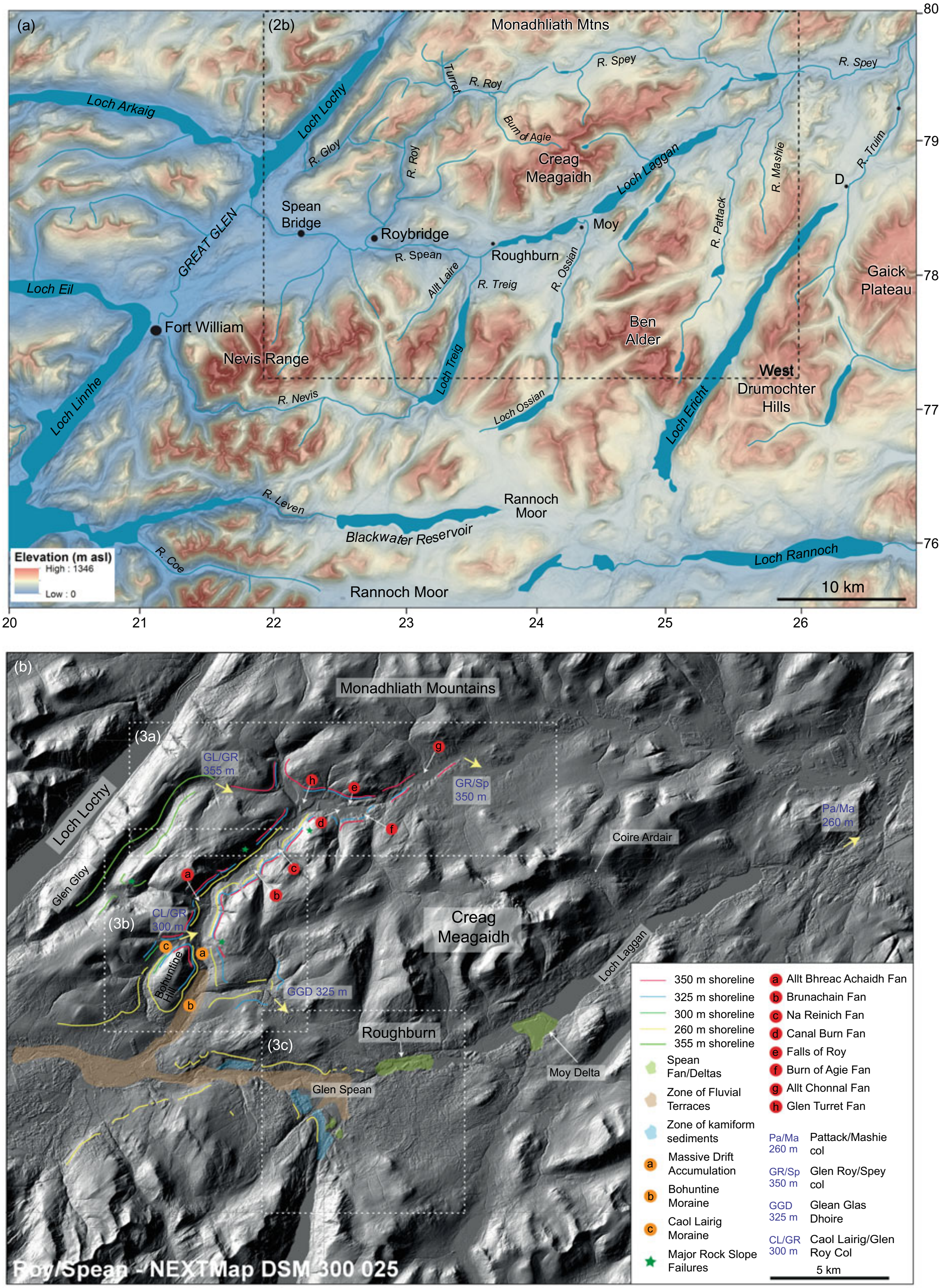

Fig. 16.2 **a** Regional topography and river systems in the area of Glen Roy and Glen Spean (Base map uses OS Terrains5 DTM data available and downloaded from Digimap.edina.ac.uk. © Crown copyright and database rights (2017) Ordnance Survey (100025252)). **b** Digital surface model of Glen Roy and Glen Spean highlighting the distribution of the shorelines and other geomorphological features discussed in the text. (DSM data supplied through Intermap Technologies Ltd)

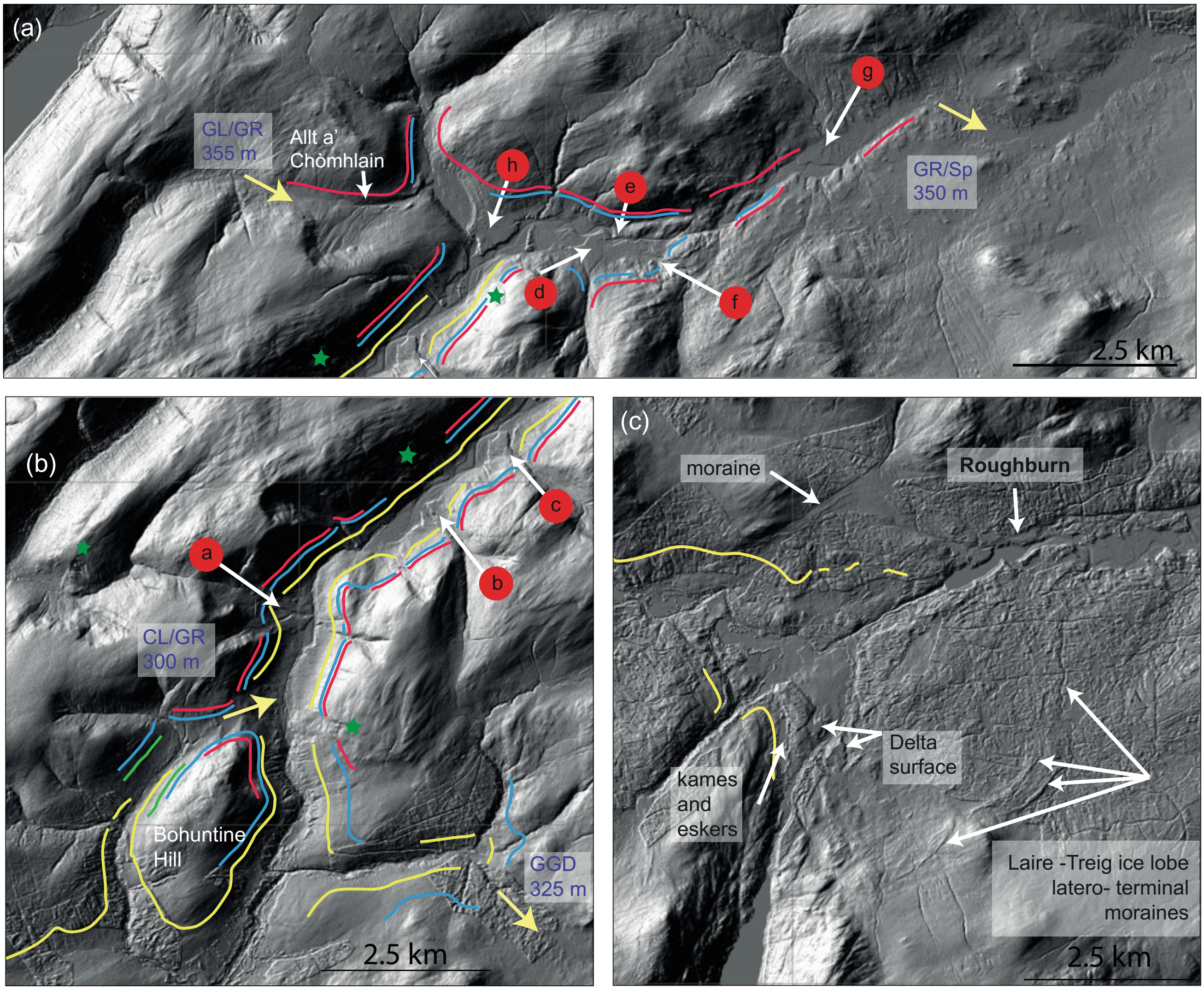

Fig. 16.3 **a** DSM of upper Glen Roy. **b** DSM of middle and lower Glen Roy in the vicinity of the viewpoint. **c** DSM of middle Glen Spean in the area of the mouth of the Treig valley and Roughburn. Key as for Fig. 16.2b. (DSM data supplied through Intermap Technologies Ltd)

mapping of the Spean and Roy catchments and the results of detailed instrumental levelling of the lake shorelines and river terraces. These data and the model for the development of the landscape that emerged from this work are still fundamental to our understanding of the glacial limits, the extent and form of the shorelines, the impact of glacio-isostatic adjustment on the shorelines and the identification of later, lower lake levels and fluvial terrace aggradation after the demise of the main lakes.

The centrepiece of this work is the paper by Sissons (1979a), which presented a reconstruction of the limits of the Loch Lomond Readvance (LLR) in the area and a model for glacilacustrine evolution in Glen Roy and Spean (Fig. 16.4). Sissons proposed that ice built up initially from the Nevis Range and the mountains farther west, coalescing in the Great Glen, and eventually damming the westwards-flowing River Spean and its Roy tributary. As a result, the lake waters rose to the height of the lowest available col, at 260 m at the eastern end of the Spean valley between the Spean and Spey catchments (Fig. 16.4a). Further ice advance saw the merging of the ice in the lower Spean with the Laire-Treig ice lobe (an outlet glacier of the West Highland Icefield), which advanced northwards into the middle Spean valley. Advance of the Spean glacier eventually severed the connection between the lakes in Glen Roy and Glen Spean, causing the lake level in Glen Roy to rise to the next lowest available outlet col, at 325 m at the head of Gleann Glas Dhoire. At this stage (Fig. 16.4b), Sissons envisaged that drainage of the 325 m lake in Glen Roy overspilled into the 260 m lake in Glen Spean, although drainage was diverted around the northern margins of the Laire-Treig ice lobe at Roughburn. He argued that thereafter the ice margin remained stationary in Glen Spean, but ice continued to advance farther into lower Glen Roy to the vicinity of the Viewpoint, as well as into Caol Lairig and Glen Gloy to the west. Eventually, this ice advance in Glen Roy prevented drainage across the Gleann Glas Dhoire col, and the lake level rose to 350 m, the altitude of the outlet at the head of Glen Roy (Fig. 16.4c). This progressive increase in lake level in Glen Roy is referred to as the 'rising' sequence.

Sissons (1979a) considered that initial ice retreat in Glen Roy resulted in a glacial lake outburst flood passing across the Gleann Glas Dhoire col into the Spean at Roughburn, causing the Glen Roy lake level to drop from 350 to 325 m (Fig. 16.4d), though he depicted the coeval margin of the

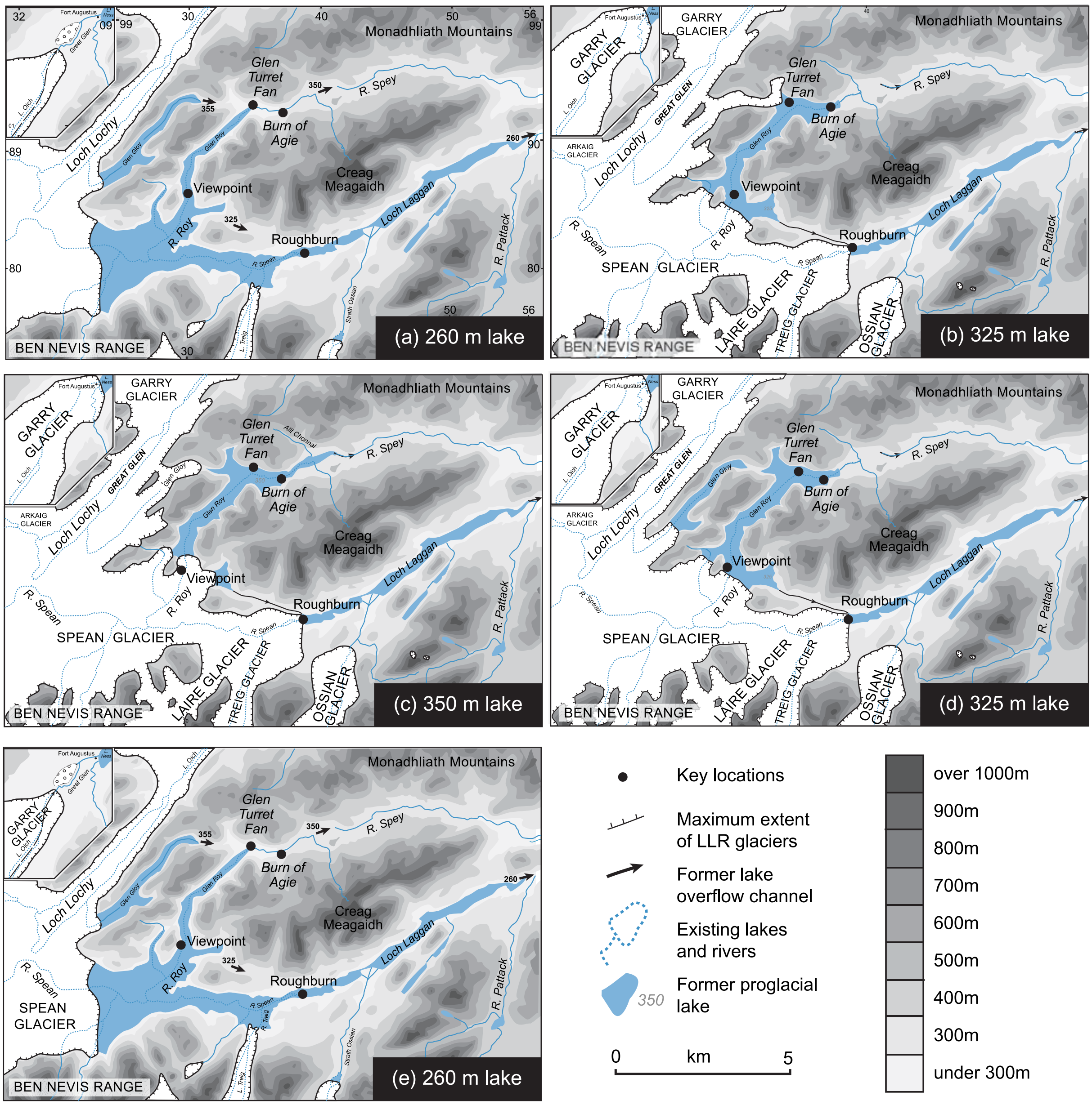

Fig. 16.4 Changing extent of LLR glaciers and evolution of the ice-dammed lakes in Glen Roy and Glen Gloy as proposed by Sissons (1979a). Stage 1 depicts ice advancing from the west and south causing the formation of the unified 260 m lake system in Glen Spean and Glen Roy. Subsequent ice advance blocked lake drainage routes, enabling the water level to rise in Glen Roy (stages 2 and 3) but to fall during ice retreat (stages 4 and 5). The extent of the glacier and timing of ice advance and retreat in Glen Gloy are debatable. Depicted here is the interpretation of Sissons and Cornish (1983), Sissons (2017) and Cornish (2017), which shows ice from Glen Gloy and the Great Glen terminating in Glen Turret (stage 2); Peacock (1986) envisaged a more restricted glacier advance in Glen Gloy at this time. More recently, the configuration of ice prior to, and during, ice advance in lower Glen Spean has also been questioned by Palmer et al. (2020). (Creative Commons CC-BY licence)

Laire-Treig lobe as remaining close to its maximum position at Roughburn. Subsequent ice retreat, he argued, resulted in severance of the Glen Roy lobe and Spean glacier from the Laire-Treig lobe, with continued retreat of the former eventually permitting lowering of the Glen Roy lake level to 260 m (Fig. 16.4e). This sequence of events is referred to as the 'falling' sequence. Sissons (1979b) argued that final drainage of the lake system was effected via catastrophic subglacial drainage of ~ 5 km^3 of water northwards along the Great Glen, with associated jökulhlaup sediments being deposited both as a massive terrace of coarse sediment at Fort Augustus (Russell and Marren 1998) and an extensive spread of subaqueous gravels in the Beauly Firth at the northern end of the Great Glen, near Inverness. Further transient, lower-altitude, ice-contact lakes continued to be dammed by the retreating ice in Glen Spean and the Great Glen as the fluvial network became re-established (Sissons and Cornish 1983).

16.5 Glacial Geomorphology

Glacially streamlined bedrock outcrops extend upslope to the summit crests (up to an altitude of 900 m) that flank Glen Spean (Fig. 16.5a). Often these outcrops are asymmetric, with steep, smoothed stoss sides and shallower, irregular lee sides and have elongation ratios averaging 5:1 (Turner et al. 2014). Most streamlined outcrops are oriented SW–NE or WSW–ENE, parallel to the trend of the valley and indicate that these hilltops were overtopped by the last and probably earlier ice sheets. Both Glen Roy and Glen Spean have a classic glacial trough profile, and the Treig valley is one of the best examples in Scotland of a glacial breach. These features reflect glacial erosion over multiple glacial cycles. The bedrock in the valley floors and lower flanks is also glacially smoothed and locally striated, although striae are rarely observed beyond the limits of the LLR.

Excellent examples of moraines are present within Glen Roy and Glen Spean. The most spectacular of these is the LLR latero-terminal moraine complex deposited by the Laire-Treig outlet glacier (Figs. 16.3c and 16.5a, b). The limit of this glacier in middle Glen Spean is marked by an almost continuous moraine, 15 km long, that descends eastwards from 500 m at the northern outlet of the Treig gap to 400 m along the southern slopes of the glen, then loops northwards across Glen Spean in the form of three massive ridges up to 20 m high. The moraine then descends to a fan-delta complex at 250 m on the valley floor at Roughburn and then rises westwards to 400 m. Shorter recessional moraine fragments within the outer moraine ridge suggest active ice retreat. Further moraines occur in Glen Spean to the west of the Laire valley at around 400 m.

A distinct, 5 m high terminal moraine in Caol Lairig occurs close to the col (Fig. 16.5c) and has a crest that runs transverse to the valley axis and then rises up the valley sides to merge with a delta associated with the 350 m shoreline on the northern valley side. This feature has been interpreted as a subaqueous moraine ridge that formed as the LLR glacier advancing northeastwards up Caol Lairig reached its maximum extent when the lake level was at 350 m (Tye and Palmer 2017).

In lower Glen Roy, two former ice limits are defined by morainic debris although they have distinctly different geomorphological expressions. The first is at Bohuntine, where a ~ 2–3 m high ridge rises from 150 to 250 m on either side of the River Roy, close to the most southerly extent of the 325 m shoreline (Fig. 16.5d). The second, about 1 km north of this point, is a so-called Massive Drift Accumulation; this extends upvalley for 1.5 km and is composed of superficial deposits up to 80 m thick containing glacial diamictons and sandy gravels that are concentrated below the 260 m contour (Fig. 16.1a). This accumulation of material is now extensively gullied by valley-side streams and therefore appears to have ridge-like forms transverse to the valley axis. It is not clear whether these gully features were initiated during glacial lake drainage or as a result of paraglacial incision after the demise of the lake systems. Sissons (1978) observed that a poorly defined ridge at the outer limit of the deposit extends up the valley sides to around 370 m on the eastern flanks of the glen, close to the entrance to Gleann Glas Dhoire.

As well as the large regional ice lobes described in Sect. 16.4, local ice masses have also been identified in the vicinity. Finlayson (2006) identified evidence for an icefield of probable LLR age over the Creag Meagaidh massif, including hummocky moraine in the lower reaches of the Burn of Agie valley to the south of upper Glen Roy. Boston et al. (2015) mapped LLR ice limits across the Monadhliath Mountains, including those on the northern margins of Glen Roy. Their reconstruction suggests that ice accumulated on the high plateau areas of the Monadhliath and flowed into the heads of some valleys, notably Glen Chonnal at the head of Glen Roy and the high-level tributary valleys feeding Glen Turret, but terminated short of the 350 m shoreline. The presence of an ice-contact slope and moraines at the proximal end of the Glen Turret fan in upper Glen Roy (Fig. 16.5e) adds a complication to our understanding of the landsystem since it indicates that either ice formerly extended to a much lower altitude during the LLR than in other areas of upper Glen Roy or that these landforms represent evidence for an earlier ice margin at the time of ice sheet retreat (Boston and Lukas 2017). The age of these features is considered in further detail in Sect. 16.6.

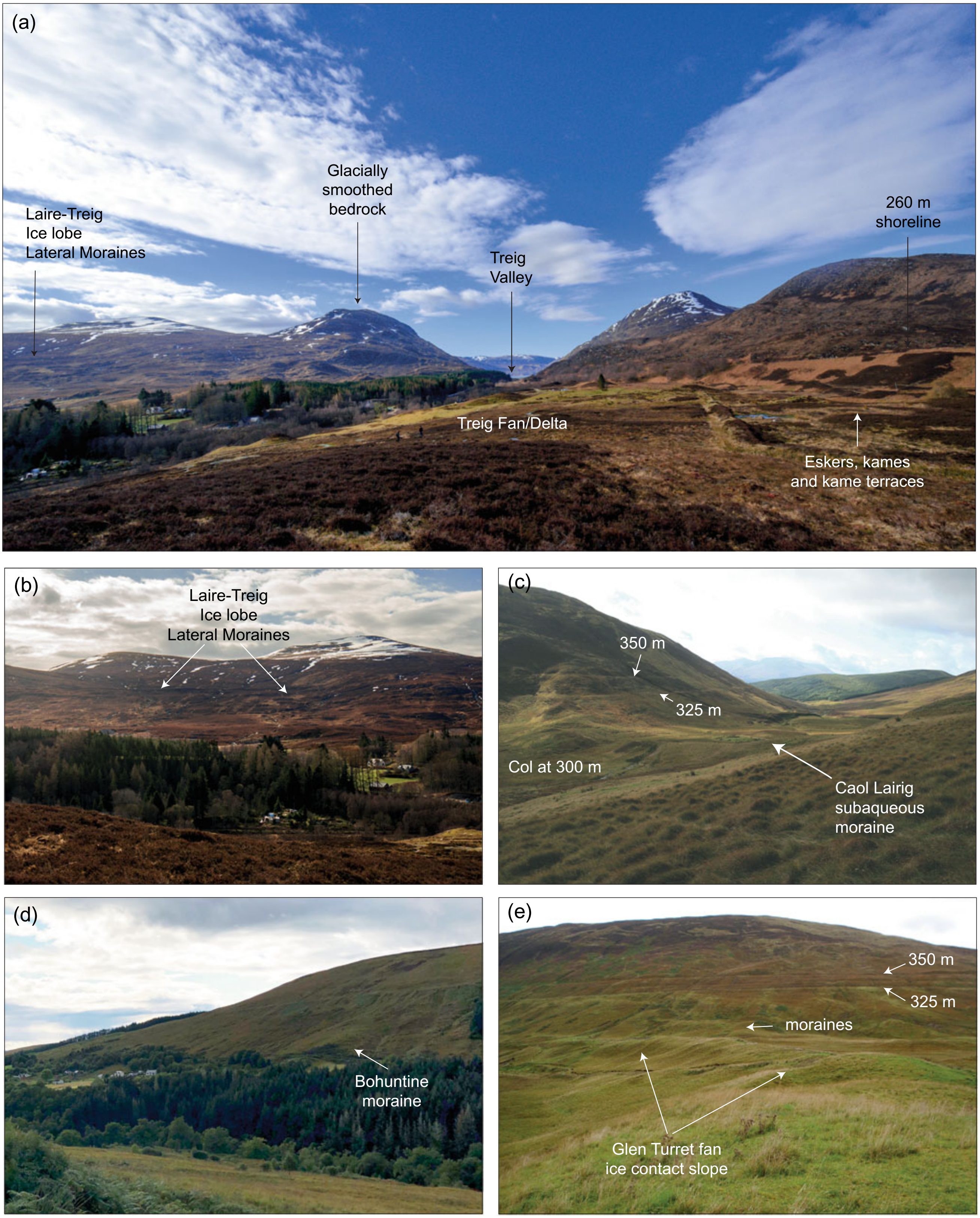

Fig. 16.5 Landforms at the mouth of the Treig valley, looking south towards the Treig glacial breach. **b** View southeast to the lateral moraines on the southern side of Glen Spean; these loops across the valley floor as the end moraine of Laire-Treig ice lobe. **c** Subaqueous moraine ridge in Caol Lairig looking southwest. **d** Bohuntine moraine in lower Glen Roy looking southeast. **e** Ice contact slope on the Glen Turret fan in upper Glen Roy looking north. (Images: **a**, **b**, **e** Christopher Francis; **c**, **d** Adrian Palmer)

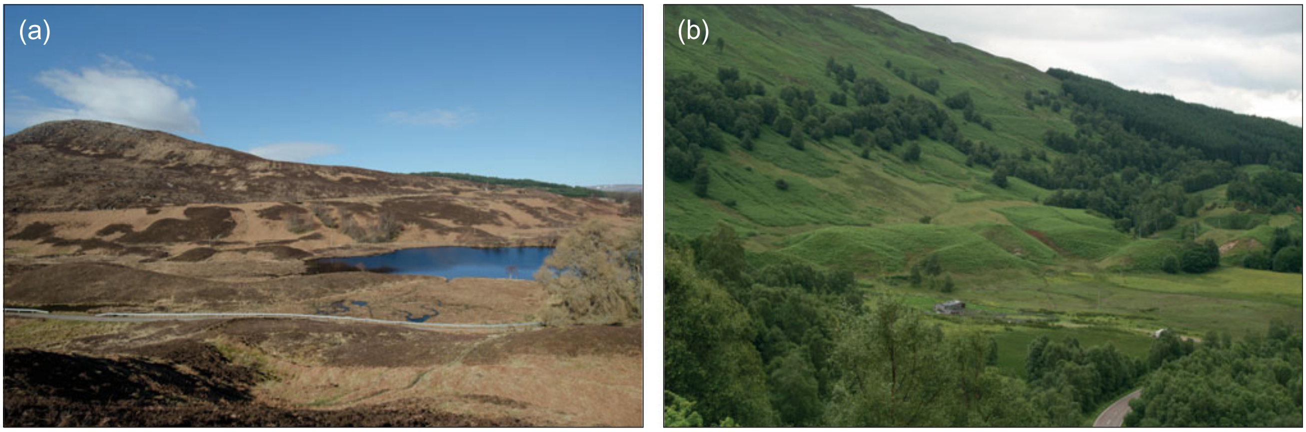

Fig. 16.6 **a** Kame and esker landscape assemblage on the western margins of the mouth of the Treig valley. A series of kame terraces occupies the foreground, and kettle depressions are either completely infilled with sediment or remain as lakes. In the centre is the end of an esker; the 260 m shoreline is cut halfway up the opposite hillside. **b** Gullied kames and kame terrace complex on the northern side of the Spean valley east of Roybridge. (Images: **a** Christopher Francis; **b** Adrian Palmer)

Kame topography is well preserved in a zone on the south side of Glen Spean and into the mouth of the Treig valley (Sissons 1978; Figs. 16.3c and 16.6a). This area contains the remnants of kettle holes that are either completely open, infilled with ~10 m of sediments (Kelly et al. 2017) or impound small, shallow lakes. Closer to the River Treig, there are more distinct sinuous, irregular ridges with crest-lines up to 10 m high parallel to the valley axis, which are likely to be eskers. A series of kame terraces that formed against a bedrock step in the centre of the valley suggest that ice downwasted during the final stages of ice retreat and that dead ice was prevalent to form the kettle holes on this side of the valley. In the areas of lower relief, continuous, sinuous, sharp-crested ridges that run parallel to the Treig valley axis on the western side of the valley are interpreted as eskers (Sissons 1979b). Gullied kame terraces and kames up to 10 m high are also present on the north side of Glen Spean to the east of Roybridge (Figs. 16.2b and 16.6b). Gullying of the terraces is probably related to the drainage of the 260 m lake system and suggests that the 260 m lake may have submerged parts of the glacier margin as it retreated.

16.6 Lake Shorelines

The shorelines were originally surveyed by the Ordnance Survey in the nineteenth century, and the resulting maps formed the basis for early reconstructions of the lake sequence. Further mapping of the features by Sissons (1978) revised the distribution of the shoreline fragments (Fig. 16.2b). The 260 m shoreline occurs throughout lower and middle Glen Roy and extends to the vicinity of the Glen Turret fan, but is absent in the lower part of the Turret valley. It is also preserved in the lower (southern) part of Caol Lairig and on the northern flank of lower and middle Glen Spean (Fig. 16.3). The 325 m shoreline extends northwards from Bohuntine in lower Glen Roy, around Bohuntine Hill into Caol Lairig and then into upper Glen Roy to the vicinity of the Burn of Agie in the east and into Glen Turret to the west. Crucially, the 325 m shoreline also extends into Glean Glas Dhoire where the 325 m col is present. The 350 m shoreline occurs in the vicinity of the 'Viewpoint' in lower Glen Roy and extends into the upper parts of Caol Lairig and through middle and upper Glen Roy and parts of Glen Turret and up to the 350 m col on the watershed to the Spey. Very short fragments of other shorelines are preserved at 334 m on the northern sides of the valley overlooking the Glen Turret fan and at 297 m close to the valley floor in Caol Lairig. A single shoreline at 355 m extends up both sides of Glen Gloy, terminating at the 355 m col at the head of the glen.

The distribution of the shorelines can be linked to landforms delimiting the former ice margin positions. The southern limit of the 325 m shoreline in lower Glen Roy coincides with the distinctive, subaqueous terminal moraine preserved on the valley floor and flanks at Bohuntine (Figs. 16.3c and 16.5d). The 350 m shoreline, however, extends southwards of the 'massive drift accumulation' that marks the LLR limit in Glen Roy, suggesting that ice retreat was initiated during the existence of the 350 m lake, enabling shorelines to be cut south of the ice limit prior to lake drainage to the 325 m level. Shorelines are also present within the limits of other former ice margins, such as the 260 m shoreline present in the lower and middle Spean and the 325 m and 350 m shorelines within the area of lower Glen Turret (Fig. 16.2b).

Sissons (1978) carried out detailed instrumental levelling of all the three main shorelines in Glen Roy and recorded their width, aspect and implied volume of material removed at regular intervals along his survey transects. There are several key points from his analysis: (i) the shorelines are mainly cut in bedrock, although fragments of rock have often accumulated on the front edge of the notch; (ii) the widest shorelines occur close to tributary streams where material from the streams accumulated as a beach on the front edge of the shoreline, particularly on the upvalley side; (iii) shoreline width is negatively related to hillslope gradient and (iv) the volume of material removed from the shorelines correlates with the length of the fetch from the southwest. He proposed that the shorelines formed through a combination of wave action, frost shattering of bedrock associated with slight fluctuations in lake level and removal of coarse debris on lake ice.

The rate of shoreline formation was also considered by Sissons (1978) using measurements of the shoreline widths inside and outside the associated LLR glacier limits. Average widths of 11.3 m, 10.6 m and 10.1 m were found outside the glacier limits for the 350 m, 325 m and 260 m shorelines, respectively, and 9.4 m, 9.4 m and 11.7 m for the shorelines inside the limits. On the assumption that glacial erosion would have removed or modified shorelines during periods of ice advance, he inferred that the shorelines initially formed rapidly to enable similar widths to be achieved during different durations of lake existence during ice advance and retreat.

16.7 Deltas and Fans

16.7.1 Deltas and Fan–Deltas in Glen Spean

Both Glen Roy and Glen Spean contain fine examples of deltas and fans associated with the glacial lake systems. In Glen Spean, the Moy Delta is a large (1.52 km^2) body of sediment that accumulated in the 260 m lake, fed by the northwards-flowing meltwaters draining the Ossian glacier terminus (Fig. 16.2b). The apex of the delta lies at 265 m with the toe at an altitude of 259 m. The feature is 8–10 m high and composed of sand and gravel facies, although exposures of the sediment architecture are poor and have not been examined in detail.

Also within Glen Spean, two ice-contact fan–deltas at Roughburn and the mouth of the Treig valley occur with surface altitudes of $\sim$260 m (Fig. 16.3c). The Roughburn fan–delta has been fragmented by erosion during the final drainage of the 260 m lake. It had an estimated extent of 1.94 km^2 and is mainly composed of sands and gravels that spread along approximately 1.5 km of the narrow western end of the present Loch Laggan. These sands and gravels display both horizontal bedding that probably represents bottomsets and topsets, and steep foreset bedding dipping towards the east and southeast, evident in exposures on the south side of the valley. On the eastern margins of the landform and distal delta bottomset varves overlie thick massive sand beds (Palmer et al. 2020). The Roughburn fan–delta is located close to three terminal moraines of the Laire-Treig ice lobe that cross Glen Spean at this point. On the northern side of Glen Spean, the surface of the fan–delta is incised by an 8-m-deep channel containing many large boulders, and, farther to the east, large boulders are distributed over an area of 800 m^2 across the fan surface; some are perched on braid bars and show imbrication indicating flow from the northwest. These bouldery deposits were deposited by glacial outburst floods that occurred when retreat of the ice margin in Glen Roy permitted drainage of the 350 m lake across the 325 m col at the head of Gleann Glas Dhoire, reducing the lake level to 325 m. These boulders comprise material washed down the valley and reworked morainic debris deposited at the margins of the Laire-Treig ice lobe. The outburst flood was directed around the northern margins of the Laire-Treig ice lobe and into the Spean glacial lake. This complex feature may reflect ice contact, subaqueous fan deposition in combination with a glacilacustrine delta complex as the ice margin actively retreated.

The Treig delta is a complex landform at the mouth of the Treig valley close to its confluence with the Spean (Figs. 16.3c and 16.5a). It is a gently sloping landform with a surface altitude of 260 m in the central and eastern parts of the valley, and is separated from the kamiform and esker deposits to the west by a bedrock step. The extant delta surface has an area of 0.9 km^2 and slopes down gradually northwards, but has been eroded to the south and east by the River Treig and originally extended farther eastwards. Sediment exposures are limited, although horizontally bedded and massive gravels occur towards the top of this feature. The size and position of this delta at the mouth of the Treig valley and its association with nearby stagnant ice features suggest that it was deposited by proglacial outwash rivers at the margins of the 260 m lake during the initial retreat of the Laire-Treig glacier (Peacock and Cornish 1989).

16.7.2 The Gravel Fans of Glen Roy

Glen Roy contains some of the most spectacular gravel fans in Scotland, rivalled only by Lateglacial alluvial and outwash fans in the Spey valley catchment (Chap. 19). All are relict landforms, perched above the level of the River Roy floodplain, truncated at their distal ends and deeply incised and terraced by their parent streams. The age and origin of these fans have proved contentious (Sissons and Cornish 1983; Peacock 1986; Boston and Lukas 2017; Lowe et al. 2017; Palmer and Lowe 2017), and some aspects of their formation remain controversial. Cornish (2017) has provided a detailed account of fan morphology and lithostratigraphy and the debates concerning their interpretation.

The most extensive fan is the Allt Chonnal fan, deposited by the river of that name at the head of Glen Roy (Fig. 16.3a). This remarkable landform consists of distinctive upper and lower fan fragments separated by a steep bluff. Although now extensively terraced, the surviving fragments of this fan indicate that it had impressive dimensions: collectively the upper and lower fans occupied an area of at least 0.8 km^2, and Cornish (2017) estimated that together they originally contained $\sim 6 \times 10^6$ m^3 of sediment, mainly coarse gravels. The upper fan has its apex at 360 m, and most of its surface lies at the altitude of the highest shoreline (350 m); the lower fan apex is at 335 m, and its surface lies around or just below the altitude of the middle shoreline (325 m).

Equally impressive are the remnants of three merging fans in upper Glen Roy, deposited by the north-flowing Canal Burn (Fig. 16.7b) and Burn of Agie (Fig. 16.7c) and linked by a fan deposited by the River Roy. The apices of the Canal Burn and Burn of Agie fans lie at similar altitudes (286 m and 288 m), and the former descends to 261 m, where it is truncated by a frontal bluff that may in part represent the former 260 m shoreline.

In middle Glen Roy, the two most spectacular gravel fans are those deposited on the valley floor by the Allt Brunachain and Allt na Reinich, which flow northwestwards from small, steep catchments. Both fans are deeply incised and terraced by their parent streams and terminate at steep frontal bluffs (Fig. 16.7d, e), and the Allt na Reinich fan consists of upper and lower surfaces separated by a 5 m high bluff at 259 m, suggesting a relationship with the lower lake shoreline (260 m).

The fans described above (and other tributary fans in Glen Roy) were initially interpreted by Peacock (1986) as paraglacial landforms deposited during the Lateglacial Interstade ($\sim$14.7–12.9 ka). Cornish (2017), however, demonstrated that the fans of upper Glen Roy (Canal Burn, Burn of Agie and Allt Chonnal) comprise a consistent tripartite stratigraphy consisting of a lower unit of coarse gravel or diamicton, in places up to 20 m thick, overlain in turn by lacustrine sediments, usually no more than 2 m thick, of laminated silts and clays (varves) or locally by stratified sands, then by an upper gravel unit typically 1–2 m thick. He proposed that the lower gravels represent deposition in shallow water during the rising lake sequence, followed by deposition of lacustrine sediments as lake levels rose higher, and finally subaerial deposition of gravels as lake levels fell. A feature of this interpretation is that it requires extremely rapid deposition of coarse sediment during the rising lake sequence: in the case of the Allt na Reinich lower fan, which has a catchment area of 2.23 km^2 and originally contained $\sim 2.5 \times 10^6$ m^3 of sediment, an average catchment denudation rate of at least 11 mm a^{-1} is implied (Cornish 2017).

The origins of the Glen Turret fan, located at the confluence of Glen Turret and Glen Roy, have proved singularly controversial. Unlike the fans described above, this fan terminates upvalley at an ice-contact slope, implying deposition when glacier ice occupied Glen Turret (Figs. 16.5e and 16.7a). The Glen Turret fan has an upper surface altitude of $\sim$270 m and slopes down southeastwards to 253 m, towards the River Roy. The bluffs at the toe of the fan are 25 m high. Most of the fan is composed of horizontally bedded and massive, crudely stratified gravel beds or massive diamicton, with a thin ($\sim$1.2 m) layer of varved glacilacustrine sediments exposed on the fan surface and locally up to 5 m of lacustrine sediment under the fan gravels. This stratigraphy implies that deposition of the fan both pre-dated and post-dated lacustrine deposition. A series of indistinct mounds in the northwestern sector of the fan surface appear to represent moraines; hollows and channels run across the fan surface parallel to the fan apex, and the 260 m shoreline is absent within Glen Turret (Figs. 16.2b and 16.5e; Cornish 2017).

Interpretation of the Glen Turret fan has polarised around two viewpoints. Peacock (1986) interpreted the fan as a proglacial ice contact fan deposited during retreat of the last ice sheet and subsequently inundated by rising lake levels during the LLR, a view supported by Boston et al. (2013) and Boston and Lukas (2017) on the grounds that morphostratigraphic evidence suggests that lower Glen Turret was ice-free during the LLR. Conversely, Sissons and Cornish (1983) and Cornish (2017) argued that the Glen Turret fan was deposited during the LLR by ice feeding into Glen Turret from Glen Gloy and the Great Glen. The source of the ice that occupied Glen Turret at the time of fan deposition is still open to debate, but Palmer et al. (2020) concluded that the fan accumulated during the early stages of the LLR, and that the subsequent rise of the Glen Roy lakes to 325 m and 350 m destabilised the ice margin and caused the ice to retreat, allowing inundation of the lower Turret valley by lake waters. An implication of this interpretation is that the single lake shoreline at 355 m in Glen Gloy formed after retreat of ice from this glen.

16.8 Fluvial Terraces and Later Lower Lakes

Excellent fluvial terraces are preserved in upper Glen Roy, providing a glimpse of the evolving landscape immediately after the 260 m lake drained. The terraces have been eroded by the River Roy, revealing mainly sand and gravel facies at the toes of the Canal Burn, Burn of Agie, Allt Na Reinich and Allt Brunachain fans. Sissons and Cornish (1983)

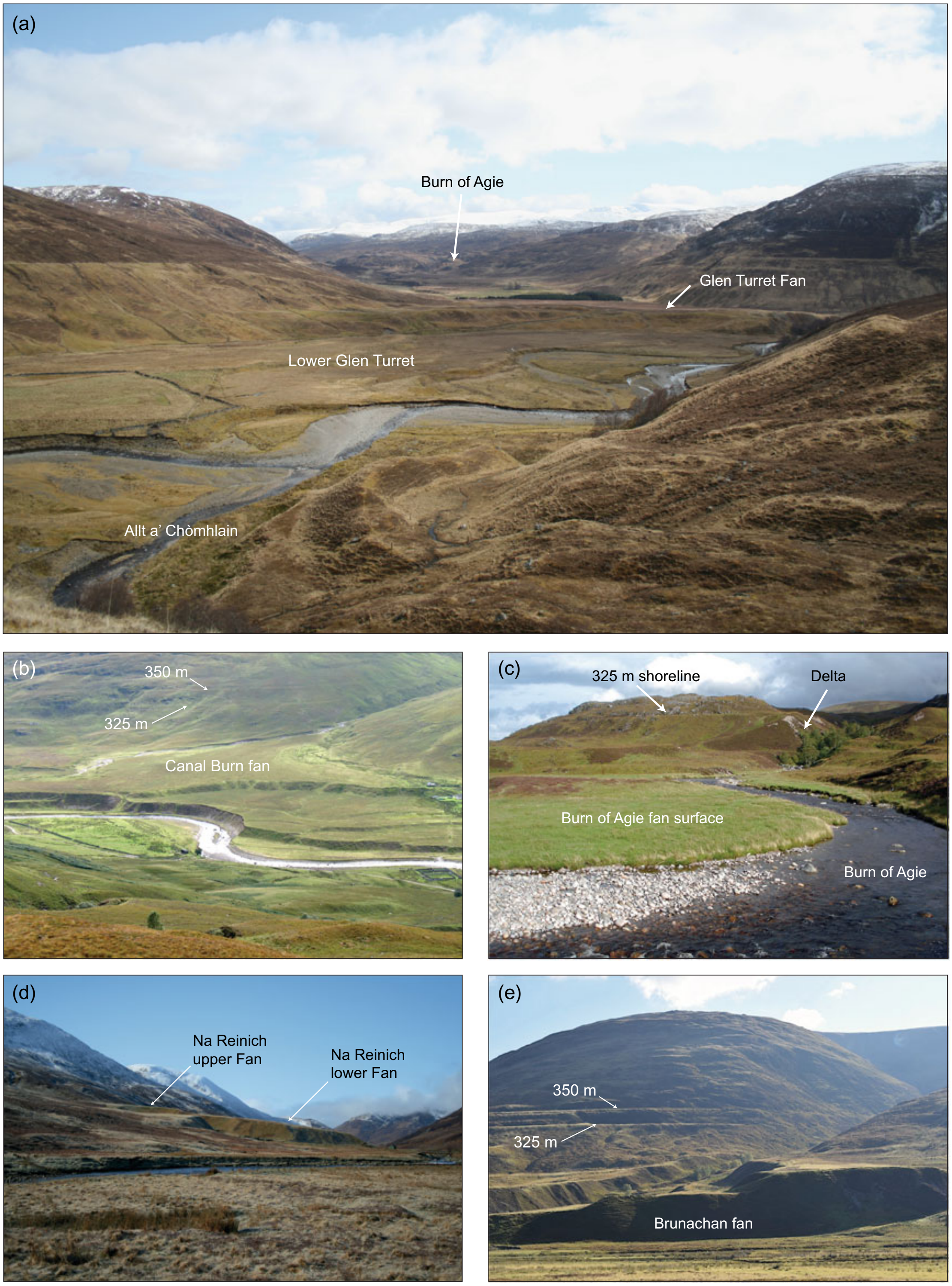

Fig. 16.7 Fan and delta landforms in Glen Roy. **a** Looking east from the Allt a' Chòmhlain towards lower Glen Turret and upper Glen Roy. The Glen Turret fan lies at the confluence of the two glens, with the proximal, ice contact slope facing northwest. **b** The Canal Burn fan with the River Roy flowing around the toe of the fan. **c** The Burn of Agie delta associated with the 325 m lake system lies above the surface of the Burn of Agie fan at 272 m altitude. **d** The Na Reinich fan in middle Glen Roy: the apex of the upper fan is at 260 m, suggesting an association with the 260 m lake. **e** The Allt Brunachain fan in middle Glen Roy, which has been dissected and terraced by its parent stream. (Images: **a**, **c**, **d** Adrian Palmer; **b**, **e** Colin Ballantyne)

projected the surface form of the surviving fan fragments across the Glen Roy valley floor, proposing that, after 260 m lake drainage, the valley floor had an undulating surface with small lakes persisting in depressions between the fans. The River Roy transformed this landscape through incision of the fan toes, causing drainage of these small lakes and the deposition of material close to the valley floor.

Both Glen Spean and lower Glen Roy also exhibit good examples of fluvial terraces (Sissons 1979c). These are distributed between the mouth of the Treig valley in the east through lower Glen Spean (Fig. 16.2b) in the west and extend into lower Glen Roy up to the viewpoint. The terrace belts are typically 0.5–1.0 km wide; individual terrace fragments are 10–300 m wide and up to ~1 km long. They are normally present up to 30–40 m above the current rivers, although in lower Glen Roy they rise up to 60 m above the river in the vicinity of the Viewpoint; locally, a suite of more than five terraces is present. Exposures in the terraces exhibit sandy gravel facies in the west close to the Treig and Laire valleys, whilst in lower Glen Spean sandier facies are exposed and in lower Glen Roy sandy facies locally overlie laminated sediments (sands, silts and clay) indicating that the terraces post-date the ice-dammed lake sequence.

Through mapping and survey of the terrace surfaces, Sissons (1979c) reached three key conclusions. First, four further lake levels post-dated the 260 m lake and were confined to lower Glen Spean. These lower lakes had altitudes of 113, 99, 96.5 and 90.5 m OD and were fed by outwash from the retreating Treig and Laire ice lobes and from Glen Roy. Second, from the distribution of the terraces in lower Glen Roy, Sissons inferred that a small lake was temporarily impounded behind the 'Massive Drift Accumulation' during the development of the terraces and the associated lower lakes. Third, the lack of fluvial terraces for 2 km in the vicinity of the kame terraces and kames in middle Glen Spean (Fig. 16.6b) is explained by the presence of dead ice at this location whilst the higher terraces formed.

16.9 Rock-Slope Failures and Palaeoseismicity

Glen Roy, Glen Gloy and intervening valleys contain a cluster of about 17 postglacial rock-slope failures (RSFs) that collectively occupy an area of ~5.7 km^2. Most are rock-slope deformations (RSDs), formed by the gradual downslope displacement of mountainside slopes (Chap. 14). These typically exhibit upper-slope tension cracks and trenches, slope bulging and bedrock benches, scarps and/or antiscarps (upslope-facing scarps) with localised superficial collapse of displaced bedrock blocks. Those in the Glen Roy–Glen Gloy area exhibit limited downslope displacement and tend to lack rupture surfaces or clearly defined lateral margins, but their extent is locally defined by springlines where groundwater percolating through fractured rock emerges at the surface. Two RSDs in Glen Roy exhibit partial collapse: a huge (~2.0 km^2) RSD opposite Brunachan on the steep (~38°) eastern slopes of Beinn Iaruinn and the Bohaskey RSD (~0.6 km^2) in the lower glen, where about one-third of the original RSD (informally termed the 'Viewpoint' RSF) has failed as a rockslide that extends to the floor of the glen (Fig. 16.8a). The area contains only two relatively simple rockslides. One of these, the Glen Fintaig RSF (0.17 km^2) represents catastrophic failure along

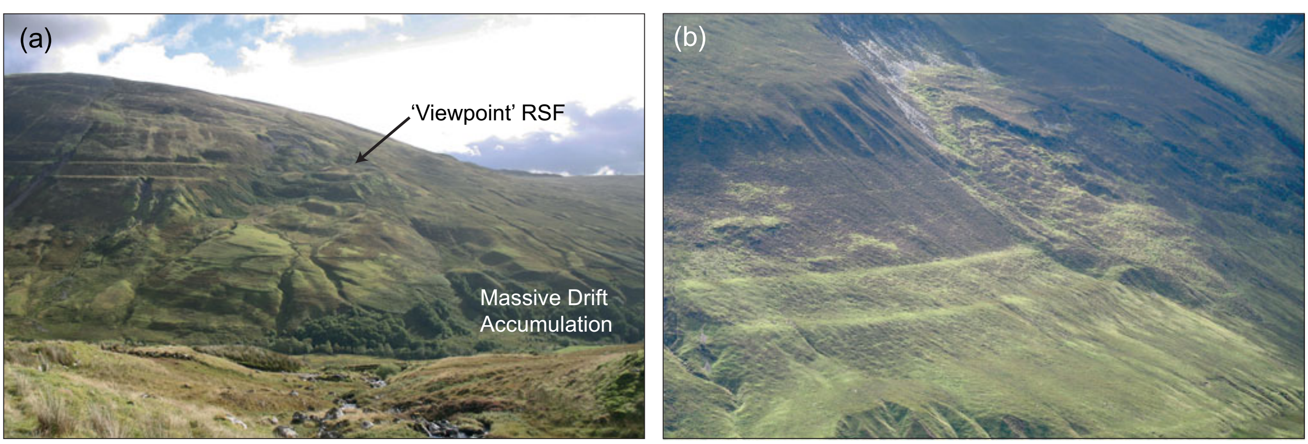

Fig. 16.8 **a** Eastern hillslope of lower Glen Roy opposite the 'Massive Drift Accumulation'. On the left is the Bohaskey rock-slope deformation (RSD), which is cut by the 325 m and 350 m shorelines, indicating that downslope displacement had ceased prior to shoreline formation. In the centre is the 'Viewpoint' RSF, a deep-seated rockslide that represents failure of the southern margin of the original RSD. This rockslide has displaced the same shorelines and therefore occurred during or after lake drainage. **b** The Braeroy arrested slide, which displaced the upper shorelines but failed to reach the 260 m shoreline, although this shoreline is poorly defined in this area. (Images: **a** Adrian Palmer; **b** Colin Ballantyne)

slope-parallel schistosity and extends to the valley floor, where the run-out exhibits brittle folding and thrusting of planar blocks. The other, the Braeroy RSF (0.12 km^2) is a small arrested rockslide on the 35° slope opposite Brae Roy Lodge (Fig. 16.8b).

The Glen Roy–Glen Gloy RSF cluster is of interest because the morphostratigraphic relationships between the RSFs and the lake shorelines permit relative dating of the RSFs (before or after shoreline formation) and also because of evidence that some (possibly many) of the RSFs were triggered or conditioned by earthquakes. At some sites, undisturbed shorelines cut across RSDs (Fig. 16.8a), implying that these developed then stabilised in the interval (~15–12 ka) between ice-sheet deglaciation and shoreline formation, consistent with the timing of RSF response following deglaciation elsewhere in Scotland (Ballantyne et al. 2014; Chap. 4). This was confirmed by Peacock and May (1993), who demonstrated that strata deformed by deep-seated rock-slope deformation in Glen Gloy are cut by the single (~355 m) shoreline in that glen. Other RSFs, including the Glen Fintaig and Braeroy landslides and the collapsed margins of the Brunachan and Bohaskey RSDs, have disturbed, buried or destroyed some or all of the shorelines (Fig. 16.8), demonstrating that they occurred during lake-level lowering or after final lake drainage. Using such evidence, Fenton (1991) concluded that about 30% of the RSFs had stabilised prior to flooding of the glens, a few occurred during lowering of lake levels and the remainder failed (or continued to move) after lake drainage.

The possibility that some or all of the RSFs were seismically triggered was first explored by Sissons and Cornish (1982a, b), who demonstrated that the Glen Roy shorelines exhibit dislocation and tilting. They attributed shoreline dislocation to differential glacio-isostatic uplift of crustal blocks and suggested that such movements were accompanied by RSF-triggering earthquakes. Ringrose (1989) reported further evidence for seismic activity in the form of soft-sediment deformation structures in varved lake sediments. He interpreted these structures as seismites produced by cyclic shaking and sediment liquefaction, showed that at least two deformation events had occurred and found that deformation intensity decreased radially from an inferred earthquake epicentre in Glen Roy. From the maximum epicentral distance (~15 km) of sediment deformation structures, he estimated that they were caused by an earthquake with $M_W = 5.9$, later revised by Fenton (1991) to $M_S = 6.3 \pm 0.5$. More controversially, Ringrose related the earthquake activity to glacio-isostatic reactivation of the Glen Gloy fault, which extends SSW from the head of Glen Gloy but is obscured by the Brunachan RSF in Glen Roy. The fault trace exhibits evidence of sinistral displacement of ~40 m and later dextral displacement of ~0.5 m. The former, however, probably reflects ancient (possibly Caledonian) fault movement (Firth and Stewart 2000), though the later dextral movement is apparently consistent with realignment of the principal horizontal stress field during glacio-isostatic uplift (Fenton 1991).

Stability analyses indicate that most RSF sites in the area are stable under gravitational loading alone (Fenton 1991), suggesting that some, most or all of the Glen Roy–Glen Gloy RSFs were initiated, conditioned or triggered by earthquake activity, including the RSDs that had stabilised prior to shoreline formation. However, the variable ages of the RSFs and the multiple dislocations of the lake shorelines imply that though there may have been a single major earthquake that produced the main suite of deformation structures in lake sediments (Ringrose 1989), it is likely that fault movements and seismic activity occurred over a prolonged period following ice-sheet deglaciation (e.g. Muir-Wood 2000; Stewart et al. 2000; Hampel et al. 2010), both before and after shoreline formation. Fluctuating lake levels (Fenton 1991) and crustal unloading by catastrophic lake drainage (Sissons 1979b; Sissons and Cornish 1982a) may also have promoted localised seismic activity. Because deglacial unloading alone may initiate RSDs (Agliardi et al. 2012; Chap. 14) and RSF response to earthquakes may be delayed (Gischig et al. 2015), establishing a definitive relationship between the Glen Roy–Glen Gloy RSFs and particular seismic events remains elusive.

16.10 The Timing of Ice Advance and Retreat in Lochaber

The reconstruction by Sissons (1979a; Fig. 16.4) provided the framework for the advance, maximum extent and retreat of glaciers in this sector of the West Highland Icefield. The Loch Lomond Stadial age of the readvance is confirmed by palynological evidence, which demonstrates that kettle holes and rock basins that infilled after the demise of the lake systems contain organic sediments no older than Early Holocene in age (Lowe and Cairns 1989). Direct dating of shorelines using cosmogenic radionuclide dating shows that these surfaces were exposed at ~11.9–11.5 ka, and the drainage of the lake systems has been dated to ~11.0–10.7 ka using samples extracted from the surface of the Glen Turret fan; though samples from the Allt Brunachain fan recorded a later age of ~9.9–9.7 ka (Fabel et al. 2010).

Detailed analysis of the record provided by glacilacustrine varves (annually layered sediments) in Glen Roy and Glen Spean has permitted the chronology of Lateglacial events in these glens to be reconstructed with unparalleled accuracy and precision. Palmer et al. (2010, 2012; Palmer and Lowe 2017) employed microfacies analysis of these laminated sediments to establish the first LLR glacilacustrine varve chronology in the UK, demonstrating that the entire lake system existed for a total of 515 years. The duration of specific lake levels was more difficult to estimate by using the varve thicknesses, and it was suggested that the falling 260 m lake system may not be fully represented within the record.

More recent glacilacustrine varve analysis employing data from ten sites within Glen Roy and Glen Spean includes varve sequences from all three lake levels. In upper Glen Roy, one of these sites could have accumulated varves only in the 325 m and 350 m lakes, whilst another site represents varves formed only in the 350 m lake (Palmer et al. 2020). The resulting Lochaber Master Varve Chronology 2019 (LMVC19) implies the following conclusions.

1. The total duration of the 260 m, 325 m and 350 m lake systems was 518 ± 18 varve years.
2. The configuration of ice advance reconstructed using the varve evidence differs from that of Sissons (1978) because it indicates that accumulation of varve sediments within Glen Roy did not commence until the 325 m lake level was reached. This was 188 years after the initiation of varve sedimentation in Glen Spean and suggests that (i) the Laire-Treig ice lobe had advanced to close to its maximum extent prior to the maximum of the Spean-Roy ice lobe; and (ii) Glen Roy was initially free-draining with the ice dam yet to form in lower Glen Spean and, when it did, the waters in Glen Roy rose to 325 m. Consequently, the 260 m lake level in Glen Roy was only created during the falling sequence of the Glen Roy lakes.
3. Ice advance to its maximum extent in Glen Roy occurred within ~300 years from the inception of the 260 m lake in Glen Spean. The interval between the onset of ice retreat in Glen Roy to complete drainage of the main glacial lake systems was less than ~200 years. Average ice advance rates through the lower Spean and into Glen Roy were ~88 m a^{-1}, and retreat rates were up to ~30 m a^{-1}.
4. Shoreline formation rates for the 260 m and 325 m lake levels averaged 0.02–0.13 m a^{-1}, although the 350 m shoreline was excavated in only 47 years at a rate of 0.32 m a^{-1}.
5. The presence of the Vedde Ash, a tephra from Iceland, in the varve sediments dates the lake systems to between 12,135 and 11,618 cal ^{14}C a BP, indicating that glacier ice persisted in the Highlands of Scotland into the Early Holocene.

The new age estimates from the LMVC19 chronology provide the most precise evidence for the timing of LLR glacial activity in Scotland, improving our understanding of how and when different outlet glaciers of the West Highland Icefield advanced and retreated. It also gives unparalleled insights into the rate that geomorphological processes were operating and, with the aid of sub-surface mapping technologies, may allow more detailed estimates of sedimentation rates in the larger fan–delta accumulations within the valleys.

The precision and accuracy of the LMVC19 chronology also permit comparison to other European palaeoclimatic and palaeoenvironmental records at Meerfelder Maar, Germany and Krakenes, Norway. Lane et al. (2013) propose that there was a time-transgressive shift in the pattern of warmer air masses moving across northern parts of Europe during the mid-Younger Dryas (Loch Lomond Stade) . Using the Vedde Ash to align records in Germany and Norway, the warming in Norway would have been delayed by ~140 years. Theoretically, this warming would also have been experienced in Scotland and may have led to the ablation of the ice mass which formed the ice-dammed lakes. However, the initial retreat of the ice in Glen Roy appears to have post-dated the inferred timing of warming in Norway by ~150 years (Palmer et al. 2020). The reason for this delay is still to be resolved and could relate to the sensitivity of the cryospheric response to atmospheric warming and/or the influence of the ice mass on local climates.

16.11 Conclusions

The Parallel Roads and associated landforms of Glen Roy and Glen Spean represent one of the most remarkable landsystems in Scotland and deservedly merit their international geomorphological acclaim. The geomorphological importance of the area was highlighted in the Geological Conservation Review (Gordon 1993; Gordon and McEwen 1997). It has justly been accorded several conservation designations and is one of the key areas for geotourism in Lochaber Geopark (Brazier et al. 2017). From its pivotal contribution in the emergence of the glacial theory in the nineteenth century to the recent reconstruction of one of the most precise and accurate chronologies for the advance and retreat of the LLR glaciers now available, the area has played a major role in our understanding of Lateglacial landscape evolution and environmental change in Scotland (Palmer and Lowe 2017, Palmer et al. 2020). Recent geochronological studies have also greatly elucidated the timing and rates of formation of certain landforms. However, there is still much to learn from the rich geomorphological archive in Glen Roy and Glen Spean. Future refinements may include constraining the spatial and temporal patterns of ice build-up and the later phases of ice retreat in the wider area, the development of glacilacustrine systems in Glen Gloy, the role of icefields on plateaux adjacent to Glen Roy and Glen Spean, the impact of glacio-isostatic adjustment on the shorelines and the rates of paraglacial adjustment after final lake drainage. The answers perhaps lie in the application of technologies such as remote mapping and ground-truthing, to geophysical profiling of sediment accumulations such as the fans, deltas and basins in Glen Roy and Glen Spean. Finally, there is great importance in conserving the landscape for future use by the scientific community and communication of research results to enhance wider public understanding and appreciation of this exceptional area.

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17 Western Grampian Highlands

John E. Gordon and D. Noel Williams

Abstract

The Western Grampian Highlands, including Ben Nevis and Glen Coe, form a rugged mountain landscape displaying classic features of glacial erosion developed across a range of metamorphic and igneous rocks. Glacial troughs radiate out from Rannoch Moor, which formed a major ice dispersal centre during successive Pleistocene glacial episodes. The area is also notable for glacier depositional landforms associated with the Loch Lomond Readvance, particularly a large valley sandur and associated landforms at the western end of Loch Etive and hummocky recessional moraines on Rannoch Moor—a key locality for establishing the timing of deglaciation of the West Highland ice cap. Excellent examples of Holocene rock-slope failures, debris flows, debris cones and alluvial fans add to the geomorphological diversity, while Glen Coe provides a geologically outstanding illustration of a caldera collapse. A large part of the area lies within Lochaber Geopark.

Keywords

Caldera collapse • Glacial erosion • Loch Lomond Readvance • Valley sandar • Hummocky recessional moraines • Debris flows and cones • Alluvial fans • Rock-slope failures • Geopark

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17.1 Introduction

The Western Grampian Highlands (hereafter Western Grampians) contain some of the most impressive glacially eroded mountain landscapes in Scotland. They are underlain by Neoproterozoic and Palaeozoic metamorphic and igneous rocks whose characteristics have influenced processes of weathering and erosion. This chapter highlights the mountain heartland of Lochaber, centred on Ben Nevis and Glen Coe, the montane basin of Rannoch Moor and the landforms of the western coastal fringe along Loch Linnhe (Fig. 17.1). Numerous summits exceed 1000 m in altitude, including Ben Nevis (1345 m), the highest mountain in Britain. To the north of Glen Coe the mountain ridges of Aonach Eagach, the Mamores and the Ben Nevis–Aonach Mòr–Grey Corries range run west–east, while to the south of Glencoe the mountain blocks are more fragmented. The mountains are heavily dissected but there is a general accordance of higher summit altitudes at ~900–1100 m, with occasional plateau remnants (Fig. 17.2a). Geologically, both Ben Nevis and Glen Coe provide sections through eroded Palaeozoic volcanoes, and the latter was the location where evidence for 'cauldron subsidence' (caldera collapse) was first identified. During Pleistocene glacial episodes, the Rannoch Moor basin formed a major ice accumulation and dispersal centre for successive Scottish ice caps and ice sheets.

Classic features of glacial erosion include troughs, breached watersheds, cirques, deep rock basins, hanging valleys and extensive ice-scoured bedrock. Along the coast, the sea has flooded the lower reaches of the glacial troughs, forming sea lochs and fjords (Fig. 17.2b). Depositional landforms associated with the Loch Lomond Readvance (LLR) during the Loch Lomond (≈Younger Dryas) Stade (~12.9–11.7 ka) include hummocky recessional moraines on Rannoch Moor and glacifluvial valley sandar formed as the glaciers retreated from the coast. Rannoch Moor is also a critical locality for dating the timing of the demise of the LLR ice cap. In the Glen Coe area, the high precipitation

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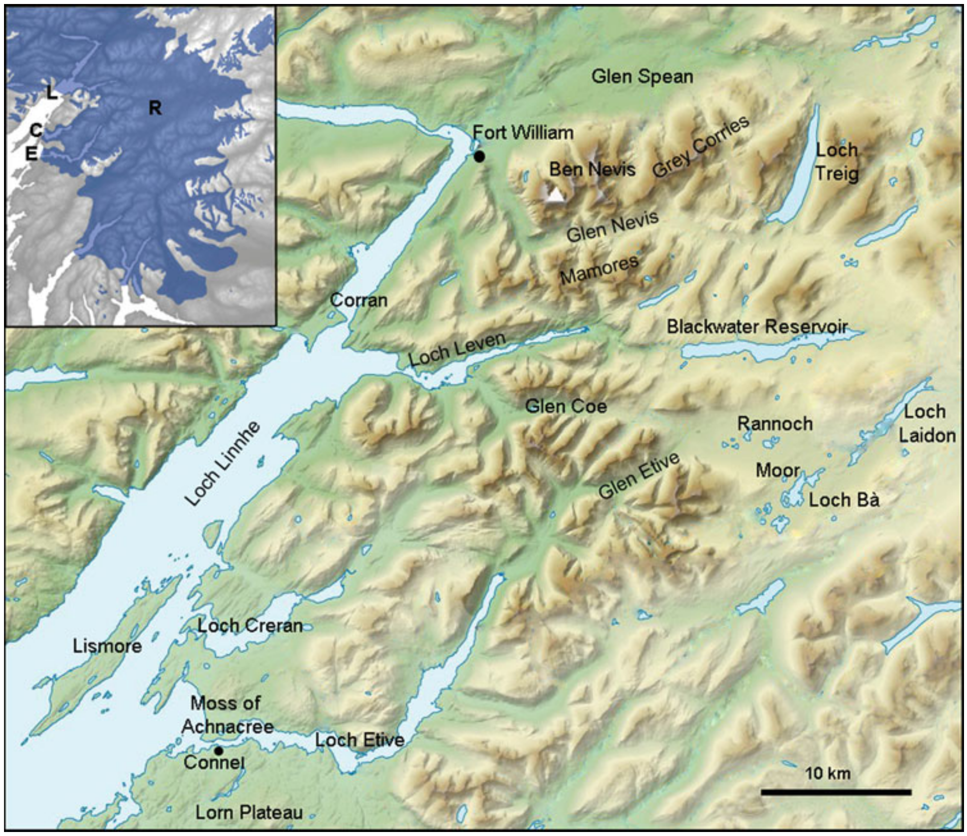

Fig. 17.1 Western Grampian Highlands: relief and main locations mentioned in the text. (Contains Ordnance Survey data © Crown copyright and database right / CC BY-SA; https://creativecommons.org/licenses/by-sa/3.0). Inset: Loch Lomond Readvance glacier limits; note that numerous nunataks protruded through the icefield complex but are not shown. (From Bickerdike et al. 2018) Glacial landsystems, retreat dynamics and controls on Loch Lomond Stadial (Younger Dryas) glaciation in Britain. Boreas 47:202–224. Attribution 4.0 International (CC BY 4.0). C: Creran glacier; E: Etive glacier; L: Linnhe glacier; R: Rannoch Moor

($\sim$4000 mm a^{-1} on the mountains) and available paraglacial sediment sources have enabled the formation of many Holocene debris flows, debris cones and fluvially modified debris cones.

17.2 Geology and Landscape

The Western Grampians are part of the Grampian Highlands terrane and represent the deeply eroded roots of the Caledonian orogenic belt (Chap. 2). They comprise a varied assemblage of metasedimentary rocks belonging to the Dalradian Supergroup, together with significant outcrops of both intrusive and extrusive igneous rocks (Stephenson and Gould 1995; Fig. 17.3). This geological diversity and the NE–SW structural grain and weaknesses imparted during the Caledonian Orogeny have strongly influenced the processes of weathering and erosion and hence the form of the present landscape.

During the Grampian phase of the Caledonian Orogeny at $\sim$470 Ma, folding occurred on both a large and small scale, with several recumbent folds extending many kilometres across the strike, including the Kinlochleven Anticline and Ballachulish Syncline. These large folds were dislocated by major thrusts or slides which resulted in an extremely complex overall structure. Striking examples of overturned rocks occur in the Mamores on the south side of Glen Nevis. Although much of the quartzite is inverted, on the summits of Sgùrr a' Mhaim and Stob Bàn the strata are folded the right way up in a large recumbent anticline (Fig. 17.2a).

Although igneous activity occurred before, during and after the Caledonian Orogeny, it is the igneous rocks formed after the main mountain-building events that dominate the Western Grampians. Fissure eruptions of large quantities of magma produced basalt and basaltic andesite lavas which at one time probably covered much of the Western Grampians. These Early Devonian rocks now form the >300 km^2 Lorn Plateau in the south of the area. Two smaller occurrences of volcanic rocks occur farther north around Glen Coe and Ben Nevis. Those at Glen Coe were probably preserved because they subsided >1 km during the piecemeal collapse of a large elliptical caldera. This phenomenon was originally discovered in the early twentieth century by Geological Survey geologists and termed 'cauldron subsidence' (Clough et al. 1909)—the first recognition in the world of caldera collapse in ancient volcanic rocks. Later work in the 1990s suggested that the collapse was not a single event, but took place in stages, involving different faulted blocks (Kokelaar and Moore 2006). The initial igneous activity at Glen Coe produced sills of andesite and basalt, which were intruded into wet sediments resting on an eroded land surface of

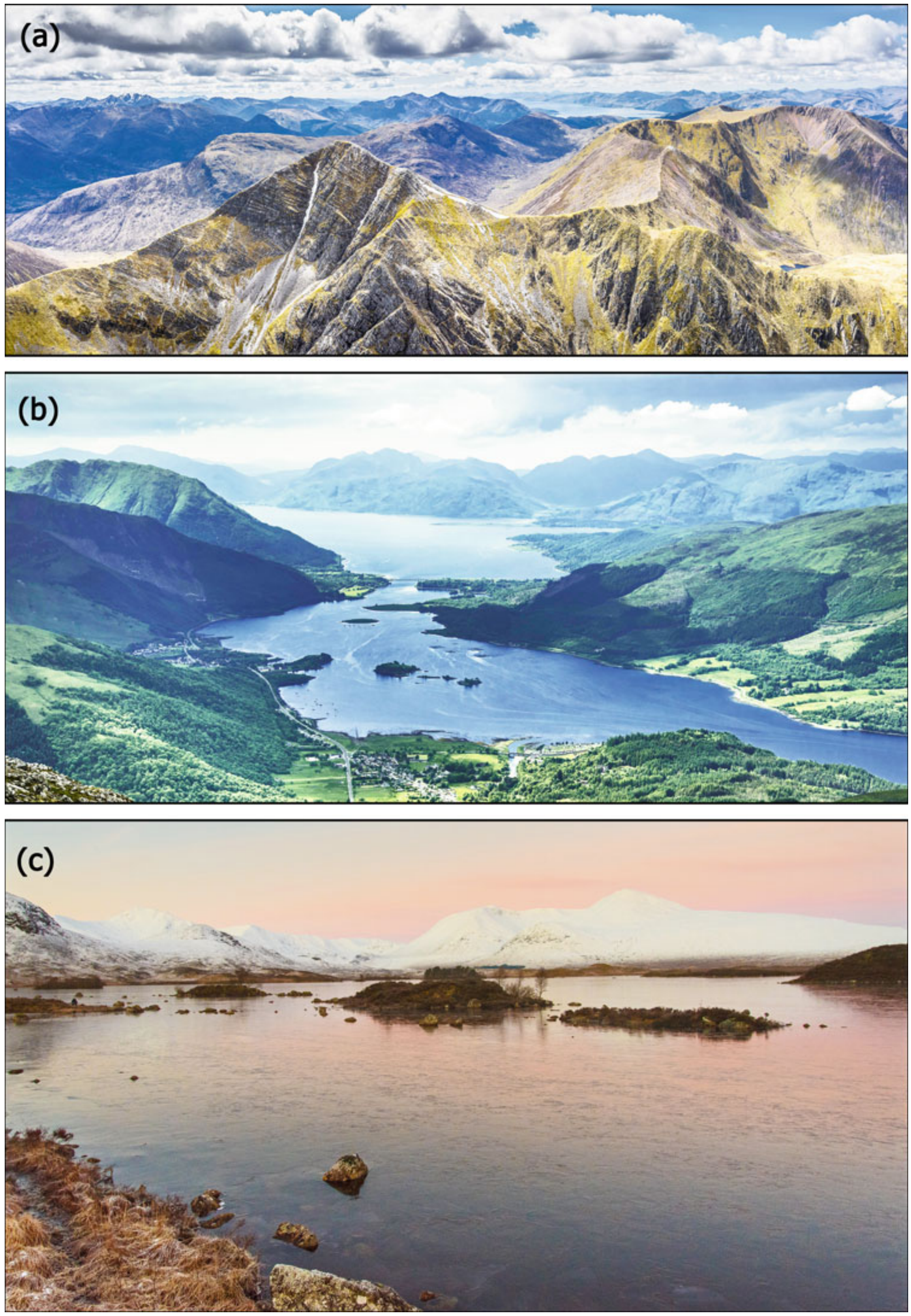

Fig. 17.2 **a** Aerial view of the Mamore range and the mountains to the south. Summits up to ∼800 m in the middle distance are ice scoured, while higher summits often support a cover of frost-weathered regolith. The top half of a large recumbent fold can be seen in the quartzite strata of Stob Bàn (centre left), exposed in the headwall of a rock-slope failure. **b** View west along the drowned glacial trough of Loch Leven. A marine-reworked outwash fan forms the narrows at its junction with Loch Linnhe. **c** The high-level basin of Rannoch Moor. (Images: **a** © Kyle Macintyre**;** **b**, **c** John Gordon)

metamorphic phyllites. A conspicuous change from dark andesite to orange rhyolite on the flank of Aonach Dubh marks a transition to more explosive eruptions (Fig. 17.4a). Seven separate eruptions in the later volcanic sequence have been interpreted as intra-caldera ignimbrites formed during different stages of collapse.

Volcanic and sedimentary rocks form the upper half of the north face of Ben Nevis (Fig. 17.4b). These brecciated volcanic rocks are thought to represent block and ash flows, rather than lava flows, on the flank of a former volcano situated to the northwest. The volcanic pile is surrounded by several concentric intrusions. The unusually close proximity of extrusive and intrusive rocks led early geologists to interpret this structure as a deeply eroded caldera, rather smaller than that in Glen Coe. Recent work suggests that the volcanic pile may instead be a large roof pendant and that no collapse took place (Muir and Vaughan 2018).

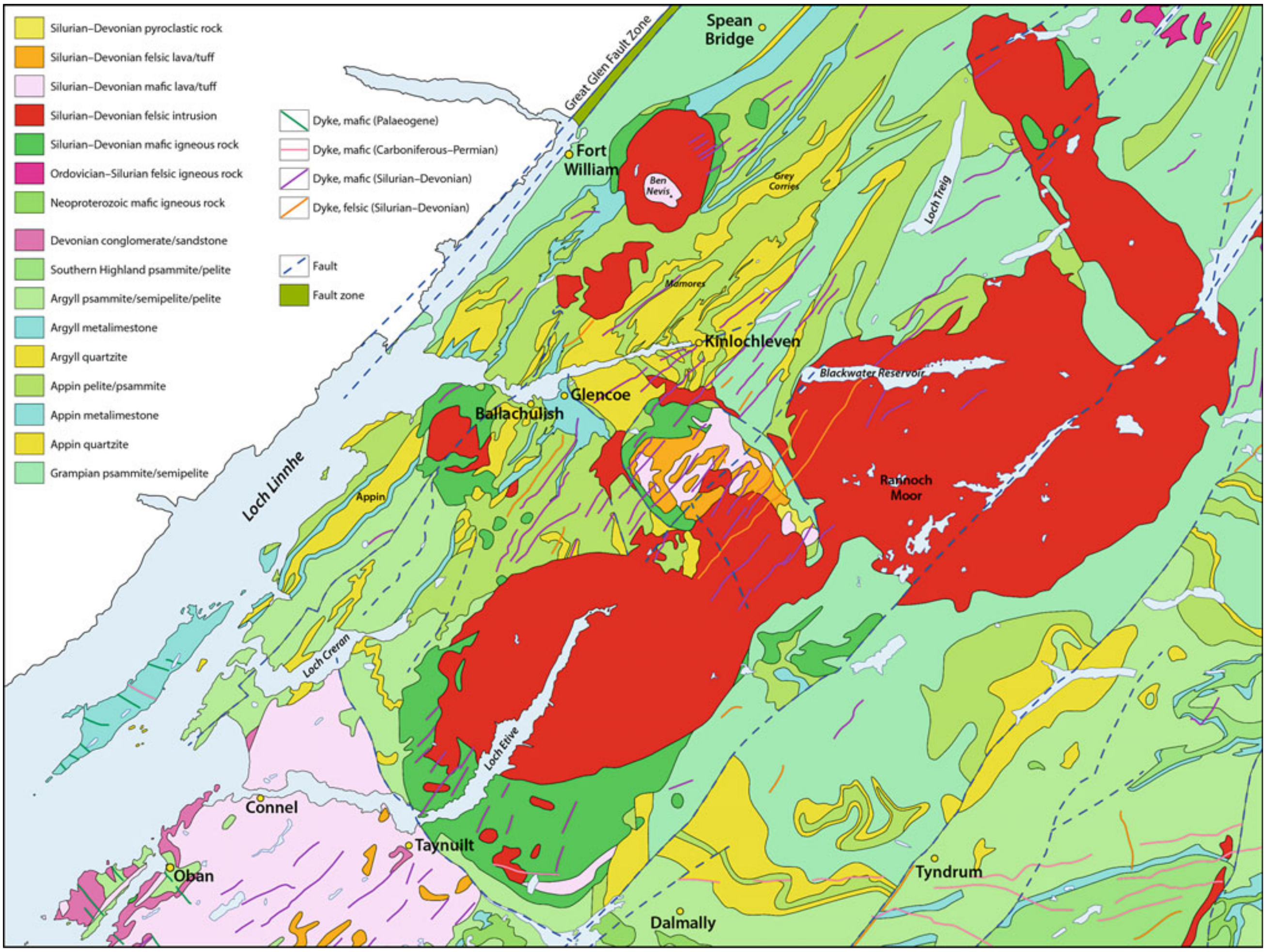

Fig. 17.3 Bedrock geology of the Western Grampian Highlands. (Contains British Geological Survey materials © UKRI 2020)

Large, mainly granitic plutons were emplaced in the Western Grampians from 425 to 410 Ma, including the Etive, Clach Leathad, Ballachulish, Mullach nan Coirean, Ben Nevis and Rannoch Moor intrusions (Fig. 17.3). Major dyke swarms running in a NE–SW direction are associated with the Etive and Ben Nevis igneous complexes. Another episode of dyke injection, in WNW–ESE and E–W directions, occurred in Permian-Carboniferous times.

The Western Grampians are cut by major NE–SW trending faults. Movement on the most significant of these, the Great Glen Fault, was initiated at the close of the Caledonian Orogeny, ∼ 430–390 Ma, when significant sinistral strike–slip movement took place. The whole fault zone is ∼ 3 km wide and includes shattered rocks that have proved susceptible to erosion and the formation of a major topographic feature, the Great Glen.

The post-Caledonian history of the Western Grampians involved deep and prolonged erosion, and by the end of the Cretaceous the Caledonian mountains had been reduced to a landscape of low relief. Subsequent episodic uplift during the Palaeogene and Neogene accompanied by differential weathering and erosion shaped the broad-scale elements of relief and drainage, resulting in a heavily dissected pre-glacial landscape (Chap. 3). Erosion exploited shatter belts associated with faults and rocks more prone to weathering. Quartzites and the more resistant igneous rocks generally form higher ground: several peaks in the Mamores and the Grey Corries (northeastwards from Sgùrr a' Bhuic to Cruach Innse) bear quartzite caps, while the high summits of Ben Nevis and Glen Coe comprise volcanic rocks. The large (∼ 130 km^2) upland basin of Rannoch Moor (Fig. 17.2c) is a product of differential weathering and erosion of biotite granite that is less resistant than the metamorphic and volcanic rocks underlying the surrounding mountains. The more erodible dykes form gullies, or where more resistant, occasionally stand proud as ridges. A fine example of erosion along two different dykes occurs on the floor of Glen Coe (Sect. 17.8).

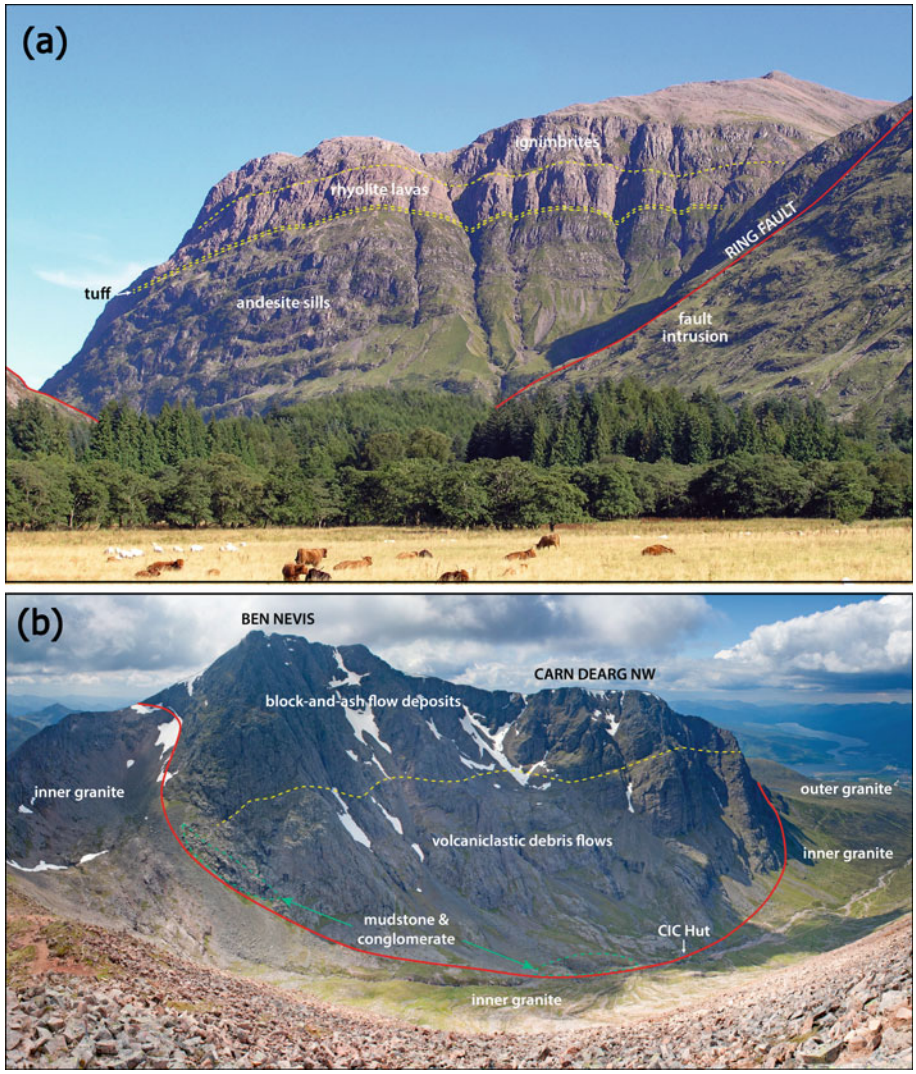

Fig. 17.4 **a** West face of Aonach Dubh, Glen Coe, showing a stack of shallow-level sills that were originally intruded in unlithified wet sediments, overlain by rhyolite lavas and ignimbrites, which indicate a change to more explosive volcanic activity. **b** The 700 m high North Face of Ben Nevis comprises a pile of volcaniclastic rocks set within a granite intrusion. (Images: **a** Noel Williams; **b** © Nick Landells, Lakeland Photo Walks)

17.3 Glaciation History and Patterns

Lochaber played a significant part in the early development of the glacial theory through the observations of Louis Agassiz in 1840, which provided evidence for the former existence glaciers in Scotland (Agassiz 1841, 1842; Chap. 1). Agassiz highlighted the similarity of 'terraced mounds of blocks' at the north end of Loch Treig to moraines in the Chamonix valley in Switzerland and noted the occurrence of polished rock surfaces and moraines at the foot of Ben Nevis and roches moutonnées and polished and striated bedrock at Loch Treig and along the shores of Loch Leven near Ballachulish (Boylan 1998). He also interpreted the Parallel Roads of Glen Roy (Chap. 16) as the shorelines of former ice-dammed lakes.

The landscape of Lochaber reflects the cumulative effects of multiple glaciations during the Pleistocene, involving ice sheets, ice caps and mountain glaciers (Hall et al. 2019; Merritt et al. 2019; Chap. 4). Stratigraphic evidence for glacial episodes pre-dating the Late Devensian is absent, however, having been removed by later glaciations. Favoured by topography and high precipitation, glaciers likely built up repeatedly in the mountains to the west of Rannoch Moor and flowed into the Rannoch basin, which then became a major ice dispersal centre (Bailey and Maufe 1916; Sissons 1967; Payne and Sugden 1990). It is probable that all summits were covered by the last Scottish Ice Sheet during the Late Devensian glacial maximum and also during earlier episodes of ice-sheet glaciation; erratics of Rannoch granite occur up to ~900 m on the Glen Coe hills and 1100 m on Ben Nevis. The patterns of striae and erratic dispersal indicate that the position of the ice divide likely migrated, at times lying across the western side of Rannoch Moor and running northwards towards Ben Nevis, and at other times, over Rannoch Moor itself (Thorp 1987; Hughes et al. 2014).

Later, Rannoch Moor and the mountains to the west were major centres of ice accumulation for the glaciers of the

LLR. Empirical evidence and numerical modelling indicate growth of a domed ice cap with a maximum surface elevation of ~900 m OD, transitional to an icefield system with outlet glaciers confined within the main glens (Golledge and Hubbard 2005; Golledge et al. 2008, 2009; Bickerdike et al. 2018; Fig. 17.1 inset). Outlet glaciers drained west along Glen Nevis, the Loch Leven trough and Glen Coe, merging with a glacier in Loch Linnhe. To the northeast, ice in Glen Spean originating west of the Great Glen merged with glaciers emerging from the cirques of the Ben Nevis range and with ice flowing through the Treig breach and extended eastwards to a limit formed by a large, multiple moraine loop near the western end of the present Loch Laggan (Chap. 16). To the east, an outlet glacier reached the eastern end of Loch Rannoch. Other prominent glaciers flowed along Glen Creran and Glen Etive to the southwest, terminating at the coast. Geomorphological evidence at Loch Etive, Loch Creran and Loch Linnhe demonstrates that glaciers retreated from their maximum limits and stabilised at topographic pinning points. Subsequent active retreat is indicated by recessional moraines on the floor of the inner basin of Loch Etive (Audsley et al. 2016) and by hummocky recessional moraines on Rannoch Moor (Sect. 17.5.2).

Evidence from the area occupied by the West Highland Icefield is critical for establishing the chronology of the LLR ice maximum and subsequent ice recession (Chap. 4). Scotland occupies a key geographical position on the Atlantic seaboard of Europe, and its glacial chronology is important for understanding the driving mechanisms of abrupt climate change in the North Atlantic region during the Younger Dryas, and particularly the influence of sea-surface temperatures and seasonality on terrestrial glacier responses (Golledge 2010; Bromley et al. 2014, 2018). The balance of available evidence (Chaps 4 and 16) indicates that the West Highland Icefield and some of its outlet glaciers survived until the end of the Loch Lomond Stade (~11.7 ka), although some smaller glaciers probably retreated from their maximum positions in mid-stade. Radiocarbon dates on plant macrofossils retrieved from the base of cores from peat-filled basins within the hummocky moraines on Rannoch Moor provide crucial limiting ages for the final disappearance of ice from the main centre of ice dispersal (Sect. 17.5.2), although glaciers likely survived longer in the cirques of the adjacent mountains.

17.4 Landscapes and Landforms of Glacial Erosion

Glacial erosion has had a profound influence in shaping the landscape of the Western Grampians. High snow accumulation fed by Atlantic airmasses and steep pre-glacial glens favoured the development of fast-flowing, warm-based glaciers, particularly during ice-cap and ice-sheet build-up stages. Numerical modelling of the LLR ice cap shows that Glen Nevis, Glen Coe and Glen Etive coincided with corridors of highest subglacial erosion potential (Golledge et al. 2009). Lochaber is a spectacular example of a 'composite' landscape of glacial erosion (Sugden and John 1976), shaped by multiple phases of cirque, valley-glacier and ice-sheet erosion. Much of the pre-glacial valley network has been comprehensively modified, compartmentalising the relief into distinct mountain massifs (Clayton 1974; Haynes 1977, 1983). Contrasting with the extensive plateau palaeosurfaces of the Eastern Grampian Highlands and the Cairngorm Mountains (Chaps 18 and 20), the high density of glacial troughs, breached watersheds, cirques and over-deepened valleys dominates the scenery of the Western Grampians (Fig. 17.2a, b).

Glacial troughs form a distinctive radial pattern centred on Rannoch Moor (Linton 1957). They mostly follow pre-existing valleys exploiting the NE–SW Caledonian structural lineaments, although some glaciers breached pre-existing watersheds (Linton 1951). Examples of the latter include the fault-controlled trough of Loch Treig to the north of Rannoch Moor, which involved ~550 m of glacial incision, Glen Coe to the west and Glen Etive to the southwest (Fig. 17.5a). Glen Coe exhibits truncated spurs and the hanging valley of Coire Gabhail (the 'Lost Valley') along its south side (see Fig. 17.9b below). Glen Nevis is a fine example of a glacial trough, with ice-moulded bedrock at Polldubh crags, roches moutonnées on the nearby valley floor and a spectacular meltwater gorge separating the lower and upper sections of the glen. Above the gorge, the Steall waterfall plunges in a single drop of 120 m from the hanging valley of Coire a' Mhàil (Fig. 17.5b).

Intermediate slopes, valley shoulders, cols and summits up to ~800 m are frequently ice-scoured, with frost-weathered regolith covering summits at higher elevations (Figs. 17.2a and 17.5c). The distal parts of western troughs were drowned by relative sea-level rise during the Early Holocene (Chap. 4), forming the fjords and sea lochs of Loch Linnhe, Loch Leven and Loch Creran (McIntyre and Howe 2010; Fig. 17.1). Glen Etive, and its continuation as Loch Etive, is partly fault controlled. The loch occupies a glacially over-deepened trough comprising several basins, the deepest reaching a maximum water depth of 145 m (Howe et al. 2001; Audsley et al. 2016). A rock bar at the mouth of the loch is the site of a tidal overfall, the Falls of Lora. Loch Laidon, on Rannoch Moor, occupies a rock basin excavated along the line of the Ericht-Laidon fault.

Numerous cirques fret the main ridges and mountain massifs (Figs. 17.2a and 17.5c). They are predominantly but not exclusively developed on the northern flanks of the mountains, with further asymmetry present in the strong northeasterly and northerly orientations of their headwalls.

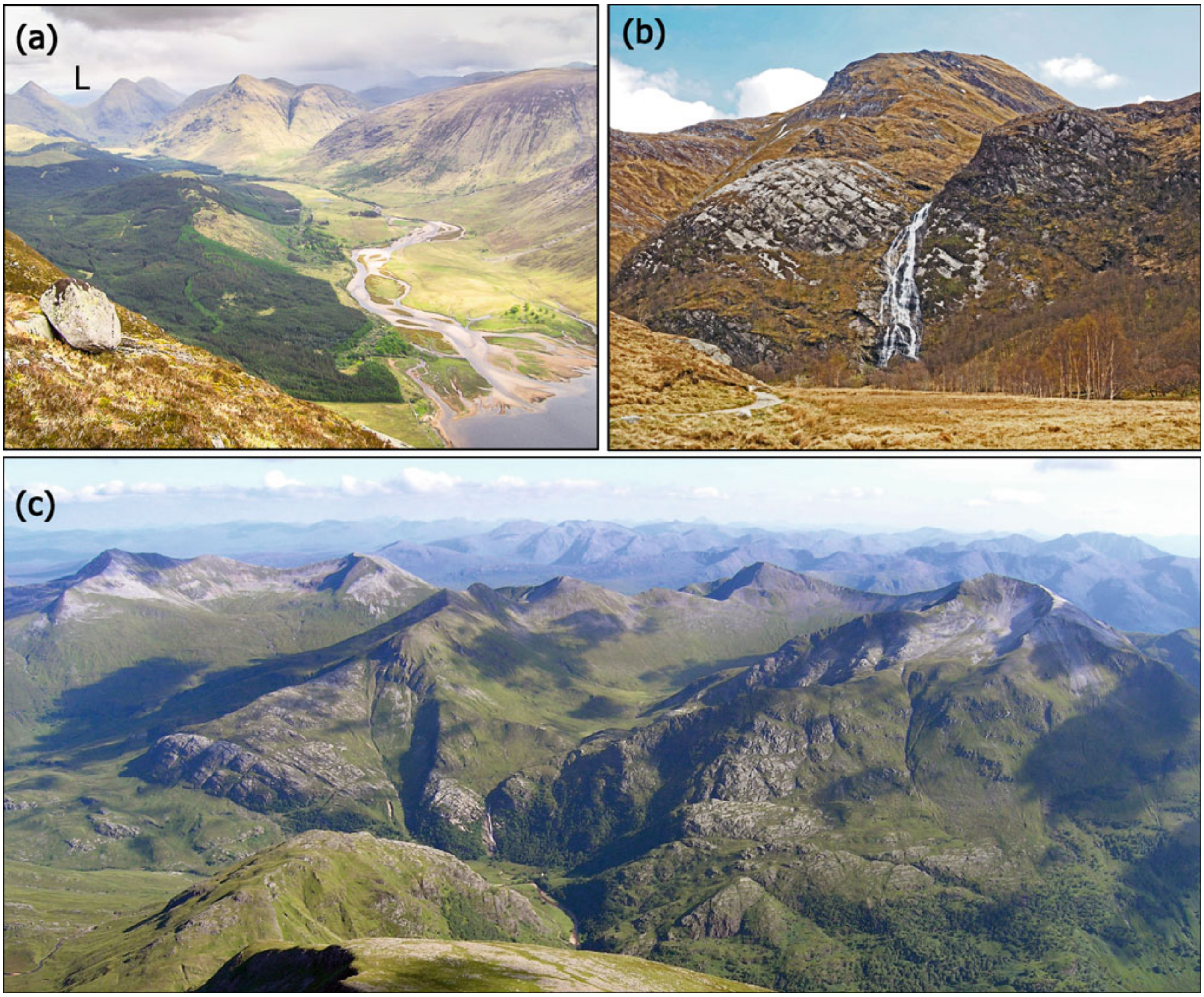

Fig. 17.5 **a** The partly fault-aligned glacial trough of Glen Etive is the product of multiple episodes of glacial and fluvial erosion. The Lairig Gartain (L) is a prominent, glacially breached watershed. The River Etive transitions from a rocky upper reach to a wandering gravel-bed and then to a low-angle gravel delta as it enters Loch Etive. **b** Upper Glen Nevis gorge and the Steall waterfall. **c** The Mamore range displays frost-weathered quartzite summit ridges and screes, with ice-scoured slopes below. (Images: **a** Adrian Hall; **b**, **c** Noel Williams)

Cirque headward erosion has contributed to the formation of the arêtes of Aonach Eagach and Càrn Mòr Dearg.

17.5 Landscapes and Landforms of Glacial and Glacifluvial Deposition

LLR glaciers removed or reworked much of the earlier glacial deposits, and thick accumulations of glacial deposits are rare within the area apart from Rannoch Moor (Thorp 1991). This contrasts with the area immediately to the south of Rannoch Moor, where extensive glacial deposits emplaced by the last ice sheet were only slightly modified during the LLR (Golledge 2006, 2007). Two distinctive assemblages of deposits are associated with the LLR glaciers: valley sandar and recessional moraines.

17.5.1 Valley Sandar and Ice Limits

Several outlet glaciers draining the LLR ice cap terminated at the west coast producing a series of striking landforms, including ice-contact features and glacial outwash deposits. The maximum extent of these glaciers was probably determined by tidewater calving and their patterns of recession by the positions of topographic constrictions or pinning points where the glacier fronts stabilised, so that outwash spreads or valley sandar occur inside the maximum glacier limits (Peacock 1971; Gray 1975; Greene 1992).

At its maximum, the Etive glacier probably extended to a large accumulation of glacial deposits 2 km west of Connel, at the westernmost extent of Loch Etive. Within this limit, on the north side of the loch, a ~4 km^2 outwash plain at Moss of Achnacree (Gray 1975, 1993, 1995; Fig. 17.6) formed across a bedrock threshold that constricts the mouth of the loch. Kettle holes and a meltwater channel occur on the outwash surface. A narrow channel, possibly cut by outflow from a freshwater ice-contact lake during glacier retreat, forms the exit of the loch. The margins of the outwash plain were later trimmed by the sea at ~13 m OD (Ordnance Datum), and a spit formed along its western margin. As relative sea level fell, Holocene raised beach deposits formed along the margins of the outwash plain. Smaller outwash terraces occur on the south side of the loch, and kame terraces, kame and kettle topography and eskers extend eastwards along both its sides (Fig. 17.6), together with the Moss of Achnacree forming perhaps the finest outwash and kame terrace system in Britain (Gray 1993).

An end moraine complex at the mouth of Loch Creran (Peacock 1971, 1993; Gray 1975; Peacock et al. 1989) comprises reworked marine clay with sand and gravel and runs for ~2.5 km southeast from South Shian parallel to the shore of the loch, with a peat-covered outwash plain beyond. On its distal side, mounds of ice-contact sediments surrounded by outwash suggest that the Creran glacier extended ~2 km further south and west. The glacier advanced over fossiliferous Lateglacial marine and glacimarine silts and clays (Clyde Beds), including high-arctic molluscan shells, laid down on the loch floor. Exposures at South Shian and Balure of Shian show contorted clay laminae and crushed shells, and at Balure of Shian the clay is glacitectonically folded into ice-contact silts overlain by gravel (Peacock 1971, 1993). Deposits at Balure of Shian are less glacitectonised and overconsolidated than those at South Shian, suggesting proximity to the ice-maximum position (Peacock et al. 1989). After reaching its maximum extent, the glacier retreated and stabilised at the South Shian moraine, possibly controlled by a rock bar. The outwash and moraines were later trimmed by the sea, and raised gravel beaches occur at ~13–14 m OD.

Radiocarbon dates on shells from the fossiliferous sediments incorporated in till at South Shian and North Shian

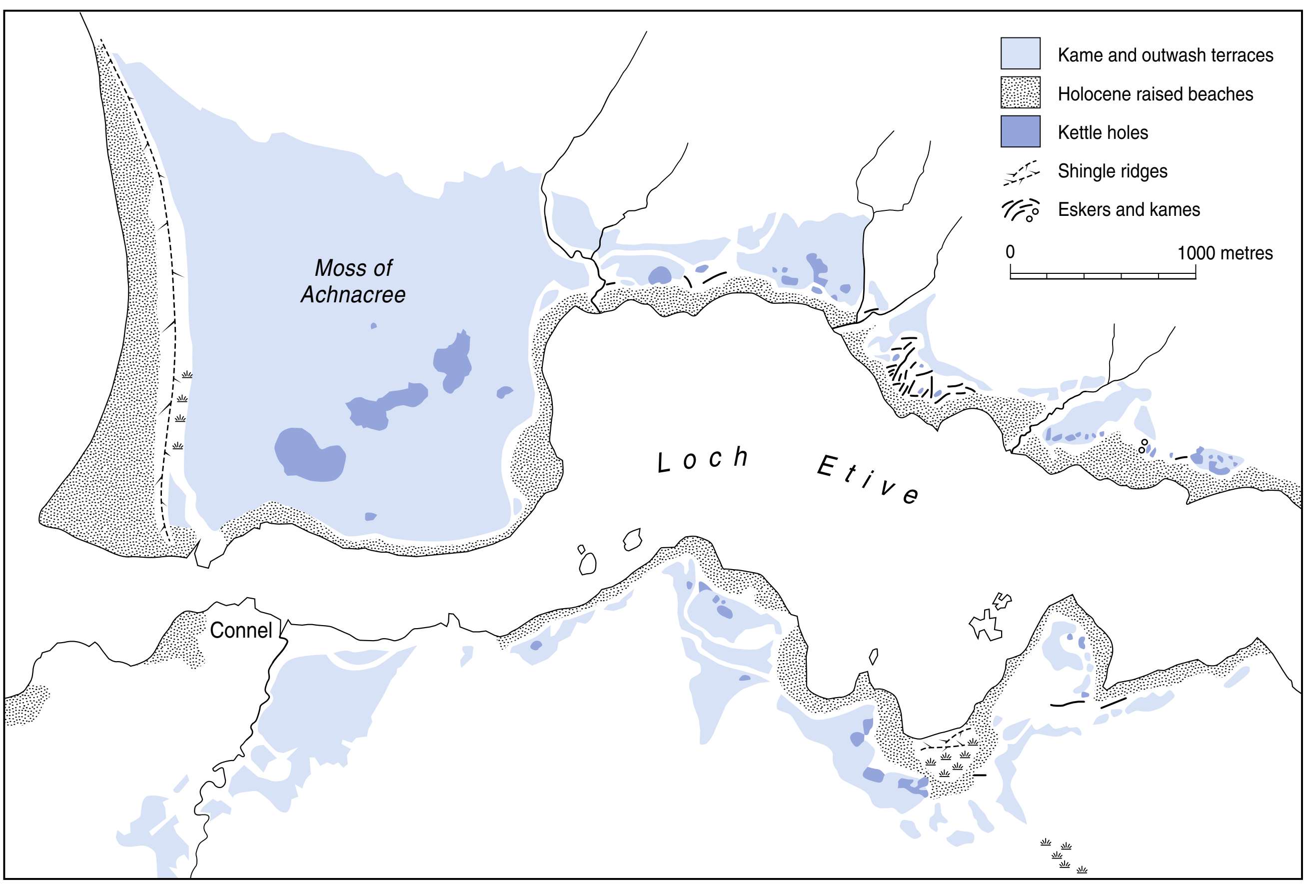

Fig. 17.6 Glacifluvial landforms at the mouth of Loch Etive. (After Gray JM (1975) The Loch Lomond Readvance and contemporaneous sea-levels in Loch Etive and neighbouring areas of western Scotland. Proc Geol Assoc 86:227–238, as adapted by Ballantyne 2019b. Reproduced with permission from the Geologists' Association)

provide a maximum age of 12.6–12.8 cal ka BP for the advance of the Creran glacier (Bromley et al. 2018). Dates obtained from the deposits at Balure of Shian suggest that the glacier remained relatively advanced until the middle or near the end of the Loch Lomond Stade (Peacock et al. 1989).

Thorp (1986) considered that the LLR limit in Loch Linnhe is marked by a small outwash fan at Kentallen, an interpretation supported by seismic evidence for the presence of ice-contact and ice-proximal sediments on the adjacent loch floor (Greene 1995). However, compacted marine sediments at the base of a vibrocore from the Shuna basin, ~14 km farther south, indicate that the Loch Linnhe glacier was possibly more extensive (McIntyre and Howe 2010), consistent with the numerical simulation of Golledge et al. (2008). Subsequently, retreat of the Linnhe glacier was interrupted by stabilisation at pinning points, indicated by a large kettled outwash terrace at Corran and a marine-reworked outwash fan at the mouth of Loch Leven (McCann 1961, 1966; Fig. 17.2b).

17.5.2 Recessional Moraines

Extensive recessional moraines are particularly well displayed on Rannoch Moor, with concentrations in the Bà valley in the southwest of the Moor, and around and to the north of Lochan Gaineamhach (Horsfield 1983; Thorp 1986; Wilson 2005; Bromley et al. 2014; Turner et al. 2014; Fig. 17.7). They are often described as 'hummocky moraine', comprising generally steep-sided mounds and ridges with numerous surface boulders of granite and a variety of constituents ranging from water-sorted silts, sands and gravels to coarse, rubbly till. Clasts are predominantly of local granite, but igneous erratics from the mountains to the west occur in the Bà valley. At ground level, the ridges and mounds appear chaotic, but many display nested alignments when mapped or viewed from the air. Such moraines occur extensively in the Highlands and are now considered to represent the successive ice-margin positions of actively retreating glaciers (Bennett and Boulton 1993; Lukas and Benn 2006). Moraine patterns on Rannoch Moor demonstrate that the last glaciers retreated actively into valley heads in adjacent mountains (Horsfield 1983; Boulton 1992; Wilson 2005). Cross-valley moraines on the north side of Blackwater Reservoir and east of Loch Bà (Turner et al. 2014) similarly indicate active glacier retreat, as do recessional moraines on the floor of Loch Etive partly controlled by topographic pinning points (Audsley et al. 2016). Possibly the extent and complex pattern of hummocky moraines on Rannoch Moor reflect the presence of earlier Late Devensian ice-sheet moraines partially preserved and reworked under the LLR ice cap (Bickerdike et al. 2018), while some ridges may represent glacial streamlining by ice flowing onto the Moor from the west (Wilson 2005).

Rannoch Moor provides critical evidence for dating the demise of the LLR glaciers in Scotland and more widely for understanding abrupt climate change and glacier responses in the North Atlantic region during the Younger Dryas (Golledge 2010; Bromley et al. 2018). A recent interpretation proposing an ice maximum early during the Loch Lomond Stade and deglaciation of Rannoch Moor by ~12.6–12.2 ka (Bromley et al. 2014, 2018) is at variance with other evidence that Rannoch Moor was not completely deglaciated until the end of the stade or the beginning of the Holocene (Small and Fabel 2016; Peacock and Rose 2017; Lowe et al. 2019; Chaps 4 and 16).

17.6 Periglacial Landforms

Periglacial features formed on exposed mountain summits and slopes as the latter became exposed to intensely cold conditions during ice-sheet deglaciation and the Loch Lomond Stade. Blockfields and frost-weathered debris, probably of pre-Late Devensian age (Fabel et al. 2012), occur on many summits, notably on Ben Nevis, Ben Starav and the quartzite ridges of the Grey Corries and the Mamores. Relict, stone-banked solifluction sheets and lobes formed where the debris moved downslope, as on the upper slopes of Aonach Beag and on Leum Uilleim (Thorp 1986). On some summits, this debris displays a clear lower limit, or 'trimline', below which the bedrock is ice scoured (Thorp 1981, 1986; Fig. 17.5c). Postglacial scree slopes are also well developed on the quartzites of the Mamores and Grey Corries (Fig. 17.5c).

In winter, high snowfall and heavy drifting from the summit plateau of Ben Nevis generate substantial snowbeds that survive late into the summer and provide important nival habitats (Manley 1971; Watson and Cameron 2010; Watson 2011; Fig. 17.4b). Below the 600-m-high cliffs on the north face of Ben Nevis a rampart of angular boulders, ~3 m high, ~75 m long and with a distal apron of avalanche run-out debris, impounds a small loch at ~910 m OD. These features are interpreted as a rare example of snow avalanche impact landforms in Britain (Ballantyne 1989). A good example of an avalanche boulder tongue also occurs below Tower Gully on Ben Nevis.

17.7 Mass-Movement Landforms

As the last ice sheet decayed, unloading of steep rock slopes resulted in propagation of stress-release joints, leading to deformation or local catastrophic collapse of mountain-side slopes. A seismic trigger associated with fault reactivation caused by differential glacio-isostatic rebound is possible (Ballantyne et al. 2014; Chap. 14). A cluster of such

Fig. 17.7 Hummocky recessional moraines, Rannoch Moor. (Image: John Gordon)

rock-slope failures (RSFs) occurs in Lochaber (Jarman 2006). They include the failure scars of Lateglacial RSFs where the run-out debris was removed by LLR glaciers; examples include Coire Sgreamhach on Beinn a' Bheithir and Stob Bàn in the Mamores (Ballantyne 2013; Fig. 17.2a). Many are arrested slides where the displaced rock mass has retained its integrity (Fig. 17.8a). Although most catastrophic RSFs in Scotland occurred during the Lateglacial period, prior to ~11.7 ka, some are of Holocene age. A striking example occurs near the mouth of Coire Gabhail in Glen Coe, where a rock avalanche involving some 0.6 Mt of rhyolitic ignimbrite has formed a massive cone of boulders that block the entrance to the hanging valley (Ballantyne 2007; Fig. 17.8b). This event occurred ~1700 years ago (Ballantyne et al. 2014), damming the glen so that floodplain deposits accumulated upvalley from the landslide debris. Except during periods of high discharge, run-off within Coire Gabhail now sinks into the floodplain and emerges below the avalanche debris.

Steep, drift-mantled slopes and talus slopes have been extensively modified during the Holocene by debris flows triggered by high-intensity rainfall events, although other factors such as land-use changes during the last few centuries may have contributed to increased slope-failure susceptibility (Ballantyne 2019a). Hillslope and valley-confined debris flows scar the flanks of many glacial troughs, while repeated debris flows along the same tracks have deposited some of the largest postglacial debris cones in Scotland, notably in Glen Etive and Glen Coe. These two glens display some of the finest examples of both active and fluvially modified debris cones in Britain (Brazier et al. 1988; McEwen 1997; Werritty and McEwen 1997). Often these are fed from deep gullies eroded along porphyritic dykes. They include the Chancellor cone, the largest debris cone in Britain, below Aonach Eagach on the north side of Glen Coe (Fig. 17.9a) and the Eas na Broige cone in Glen Etive. Lichenometric dating suggests increased debris-flow activity during the last ~250 years associated with land-use changes (Innes 1983). However, burial of older debris by later events skews the lichenometric record towards recent activity (Ballantyne 2019a). Debris cones are often polygenetic; for example, that below Coire nan Lochan (Glen Coe) has been modified by mountain torrents and comprises coarse, bouldery deposits with abandoned channels on its surface (McEwen 1997; Fig. 17.9b). At the Eas na Broige cone below Dalness Chasm, exposures have provided evidence for successive phases of Holocene debris-flow and alluvial fan development (Brazier et al. 1988; Werritty and McEwen 1997). The landform comprises a debris cone with an inset alluvial fan. Radiocarbon dating of buried palaeosols indicates an earlier, active phase during the Early-to-Middle Holocene when glacial debris was available for reworking, and a later, recent phase initiated by climatic deterioration or grazing pressure. Debris cone aggradation ceased prior to ~5.2 ka, possibly as a result of sediment exhaustion. This was followed by a prolonged period of relative slope stability and soil development until ~0.6 ka, after which fluvial reworking of debris cone sediments produced the inset alluvial fan. Pollen and charcoal evidence indicates that fluvial incision was in part triggered by burning for arable and pastoral agriculture (Brazier et al. 1988). Evidence of human impacts on the landscape is also present in lake-floor sediments in Loch Etive. Accelerated rates of sedimentation during the last thousand years (Nørgaard-Pedersen et al. 2006; Cundill and Austin 2010) likely record enhanced soil erosion arising from anthropogenic deforestation. This was part of a wider pattern of landscape transformation during the last millennium (Macklin et al. 2000) and in particular during the eighteenth and nineteenth centuries when there was extensive woodland removal to produce charcoal for the Bonawe iron furnace at Taynuilt.

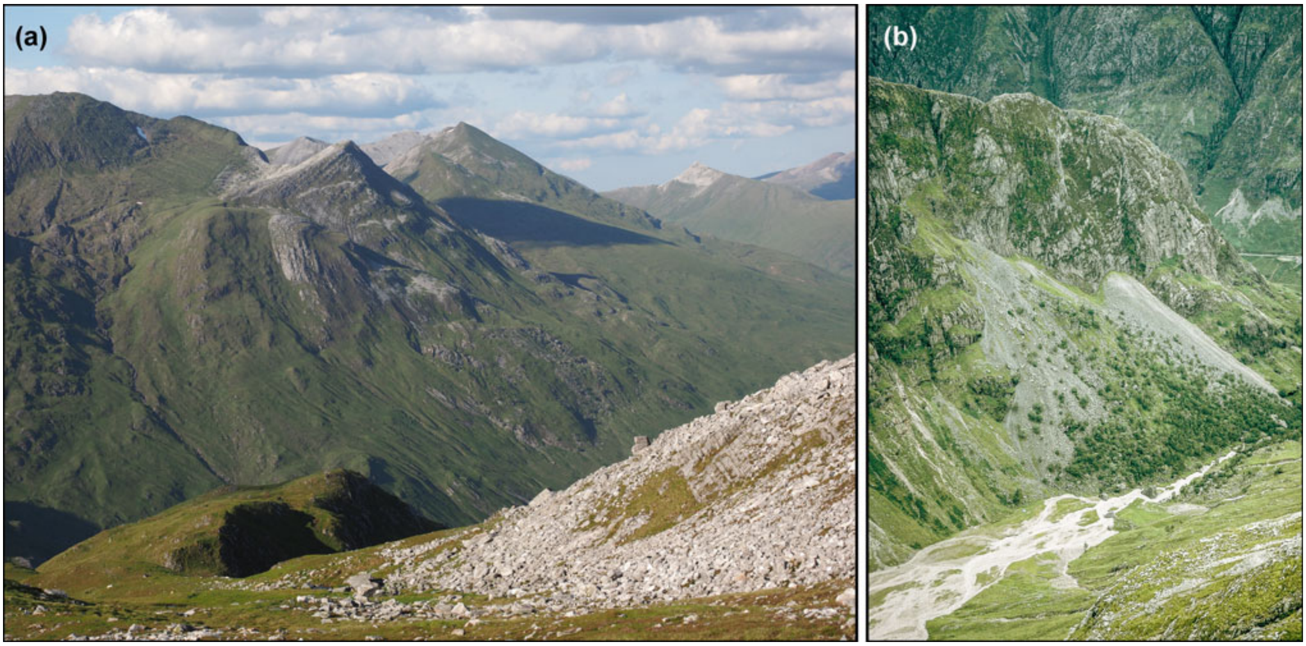

Fig. 17.8 Rock-slope failures. **a** Arrested slide (centre) in folded quartzite bedrock, Sgùrr a' Bhuic, Glen Nevis. **b** Catastrophic rock-slope failure, Coire Gabhail, Glen Coe. The height of the slope from the valley floor to the top of the rock ridge is ~300 m. (Images: **a** Noel Williams; **b** John Gordon)

17.8 Fluvial Landforms

The fluvial geomorphology of the area is characterised by steep mountain torrents in the upper reaches of many catchments and gravel-bed rivers in the lower reaches. Tributaries from higher valleys and cirques form steep alluvial fans at their junctions (e.g. below Stob Coire nam Beithe in Glen Coe). The River Coe provides an excellent example of a high-energy fluvial system with an unusually rapid transition in channel form downstream (McEwen 1994, 1997). The upper reach of the river at the head of Glen Coe is a steep boulder-bed mountain torrent incised into a gorge excavated along porphyritic dykes (Bailey and Maufe 1916; Fig. 17.9b); the River Coe and two tributary streams converge at the 'Meeting of Three Waters', where a NE–SW Caledonian dyke which runs the length of Coire Gabhail intersects a WNE–ESE Carboniferous dyke running along the floor of the main glen. Within a distance of 1.5 km the River Coe transitions to a sinuous, wandering gravel-bed channel with adjacent palaeochannels of different ages, then to a low-angle gravel delta where it enters a small loch (Chap. 4). The post-nineteenth century history of channel adjustment reveals a complex record of braiding and channel shifting across the valley floor (McEwen 1994). This dynamism reflects high discharges, together with sediment inputs from mountain-torrent tributaries, from undercutting of adjacent alluvial fans and debris cones and from paraglacial sediment stores on the floodplain. The River Etive displays a similar change in character from a bedrock river in its upper reaches to a wandering gravel-bed river in its lower reaches (Fig. 17.5a). The Allt Coire Gabhail, a tributary of the River Coe, is unique in Scotland as the location of an alluvial basin that has developed upstream of the rock avalanche that dammed the mouth of the glen (Werritty 1997; Fig. 17.8b).

17.9 Coastal Landforms

The Lateglacial Main Rock Platform (Chap. 4) is well displayed along the fjordic southeast coastline and adjacent islands of Loch Linnhe and the Firth of Lorn. The raised platform and its cliffline fringe much of the Isle of Lismore, carved in predominantly limestone bedrock. Uranium-series disequilibrium dates on speleothems from several of the many caves and undercut notches raise the possibility that the origins of the platform may pre-date the Late Devensian and that it is polycyclic in origin (Gray and Ivanovich 1988), though cosmogenic ^{36}Cl exposure dating of the platform on Lismore is consistent with its formation under periglacial conditions during the Loch Lomond Stade (Stone et al. 1996). An excellent example of a raised sea-arch occurs along a steeply dipping fault at the rear of the platform at Clach Toll on the adjacent mainland.

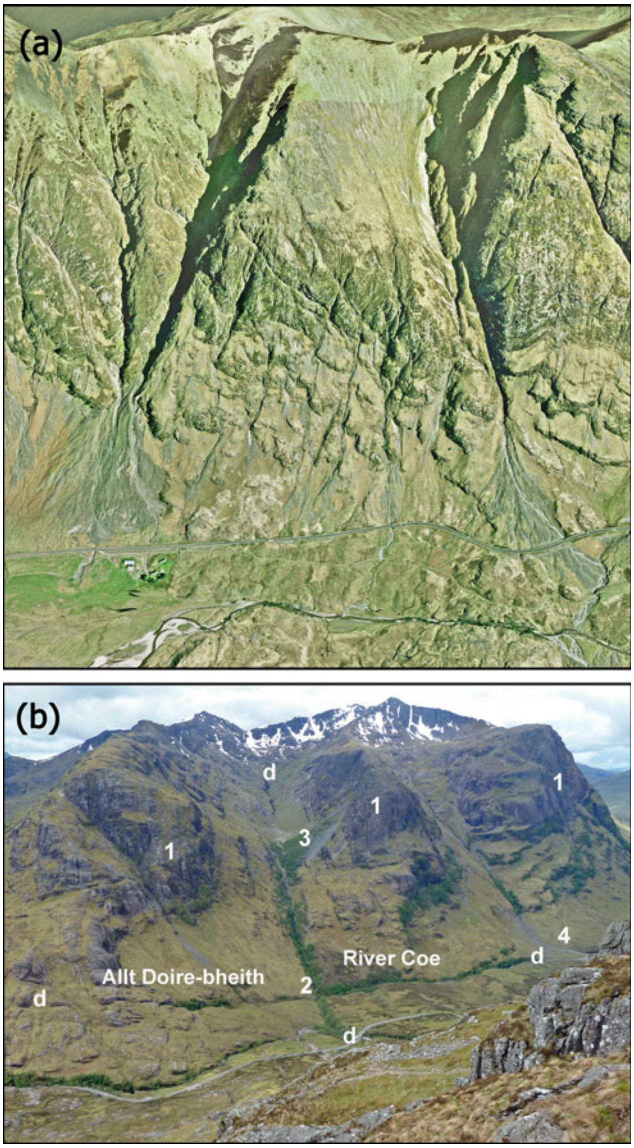

Fig. 17.9 Hillslope and fluvial landforms, Glen Coe. **a** Debris cones below Meall Dearg (953 m) on the Aonach Eagach ridge on the north side of Glen Coe are fed from gullies controlled by porphyritic dykes. The Chancellor cone on the left is named after a prominent rock tower on the ridge. Debris-flow activity periodically blocks the A82 road, and it is recorded that the settlement of Achtriochtan (bottom left) was destroyed by debris flows in the eighteenth century. **b** The Glen Coe glacial trough viewed from the north. 1: Truncated spurs. 2: 'The Meeting of Three Waters' at the intersection of two dykes (d). The middle reach of the River Coe and its headwater, the Allt Doire-bheith, follow a rocky gorge controlled by the ESE-WNW aligned dyke. 3: Catastrophic rock-slope failure, Coire Gabhail. 4: Fluvially modified debris cone below Coire nan Lochan. (Images: **a** Google Earth™ image; **b** Lochaber Geopark)

17.10 Geoheritage and Geoconservation

The natural heritage value of many of the landforms described in this chapter is recognised through a variety of statutory and non-statutory conservation designations. Large parts of Ben Nevis and Glen Coe, together with Moss of Achnacree, South Shian and parts of Rannoch Moor, are protected as Sites of Special Scientific Interest for their geoheritage features. Ben Nevis and Glen Coe are managed by non-governmental organisations, respectively, the John Muir Trust and National Trust for Scotland, which promote landscape conservation, public understanding and enjoyment. Geology and geomorphology underpin the landscape qualities of the Ben Nevis and Glen Coe National Scenic Area. Ben Nevis, Glen Coe, Rannoch Moor and Glen Etive are incorporated within Lochaber Geopark which, *inter alia*, promotes geoconservation, sustainable development, geotourism and educational activities in partnership with local communities. The Geopark has played a significant part in raising wider awareness of the links between geology, geomorphology and landscape through publications, educational activities and public engagement.

The connections between geology, landscape and culture have formed the basis of visitors' appreciation of the West Highland landscape since the latter part of the eighteenth century and were popularised in the journals of early travellers such as Thomas Pennant, Dorothy Wordsworth and Samuel Taylor Coleridge (Crocket 1986). Today, the area is an attractive visitor destination for its scenic qualities and opportunities for outdoor recreation in a variety of landscape settings shaped by the geology and geomorphology. The mountain ridges and rock faces of Ben Nevis and Glen Coe are renowned for their mountaineering and hillwalking challenges, both in summer and winter, and downhill skiing facilities are located on Aonach Mòr and on the slopes of Meall a' Bhùiridh above Rannoch Moor. Glen Coe is a major stopping point for visitors to view the scenic qualities of the huge rock buttresses and truncated spurs formed by glacial erosion. It is has been widely celebrated by artists for its 'sublime' scenery (e.g. in paintings by Horatio MacCulloch and J.W.M. Turner), as well as for its cultural heritage and associations with Celtic mythology, its notorious place in Scottish clan history and as a film-set location (e.g. see https://www.nts.org.uk/visit/places/glencoe).

17.11 Conclusions

The Western Grampians bear a strong imprint of the Caledonian Orogeny and post-orogenic volcanism in terms of lithological and structural influences on the form of the landscape and topographic features at different scales. Rannoch Moor and the adjacent mountains formed the principal ice-accumulation centre in Scotland for repeated ice-cap and ice-sheet growth and dispersal during the Pleistocene. The landscape displays many classic landforms of glacial erosion and excellent examples of LLR outwash plains and hummocky recessional moraines, the latter indicating active glacier recession. Topographic pinning points played an important part in determining the style of glacier retreat within the sea lochs and controlled the positions of large outwash accumulations. The area is particularly important in the context of dating the culmination of the LLR and the final disappearance of the ice, currently a matter of debate, which has a significant bearing on understanding the drivers of abrupt climate change and the response of terrestrial glaciers during the Younger Dryas in the North Atlantic region. The array of rock-slope failures and Holocene debris flows and debris cones is also outstanding. The geodiversity of the region is reflected in numerous conservation designations and the inclusion of a large part of the area within Lochaber Geopark.

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18 The Cairngorm Mountains

John E. Gordon and Vanessa Brazier

Abstract

The Cairngorm Mountains represent a classic landscape of selective linear glacial erosion formed in a largely granite massif. Relict, non-glacial landforms include palaeosurfaces with tors, weathered rock and broadly convex summits that have generally been little modified by glacial erosion. These features contrast sharply with glacial troughs and breached watersheds that are deeply incised into the massif. This remarkable geomorphological diversity is enhanced by the presence of cirques, periglacial landforms on the upper slopes and plateaux, ice-marginal deglacial landforms in the valleys and on the lower slopes, paraglacial reworking of debris slopes and glacial deposits, and dynamic river systems. The landforms and geomorphological processes of the Cairngorms provide the foundations for internationally important biodiversity and form a highly sensitive geo-ecological system maintained by the interactions of geology, geomorphology, soils, vegetation and climate.

Keywords

Selective linear glacial erosion • Palaeosurfaces • Tors • Glacial troughs • Ice-marginal landsystem • Periglacial landforms • Debris flows • Gravel-bed rivers

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18.1 Introduction

The Cairngorm Mountains, in the NE Grampian Highlands, comprise a broad, dome-shaped, predominantly granite massif ~30 km by 20 km in extent (Fig. 18.1), including the largest expanses of high ground above 1000 m in Britain and with several summits above 1200 m in altitude. With their expansive stony ground, late-lying snowbeds, montane flora and severe weather, the high plateaux are comparable in character to sub-arctic or montane fell-field landscapes. Geomorphologically, the Cairngorms represent one of the most distinctive mountain landscapes in Britain, renowned for an exceptional diversity of landforms that provides insights into granite mountain evolution and environmental change over the last ~400 Ma under a range of tropical, temperate-humid, glacial and temperate climates. In particular, the Cairngorms support the finest assemblage of glaciated granite mountain landforms in Britain, noted for their distinctive palaeosurfaces, broadly convex summits, tors, weathered bedrock and glacially sculptured glens (valleys) and corries (cirques) that collectively constitute a classic landscape of selective linear glacial erosion (Sugden 1968; Rea 1998). The adjacent slopes and glens support a diverse assemblage of periglacial landforms, glacial meltwater features, glacial deposits, rock-slope failures, debris flows and related features, river terraces and wandering gravel-bed rivers (Gordon 1993; Brown and Clapperton 2002; Gordon and Wignall 2006; Kirkbride and Gordon 2010). The geodiversity of the Cairngorms, in conjunction with a montane climate that combines oceanic and continental influences, underpins internationally important biodiversity, evident both in the spatial and altitudinal variations in habitats and species and in the fine-scale mosaics of plant communities (Gimmingham 2002; Shaw and Thompson 2006).

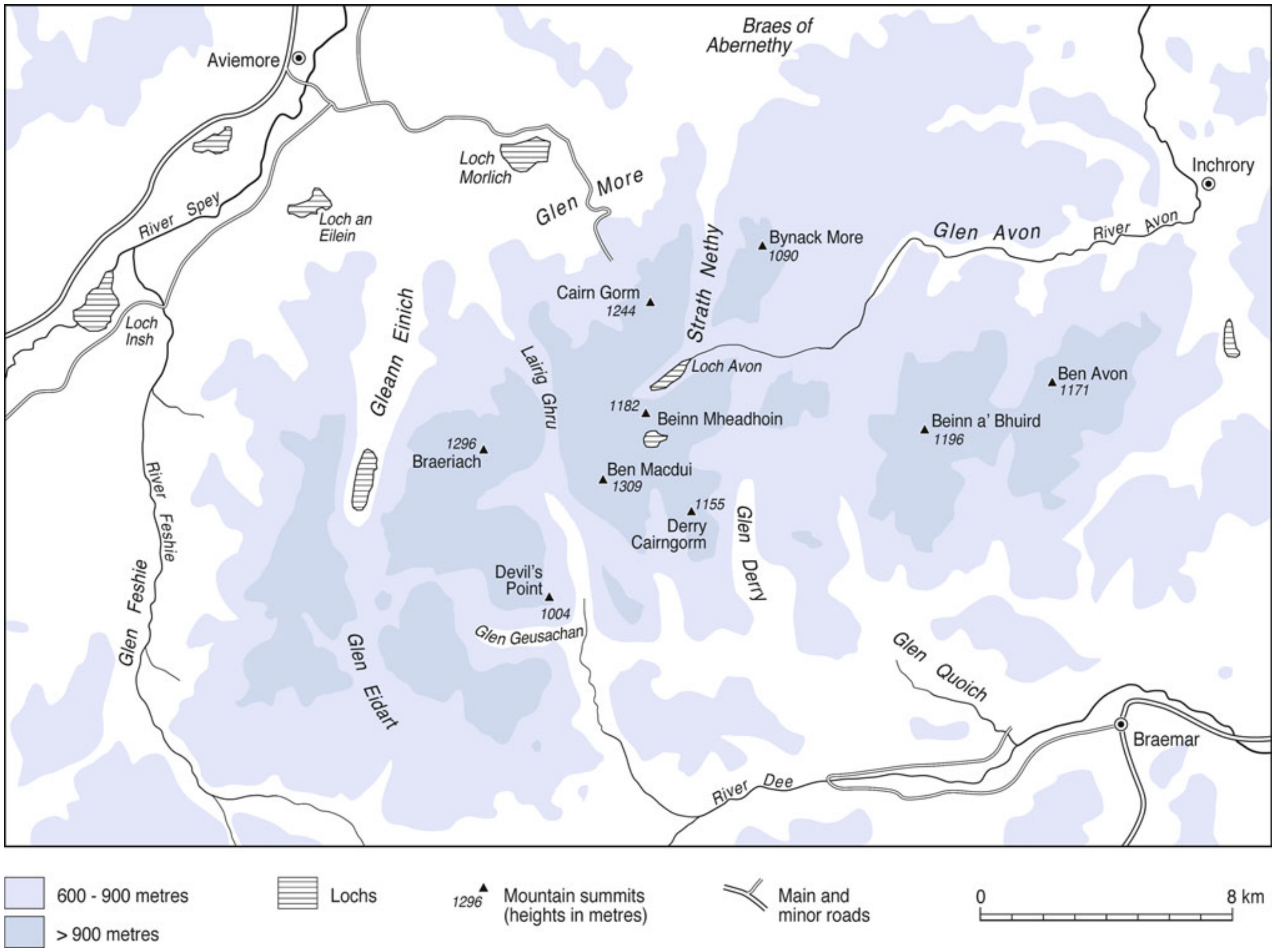

Fig. 18.1 Cairngorm Mountains: relief and main locations mentioned in the text

18.2 Geology and Landscape Evolution

The Cairngorm Mountains form a predominantly granite massif surrounded by metasedimentary rocks. The geomorphology of the massif strongly reflects the form and characteristics of the underlying Cairngorm Granite and the processes that have shaped the landscape during the last ~400 Ma. The traditional Gaelic name of the area, Am Monadh Ruadh ('The Red Hill-range'), derives from the colour of the granite.

The Cairngorm Granite pluton was intruded into rocks of the Neoproterozoic Dalradian Supergroup, which consists of predominantly metamorphosed and structurally deformed marine sediments and igneous rocks. The Dalradian sediments were deposited between ~700 and 600 Ma and were subsequently deformed and metamorphosed during the Grampian event of the Caledonian Orogeny at ~475 Ma (Chap. 2). The uplift of an alpine-scale mountain belt, the Caledonides, extending from the Appalachians to northern Scandinavia, was accompanied by the emplacement of syn-, late- and post-tectonic intrusions of siliceous to ultrabasic igneous rocks. The Cairngorm pluton (~365 km^2) is the largest of a number of siliceous plutons that lie along the line of an ancient lineament followed today by the Dee valley. It was emplaced at ~427 Ma (Thomas et al. 2004).

The Cairngorm Granite pluton was unroofed rapidly in the Early Devonian and may have remained as an area of positive relief until the present without being covered by younger rocks. The present land surface lies within the original roof zone of the pluton (Thomas et al. 2004). The form of the pluton and the presence of lines of weakness have strongly influenced the patterns of subsequent weathering, erosion and landscape evolution. At a macro-scale, the overall size and dome-like form of the mountain massif reflect the general form of the pluton. The major glens are aligned along sets of quartz veins and linear zones of hydrothermal alteration and hydro-fracturing, where hot fluids circulated through fractures as the granite solidified (Thomas et al. 2004; Hall and Gillespie 2017). These altered zones, up to 200 m wide and several kilometres long, and the fractured rock along faults, provided lines of weakness that were selectively exploited by weathering and river erosion over many millions of years after the granite was unroofed to form the precursors of the present valleys. The valley pattern today essentially represents 'a Devonian drainage system that was subsequently modified through capture by eastwards-flowing rivers in the Palaeogene and by glacial diversion during the Pleistocene' (Hall and Gillespie 2017, p. 2215). The persistence of the main valley pattern for ~400 Ma reflects the geological controls and the relatively low rates of denudation (probably less than 500 m) since the Devonian (Macdonald et al. 2007). At a more local scale, variations in the type and density of jointing in the granite have affected differential weathering and erosion of the bedrock (see below).

By the end of the Cretaceous, much of Scotland had been reduced by erosion to a landscape of low positive relief near to sea level. During the early Palaeogene, uplift of the Scottish Highlands was associated with the opening of the North Atlantic Ocean and further uplift continued intermittently during the Neogene (Hall and Bishop 2002; Hall et al. 2019; Chap. 3). Regional uplift in the Cairngorms was of the order of 1 km and was accompanied by tilting of the Grampian Highlands towards the east. During the Palaeogene and Neogene, the landscape evolved under subtropical then warm-temperate humid climates, favouring chemical weathering of the bedrock, the formation of deep saprolite cover and accelerated erosion of the weathered material during periods of more extreme climate and as the land was uplifted. Pulsed uplift, weathering and erosion produced a series of tilted palaeosurfaces rising in steps inland from the coast of NE Scotland and culminating in the high plateaux and gently rolling summits of the Cairngorms (Hall 1991; Ringrose and Migoń 1997). Cooling of the global climate during the later Palaeogene and throughout the Neogene led to the onset of glaciation in Scotland at the start of the Quaternary at 2.59 Ma, by which time the broad outlines of the present landscape were essentially in place.

18.3 Relict Non-Glacial Landforms and Landscapes

Although successive ice sheets covered the Cairngorms during the Pleistocene, some relict non-glacial landforms have persisted with only limited glacial modification, thus illustrating the longer-term processes of mountain landscape evolution in the Scottish Highlands. The selective nature of glacial erosion has ensured that many plateau surfaces and landforms show little or no signs of glacial modification, enabling reconstructions of what was considered to be the pre-glacial relief (Sugden 1968; Thomas et al. 2004). Although the plateau surfaces have often been interpreted as representing relatively unmodified pre-glacial landscape elements, subsequent research has demonstrated that they have continued to evolve through weathering, erosion and surface lowering during the Quaternary (Hall 1996; Hall and Glasser 2003; Phillips et al. 2006; Goodfellow 2007; Goodfellow et al. 2014).

18.3.1 Palaeosurfaces

The Cairngorms comprise the highest in a series of palaeosurfaces that dominate the macro-scale geomorphology of NE Scotland (Sugden 1968; Hall 1991; Figs. 18.2 and 18.3a). The Summit Surface of the Cairngorms (~1070–1220 m) is separated from the Eastern Grampian Surface (~760 m) by a well-defined break of slope. The Eastern Grampian Surface is assigned to the late Palaeogene and cuts across a range of lithologies over a wide area of the Eastern Grampian Highlands (Chap. 20). These surfaces are believed to have formed by processes of dynamic etchplanation, involving deep weathering, slope retreat and fluvial erosion controlled by variations in rock resistance (Hall 1991; Thomas 1994; Hall et al. 2013).

The rolling topography of the Summit Surface of the Cairngorms has a relative relief of up to 200 m. It comprises broadly convex summits, often surmounted by tors, and shallow headwater stream valleys (Fig. 18.3b, c). The latter are aligned along alteration zones within the granite and have been little modified by glacial erosion (Hall and Gillespie 2017). The domed summits follow curvilinear sheet joints associated with the unloading of the granite as the pluton was unroofed. These sheet joints are sometimes clearly truncated at plateau edges along the rims of cirques and glacial troughs, indicating their relict, non-glacial origin (Sugden 1968; Goodfellow et al. 2014). A second set of steeply dipping sheet joints subsequently developed parallel to the flanks of the glacial troughs (Glasser 1997). Along the margins of the massif, valley benches at 600–700 m OD represent remnants of pre-glacial valleys later deepened by glacial erosion (Hall et al. 2013; Fig. 18.2).

18.3.2 Topographic Basins

Topographic basins are a feature of the macro-scale geomorphology of the Scottish Highlands (Hall et al. 2019). In the Cairngorms, the floors of the high-level topographic basins of the Mòine Mhòr and Mòine Bhealaidh occur at 920–950 m OD and 850–870 m OD, respectively, while low-level basins occur to the north of the massif in Glen More and Strath Nethy at 240–320 m OD (Fig. 18.2). The formation of both sets of basins is the result of long-term

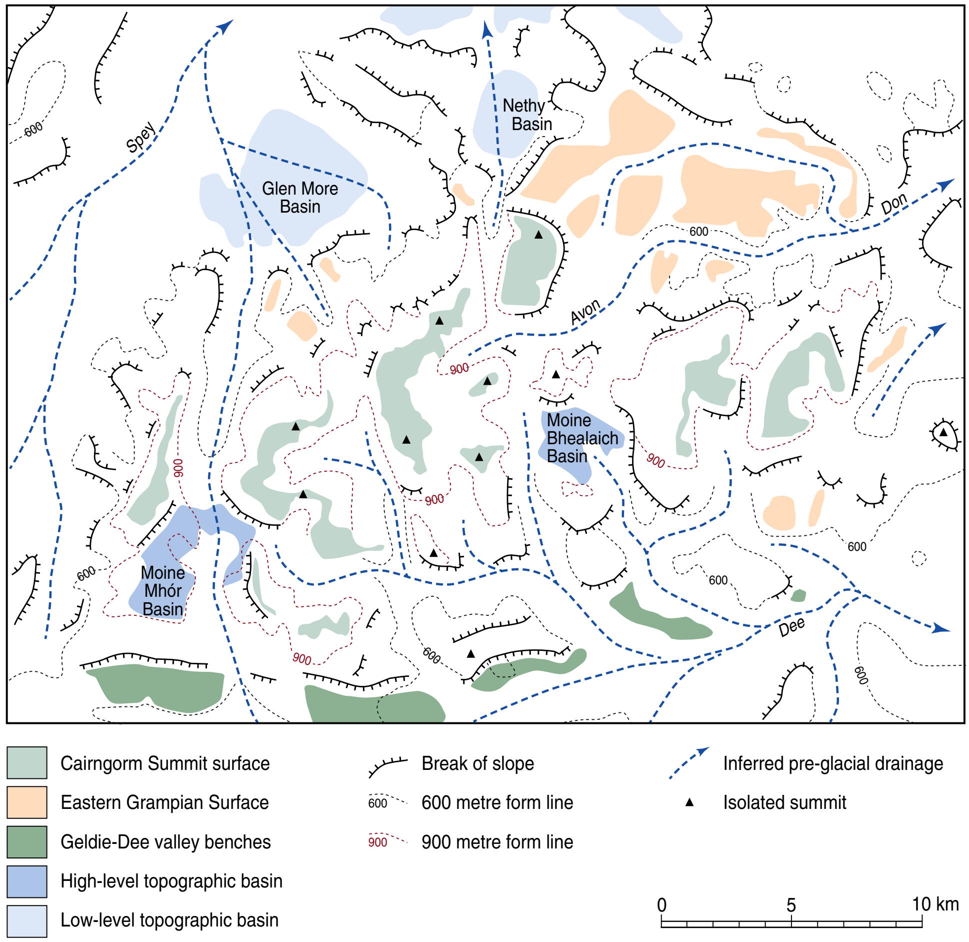

Fig. 18.2 Pre-glacial geomorphology of the Cairngorm Mountains. (Adapted from Hall AM, Gillespie MR (2016) Int J Earth Sci 106:2203–2219 © the authors)

differential weathering and erosion, the underlying psammites and other rocks having lower resistance to chemical weathering than the granite (Hall et al. 2013).

18.3.3 Weathered Granite

Pockets of weathered granite are exposed in the headwalls of several cirques and along headwater streams and lines of weakness determined by hydrothermal alteration (Hall 1996). The weathered rock is generally shallow (<2.5 m deep), with a limited degree of chemical alteration comparable with that of the geochemically immature, grus-type weathering profiles in the lowland granites of NE Scotland, which are thought to have formed under humid-temperate conditions after the end of the Miocene (Hall 1986; Hall et al. 1989; Chap. 21). Such weathering has been considered to represent the basal remnants of formerly thicker, relict pre-glacial weathering profiles (Linton 1955; Sugden 1968). However, it is probable that several styles and generations of alteration have been superimposed, including hydrothermal alteration during late stages of granite cooling, deep chemical weathering during the Palaeogene and Neogene and shallow weathering during temperate Quaternary interglacials (Hall 1996; Hall et al. 2019). The limited chemical alteration of primary minerals in the weathered granite suggests that little, if any, of the present plateau regolith predates the Pleistocene (Hall et al. 2013).

18.3.4 Tors

Granite tors are a distinctive feature of the Cairngorm Mountains, occurring above ~600 m OD on ridge crests and summits (Fig. 18.3b, c). The largest and most impressive forms are located on the eastern summits and upper slopes of Bynack More, Beinn Mheadhoin and Ben Avon, where they rise up to ~25 m above the adjacent ground (Ballantyne 1994; Phillips et al. 2006; Hall and Phillips 2006a, b). According to Phillips et al. (2006, p. 225), the 'Cairngorms contain perhaps the best example of a glaciated tor field in the world, with 38 groups of tors'. Differential weathering modulated by varying densities of jointing in the granite has produced a variety of tor forms. These range from massive, blocky monoliths up to 24 m high to stacks of thin, sheet-jointed layers of rock, the sheet jointing usually

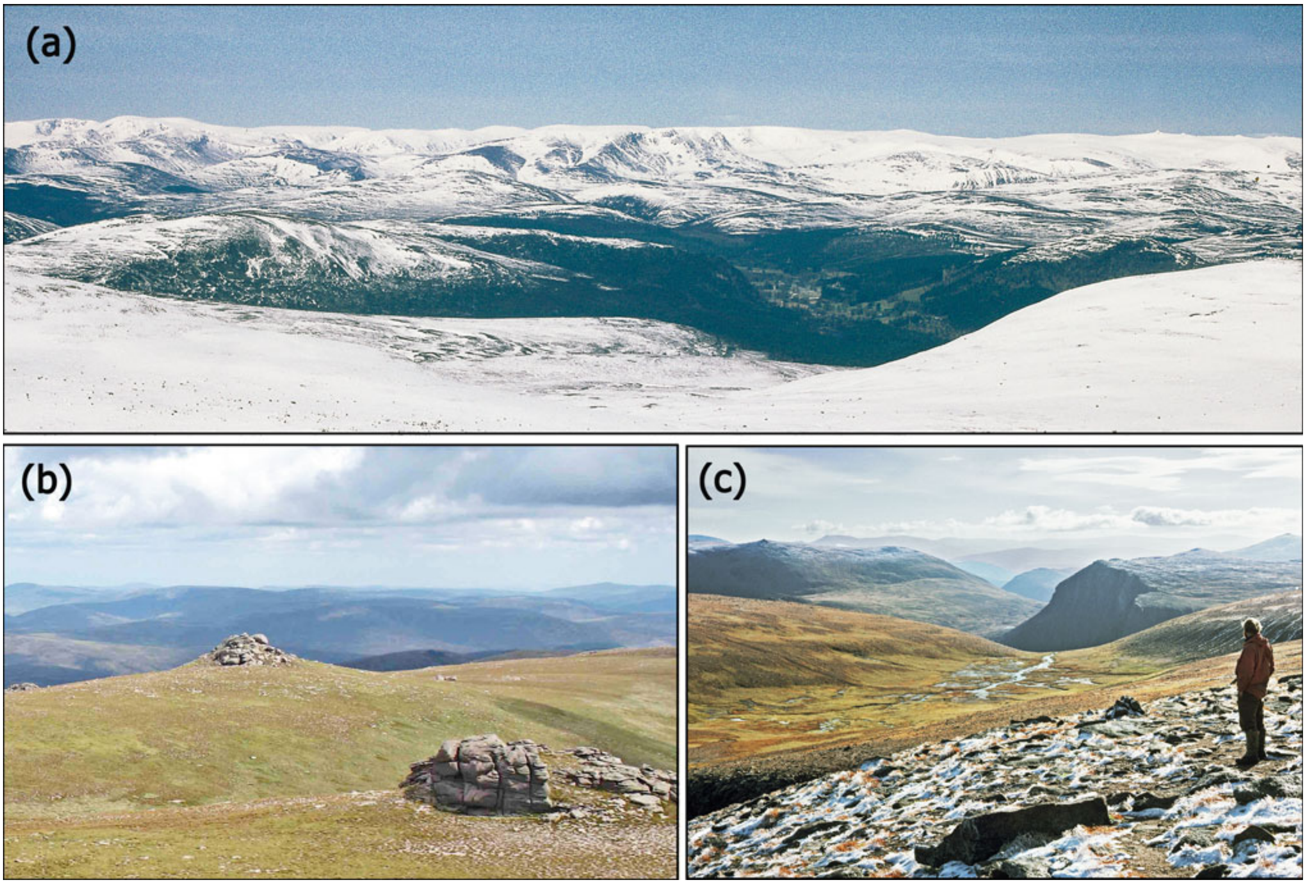

Fig. 18.3 **a** Cairngorm Summit palaeosurface viewed from the southeast, with the tor-capped summit of Ben Avon on the right. The Dee glacial trough (centre) is incised into a lower pre-glacial surface separated from the summit surface by a clear break of slope. **b** The summit plateau of Ben Avon comprises a series of broadly convex summits with tors and intervening headwater valleys. Blockfield aprons surround the tors. **c** The shallow headwater valley of Coire Domhain southwest of Cairn Gorm summit is truncated by the Loch Avon glacial trough. (Images: **a**, **c** John Gordon; **b** Colin Ballantyne)

being aligned parallel to the adjacent plateau surface (Goodfellow et al. 2014; Fig. 18.3b). Many of the tors, particularly those in the eastern Cairngorms that show little or no glacial modification, display well-developed weathering pits exceeding 1 m in depth, indicative of a considerable age of surface exposure to weathering (Hall and Phillips 2006a). In contrast, others have been glacially modified to varying degrees indicating the former presence of ice cover on the plateaux (Hall and Phillips 2006b).

The origin of the Cairngorms tors is related to the density of vertical jointing in the granite: they occur in areas of relatively coarse-grained granite where vertical joints are more widely spaced than in the surrounding rock (Goodfellow et al. 2014). They were formerly interpreted as relics of a deeply weathered, pre-glacial (Neogene) landscape (Linton 1955; Sugden 1968), but terrestrial cosmogenic nuclide (TCN) exposure dating now indicates that the oldest surviving features were first exhumed during the Middle Pleistocene (Phillips et al. 2006). They have evolved through repeated shallow chemical weathering of the granite during milder phases early in the Pleistocene, combined with recurrent phases of removal of the weathered material, leaving the present forms as upstanding bodies of locally more resistant rock that progressively emerged from relatively shallow regolith (a metre or two thick) during and after the Middle Pleistocene (Hall and Phillips 2006b). Therefore, although not pre-glacial landforms, their survival through several glacial cycles indicates that the summit plateaux were, in general, not heavily eroded by the Pleistocene glaciers.

The Cairngorms plateaux are extensively covered in regolith that comprises granite blocks embedded in a sandy matrix, and several of the plateau summits support openwork blockfields, notably Ben Macdui and Derry Cairngorm, while blockfield aprons occur around many of the tors (Fig. 18.3b). Some of the blocks may have originated as chemically weathered corestones from disintegrating tors, but breakdown of jointed rock by frost wedging seems more likely (Ballantyne 1996). The present regolith cover probably developed prior to the last glacial stage and survived, largely intact, under a cover of cold-based glacier ice (Hopkinson and Ballantyne 2014), with limited modification by frost weathering and wind erosion since the disappearance of the last plateau ice caps.

18.4 Glacial Landforms and Landscapes

Palaeoclimatic records from deep-sea sediments suggest that the Scottish Highlands were affected by numerous glacial cycles during the Pleistocene, so that the larger features of glacial erosion must have evolved through multiple episodes and different styles of glaciation over prolonged periods (Phillips et al. 2006; Hall et al. 2019; Chap. 4). Throughout the Pleistocene, local mountain glaciers would have been frequently present in the cirques and upper glens, with cold-based icefields on adjacent plateaux, while large ice sheets periodically covered the entire landscape during the Middle and Late Pleistocene (Hall et al. 2019). The onset of glaciation in the Cairngorms probably occurred near the start of the Quaternary at 2.59 Ma, as indicated by the presence of glacial indicators in the North Atlantic Ocean and evidence for grounded ice sheets in the northern and central North Sea Basin (Thierens et al. 2012; Rea et al. 2018). The last Scottish Ice Sheet expanded during the Late Devensian sometime after ~35 ka and at its maximum extent covered all of Scotland, including the Cairngorms, as indicated by glacially modified tors and the presence of granite erratics on the high plateaux (Phillips et al. 2006). The absence of far-travelled allochthonous erratics within the massif and the confinement of metasedimentary erratics to flanking slopes below ~800 m OD indicate that throughout the lifetime of the last ice sheet the Cairngorms formed an independent centre of ice dispersal that merged with externally sourced ice and diverted ice flow northeastwards along Strathspey and eastwards along the Dee valley (Sugden 1970; Hall et al. 2016). TCN exposure dating indicates substantial recession of the ice sheet before the rapid climate warming at the beginning of the Lateglacial Interstade at ~14.7 ka (Ballantyne and Small 2019) and that externally sourced ice retreated across the northern flanks of the Cairngorms within the period 16.5–15.5 ka (TCN exposure ages from Everest and Kubik (2006) and Hall et al. (2016) recalibrated by Ballantyne and Small (2019)). During the Loch Lomond (≈Younger Dryas) Stade of ~ 12.9–11.7 ka, small glaciers and plateau ice caps formed in the Cairngorms (Sissons 1979; Standell 2014; Bickerdike et al. 2018) and periglacial processes modified the adjacent slopes.

18.4.1 Landforms and Landscapes of Glacial Erosion

Large features of glacial erosion in the Cairngorms, such as glacial troughs, glacially breached watersheds and cirques, represent the cumulative imprint of successive episodes of mountain and ice-sheet glaciation throughout the Pleistocene (Fig. 18.4). These landforms are closely juxtaposed with the relict, non-glacial landforms and collectively form a classic landscape of selective glacial erosion (Sugden 1968; Rea 1998), which displays similarities with parts of northern Scandinavia, Arctic Canada, Greenland and Antarctica (Goodfellow et al. 2008; Jamieson et al. 2010). The contrast between the preservation of the rolling plateau landscape, with tors and pockets of weathered granite, and the deeply eroded, steep-sided glens is striking. It is explained by the form of the pre-glacial topography, which is related to the characteristics and linear weathering zones of the granite bedrock, and the basal thermal regime of former ice masses (Hall and Glasser 2003). Glacier ice occupying the plateau during successive glacial periods appears to have been cold-based, slow-moving and frozen to the underlying substrate, enabling the survival or limited modification of relict, non-glacial features. In contrast, convergent flow of ice into already deeply incised fluvial valleys, formed during pre-glacial times, reached sufficient thickness and velocity to initiate basal melting and to erode the underlying hydrothermally weakened rocks (Hall and Glasser 2003).

A prominent feature of the Cairngorms is the presence of glacial troughs cut deeply into the massif, notably Glen Avon, Gleann Einich and Glen Geusachan, all of which terminate abruptly upvalley at steep, cliffed headwalls (Figs. 18.4 and 18.5a). These troughs are thought to follow the lines of pre-glacial valleys, which in turn follow linear zones of hydrothermal alteration of the granite. The latter are aligned along lines of structural weakness conditioned by regional stresses during the Caledonian Orogeny (Thomas et al. 2004). In places, the glaciers also carved through the pre-existing watersheds to form spectacular glacial breaches, notably the Lairig Ghru, Pass of Ryvoan, upper Glen Feshie, Inchrory and the Lairig an Laoigh (Linton 1951). Some of these breached watersheds may have been cut by ice streams moving in an east–northeast direction during phases of ice-sheet glaciation (Sugden 1968). Others, such as the Lairig Ghru and Lairig an Laoigh (Fig. 18.4), lie athwart the direction of regional ice flow and may in part have been cut or enlarged by headward erosion by local mountain glaciers or by local ice spilling northwards when the outflow of ice to

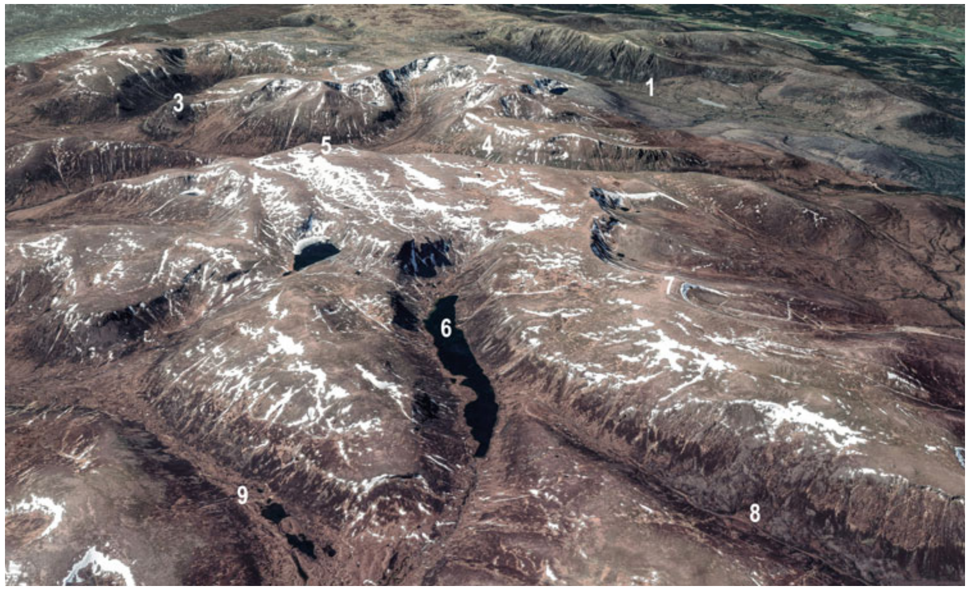

Fig. 18.4 Central Cairngorms viewed from the east, showing glacial troughs, cirques and glacially breached watersheds incised into the Cairngorm Summit palaeosurface. 1: Gleann Einich. 2: Braeriach. 3: Devil's Point. 4: Lairig Ghru. 5: Ben Macdui. 6: Loch Avon. 7: Cairn Gorm. 8: Strath Nethy. 9: Lairig an Laoigh. (Google Earth™ image)

the south was blocked by ice along the valley of the Dee. The high rock-mass strength of the Cairngorm Granite probably explains the relatively deep, narrow form of the glacial troughs and glacial breaches (Brook et al. 2004). As a consequence of the watershed breaching, the headwaters of the River Feshie, River Avon and Water of Caiplich were diverted from their pre-glacial to their present courses (Linton 1949, 1954; Figs. 18.1 and 18.2).

Glacial erosion had a variable but limited impact on the plateau surfaces (Hall and Glasser 2003). Areas of ice-scoured bedrock occur near the heads of some glacial troughs, such as Gleann Einich and Glen Avon, where there was convergent flow of ice from plateau ice caps into the troughs. Conversely, the survival of tors, particularly on the more easterly plateaux of the Cairngorms, suggests that in these areas glacier ice has not significantly modified the landscape. Some tors, such as those on Ben Avon, exhibit little or no evidence of glacial modification; others, particularly in the western hills and on Cairn Gorm, have been moulded by glacier ice; and some have been partially or totally destroyed, producing more subdued forms and, in extreme cases, stumps remodelled into roches moutonnées (Hall and Glasser 2003; Hall and Phillips 2006b; Phillips et al. 2006). Toppled tors and trains of boulders on the lee sides of some tors (Fig. 18.5b) and on blockfields also indicate modification by glacier ice. On Ben Avon, such modification occurred during at least the last three major glacial stages. Nevertheless, the survival of many tors and the absence of extensive areas of glacial erosional features strongly indicate the limited role of glaciation in shaping the plateaux, particularly on the easternmost hills. The average depth of Quaternary glacial erosion on the plateaux, through variable modification of tors and local removal of regolith, is estimated as < 5 m, and rates of non-glacial erosion during interglacials may have been greater than those of glacial erosion during glacial stages (Hall and Phillips 2006b).

Numerous cirques were excavated along plateaux margins and at trough heads (Figs. 18.4 and 18.5c) principally by mountain glaciers, although they may also have continued to evolve under ice-sheet glaciation. The cirques occupy mainly north- and east-facing slopes on the lee sides (in relation to prevailing winds) of the high plateaux, where small glaciers formed in shaded sites (pre-glacial valley heads) that received wind-blown snow from the adjacent plateau during the course of multiple glacial cycles (Sugden 1969; Gordon 2001; Thomas et al. 2004). Most are arcuate in planform, reflecting the relatively uniform bedrock. However, structural controls have locally influenced cirque long profiles (Haynes 1968). Some cirques have multiple basins, possibly in response to changing sources of wind-blown snow on the plateaux, and others appear to have been over-ridden by ice from either plateau icefields or ice sheets (Sugden 1969), which may account for some degree of asymmetry and degraded cirque forms in the Glen Feshie Hills and along the eastern flank of upper Gleann Einich. Cross-cutting relationships suggest that some cirques post-date the excavation of glacial troughs.

Other features of glacial erosion include truncated spurs, such as the Devil's Point, where pre-glacial valleys were straightened during recurrent episodes of glacial erosion. Large roches moutonnées, up to 200 m high, are present in Strathspey at Ord Ban, north of Loch an Eilein, and Farleitter Crag, south of Loch Insh, and along both sides of the Dee valley between Glen Derry and east of Braemar (Sugden et al. 1992; Glasser 2002).

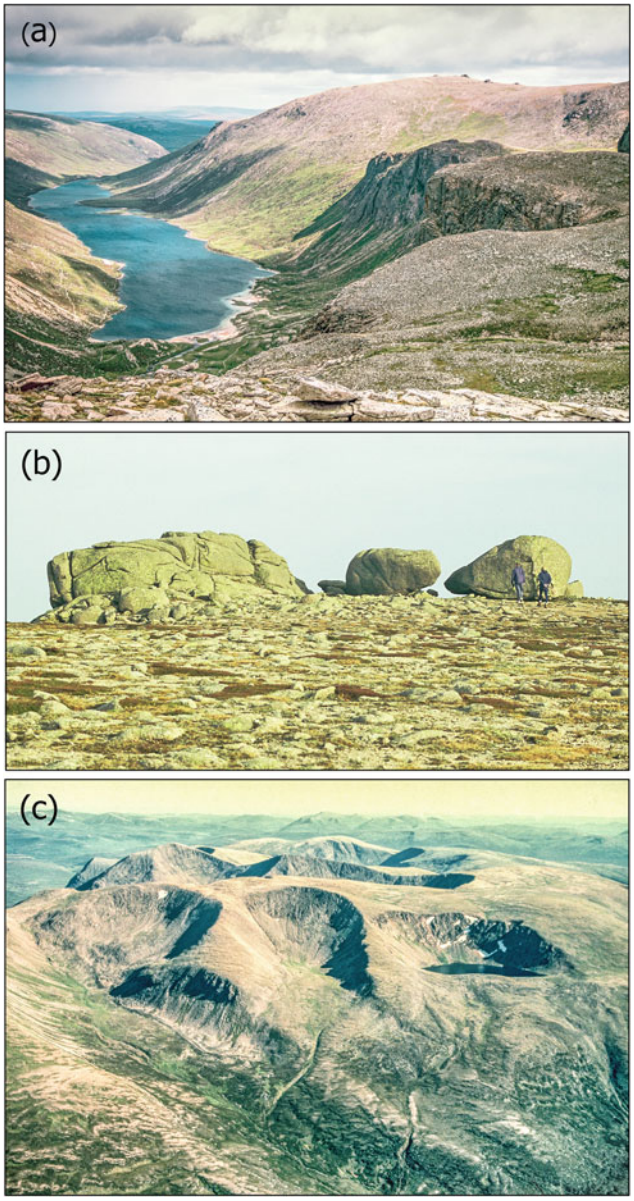

Fig. 18.5 **a** Head of the Loch Avon glacial trough viewed towards the tor-capped summit of Beinn Mheadhoin. Note the hummocky moraine at the head of the loch. **b** Glacially displaced tor blocks on the summit plateau of Ben Avon. **c** Aerial view looking south towards Braeriach (centre) across the Cairngorm Summit palaeosurface; several cirques scallop the plateau edge. (Images: **a**, **b** John Gordon; **c** © Jane Poole/Idris Consulting Ltd)

18.4.2 Landforms and Landscapes of Glacial Deposition

As the last Scottish Ice Sheet expanded, glaciers from the Cairngorms merged with ice flowing northeastwards from the Western Grampians along Strathspey and into Glenmore and upper Deeside (Merritt et al. 2019). Thinning and retreat of the ice sheet between ~18 ka and ~15 ka were marked by the deposition of assemblages of glacifluvial and glacial landforms that record active retreat of the ice. These features are exceptionally well developed along the northern flanks and lower slopes of the Cairngorms from lower Glen Feshie northeastwards through Glenmore to the Braes of Abernethy, delineating an outstanding example of an ice-marginal landsystem (Sugden 1970; Young 1974, 1975; Brazier et al. 1998; Standell 2014; Hall et al. 2016; Fig. 18.6). The main features include lateral moraines, meltwater channels and kame terraces that cut across the grain of the topography or run parallel to the contours. They indicate that a significant advance of Spey ice interrupted the recession of the last ice sheet ~16 ka (Hall et al. 2016). Large meltwater channels cut into bedrock now occur as dry valleys, including those at Creag a' Chalamain and Airgiod-meall in upper Glenmore (Fig. 18.6), and probably reflect repeated erosion by meltwaters over several glacial cycles. Glaciers retreating back into the mountains became decoupled from the Spey ice lobe in Glenmore, which ponded ice-dammed lakes in the lower parts of Gleann Einich, the Lairig Ghru and Strath Nethy. The deltas formed by glacial meltwater rivers flowing into some of these lakes remain as distinctive flat-topped terraces up to 35 m high (Brazier et al. 1998; Fig. 18.7a). Small lakes were also ponded against the receding ice margin to the northeast in Abernethy Forest (Hall et al. 2016). Radiocarbon dates from Loch Etteridge and TCN exposure ages from Glen Banchor in the Monadhliath Mountains to the west indicate retreat of Spey ice from the margins of the Cairngorms by ~15.5 ka (Ballantyne and Small 2019).

The final phase of ice-sheet deglaciation was marked by extensive stagnation of the retreating ice margin. This is evident in the increased topographic conformity of meltwater channels and eskers (e.g. in lower Glen Feshie), the

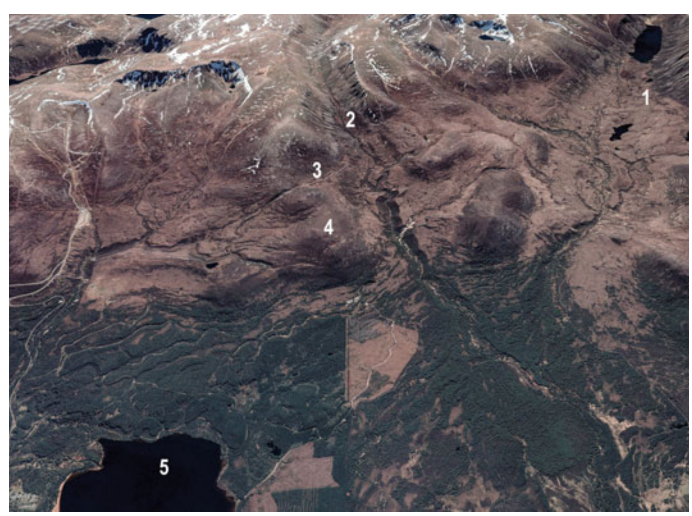

Fig. 18.6 Ice-marginal landsystem on the northern flanks of the Cairngorms. 1: Gleann Einich. 2: Lairig Ghru. 3: Creag a' Chalamain. 4: Airgiod-meall. 5: Loch Morlich. (Google Earth™ image)

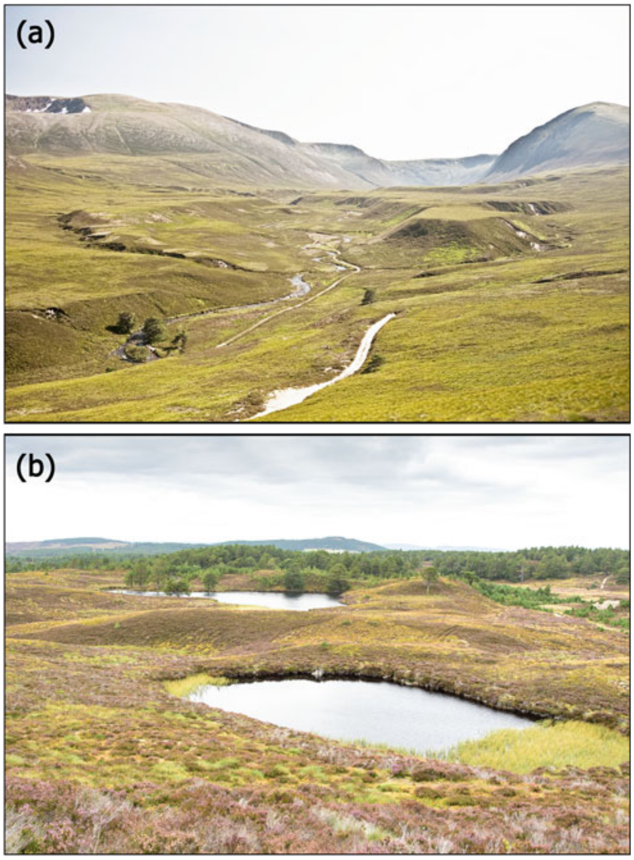

Fig. 18.7 **a** Delta terraces in Gleann Einich, formed in an ice-dammed lake and later dissected by river erosion. **b** Kame and kettle topography northeast of Strath Nethy. (Images: John Gordon)

presence of kame and kettle topography indicative of stagnant ice (e.g. at the southwest margin of Loch Morlich, at Achlean in Glen Feshie and in Glen Quoich) and extensive suites of terraced outwash deposits (e.g. throughout the middle and lower parts of Glen Feshie), some pitted with kettle holes (Young 1974, 1975, 1976). Kettle holes also developed where large masses of ice became stranded and slowly melted out, as at Loch Morlich, Uath Lochs and northeast of Strath Nethy (Fig. 18.7b).

The glacial landforms in the valleys and on the lower hillsides bordering the River Dee on the south and east side of the Cairngorms record a similar pattern of ice-sheet wastage (Kirkbride and Gordon 2010). Initially, local Cairngorm ice was confluent with the Dee ice mass, with regional flow in an eastwards direction. Thick ice-marginal moraine debris, much of it moundy but aligned in the direction of ice flow, and eastwards-directed meltwater channels such as Clais Fhearnaig, on the west side of Glen Quoich, indicate a progressively thinning ice mass. As the local Cairngorm ice separated from the Dee glacial system, meltwater was temporarily ponded in ephemeral ice-dammed lakes in Glen Quoich and Glen Derry. Active glacier retreat back into the mountains is indicated by recessional moraines in Glen Dee near Derry Lodge, upper Glen Derry, the Lairig an Laoigh and Glen Avon.

During the Loch Lomond Stade of ~12.9–11.7 ka, small glaciers and plateau icefields again formed in the Cairngorms. Boulder moraines in several cirques (e.g. Coire an t-Sneachda, Coire an Lochain (Cairn Lochain) and Coire an Lochain Uaine (Ben Macdui)) and hummocky recessional moraines in upper glens (e.g. Glen Eidart, Glen Geusachan, Glen Dee and Glen Avon) indicate the extent of the readvance glaciers and their subsequent retreat (Sugden and Clapperton 1975; Sissons 1979; Figs. 18.5a and 18.8). However, recognition of the possibility that some of these glaciers were fed by thin, cold-based plateau icefields that left little geomorphological evidence (Rea et al. 1998; Standell 2014) suggests that ice cover during the Loch Lomond Stade may have been more extensive than previously believed, an interpretation also supported by numerical modelling (Golledge et al. 2008).

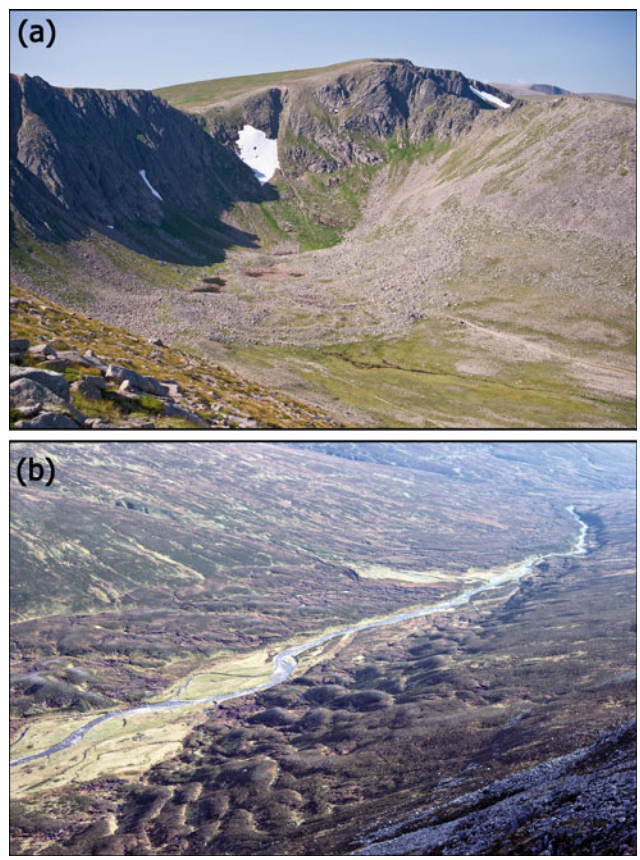

Fig. 18.8 Loch Lomond Readvance moraines. **a** Cirque glacier boulder moraines, Coire an t-Sneachda. **b** Nested hummocky recessional moraines, Glen Eidart. (Images: John Gordon)

18.5 Periglacial Landforms and Landscapes

Periglacial landforms on the higher slopes and plateau surfaces include both relict and active forms (Sugden 1971; Ballantyne 1996). While some of the relict landforms may have survived beneath cold-based parts of the last ice sheet, most probably formed during the period of ice-sheet downwastage and during the subsequent Loch Lomond Stade when the climate was intensely cold. Parts of the high plateau surfaces support relict blockfields (e.g. on Ben Macdui and Derry Cairngorm) similar to those that have been shown elsewhere in Scotland to have survived under cold-based ice throughout the last ice-sheet glaciation (Fabel et al. 2012). Locally, the plateau debris has been frost-sorted into patterned ground, including large-scale, relict sorted circles and stripes. Some upper slopes with low-to-moderate gradients are festooned with relict, bouldery ('stone-banked') lobes and terraces, commonly with risers up to 3 m high. These are particularly well developed on the northwest slope below Creag an Leth-choin (Lurcher's Crag), on the slopes of Coire Domhain and Coire Raibert southwest of Cairn Gorm, and on Ben Avon (Figs. 18.3c and 18.9a). Some authors have interpreted these landforms as the product of permafrost creep (deformation of interstitial ice), but others have suggested that the surface boulders have been rafted downslope by solifluction operating in an underlying layer of frost-susceptible fine sediment (Ballantyne and Harris 1994). Bouldery lobes descend locally to 540 m in the Cairngorms but are absent from areas occupied by Loch Lomond Readvance glaciers, suggesting that they were last active under the permafrost conditions of the Loch Lomond Stade.

Active periglacial features reflect the prevailing wet and windy climate conditions rather than extreme cold and also the availability of fine-grained, frost-susceptible regolith above ~800 m. They include small-scale sorted circles and sorted stripes on areas of bare ground, turf-banked terraces, active solifluction sheets and lobes up to about a metre thick and ploughing boulders (Ballantyne 1996). The presence of buried soils over-ridden by small solifluction lobes indicates that some solifluction features have been mobile during the last few thousand years (Sugden 1971). Much of the plateau area supports fragile mountain plant communities established on friable mountain soils that are especially susceptible to erosion by frost and wind, resulting in the formation of outstanding examples of wind-patterned vegetation, deflation scars and deflation surfaces (Bayfield 1974; Ballantyne 1996; Haynes et al. 1998; Fig. 18.9b). These features are extensive in the western Cairngorms and on plateau margins, cols and exposed upper spurs, where wind speeds are accelerated. It is likely that erosion is being accelerated by increased pressure on vegetation through human and animal trampling.

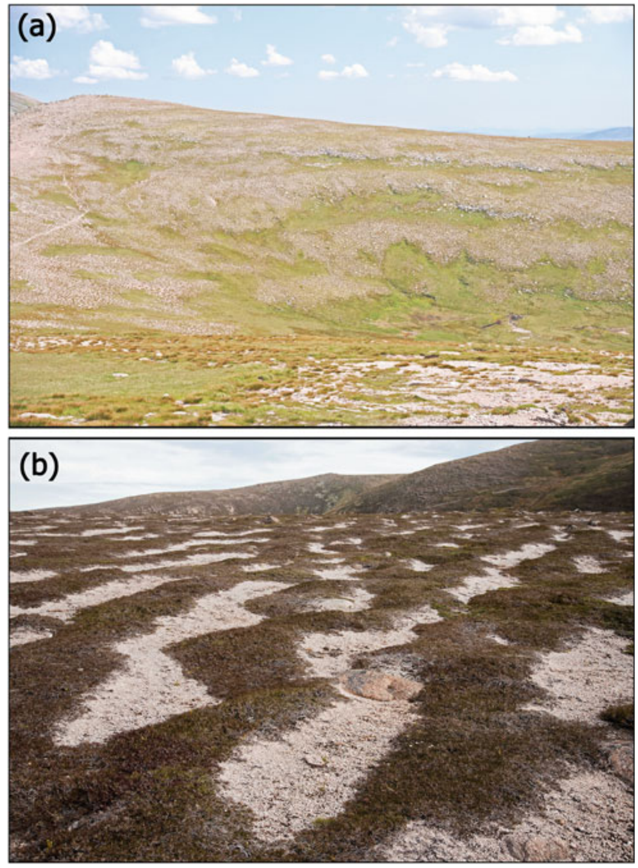

Fig. 18.9 **a** Bouldery, stone-banked lobes, Coire Domhain. **b** Wind-patterned vegetation, Sròn an Aonaich, north of Cairn Gorm summit. (Images: John Gordon)

The Cairngorm Mountains also support a number of late-lying snow beds which survive through much of the summer months and sometimes the entire summer (Watson 2011). Nival weathering processes in the past may have enlarged shallow valley heads developed in alteration zones in the granite, but the current role of nivation processes in the Scottish uplands appears to be limited to erosion and deposition of fine sediment by surface wash (Ballantyne 1985) and possibly enhanced chemical weathering under late-lying snowbeds (Ballantyne et al. 1989). During the Little Ice Age, from the middle of the fifteenth to the middle of the nineteenth centuries, semi-permanent snow cover was much more extensive in the Cairngorms, as reported by early travellers (Sugden 1977), and a small niche glacier may have formed high up in the north-facing Coire an Lochain at this time (Kirkbride et al. 2014; Harrison et al. 2014; Kirkbride 2016).

18.6 Debris Flows and Related Features

The steep middle and lower slopes of the Cairngorm Mountains are partially or wholly mantled by rockfall debris (talus accumulations) and glacial sediments. Following deglaciation, these sediment-mantled slopes have been extensively modified by debris flows and snow avalanches.

Debris-flow activity is the main agent of sediment redistribution on hillslopes under present conditions, with multiple debris-flow deposits forming steep debris cones at the foot of gullies (Fig. 18.10a). Intense summer convectional rainstorms, combined with wet antecedent soil moisture conditions, are the principal triggers of debris-flow activity. The stratigraphy of coalescing debris cones in Glen Feshie indicates that debris-flow activity in the past has occurred at similar intensities to that experienced under present climatic conditions, and that the timing of individual debris-flow events largely reflects storm magnitude and frequency (Brazier and Ballantyne 1989). Debris flows have occurred intermittently at susceptible sites in the Highlands, including the Cairngorms, throughout the Holocene, with evidence for rapid accumulation of some debris cones in the centuries following deglaciation. There is also some evidence for enhanced debris-flow activity within the past few centuries, but the causes remain unclear (Ballantyne 2019).

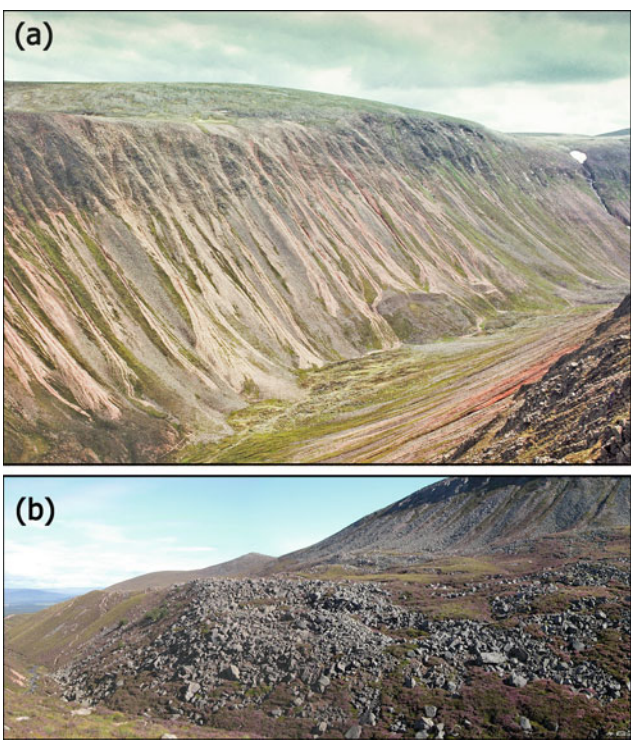

Fig. 18.10 **a** Multiple debris-flow tracks scar the flanks of the Lairig Ghru. **b** Rock-slope failure debris accumulation, Lairig Ghru. (Images: **a** John Gordon; **b** Colin Ballantyne)

The current geomorphological role of snow avalanching in the Cairngorm Mountains is locally recorded by a variety of landforms, including avalanche boulder tongues in the Lairig Ghru (Luckman 1992). Other indicators of debris slopes modified by recent snow avalanches include surfaces swept flat and sometimes stripped of debris and the presence of angular chips of rock littered and perched on older deposits downslope (Ward 1985). In Coire an Lochain, slab avalanches from the steeply dipping granite bedrock below the cirque headwall, both in the present and historically, have produced spreads of reworked debris across the cirque floor (Kirkbride et al. 2014; Kirkbride 2016). Relict avalanche boulder tongues of uncertain age occur in a few locations.

18.7 Rock-Slope Failures

Paraglacial rock-slope failures (RSFs) during the Lateglacial and Holocene occur widely across the Scottish Highlands, principally on metasedimentary rocks (Jarman 2006), but only a few examples occur in the Cairngorms (Fig. 18.10b). TCN exposure dating of RSF runout debris in Strath Nethy, the Lairig Ghru, Coire Beanaidh and Coire Etchachan has yielded ages of 18.1 ± 1.2, 17.2 ± 1.0, 14.3 ± 0.9 and 13.6 ± 0.8 ka, respectively (Ballantyne et al. 2009, 2014). Although the two oldest ages may be compromised by nuclide inheritance, all of these ages indicate failure during the Lateglacial period, within a few millennia following ice sheet deglaciation; the timing of these RSFs suggests that they were responses to progressive rock-mass weakening through deglacial stress release, with failure possibly being triggered by seismic activity during the period of most rapid glacio-isostatic uplift (Ballantyne et al. 2014).

18.8 Fluvial Landforms and Landscapes

The Cairngorms are noted for a variety of fluvial landforms including river terraces and active alluvial fans and floodplains characterised by channel migration and avulsion. Active channel systems are constantly readjusting to fluctuations in climate, vegetation changes, soil development, and perhaps most importantly, the growing impact throughout the later Holocene of human activities. The main forcing agent of change at present, however, is extreme weather conditions, especially flood-producing storms or longer periods of wet weather.

Glacial and glacifluvial deposits in the glens constitute a large supply of sediments that continue to be fluvially reworked. The history of modification and redistribution of these sediments is recorded in flights of river terraces (e.g. in Glen Feshie and along the River Dee) and in large, complex,

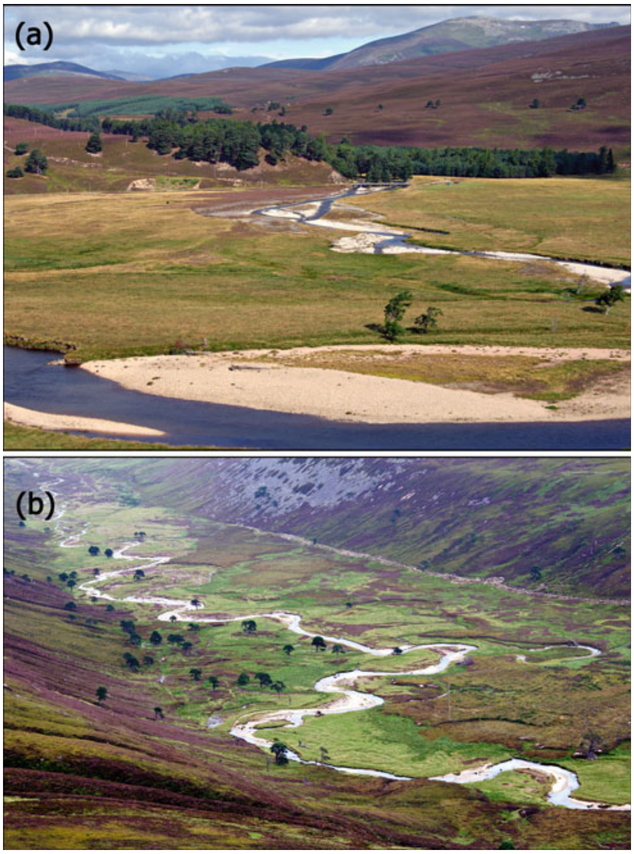

Fig. 18.11 **a** Quoich Water alluvial fan at the confluence with the River Dee. **b** Meanders of the Derry Burn on the floor of Glen Derry (Images: © Lorne Gill/NatureScot)

river-confluence alluvial fans, some of which are actively aggrading today, as at the Feshie–Spey confluence (Brazier and Werritty 1994), the Quoich Water fan at the confluence of the Quoich Water with the River Dee (McEwen 1997a, b) and the Allt Lorgaidh fan in Glen Feshie (Fig. 18.11a; Chap. 19). These terraces and fans and their associated palaeochannel patterns may imply periodic changes in the amount and type of sediment being transported by the rivers and changes in runoff regime (flood magnitude and frequency) that reflect changes in climate.

Channel-reach morphology reflects geological controls and the particular geomorphological evolution of the region, including paraglacial influences (Addy et al. 2014). Mountain torrents with coarse bed material are typical of the steeper valleys, as exemplified by the Allt Mòr (River Druie) which drains the northern cirques of Cairn Gorm (Chap. 19), and the Luibeg Burn, on the west side of Glen Derry (McEwen 1997a). Wandering gravel-bed rivers that exhibit both braiding and meandering characteristics are well developed on the floors of glacial troughs. The River Feshie is a particularly notable example, comprising one of the most active braided river sections in Britain in its upper reaches, a wandering gravel-bed river in its middle reaches and a highly dynamic confluence fan at its junction with the River Spey (Werritty and McEwen 1997b; Chap. 19). The Derry Burn occupies an alluvial basin which has developed due to local base-level control exerted by an outcrop of resistant bedrock. The river is noted for the tortuous meander bends, the presence of cut-offs and former channels of varying age on the valley floor and features such as meander scrolls (Fig. 18.11b).

Studies of the sensitivity of Scottish upland rivers to environmental change over the last 250 years show that rivers have episodically and extensively reworked their valley floors through rapid rates of lateral channel migration during 'flood-rich' periods, but have otherwise not undergone significant changes in behaviour in response to natural processes, exhibiting neither net floodplain aggradation nor net floodplain incision (Werritty and Leys 2001). McEwen and Werritty (1988) observed that the Allt Mòr is only infrequently geomorphologically active, reflecting the high magnitude of storms required to generate floods of sufficient competence to mobilise the bulk of the coarse sediment derived from adjacent glacifluvial deposits. Sands and finer gravels are winnowed away by the river during normal flow conditions, but this leaves behind a lag of larger cobbles and boulders that effectively armour the channel and limit channel dynamics. Similarly, studies of the braided reaches of the River Feshie have shown that major flood events tend to obliterate the former channel pattern and replace it with a chaotic braided system that gradually regains stability during inter-flood periods (Werritty and Ferguson 1980; Rumsby et al. 2008; Chap. 19). High-intensity convective storms and rapid snowmelt tend to trigger a flashy response in streams draining the catchments in the Cairngorms, notable examples being the 'Muckle Spate' of 4 August 1829 and floods in 1956 and 1978 (McEwen and Werritty 1988, 2007).

18.9 Geomorphology and Nature Conservation

18.9.1 Natural Heritage Values

The Cairngorm Mountains are highly valued for their natural heritage (Gimmingham 2002; Shaw and Thompson 2006). Much of the area is incorporated within national Site of Special Interest (SSSI) and National Nature Reserve (NNR) designations, while at an international level, a large part of the area is notified under the European Habitats Directive and the European Birds Directive. The Cairngorm Mountains also form the heartland of Scotland's second National Park, established in 2003. The geomorphological importance of the Cairngorms is reflected in their inclusion in the Quaternary of Scotland and Fluvial Geomorphology volumes of the Geological Conservation Review, the assessment of key geoheritage sites in Great Britain (Gordon 1993; Werritty and McEwen 1997a; Chap. 29). The landforms and soils also provide the abiotic 'stage' supporting nationally and internationally valued habitats and species (Gordon et al. 1998). For example, the plateaux are noted for their montane vegetation, including lichen-rich montane heath and specialised snowbed communities, while cliffs, gullies and screes provide habitats for several rare alpine plants. There is close dependence of vegetation communities on altitude, wind exposure, snow duration, regolith properties, moisture availability, ephemeral landforms and micro-topography (Watt and Jones 1948; Gordon et al. 1998; Haynes et al. 1998). Within each broad vegetation zone, minor variations in the type of vegetation as well as the vigour and habit of species within the same community are determined by variations in microclimate and substrate. These are often themselves controlled by the presence of geomorphological features such as solifluction lobes and terraces. These factors create a mosaic of vegetation, which is closely associated with the geomorphology at all scales. The geo-ecological links between soils, geomorphology and vegetation also have a strong influence on the terrain sensitivity to natural and anthropogenic disturbance (Gordon et al. 1998, 2002; Haynes et al. 1998; Morrocco et al. 2016).

The glacial and glacifluvial deposits of the lower ground form well-drained ridges of sands and gravels interspersed with wet hollows. They support a mosaic of habitats for birds, insects and plants and the largest areas of native Caledonian pinewoods (*Pinus sylvestris*) in Scotland. Peat bogs and kettle-hole deposits provide detailed palaeoenvironmental records of pine woodland history, the spread of blanket peat, the history of arctic/alpine plant species, tree-line dynamics, past floristic diversity and Lateglacial and Holocene climate change (e.g. Bennett 1996; Allen and Huntley 1999; Nagy et al. 2013; Rydval et al. 2017).

Similarly, the geomorphological characteristics of the rivers are a fundamental influence on riverine habitats and species of high conservation value. The Rivers Spey, Dee and Don have headwaters in the Cairngorms. The catchment characteristics, including geology, soils, hydrological pathways, river channel geomorphology, sedimentary properties and flow characteristics, have a fundamental influence on water quality and habitat availability. For example, gravel-bed rivers provide spawning grounds for Atlantic salmon (*Salmo salar*), which depend on the availability of geomorphological features such as pools, riffles and glides and on local sediment characteristics (Moir et al. 2004).

The rocks, landforms and soils of the Cairngorms have exerted a powerful influence on the landscape, wildlife and land use and are celebrated by many writers, artists and photographers, and notably in the poetry and prose of Nan Shepherd (1934, 1977). The geomorphological diversity of the Cairngorms landscape also provides many opportunities for outdoor recreation, including summer and winter walking as well as more specialised rock and ice climbing, downhill skiing, ski-touring and training in winter mountaineering skills.

18.9.2 Pressures and Impacts

The principal activities which have impinged, or potentially impinge, on the geomorphology of the Cairngorms are afforestation, river management, recreation, changes in grazing intensity and changes in estate management (Gordon et al. 1998). In Glen More and lower Glen Feshie, commercial afforestation has obscured landform assemblages. Footpath expansion has locally disturbed periglacial features on the higher slopes and plateaux, where the fragile soils are vulnerable to enhanced erosion through overgrazing and increased recreational pressures. Overgrazing by red deer has locally accelerated slope instability and erosion in Glen Feshie, although deer numbers are now much reduced. Peat erosion is extensive on blanket bog surfaces across the area. There are potential threats from climate change and atmospheric pollution (nitrogen deposition) to periglacial features and soils on the plateau, both through direct changes in processes (frost action and erosion by wind and surface runoff) and through possible changes in vegetation cover. River engineering for flood protection, fisheries management and local gravel extraction continue gradually to impede the natural evolution of rivers (e.g. the Allt Mòr, the Feshie–Spey confluence and the Quoich Water fan). Such activities may damage landforms, disrupt natural processes and have adverse effects downstream, particularly where engineered structures for flood or fisheries management divert river flow, or river channels are relocated by bulldozing. The recent reduction in the red deer population in Glen Feshie has enabled natural regeneration of the Caledonian pinewood, which may reduce flood peaks and thus the instability of the River Feshie channel system.

18.10 Conclusions

The Cairngorm Mountains display an exceptional assemblage of geomorphological features illustrating the long-term evolution of a glaciated, mid-latitude, granite mountain landscape. They represent a classic landscape of selective glacial erosion: relict, non-glacial landforms such as tors, weathered granite and sheet jointing are exceptionally well developed on palaeosurfaces that are essentially pre-glacial in origin and are closely juxtaposed with glacial troughs, glacially breached watersheds and cirques. The adjacent glens support a diverse assemblage of glacial and glacifluvial landforms, notably meltwater channels, eskers, kames, kettle holes, outwash terraces, ice-dammed lake deposits and moraines. On the northern flanks of the massif, there is evidence for active recession of the last Scottish Ice Sheet, while several cirques contain excellent examples of moraines formed during the Loch Lomond Stade. Both relict and active periglacial landforms occur extensively on the high slopes and plateau surfaces and add further to the landform diversity, as do several rock-slope failures and other slope landforms. Postglacial geomorphological modifications of the landscape are represented by a variety of fluvial landforms and process systems, including those of mountain torrents, alluvial fans and wandering gravel-bed rivers.

The occurrence of such a diverse assemblage of features in a relatively compact area is exceptional and of high conservation value. Some of the landforms are among the best examples of their kind in Britain, while others rank on an international scale for their clarity of development and inter-relationships. Not only do the Cairngorms provide an invaluable record of long-term landscape evolution, but they also represent a potentially important natural laboratory for the study of the geomorphological sensitivity of the landscape to environmental change and the role of geomorphology in ecosystem support and functioning. Together, the different elements of the geodiversity and biodiversity form 'a landscape tapestry of many and rare distinctions' (Crumley 1991, p. 16).

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Fluvial Landforms of Glen Feshie and the Spey Drainage Basin

19

Alan Werritty

Abstract

The Spey drainage basin incorporates a classic assemblage of fluvial landforms recording both their evolution during the Lateglacial and Holocene and the operation of present-day fluvial processes. Noteworthy examples of relict and active alluvial fans occur alongside major sequences of river terraces. Although most reaches of wandering gravel-bed rivers within the drainage basin are relatively stable today, active meanders and braided reaches (notably in Glen Feshie) provide an opportunity to assess the operation of present-day processes. Sediment reworked by fluvial processes from glacigenic sources—notably undercut terraces and fans—is often boulder-sized, creating a high threshold for subsequent re-entrainment. Whilst relatively frequent floods generate channel change in some wandering gravel-bed rivers, other fluvial assemblages such as mountain torrents and large fans are only reworked during rare floods.

Keywords

Braided rivers • Meanders • Wandering gravel-bed rivers • Alluvial fans • Terraces • Floods • Sediment storage

19.1 Introduction

The Spey is Scotland's third largest river, flowing north-east from its headwaters in the Central Grampians to its outlet in the Moray Firth, with a drainage area of 3011 km^2. Bounded to the north and west by the Monadhliath Mountains and to the south and east by the Cairngorm Mountains and the Drumochter Hills, for much of its course the main stem of the river is nearer to its northern and western boundary (Fig. 19.1). The bedrock geology of the Spey drainage basin is dominated by crystalline metasedimentary rocks of the Neoproterozoic Dalradian Supergroup and the Late Silurian Cairngorm granitic pluton (Chap. 2). The oldest surviving landforms are high-level Cenozoic palaeosurfaces, with the present-day River Spey reflecting a drainage system of SW–NE trending valleys, partly modified via capture by eastwards-flowing rivers in the Palaeogene and by glacial diversion during the Pleistocene. Following repeated glaciations during the Pleistocene, as the last Scottish Ice Sheet thinned and retreated, a complex assemblage of glacial and glacifluvial landforms was deposited throughout the Spey valley. During the Loch Lomond Stade of ~12.9–11.7 ka, small glaciers and plateau icefields again formed in the Cairngorms, the Drumochter Hills and the Gaick Plateau.

19.2 Lateglacial and Holocene Fluvial Landforms and Landscapes

The reworking of slope glacigenic sediments by various forms of mass movement in the Cairngorms is documented in Chap. 18; here the focus is on fluvial landforms and the postglacial reworking of valley-floor glacigenic sediments, resulting in many examples of alluvial fans and terraced valley floors plus some of the best locations in the UK for the study of present-day fluvial processes and gravel-bed rivers. The older landform assemblages of river terraces and alluvial fans are relict features left in situ as deposition gave way to erosion and incision during the Lateglacial and Holocene.

During and immediately after ice-sheet deglaciation, valley floors across the Spey drainage basin aggraded as glacigenic sediments and those supplied by paraglacial slope processes were reworked by proglacial rivers, the manner and rate of aggradation depending on the interplay between dominant stream discharge and sediment supply (Ballantyne

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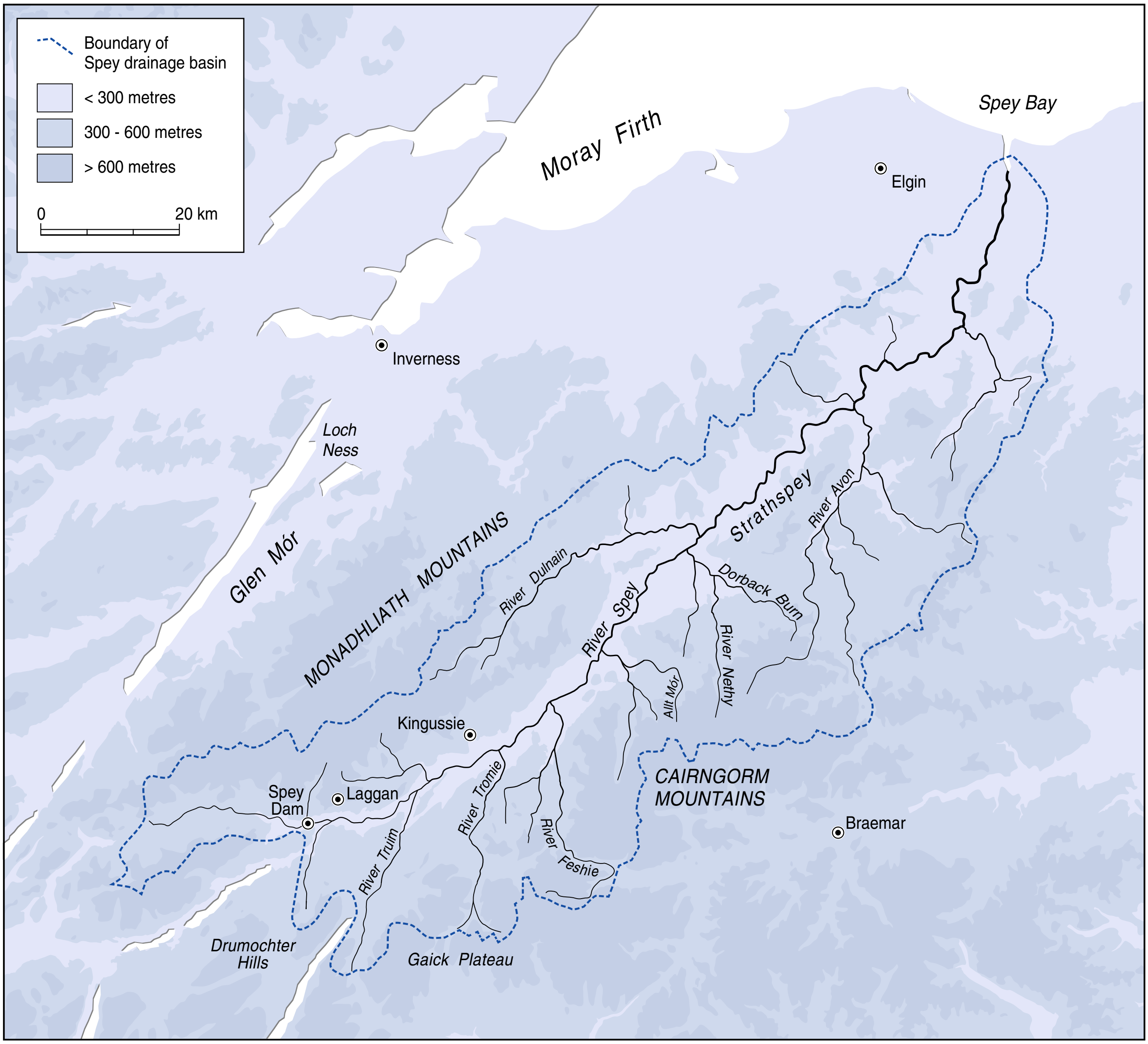

Fig. 19.1 The Spey catchment, showing the location of sites mentioned in the text

2002). When the former exceeded the latter, incision occurred: when the reverse obtained, aggradation took place. Both controls fluctuated during the Lateglacial, ultimately resulting in a single cycle of aggradation and incision that extended into the Holocene (Ballantyne 2019). This has resulted in a typical cut-and-fill sequence in valley floors with terraces, or fan surfaces of decreasing age inset within each other.

19.2.1 Alluvial Fans

Alluvial fans of contrasting ages and origins are widespread throughout the Spey drainage basin, with the oldest being of Lateglacial age. At some sites, such as the Feshie confluence fan (Fig. 19.2), Holocene and present-day fan surfaces are inset into these older alluvial fans. Smaller fans occur where steep tributary streams meet the main channel. At these sites, aggradation resulted from the episodic erosion of drift-covered slopes by channelized debris flows and subsequent transport by flood torrents (Ballantyne 2019). Particularly good examples of such alluvial fans occur in Glen Feshie where steep-side tributary valleys, most notably those of the Allt Garbhlach and the Allt Lorgaidh (Fig. 19.2), contain abundant glacigenic sediments. The distal surface of the Allt Garbhlach fan where it grades into the highest of the Feshie terraces (Fig. 19.3) comprises an outwash fan which, having been subject to an abrupt decline in discharge during the Lateglacial, is embroidered with remnants of kames and palaeochannels. The Allt Lorgaidh fan (Fig. 19.4a) consists of three nested inset fan surface fragments. These were assigned ages of ~10 ka, ~3.6 ka and ~1.0 ka by Robertson-Rintoul (1986) on the basis of soil-stratigraphic evidence, assuming that the upper fan surface formed at the end of the Loch Lomond Stade, and a radiocarbon age of 3620 ± 50 ^{14}C a BP for charcoal buried by fan gravels on the middle fan surface. Subsequent recalibration of these ages suggests that the upper fan surface may have formed at ~11.6 ka, the middle fan surface at or after 4.1–3.9 ka and the lowest fan surface at ~1.0 ka. The Allt Lorgaidh fan surfaces abut the terrace surfaces of the main channel of the River Feshie, to which Robertson-Rintoul assigned ages of 4.1–3.9 ka (recalibrated), ~1.0 ka and AD 1869–1902.

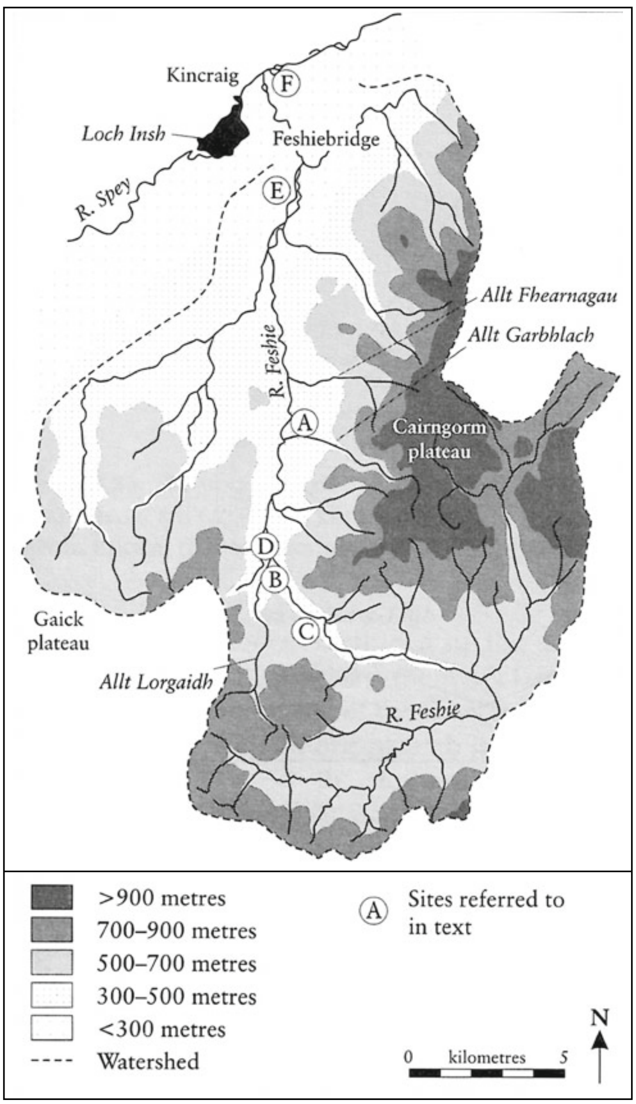

Fig. 19.2 Glen Feshie location map. **a**: Allt Garbhlach Fan; **b**: Allt Lorgaidh Fan; **c**: Debris Cones; **d**: Upper Braided Reach; **e**: Lower Braided Reach; **f**: Confluence Fan. (From Werritty and McEwen 1997)

Fig. 19.3 Allt Garbhlach fan with palaeochannels and kame remnants on the fan surface; the distal margin is cut by the highest Holocene terrace. (Google Earth™ image)

Alluvial fans are found at the confluences of the major tributaries of the Spey with the main channel but, with the exception of the Feshie confluence fan (Sect. 9.3.1), all are now relict and heavily modified landforms only exhibiting a single thread channel, having been stabilized by river engineering to support either arable cultivation (the Tromie, Dulnain and Nethy confluence fans) or forestry (the Avon confluence fan).

19.2.2 Terraces

Lateglacial terraces are widely developed across the valley floors of the Spey and its main tributaries and take the form of kame terraces, outwash terraces and postglacial alluvial terraces. Kame terraces formed by meltwater deposition at the lateral margins of valley glaciers were subsequently abandoned during deglaciation: particularly good examples occur along the lower valley-side slopes upstream of Kingussie (Young 1978). Outwash terraces, up to 30–40 m above the present-day floodplain, represent abandoned sandur surfaces and are especially well-developed in Glen Feshie.

The exceptionally well-preserved terrace sequence in Glen Feshie extends from the bedrock reach at Feshiebridge to the glacial breach beyond the Allt Lorgaidh fan (Fig. 19.2). Height-distance diagrams compiled by Young (1976) led him to identify five distinct groups of terraces, all of which support braided palaeochannels. The uppermost three groups (1–3) begin abruptly at the Allt Garbhlach (A in Fig. 19.2) in the central part of the glen, 25–40 m above the

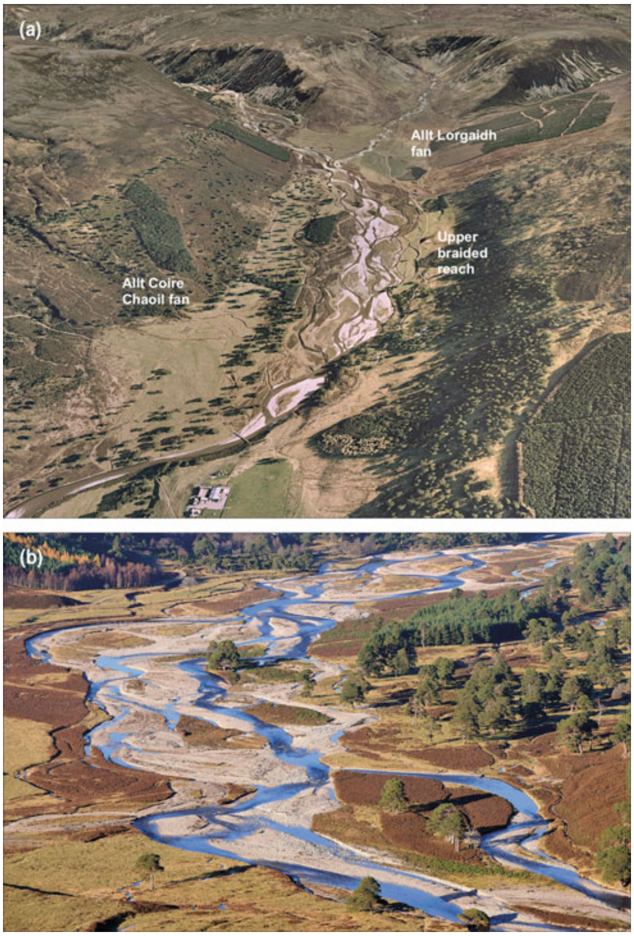

Fig. 19.4 The River Feshie. **a** The upper braided reach between the Allt Lorgaidh and Allt Coire Chaoil fans. **b**. Islands and lateral, mid-channel and overbank bars in the upper braided reach. (Images: **a** Google Earth™; **b** © Lorne Gill/NatureScot)

river, and decline steeply northwards towards the Feshiebridge bedrock reach. The uppermost two groups are pitted with kettle holes, confirming a glacifluvial origin, and have been variously interpreted as either kame terraces or, more plausibly, remnants of a proglacial palaeosandur surface that formed as glacier ice retreated up Glen Feshie at ~15.5 ka (Young 1976; Ballantyne 2019; Chap. 4). The absence of high terraces from the upper glen was attributed by Young to their subsequent removal by erosion. The abrupt upvalley termination of all high terraces at the same location, however, suggests occupation of the upper glen by glacier ice during the period of outwash deposition and subsequent high-terrace formation. If this interpretation is valid, it implies rapid fluvial incision of the initial palaeosandur surface and terrace formation before interstadial warming at ~14.7 ka forced withdrawal of glacier ice onto high ground. On the basis of soil-stratigraphic contrasts associated with the three highest terrace groups, however, Robertson-Rintoul (1986) inferred much slower high-terrace formation over the period separating deglaciation (~15.5 ka) from the Lateglacial–Holocene transition (~11.7 ka).

The lower two groups of terraces (4 and 5) in Glen Feshie are present throughout the entire length of the valley roughly parallel to, and less than 5 m above, the present floodplain. The radiocarbon age of 4.1–3.9 cal ^{14}C ka recalibrated obtained by Robertson-Rintoul (1986) for a low terrace on the Allt Lorgaidh fan (Sect. 19.2.1) implies that these terraces represent Late Holocene floodplain aggradation and subsequent incision, a pattern that has been observed at several sites in the Scottish Highlands (Ballantyne 2019). Similar low river terraces are widespread throughout the Spey drainage basin, being especially well-developed in Glen Truim, but their contemporaneity remains to be established.

19.3 Present-Day Fluvial Processes and Landforms

The main stem of the Spey and its tributaries comprise low-sinuosity, relatively wide, shallow channels in which sandy overbank facies rest on basal gravels. Where local controls permit, active meanders can develop with chute channels over bars triggering avulsion, as well-illustrated on the lower Dulnain (c.f. the active meanders described by Carson (1986)). Where not confined by terraces or proximity to the valley sides, the resulting anabranches and associated palaeochannels are common, as evidenced by the Avon (Werritty and Leys 2001), the Allt Lorgy in the lower Dulnain catchment (Williams et al. 2020) and the upper Spey above Spey Dam (Fig. 19.1). On larger rivers, the bars and islands generated by avulsion create an island/braided or anastomosing planform as illustrated by the lower braided reach in Glen Feshie (E in Fig. 19.2) and parts of the lower Spey before it flows into Spey Bay. Short bedrock reaches on the Avon, Feshie, Truim, Tromie and Dulnain locally punctuate alluvial valley floors: initially excavated by glacial meltwaters, these are now maintained by high flows.

19.3.1 Glen Feshie

The River Feshie is a north-flowing tributary of the Spey draining a catchment of 240 km^2 (Fig. 19.2). The drainage basin is mainly underlain by Dalradian schist, which supplies most of the coarse bedload, with the Cairngorm granite only present in the east. The drainage basin extends into the Cairngorm plateau to the east and the Gaick plateau to the south. From headwaters in the Gaick plateau, the Feshie initially flows eastwards before abruptly turning northwest into a steep-sided glacial trough (Linton 1949), thereafter flowing north confined by the lowest Holocene terraces and occasionally incised into bedrock. In two locations, the channel planform is braided (D and E in Fig. 19.2), and at the confluence with the Spey it incorporates braiding within rapidly evolving distributary channels (F in Fig. 19.2). These reaches comprise some of the most active assemblages of fluvial landforms in Great Britain and result from a flashy runoff regime, a steep ($\sim$0.01) channel slope and bankfull discharges (28–42 m^3s^{-1}) capable of mobilizing bed and bank materials (Werritty and Ferguson 1980). Although generally described as 'braided', the close proximity of these reaches to the braiding–meandering threshold implies that they are more accurately defined as 'wandering gravel-bed rivers' (Church 1983). This means that the river occupies a wide active area, sometimes displaying multiple channels and subject to avulsion (Fig. 19.4), but in other reaches maintains a more stable sinuous planform.

Comparison of the configuration of the upper braided reach as recorded in the AD 1750 Roy Map and large-scale Ordnance Survey maps (AD 1869, 1899 and 1971) shows that this reach has been characterized by a braided channel system since at least the mid-eighteenth century (Werritty and Ferguson 1980). Aerial photographs from 1946, 1955 and 1966 provide greater detail: an increasingly stable eastern channel inactive over two decades, but both braiding and meandering dominant in the active western channel. In 1961 a >200 m^3s^{-1} flood reactivated braiding, but increasingly sinuous channel planforms characterized post-flood recovery (Werritty and Ferguson 1980). Between 1946 and 1966, channel switching and reoccupation of abandoned channels were more significant than lateral migration, generating a series of relatively stable nodes. Ferguson and Werritty's (1983) model for the sequence of bars, channel switching and avulsion operating between these nodes has since been confirmed and elaborated by Wheaton et al. (2013). Time intervals between topographic surveys and sequential aerial photographs initially made it impossible to relate detailed morphological change to specific floods, but recent photogrammetric and high-density and GPS-based topographic surveys (Brasington et al. 2000, 2003; Wheaton et al. 2013) have made such analysis possible, despite significant technical challenges (Rumsby et al. 2008).

Synoptic changes in braiding mechanisms and sediment storage in the upper braided reach at an annual timescale have been determined by Wheaton et al. (2013) from digital elevation models (DEMs) derived from repeat high-resolution GPS surveys compiled annually over the period 2003–2007. From the difference between sequential DEMs, the evolution of the primary geomorphic units of both bars (mid-channel, lateral and overbank; Ferguson and Werritty 1983) and islands has been determined, together with annual changes in sediment storage (Fig. 19.5). The four classic braiding mechanisms of central bar development, transverse bar conversion, chute cutoff and lobe dissection (Ferguson 1993) account for 61% of the volumetric change in storage, with chute cutoff the most common braiding mechanism. Whilst central bar development, transverse bar conversion and lobe dissection are largely aggradational, bank erosion, channel incision and bar sculpting are largely erosional in nature. Bank erosion (17%) is a more important driver of changes in sediment storage than most of the other individual braiding mechanisms. This study by Wheaton et al. (2013) is significant in being the first field-based confirmation of braiding mechanisms reported in numerous earlier flume-based studies. Recent culling of red deer has enabled the native Scots pine (*Pinus sylvestris*) to recolonize the valley floor. This will provide a unique opportunity to determine whether over time, with forest regeneration and increased bank shear strength, the braided reach stabilizes, or whether an increase in flood events associated with climate change maintains a braided channel planform.

At the confluence of the Feshie and Spey, an alluvial fan with active distributary channels (Fig. 19.6) is inset within a much larger relict fan of Lateglacial and Holocene age (Brazier and Werritty 1994). As with the upper braided reach, periods of comparative stability punctuated by sudden change have been reconstructed from comparable large-scale maps, documentary sources and aerial photographs. These provide evidence of classic behaviour of channel switching: major distributaries locally aggrade until they have become locally perched above the rest of the fan, then avulsion during floods forms a new distributary (Werritty et al. 2005). The main distributary channel migrated eastwards during the 1860s to occupy what was to prove a stable corridor in the centre of the fan until 1989. Between 1989 and 1990, in response to major floods in February and June 1990 (Gilvear et al. 2000), this distributary reverted to its pre-1862 position on the western edge of the fan before switching to its pre-1989 position in the centre of the fan following reinstatement of breached flood banks in 1991 (Fig. 19.6a). Since then, the main distributary has migrated eastwards by repeated small-scale avulsions to a relict part of the fan unoccupied since 1750. Here, steered by a nineteenth-century flood bank, it has developed a complex, highly divided pattern before flowing into the Spey around 200 m downstream of the pre-1989 confluence (Fig. 19.6b). Former flood banks and bank stabilization schemes dating back to 1814 point to numerous attempts to stabilize the flow across the fan surface, the most recent being in 1991. All have failed, although the complex flow pattern in the 1999 map (Fig. 19.6a), in which the main distributary reoccupied much older palaeochannels, reflects the alignment of two historic flood banks: a flow pattern that persisted to 2020.

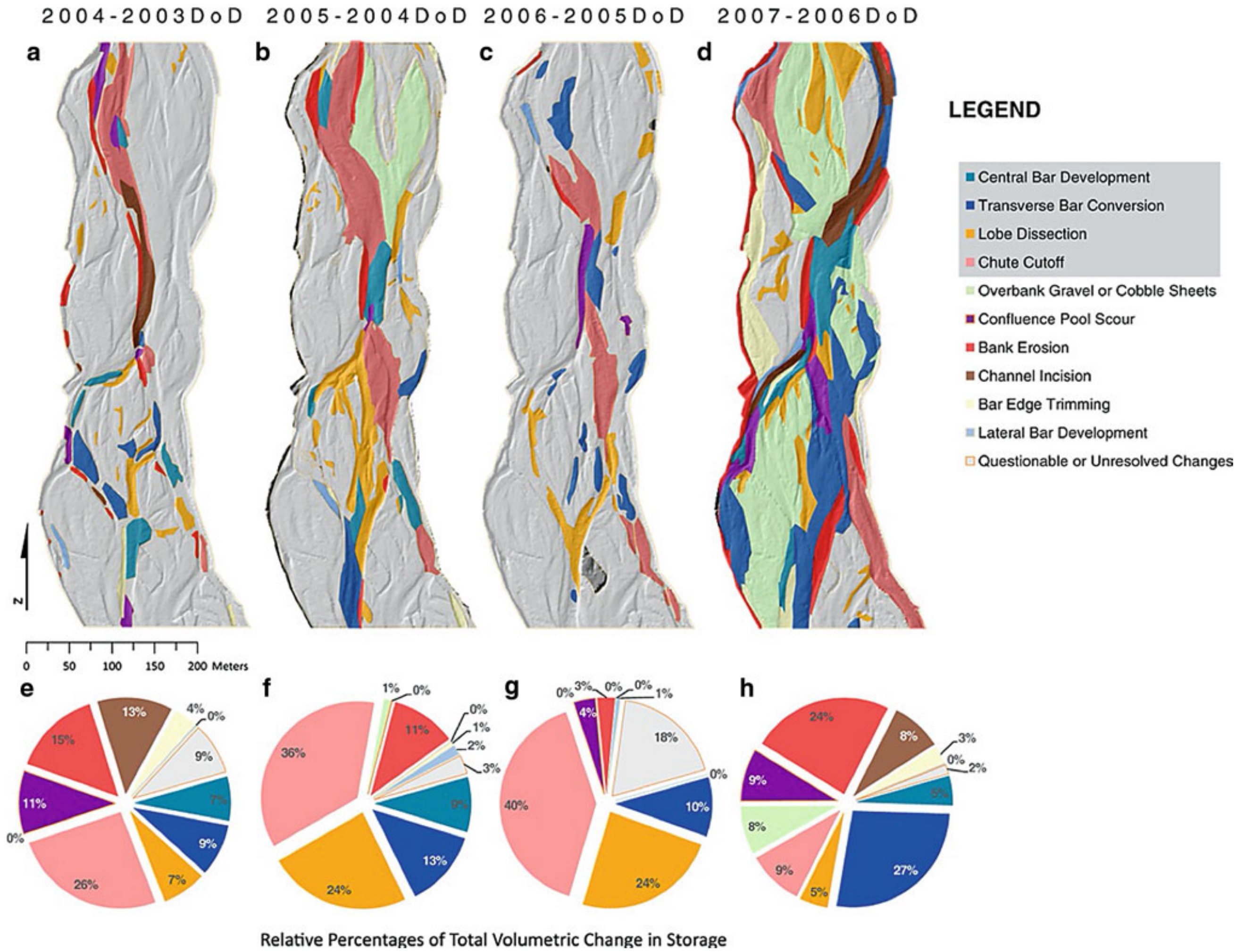

Fig. 19.5 Changes in sediment storage in the upper braided reach of the River Feshie: 2003–2007. (From Wheaton et al. 2013. © 2013. American Geophysical Union. All rights reserved. Copyright (2013) Wiley, with permission from Wiley conveyed through Copyright clearance center, Inc)

19.3.2 The Spey Valley

Much of the valley floor of the Spey comprises relatively high-value arable land and pasture locally protected by low floodbanks, or wetlands drained to provide seasonal grazing for sheep. Despite these engineering works, the floodplain and associated palaeochannels (e.g. near Kingussie) are locally inundated by floodwaters most winters, but with minimal geomorphic impact. High rates of local bank erosion and many new channels were recorded following the 1829 'Muckle Spate' (or 'Great Flood'; McEwen and Werritty 2007), but subsequent agriculture and river engineering have largely erased their imprint from the valley floor.

In the upper Spey upstream of Laggan (Fig. 19.1), hydro-power development has significantly impacted on fluvial process and landforms. Following the completion of the Spey Dam in 1942, 50% of the Spey catchment area upstream of the gauge at Invertruim (9 km southwest of Kingussie) is now regulated. As a result, for floods with return periods of up to five years, the inflow volume would be reduced by at least 60%. For floods with return periods of less than two years, the whole volume would be stored by the reservoir under normal operating practices (Gilvear 2004). This reduction in flood magnitudes has triggered aggradation downstream of unregulated tributary confluences as the channel morphology of the upper Spey adjusted to a new runoff regime. By 1989 all formerly active gravel lateral bars and point bars had been stabilized by vegetation and, immediately upstream of Laggan, an alluvial bench formed on the south bank registering overall aggradation equivalent to between 33 and 80% of the original channel depth. During the period 1946–1995, channel widths along the Spey downstream of the dam contracted by 50–80%, most notably via constrictions occurring at and immediately below tributary confluences. Given this degree

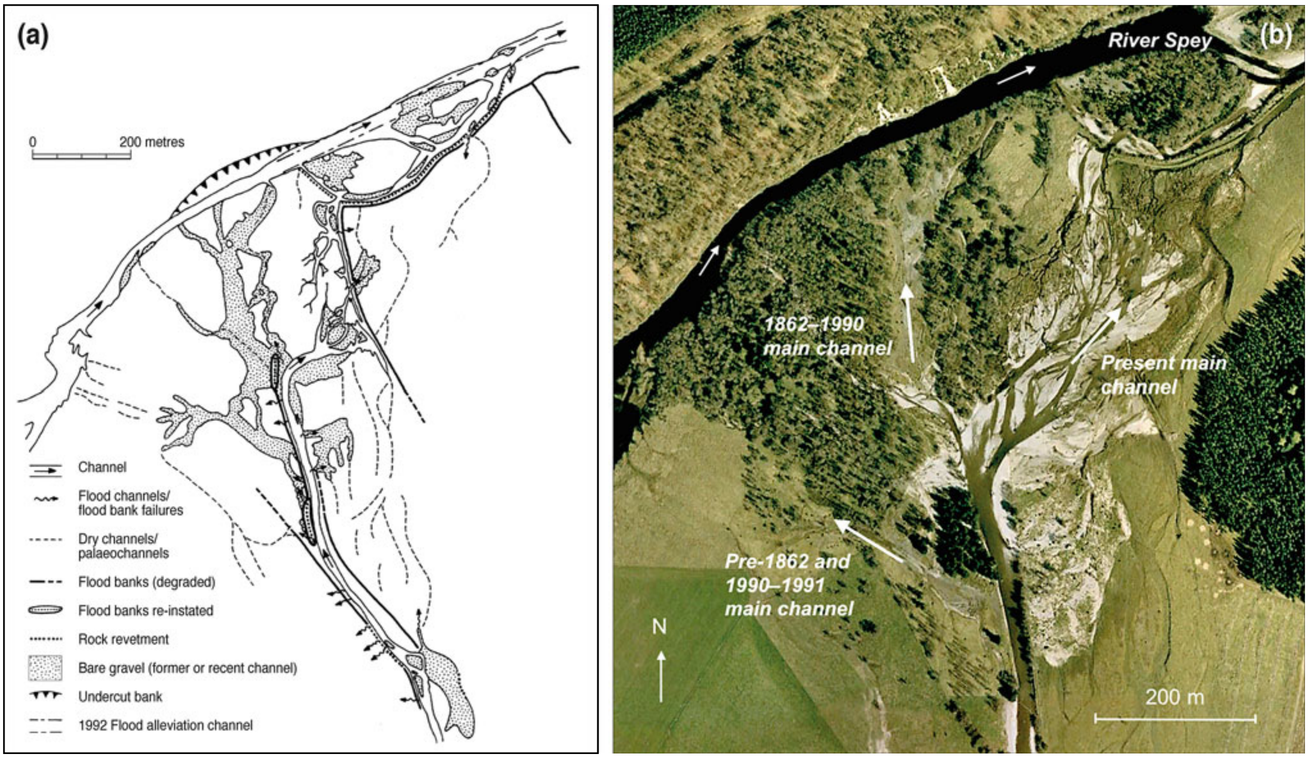

Fig. 19.6 **a** Geomorphological map of the Feshie confluence fan: June 1999. (From Werritty et al. 2005. © Elsevier Science 2005, reproduced with permission of Elsevier through PLSclear). **b** Changing distributaries on the Feshie confluence fan: March 2006. (Google Earth™ image)

of channel narrowing, Gilvear (2004) estimated that 25,000 m^3 of sediment had been deposited within 7.5 km downstream of the Spey Dam since 1942 as result of flow regulation. Gilvear's reconstruction not only explains the pattern of adjustment in fluvial processes and landform assemblages following impoundment of the upper Spey tributaries, but also validates Petts' (1979) spatio-temporal model of channel adjustment downstream of a reservoir.

In contrast to increased stability in some fluvial landform assemblages in the upper Spey, another active reach occurs where the Spey flows into the Moray Firth at Spey Bay (Fig. 19.7). Although generally described as braided, this reach also incorporates many of the features of the Feshie (Sect. 19.3.1) where migrating meanders develop chute channels over bar heads resulting in avulsion. The combination of a wide floodplain permitting channel change and highly variable flow, including catastrophic floods in 1829, 1956, 1970 and 1993 (McEwen 1997), results in a wandering gravel-bed river.

Lewin and Weir (1977) distinguished three zones within a floodplain incorporating active and inactive channels, channel bars and overbank deposits. In the first zone, the active channel comprises a reach close to the braided-meandering threshold. At low flows, lateral bars with lobate migratory avalanche faces create a transient

Fig. 19.7 The lower River Spey and its mouth at Spey Bay. The disused railway bridge lies in the centre of the image. (Google Earth™ image)

braided planform repeatedly erased and reformed during subsequent higher flows. During such flows bar heads are dissected by chutes and, in response to local bed lowering, channel switching occurs (as in the upper braided reach in Glen Feshie) adding to the complexity of the transient braided form. Over the period 1953–1995, the most highly developed braiding systems occurred in the reach north of the abandoned railway bridge (Fig. 19.7). By contrast, channel change south of the bridge was dominated by steady migration of large meander bends with point bars dissected by cutoffs and bar heads by chutes (Gemmell et al. 2001). The second zone of scrub and woodland with abandoned former channels is fluvially inactive, but channel subdivision again increases during major floods when old channels proximal to the active channel are temporarily reoccupied, notably during the 1829 Muckle Spate (McEwen 1997). This second zone is included in the braiding indices of 7.4 (1760), 5.32 (1882), 6.5 (1889) and 2.26 (1967) calculated by Lewin and Weir (1977) from large-scale maps and plans. The low braiding index for 1967 reflects a 60% reduction in the active channel width between 1870 (266 m) and 1971 (108 m), during which period afforestation and bank protection in the second zone restricted channel mobility during floods. Although at a similar elevation to the other zones, and still liable to inundation during major floods, the third zone has not been fluvially active for 200 years. Here, subsequent farming operations have erased the geomorphic signature of floods, although former traces of palaeochannels and ridges can still be identified in aerial photographs (Gemmell et al. 2001).

The overall mean level of the sedimentation surface across the first and second zones differed little over the period 1889–1967, despite the river having reworked half of its valley floor over the previous century (Lewin and Weir 1977). Sediment storage of 7.96×10^6 m^3 (residence time of ~66 years) coupled with modelled bedload transport of ~8,000 m^3 a^{-1} implies a rapid turnover of sediment within the active channel zone, necessitating regular dredging offshore. By contrast, the much greater residence times for presently inactive sediment stores imply that the entire floodplain has been reworked ~10–20 times during the Holocene (Gemmell et al. 2001). Overall this reach provides insight into one of the most active wandering gravel-bed rivers in Great Britain, inset within a Holocene valley floor that retains evidence of palaeochannels reoccupied during historic floods and palimpsests of much earlier floods.

19.4 Floods and Climate Change

In assessing the role of floods and applying magnitude and frequency concepts to fluvial processes and landforms in the Spey drainage basin, two issues emerge: the role of thresholds and the frequency of floods capable of registering long-term impacts on the fluvial landscape. Localized flood erosion/deposition regularly occurs in response to significant floods across the Spey catchment, driving channel change in the more active reaches of the Spey and its tributaries (Sect. 19.3). Such events frequently occur but are rarely documented unless they impact on transport infrastructure. One such event was Storm Frank in December 2015, which eroded the Garbhlach fan in Glen Feshie (Fig. 19.3; Chap. 5) and generated a large bar 1 km downstream. The geomorphic signature of such events depends on the local context, either providing incremental contributions to channel change or short-lived changes that are subject to reworking by later floods. This contrasts with other sites where higher entrainment thresholds maintain more stable channels, with bed and bank sediments only being mobilized by rarer floods. Two small streams on the margins of the Cairngorm massif—the Allt Mòr in Glenmore and Dorback Burn—provide contrasting examples of the geomorphic impact of such rarer floods.

The Allt Mòr (Fig. 19.1) is a small mountain torrent excavated into thick glacigenic deposits before emerging from a highly constricted course onto a palaeofan with an inset small present-day fan. Two major summer storms, in 1956 and 1978, constitute the formative events which determine the present-day course of the channel, the later storm generating an estimated discharge of 65 m^{3}s^{-1}, with a return period of 50–100 years. The geomorphic impact of these storms has been summarized by McEwen and Werritty (1988). First, boulder-sized sediment at the foot of undercut glacigenic deposits and across much of the narrow valley floor produces a binary bedload transport system that is generally inactive and only briefly operative during periods of intense flood runoff. Second, the impact of flood runoff varies according to where the stress is applied and the type of planform present. Where the channel is incised into glacigenic deposits on the outside of meanders, undercutting and rapid mass movement provide a sediment source for bedload transport during the flood (Fig. 19.8a) and potentially a source of coarse sediment for reworking in the next major flood. This close coupling of the slope-channel system operates cyclically: the larger 1956 flood released more glacigenic sediment onto the valley floor and thus provided bedload for the 1978 flood. Third, inter-arrival times of formative events are important. The 1978 flood had a lesser geomorphic impact as the channel system was already partially adjusted to extreme discharges: channels excavated in the active part of the fan in 1956 were reoccupied in 1978 but only one new distributary was formed (McEwen and Werritty 1997).

Dorback Burn (Fig. 19.1), a tributary of the River Nethy, is a gently sinuous undivided channel, except where it undercuts two major kame deposits. Here, in response to a sequence of major floods, the river displays both braided and meandering planforms characteristic of a wandering

Fig. 19.8 **a** The Allt Mòr incised into glacigenic sediments undercut on meander bends during flash floods. **b** Dorback Burn wandering gravel-bed river flowing between kames with relatively stable, but locally divided, sinuous channels. (Images: Alan Werritty)

gravel-bed river (Werritty 1984). A formative flood event in 1978 (estimated at 23 m^3s^{-1}) resulted in the whole of the active area becoming highly braided, but larger floods in 1980 and 1981 registered only a modest reworking of the active area, and from 1982 onwards the braided planform has been replaced by three relatively stable, but locally divided, sinuous channels. Subsequent annual monitoring has revealed a cyclical sequence (akin to the Feshie) as the main flow switches between channels in response to local incision and aggradation immediately upstream (Fig. 19.8b). Two main conclusions arise from over 40 years of mapping and direct observation. First, although a rare flood is necessary to initiate large-scale braiding, in this case the resulting planform was unstable and promptly reverted to a set of relatively stable sinuous channels, confirming this reach's proximity to the braiding–meandering threshold. Second, had the site only been monitored between 1978 and 1981, an erroneous conclusion on the role of rare floods might have been made. Only after 40 years of observation, during which the planform has proved to be relatively stable, could the true geomorphic impact of the 1978 flood be accurately determined. This exemplifies the value of long-term monitoring of such sites if valid geomorphological inferences are to be drawn (Leopold et al. 2005).

In both case studies, the formative events were highly localized convective summer storms which, given the small catchment sizes (16–18 km^2) and high rainfall intensities, generated floods capable of exceeding local thresholds, but with contrasting outcomes. For the Allt Mòr, significant morphological change on a locally confined, stable mountain torrent is rare: only twice in the last 60 or so years. By contrast, at Dorback Burn a storm similar in rainfall intensity and flood runoff generated short-term braiding that reverted to a persistent and more stable meandering planform as subsequent floods, and lesser flows capable of triggering local incision and aggradation, initiated a pattern of channel switching between discrete sinuous channels over the next 40 years.

At a larger scale, the 1829 Muckle Spate caused widespread inundation of the Spey valley floor with localized geomorphic impacts, but these are now palimpsests with palaeochannels erased by agriculture and engineering and active channel reaches reworked by later floods (McEwen 1997). Addressing the wider stability of valley floors during 'flood rich' and 'flood poor' periods over the past 250 years, Werritty and Leys (2001) reported high rates of bank erosion and reworking of the Feshie and Avon floodplains but no evidence of a long-term change in channel state. This implies that even projected future climate change (Chap. 5) may have only a modest impact on the stability of valley floors and fluvial landforms in upland Scotland.

19.5 Conclusions

The assemblage of fluvial landforms in the Spey drainage basin, most notably in Glen Feshie, provides a remarkable record of the Lateglacial and Holocene evolution of fluvial systems over a wide range of timescales. In terms of geodiversity, especially noteworthy are relict and active alluvial fans at major confluences on the Spey and upstream tributary valleys; the exceptional sequence of Lateglacial and Holocene river terraces in Glen Feshie; extensive wandering gravel-bed rivers with both active meanders and braided reaches; and sites where the geomorphic impact of rare floods can be compared and contrasted. Throughout the Spey drainage basin, the glacial legacy is writ large in determining the role of paraglacial processes in shaping fluvial landforms following deglaciation, in providing a source of sediment for fluvial transport, and in locally determining the size of material either reworked from undercut fans and terraces or released by bank erosion. The thresholds for mobilizing such sediment sources mean that transfer from sediment sources to sinks is both episodic in time and highly localized in space. This is particularly well-illustrated in the upper braided reach in Glen Feshie, an outstanding field laboratory for testing contrasting models of braiding and one of the most intensely studied valley floors in Great Britain.

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20 Central and Eastern Grampian Highlands

Martin P. Kirkbride

Abstract

The Central and Eastern Grampian Highlands comprise a tract of extensive mid-level to high-level palaeosurfaces unlike anywhere else in the Scottish Highlands, forming a major west–east drainage divide between the Dee and Tay catchments. This undulating upland comprises broad gentle slopes, shallow plateau valleys and upstanding rounded hills with a remarkable accordance of summit elevations. The area was overrun by successive Quaternary ice sheets but the erosional imprint is selective and linear, with well-defined glacial troughs alongside valley systems showing only limited glacial modification. The plateaux were occupied by cold-based ice within the last Scottish Ice Sheet, feeding into faster-flowing glaciers in the surrounding Dee, Tay and Strathmore valleys. During the Loch Lomond Stade (~12.9–11.7 ka), local plateau icefields sourced outlet glacier lobes in many valleys, while cirque glaciers occupied topoclimatically favoured sites. Most end, lateral and recessional moraines were deposited at this time. Solifluction lobes and terraces festoon upper slopes: those on granite-derived regolith are relict, Lateglacial stone-banked lobes; those on metasedimentary rocks are active under present conditions. In the Eastern Grampians, granite intrusions form prominent hills, some bearing summit tors and blockfields. Large-scale paraglacial rock slope failures are located on steeper slopes at trough heads and margins. Localised debris-flow activity continues to rework the drift mantle, triggered by exceptional rainstorms.

Keywords

Landscape evolution • Plateaux • Selective linear glacial erosion • Scottish ice sheet • Loch Lomond Stade • Paraglacial landforms • Periglacial landforms • Rock slope failures

20.1 Introduction

The Central and Eastern Grampian Highlands comprise an extensive tract of dissected plateau north of the Highland Boundary Fault. The Central Grampians lie between the transecting valleys of Glen Garry and Glen Truim in the west and Glen Shee and Glen Clunie in the east (Fig. 20.1). The Eastern Grampians (traditionally called The Mounth) extend from the latter as far as the east coast near Stonehaven and are bounded by the fault-line scarp of the Highland Boundary Fault to the southeast. The major valley of the River Dee separates the Mounth from the Cairngorm Mountains to the north. The 500 m contour line divides the upland into five sectors: the east Drumochter, Gaick, Tilt-Shee, Lochnagar and Glen Esk massifs, with a few outlying uplands. Together, these comprise an area of 995 km^2 above 500 m elevation, with extensive ground above 800 m.

The area has received relatively little recent attention from geomorphologists, yet its high, undulating plateaux are distinctive large-scale landforms that were the subject of early speculative research into long-term landscape evolution. They are incised by several major valleys, which are enigmatically variable in their degree of glacial modification. The gross erosional form of the region has been imposed on bedrock of complex structure and variable lithology, across which the main erosional landforms do not always correspond to geological lineations.

The Grampian Highlands terrane is dominated by thick Neoproterozoic siliciclastic psammitic and semipelitic

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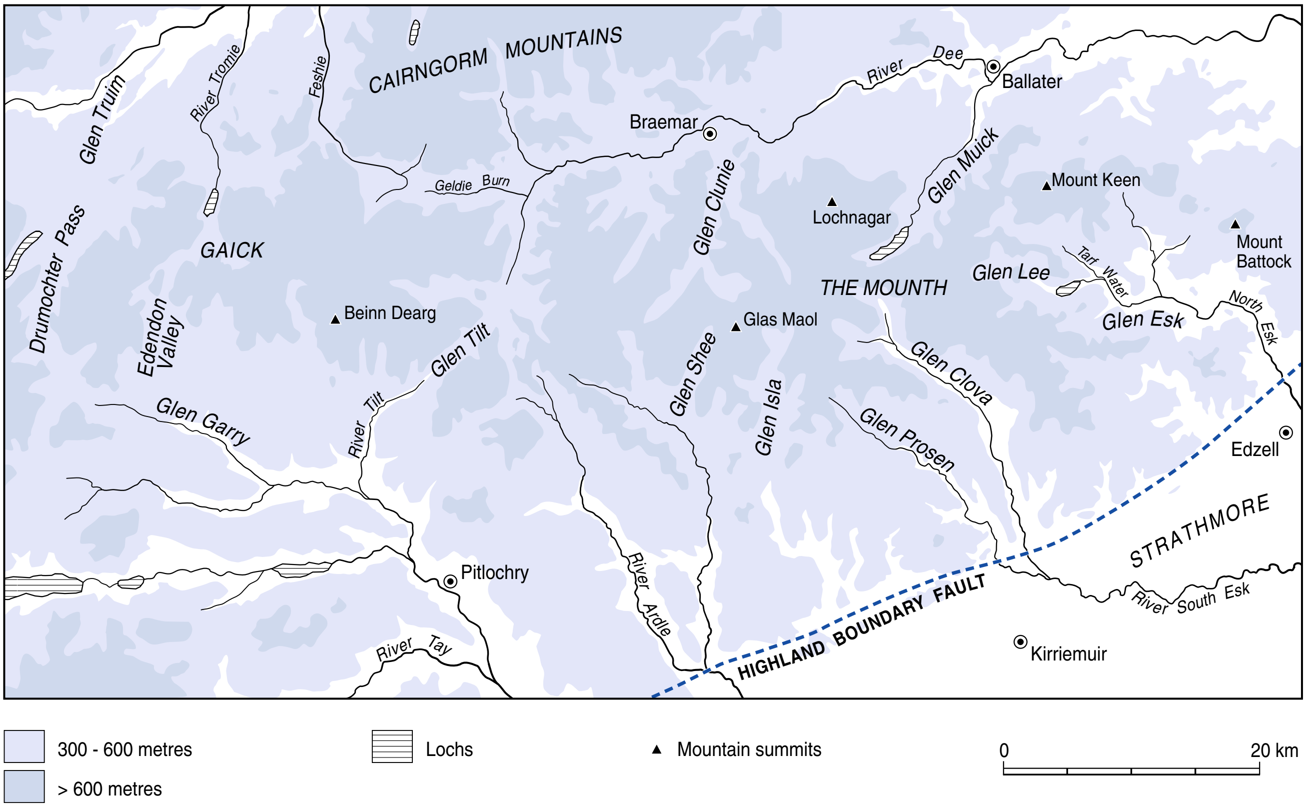

Fig. 20.1 Central and Eastern Grampian Highlands, showing key locations mentioned in the text

metasediments forming the Grampian Group, the lowermost part of the Dalradian Supergroup (Stephenson and Gould 1995), which reaches up to 5.5 km in aggregate thickness. Early in the Caledonian Orogeny (at ~470 Ma), compression created a regional-scale, southeast-facing recumbent fold with components referred to in this area as the Tarfside, Atholl and Tay Nappes. The fold limb dips regionally at up to 20–30°, steepening towards the Highland Boundary Fault. The latter was created by lateral shear and compression during the orogeny and represents one of the major ancient fault lines of Scotland, separating the Grampian Highlands terrane to the north from the Midland Valley terrane to the south (Fig. 20.1; Chap. 2). Regional metamorphic grade increases northwards from the Highland boundary from greenschist to amphibolite facies, indicating post-orogenic exhumation of up to 25 km (Gordon 2010).

Most uplift of the ancient Caledonian Mountains took place towards the close of the mountain-building phase, as large granitic plutons and dyke swarms were intruded into the Dalradian metasedimentary rocks during the Scandian orogenic event of ~435–425 Ma. These plutons include the Beinn Dearg granite and the voluminous Lochnagar and Mount Battock granites. Several smaller intrusions around the head of Glen Clova include the Glen Doll diorite, associated with major rock slope failure. The greatest relief of the Grampian Highlands occurred in this period. Between ~420 and 390 Ma, brittle deformation formed a series of NE–SW to NNE–SSW trending faults that were subsequently exploited by erosion to form valleys, including upper Glen Shee, the glacial breach of Glen Tilt and the Glen Muick trough. The present-day Grampian Highlands represent the roots of the eroded and uplifted Caledonian Mountains, which may have been affected by Cenozoic uplift of up to 1000 m (Hall 1991) and tilting associated with passive margin evolution to the west and subsidence of the North Sea Basin to the east.

20.2 Landscape Evolution: A Classic Area of Palaeosurfaces

20.2.1 Upland Morphology

The Grampian upland comprises several large-scale morphological elements. Foremost are the gently undulating to almost level plateaux of the Gaick palaeosurface in the west (800–900 m) and the Eastern Grampian palaeosurface (500–800 m) on the Mounth (Hall 1991, 2017). Within these are isolated higher hills and valleys of variable (but often slight) degrees of glacial modification (Fig. 20.2). Some

Fig. 20.2 **a** Mounth Plateau viewed from the slopes of Lochnagar to the west. The prominent domed hill to the left is Mount Keen (939 m), behind the glacial trough of Glen Muick. **b** The reverse view from the peat-mantled plateau surface to the high ground of the Lochnagar granite. Note the summit tor of Cac Carn Mór (1151 m). (Images: Martin Kirkbride)

less-modified valleys contain broad basin-like depressions. This distinctive geomorphological character is the cumulative outcome of long-term post-Caledonian weathering and erosion interspersed with phases of Cenozoic uplift and limited fluvial incision, followed by selective linear erosion during successive Pleistocene glaciations.

While the uniformity of the plateaux is remarkable, attempts to identify stepped erosion surfaces at different elevations and of different relative ages are unconvincing. Most likely, the long-term evolution of the uplands has been one of regional-scale contemporaneous development of a unified summit surface (Hall 1991), following broad crustal warping to create the initial high relief (Ringrose and Migoń 1997), rather than focussed bevelling of discrete surfaces. The underlying geology appears not to exert a primary control on the relief pattern, as upland surfaces indiscriminately truncate lithologies and structures of variable resistance to erosion. The area was one of low relief by the end of the Cretaceous and subsequently uplifted rapidly in the Palaeogene and more slowly through the Miocene (Hall 1991; Chap. 3). Unroofing of granite plutons emplaced at depths of 5–10 km implies very rapid rates of post-orogenic denudation.

The few summit tors in the Grampians are restricted to granite summits, notably Cac Carn Beag on Lochnagar (Fig. 20.2b) and Clachnaben in the far east of the region. They are fewer and smaller than the Cairngorm tors, and their survival indicates very limited glacial erosion of plateau surfaces by successive Pleistocene ice sheets.

20.2.2 Pre-Quaternary Landscape Evolution

Evidence of pre-Quaternary landscape evolution is tantalisingly fragmentary. Chemical decomposition of psammitic rocks in the Gaick area reaches preserved depths of up to ~30 m exposed in gullies around the plateau edge. Elsewhere, the saprolite has been removed by erosion. The degree of chemical alteration is modest, and the presence of halloysite clay in the regolith implies waterlogged conditions at sites that are now free draining due to later slope cutting (Hall and Mellor 1988; Hall 2004). The age of the chemical etching of the Grampian surface is unknown, but pre-dates at least the last Devensian glaciation and may indeed be older than the Quaternary glaciations as a whole. Preservation of saprolites indicates low denudation rates during the Quaternary (Hall and Sugden 1987), assuming that 'sandy grus' saprolites are of Pliocene or earlier age.

The Central and Eastern Grampians contain evidence of some of the most dramatic large-scale river captures in Scotland, resulting from glacial modification of watershed topography. Several pre-glacial rivers that formerly followed a west-to-east regional slope have been diverted into the headwaters of other rivers incised into the flanks of the plateaux. A famous example is the glacial breaching across the Feshie–Dee watershed, which diverted the headwaters of the Geldie Burn into the River Feshie, beheading the River Dee. Ten kilometres to the south, the upper Tarf Water, formerly within the Dee catchment, was diverted southwards into the River Tilt, crossing the regional watershed in the process. Such captures are often manifest as prominent knickpoints (e.g. the Falls of Tilt) or acute elbow-like changes of channel direction.

20.3 Landforms of Glacial Erosion

The last (Late Devensian) Scottish Ice Sheet existed between ~35 and 14 ka (Ballantyne and Small 2019) as a dynamic multi-centred ice sheet comprising accumulation domes, migratory ice divides and ice streaming that was probably variable in time and space (Hubbard et al. 2009). Because the main north–south ice divide lay in the Western Grampian Highlands, regional ice movement across the Central and Eastern Grampians during the Last Glacial Maximum was eastward and focussed along Strathspey and the Dee valley to the north and Strathmore to the south. The Central and Eastern Grampian plateau nevertheless operated as a persistent subsidiary ice divide for much of the ice sheet's existence (Hughes et al. 2014), separating and feeding the Dee and Strathmore glaciers. Within the eastern Highlands as a whole, however, it appears that only the high Cairngorm plateau maintained an independent accumulation centre throughout the lifetime of the last ice sheet (Chap. 18). Easterly ice movement across the central and eastern plateaux is demonstrated by the occurrence of granite erratics from the Rannoch Moor ice centre across the high ground of the Gaick.

The geomorphological signature of successive Quaternary ice sheets is more limited than in other parts of the Scottish Highlands. Large-scale glacial landforms evolved over multiple episodes of ice sheet and valley glaciation since the earliest Quaternary (Chap. 4), and the erosional topography was established before the last Devensian ice sheet. Situated in the snow shadow of the west Highlands ice divide, ice accumulation in the Central and Eastern Grampians was more gradual, and relatively thin, cold-based ice cover developed over high ground (Hubbard et al. 2009). In consequence, large-scale landforms of ice sheet erosion are concentrated where convergent flow into valley heads crossed the thermal boundary into warm, sliding basal conditions (Hall and Glasser 2003). Where plateau ice descended the trough heads, these are roughly ice scoured and stepped, as at the head of Glen Clova (Fig. 20.3a). This selectivity of glacial erosion has had the dual effects of preserving the pre-glacial topography at high elevation and imparting varying degrees of glacial topography to the main valleys.

The deepening and widening of valleys were highly selective and appear unrelated to the initial scale or orientation of pre-glacial valleys relative to regional ice dispersal (Jarman 2019a). Most troughs are incised 300–500 m below surrounding plateaux, and several (Glens Callater, Muick, Lee and Clova) contain rock basins immediately downvalley from tributary valley confluences, suggesting that accelerated glacier flow at such locations resulted in overdeepening of the trough floors. Most of the rock basins are less than 40 m deep, but Loch Muick reaches a maximum depth of nearly 80 m (Lowe et al. 1991), while a likely overdeepening in Glen Clova is infilled with sediment. Elsewhere, glacial deepening of valley floors has been modest, and well-developed troughs are limited to the Gaick-Edendon breach and Glen Tilt, which are both discordant to regional ice flow, and Glens Shee, Isla, Clova and Muick. The largest valley system in the area, Glen Esk, is also the most easterly and shows little sign of significant glacial excavation downvalley from Loch Lee. Glen Prosen similarly exhibits limited evidence of glacial modification beyond its upper reaches, though both glens are favourably oriented with

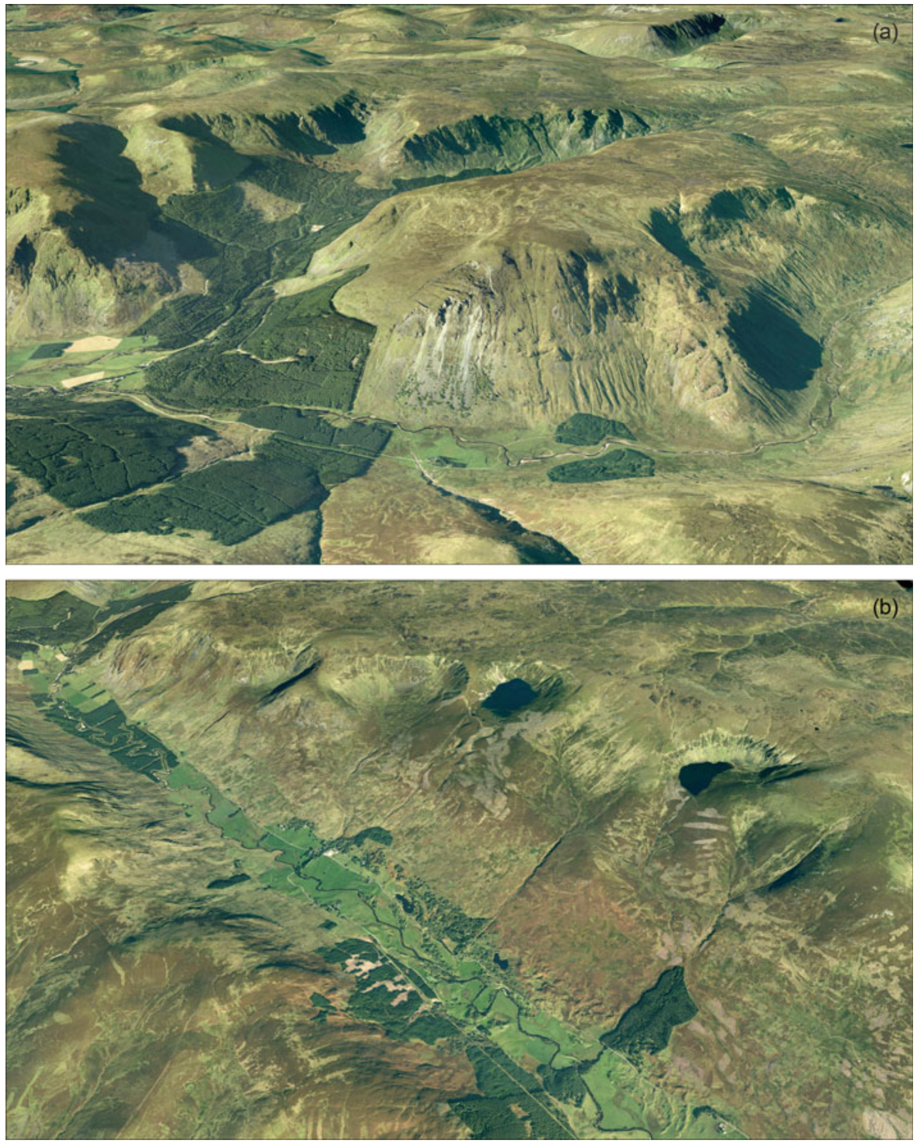

Fig. 20.3 **a** Trough heads and corries (cirques) excavated by successive Pleistocene glaciers in the Eastern Grampian palaeosurface at the head of the Glen Clova trough. In the centre of the image is the Cairn Broadlands rock slope failure. **b** The Glen Clova trough from the south, illustrating selective linear glacial erosion of the Eastern Grampian palaeosurface. Corrie Brandy is the central cirque of the three scalloped into the upper margins of the trough. Unusually, all three cirques face south. (Images: Google Earth™)

regard to former directions of ice flow. The classic trough of Glen Clova is straight, steep-sided and bordered by hanging cirques (Fig. 20.3b).

20.4 Landforms Associated with Wastage of the Last Scottish Ice Sheet

The timing of deglaciation of the upland Eastern Grampians is unknown and contested, and parts of the present east coast may have been ice free as early as ~20 ka (McCabe et al. 2007; Phillips et al. 2008). Terrestrial cosmogenic nuclide and radiocarbon ages from Strathspey to the northwest and the Highland boundary to the south imply deglaciation of low ground by ~15.5 ka, suggesting that ice margins had retreated to the glens and high ground of the Central and Eastern Grampians several centuries before the rapid warming that marked the onset of the Lateglacial Interstade of ~14.7–12.9 ka (Ballantyne and Small 2019).

Ice-marginal landforms relating to deglaciation include the Glenarm moraine in Glen Clova (Mitchell et al. 2019), a ridge complex associated with thick glacifluvial/glacilacustrine deposits where the Glen Clova glacier separated from the Strathmore Ice Stream. The 'moraine' is a complex of sandy, pebbly, subaqueous ice-contact ramps and mounds formed where the Glen Clova glacier, having detached from the Strathmore Ice Stream and retreated upvalley, readvanced into an ice-dammed lake ponded in the lower valley. These landforms are important in revealing the dynamics of detachment of Highland and lowland ice during ice sheet decay. They indicate that the northeastwards flow of the ice stream along the mountain front maintained its activity during a phase of wastage and detachment of locally nourished valley glaciers, consistent with precipitation starvation in the Eastern Grampians, while relatively vigorous ice flow was maintained from the Tay valley.

The evolution of the confluent suture between local Grampian glaciers and the Strathmore Ice Stream is marked by extensive meltwater channels and associated eskers along the Highland Boundary Fault zone (Chap. 25). Some of these meltwater channels extend for significant distances and rock-cut subglacial meltwater channels up to 60 m deep extend along the northwestern margin of the fault zone. One channel system extends ~20 km from the mouth of Glen Clova, discordantly crossing the present drainage network, and is followed by the Paphrie Burn until being buried by a glacifluvial fan system at the mouth of Glen Esk. Later during deglaciation, shorter but equally impressive systems of channels exploited the fault zone. The Quharity Den and Audallan channels are fine examples (Mitchell 2019; Mitchell and Russell 2019). The Audallan channel extends >4.5 km and reaches depths of up to 100 m. These channels probably operated subglacially, possibly under successive ice sheets, where englacial drainage converged along the suture between Highland and lowland ice, later forming antecedent routes for ice-marginal drainage. Channels were often associated with large masses of decaying ice, where kame, kame terrace and esker assemblages were formed during late-stage decay and westward retreat of the Strathmore ice margin.

20.5 Glacial Landforms of the Loch Lomond (Younger Dryas) Stade

During the Loch Lomond Stade (~12.9–11.7 ka), mean July sea-level temperatures fell to 8.5 °C or slightly lower (Brooks et al. 2012) and permafrost extended onto low ground (Ballantyne 2019), implying mean annual air temperatures no higher than −4 °C (Chap. 4). Glacier ice developed over much of the Western Highlands but was less extensive in the Central and Eastern Grampians because of limited snowfall. Former glacier distributions demonstrate that equilibrium line altitudes of stadial glaciers rose eastwards across the Grampian Highlands as a result of snow scavenging by the West Highland Icefield (Chandler et al. 2019). However, equilibrium lines in the Central and Eastern Grampians also rose northwestwards as a result of snow nourishment from the North Sea. This is consistent with palaeowind directions inferred from aeolian abrasional features in the region (Christiansen 2004). In this area, glacier cover diminished with distance from the contemporary coastline, which in the northern North Sea lay ~10 km beyond the modern coast (Stoker et al. 2009). Palaeoclimatic reconstructions based on the altitudes of stadial glaciers (Sissons and Sutherland 1976; Kirkbride et al. 2015) indicate summer temperatures ~7 °C below present and an arctic-maritime climate. Eastern glaciers may have received most snow accumulation in summer, when the North Sea was unfrozen (Golledge 2010).

In the Eastern Grampians, the glaciation limit rose northwestwards from the Highland boundary by ~10 m km^{-1} based on the equilibrium line altitudes of former glaciers (Sissons and Sutherland 1976). Inland aridity meant that the highest terrain around Lochnagar was able to support only small cirque glaciers, which have left well-preserved terminal moraines, while lower eastern parts of the Eastern Grampian plateau were able to sustain thin but extensive mountain ice caps that fed marginal outlet glaciers (Sissons 1972). The geomorphological evidence for these ice caps is fragmentary, because they were probably cold-based, slow-moving ice bodies which were able to create recognisable landforms only where flow converged into the trough heads of larger glens and cirques.

Glaciation of the Drumochter and Gaick hills at this time was limited to a small icefield west of Drumochter Pass (Benn and Ballantyne 2005) and an ice cap on the western Gaick Plateau (Chandler et al. 2019). Well-preserved moraines exist in the Edendon valley and Gaick Pass area, where glaciers descended steeply down connecting valleys, creating a double ice dam to impound a 1.2-km-long lake at the location of Loch an Duin in the Edendon-Tromie breach. Recessional ice-marginal moraines in the Gaick area typify the landsystems of many Scottish glens (Lukas 2006; Bickerdike et al. 2018; Chandler et al. 2018; Chap. 4), with numerous hummocky ridges forming arcuate nested sets which reveal a pattern of recession of valley glaciers towards their high-level ice cap accumulation zones (Chandler et al. 2020).

The terminal geomorphology of stadial glaciers in the area is varied (Fig. 20.4) and includes single terminal moraine ridges in cirques, distal rims of bouldery till sheets, multiple nested small ridges, small piedmont lobe moraines (Fig. 20.4c) and larger valley moraine and meltwater channel complexes. The variety reflects the range of glacier types, from independent cirque glaciers that received abundant rockfall debris from steep headwalls, to debris-starved cold-based plateau outlet glaciers. This fragmentary landform record makes inter-valley correlation difficult, especially when more than one moraine system exists, as in Glen Callater and Glen Muick. In Corrie Fee in Glen Clova and Caenlochan Glen in Glen Isla, basal radiocarbon dates from peat inside terminal moraines are consistent with moraine formation during the stade (Huntley 1981), but more direct dating of Loch Lomond Stadial deposits in the Grampian Highlands is lacking. The assignment of ice-marginal

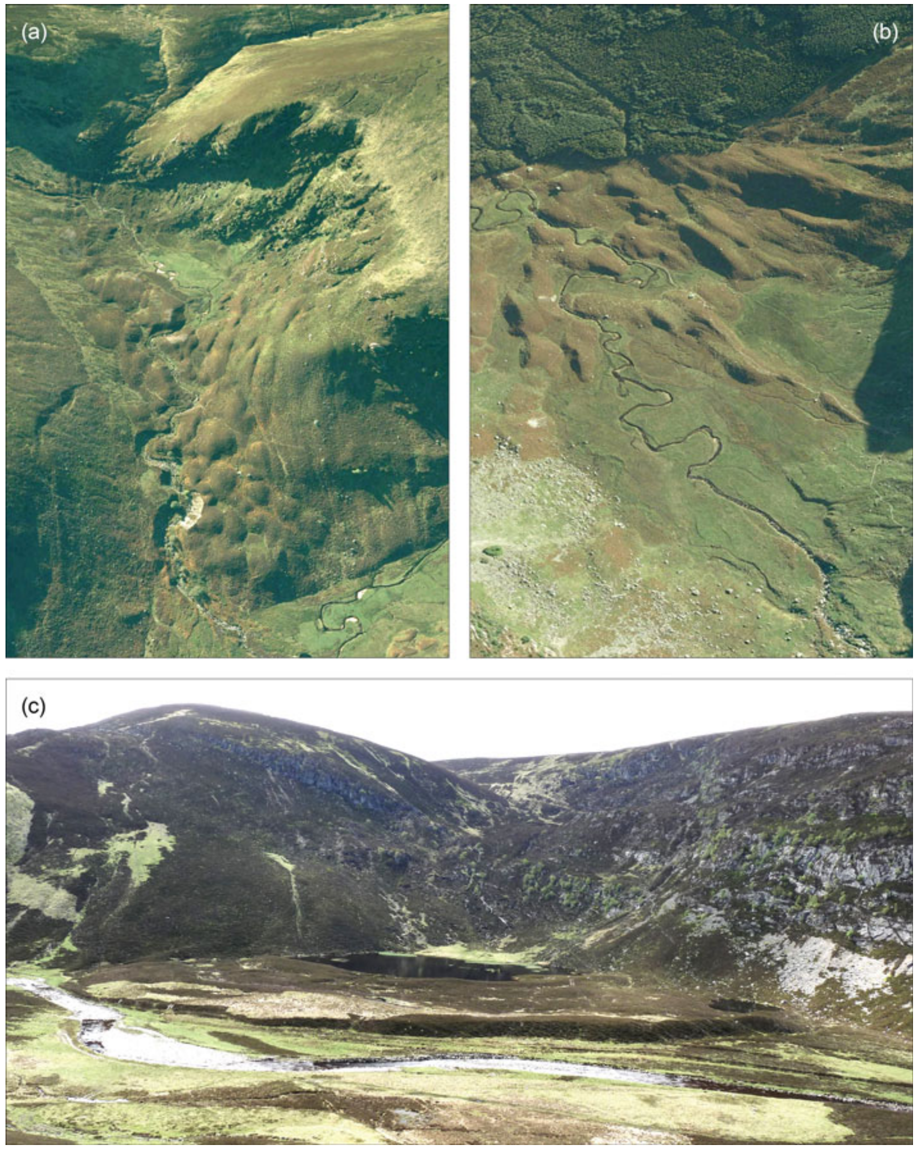

Fig. 20.4 Moraines of Loch Lomond Stadial age in the Eastern Grampian Highlands. **a** Longitudinally oriented hummocky moraines in Glen Effock. **b** Overridden, streamlined arcuate moraines in Corrie Fee. **c** Arcuate piedmont lobe moraines deposited by a glacier fed by a plateau icefield at Carlochy, Glen Esk. (Images: **a**, **b** Google Earth™; **c** Martin Kirkbride)

landforms to this period is based on inference from dated deposits in adjacent areas, notably the Cairngorms (Kirkbride et al. 2014; Chap. 18).

Two cirque floors are of particular interest, being mantled by hummocky morainic mounds streamlined in the direction of former ice flow. In Glen Effock, a tributary of Glen Esk, morainic mounds decrease in elongation from the back to the lip of the Corriedoune cirque (Fig. 20.4a). In Corrie Fee, moraines take the form of nested arcs suggesting ice-marginal moraine ridges, upon which streamlining and stoss-lee forms have been superimposed (Kirkbride 2019; Fig. 20.4b). A single unambiguous interpretation of each is difficult: they may involve deposition of supraglacial mass movement debris at the former glacier margins followed by overriding, or a two-stage Loch Lomond Stadial glacier advance.

20.6 Postglacial Landforms

20.6.1 Rock Slope Failures

Deep-seated paraglacial rock slope failures (RSFs) are sporadically distributed throughout the region. Though most of the Mounth region lacks large RSFs, local concentrations occur in steep trough-head and trough-flank locations in Glen Clova (Fig. 20.3a) and in the glacial breach of Glen Tilt (Jarman and Harrison 2019) where individual RSFs cover areas of up to 1.5 km^2. Over the Grampian Highlands as a whole, $\sim$0.26% of the ground area has been subject to deep-seated mass movement, rising to $\sim$5.4% where local concentrations occur. The low RSF density compared to some parts of Scotland is likely due to the generally shallower valleys and lower valley-side slope angles of the Central and Eastern Grampians.

Cosmogenic nuclide exposure dating of RSFs in the Scottish Highlands has shown that most are paraglacial features of Lateglacial or Early Holocene age (Ballantyne et al. 2014). Local examples of such paraglacial slope failures include the Cairn Broadlands RSF in Glen Clova (Figs. 20.3a and 20.5a, b), which has left spectacular lateral rifts and antislope scarps where slope failure has exploited the boundary of the Glen Doll diorite intrusion. In contrast, a subdued RSF in Glen Esk (Fig. 20.5c), and possibly the Loch Brandy RSF (Jarman 2019b; Fig. 20.5d), may pre-date the Loch Lomond Stade, during which they may have been modified by small local glaciers. A translational RSF in Glen Shee appears to be of a type unique in the Eastern Highlands (Ballantyne 2007). The mass movement involved $\sim$ 0.6 Mt of interbedded quartzite and semipelitic schist which during motion transitioned from slab failure to viscous flow to produce two elongated debris tongues up to 380 m long (Fig. 20.5e). The unusual rheology may be due to the mobile rock slab loading saturated till downslope.

20.6.2 Periglacial Landforms

Tors and blockfields on plateaux in the Grampian Highlands (Fig. 20.6) are essentially pre-Devensian features that were preserved largely intact under a cover of cold-based ice during periods of Devensian glaciation (Phillips et al. 2006; Hopkinson and Ballantyne 2014). On granite terrain, the most spectacular postglacial periglacial landforms are suites of bouldery stone-banked solifluction terraces and lobes with risers up to 2 m high, such as those that festoon the upper slopes of Beinn Dearg, Lochnagar and Mount Keen. These lobes descend locally to altitudes of 580 m but are absent inside the limits of Loch Lomond Stadial glaciers, implying that they were last active under the extreme periglacial conditions during the Loch Lomond Stade. They are thought to have formed through downslope rafting over Lateglacial permafrost of a surface layer of boulders upon a deforming layer of finer sediment (Ballantyne and Harris 1994). Sandy regoliths developed on granite have been susceptible to aeolian reworking in the Holocene, giving rise to stony deflation surfaces and localised high-level sand beds such as on the Lochnagar plateau.

On the frost-susceptible, silt-rich regolith produced by weathering of Dalradian schists, solifluction has been active on upper slopes (over $\sim$700 m) throughout much or all of the Holocene, forming suites of vegetation-covered solifluction terraces and lobes with steep risers up to a metre high. Such features are prominent on the upper slopes of Glas Maol and the Gaick and Drumochter hills, and excavations through individual lobes have revealed Holocene soils buried by lobe advance, though present movement appears to be limited to the uppermost 40 cm of the soil (Ballantyne and Harris 1994). Intimately associated with areas of active solifluction are ploughing boulders, moving at rates of up to $\sim$10 mm a^{-1} on the steepest slopes. Active frost-sorted patterned ground is rare on plateau areas, most of which are mantled by hill peat and complete vegetation cover, but on areas of vegetation-free ground there are occasional pockets of apparently active sorted nets and sorted stripes up to metre wide. Excellent examples occur near the summit of Cairn Bannoch (1012 m) at the head of Glen Muick. Larger, relict forms in coarse detritus occur around the summits of Glas Maol (1068 m) and nearby Cairn of Claise (1064 m).

20.6.3 Lateglacial and Holocene Alluvial Landforms

In common with many Highland glens, Grampian valleys contain fills of glacifluvial and Holocene alluvial sediment, flanked by lower slopes mantled by till or soliflucted till. The sediment mantle on the lower valley sides is frequently incised by hillslope gullying. Gullies are usually incised to bedrock and distributed tens to hundreds of metres apart across the hillside. Gully depths and densities are lower than in comparable catchments in the western Highlands, and the great majority of the reservoir of glacial sediment remains on the slopes, limiting the potential for fluvial construction of large alluvial forms in the area.

Where slope channel coupling has been high, debris discharged from valley-side gullies has supplied the fluvial system directly. Well-coupled systems tend to be steep, low-order headwater streams. Many slopes, however, are not coupled to river channels, and colluvial cones and aprons are distributed below sites of paraglacial sediment reworking by channelized run-off (Figs. 20.7a, b). The timing of Holocene alluviation and fan or cone accumulation has been established for only upper Glen Feshie on the northern margin of the Gaick and the Edendon valley on its southern margin. In both glens, there is radiocarbon-dated evidence for Late Holocene (<4.2 ka) floodplain aggradation followed by river incision, producing a low terrace $\sim$3–4 m above the active floodplain (Robertson-Rintoul 1986; Ballantyne and Whittington 1999), and Ballantyne (2008) has suggested that this Late Holocene terrace is a widespread feature in the Grampian glens. Large debris cones in upper Glen Feshie are also of Late Holocene age, with most sediment accumulating in the past 350 years (Brazier and Ballantyne 1989), and a small alluvial fan in the Edendon valley accumulated in just three storm-generated run-off events after $\sim$2.2 ka (Ballantyne and Whittington 1999). The causes of Late Holocene colluvial and alluvial events are uncertain. Although

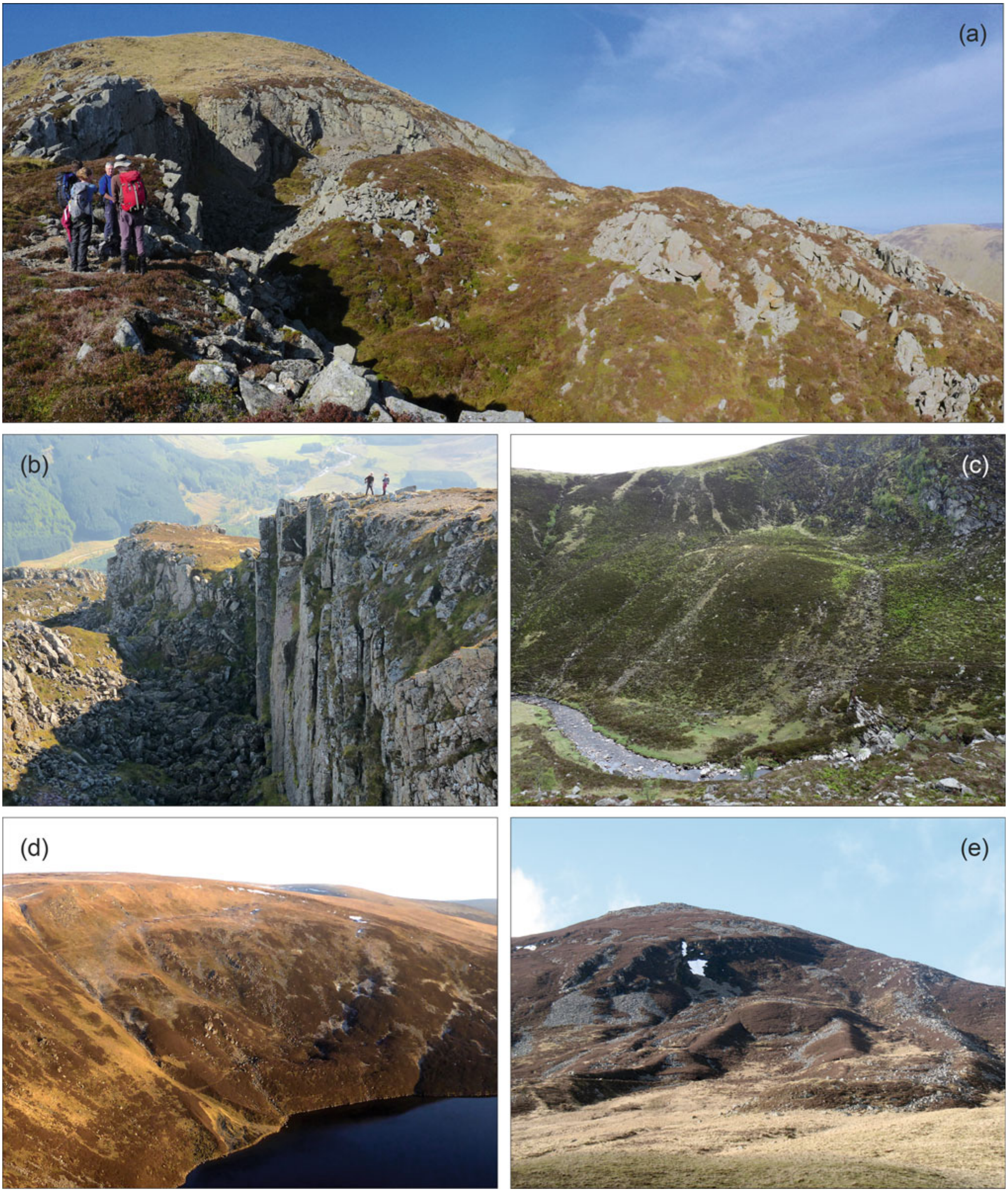

Fig. 20.5 Deep-seated rock slope failures. **a** Headscarp of the Cairn Broadlands RSF with transverse antislope scarps to right of picture. **b** Right-lateral rift of the Cairn Broadlands RSF. **c** Arrested translational slide in upper Glen Esk. The hollow contains small, relict moraine ridges, suggesting paraglacial slope failure prior to the Loch Lomond Stade. **d** Large, relict RSF in the Corrie Brandy sidewall. **e** Complex rockslide/debris flow at Ben Gulabin, Glen Shee. (Images: **a–d** Martin Kirkbride; **e** Colin Ballantyne)

Fig. 20.6 Granite tor of Cac Carn Beag rising from the deflated summit plateau of Lochnagar, at 1155 m the highest point in the Eastern and Central Grampian Highlands. (Image: Martin Kirkbride)

elsewhere in Scotland Late Holocene alluvial fan accumulation has been attributed to anthropogenic land use changes (Ballantyne 2019), the pollen record for the Edendon valley fan provides no evidence of vegetation change prior to the onset of fan accumulation. It is possible that these Late Holocene events are linked to general climatic deterioration and an increased frequency of exceptional rainstorms, but the available dating evidence is inadequate to substantiate this interpretation.

20.7 Recent Slope Activity

20.7.1 Debris Flows

Most hillslope gullies in the region are vegetated and lie above relict colluvial cones. Only very occasionally do some gully systems become activated and produce debris flows in response to heavy or prolonged rainfall (Chap. 5). Sediment sources include the flushing of rockfall debris that has accumulated in cliff-bounded gullies in very steep terrain, or shallow translational slides within till cover, or full-depth regolith failures over bedrock. A recent example of rainstorm-triggered debris-flow activity occurred on 17 July 2015, when up to 60 mm of rain fell along the Highland boundary within 6 h, an event with an estimated return period of $\sim$140 years (SEPA Flood Unit 2015). This triggered widespread debris flow activity from debris-charged gullies (Figs. 20.7a, b) as well as flooding the town of Alyth, close to the Highland boundary. This example highlights that geomorphological thresholds in the Grampian Highlands are occasionally crossed and formative events may occur as part of ongoing paraglacial landscape adjustment.

20.7.2 Rockfall Activity

Rockfall talus slopes are common below degraded cliffs in glacial cirques and trough walls, but they are not ubiquitous. Cliffs of high rock mass strength in Corrie Fee, for example, have only small talus cones accumulated at the cliff base, and in some cases the postglacial response of the rock mass has been to generate infrequent large rockfalls (Fig. 20.7c). Most talus accumulations are relict, mantled with a thick soil and vegetation mat and seamed with terracettes and animal tracks. Inactivity is a consequence of a drastic reduction in rock wall retreat rates since the climatic warming at the start of the Holocene. The character of relict talus slopes across the Grampian Highlands is consistent with Ballantyne and Harris's (1994) estimate of a 100-fold decrease in activity between the Loch Lomond Stade and the present day, and many large talus accumulations probably were most active at the close of the stadial period.

There is, however, localised evidence of recent rockfall activity. In the North-East Corrie of Lochnagar, a series of large granite flakes collapsed in the years following 1995 (Kirkbride 2005). In Corrie Brandy, rockfalls since 2017 form part of a progressive long-term failure of the cirque wall (Jarman 2019b, Fig. 20.7d). The rockfalls have been generated from the collapsing edges of slumping flakes of weak schistose bedrock.

20.8 Conclusions

The Central and Eastern Grampian Highlands form a major relief feature and a distinctive landscape within upland Scotland. They are characterised by extensive dissected plateau surfaces whose origins are related to Cenozoic uplift, deep chemical weathering, stripping of saprolite covers and

Fig. 20.7 Landforms of Holocene slope activity in Glen Clova. **a** Holocene debris cone and rock slope gully in the Burn of Kilbo, with recent (July 2015) debris-flow deposits in the cone centre. **b** Recent debris-flow deposits in Corrie Fee, triggered by an intense summer rainstorm in July 2015 and deposited on Holocene debris cones. **c** Relict, fragmented rockslide from a failure niche in Corrie Fee, covering vegetated moraines of Loch Lomond Stadial age. **d** Rare recent (2017) rockfall in Corrie Brandy, entirely resurfacing a talus slope (Source in shadow). (Images: Martin Kirkbride)

dissection by rivers radiating from a west–east drainage divide. Modification of the end-Neogene landscape during successive Pleistocene glaciations has taken the form of selective linear glacial erosion operating along the pre-existing valley system, with cold-based ice occupying plateaux and preserving ancient blockfields and tors. In comparison with the Western Grampians, the effects of long-term glacial modification have been muted, except at trough-head locations. During the Loch Lomond Stade, plateau ice caps fed outlet glaciers that extended only short distances along adjacent troughs, and some high cirques were reoccupied by small glaciers. Retreat of the last ice sheet was accompanied by the formation of meltwater channels and associated glacifluvial deposits; most of the moraines in the region were deposited during the Loch Lomond Stade. Both Lateglacial and Holocene solifluction landforms occupy the highest slopes. Since deglaciation, the landscape has been dominated by paraglacial activity in the form of scattered rock slope failures, accumulation of talus, and reworking of glacial drift by rivers and debris flows. There is evidence, however, for floodplain aggradation and incision and the deposition of alluvial fans and debris cones during the Late Holocene. Debris-flow and rockfall activity are locally and intermittently active at present.

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21 The Buchan Palaeosurface

Adrian M. Hall

Abstract

Buchan displays some of the clearest evidence for long-term regional geomorphological development in Scotland. The metamorphic and igneous basement was first exposed in the Devonian and repeatedly buried and re-exposed later in the Palaeozoic and Mesozoic. Greensand, chert and flint remnants attest to a former Cretaceous sedimentary cover. The highest terrain in eastern Buchan supports remnants of deep kaolinitic weathering profiles and flint and quartzite gravels. Oxygen isotopes indicate that kaolinization occurred in two phases at mean annual temperatures of 23 and 15 °C, indicating a Palaeogene age for the oldest flint gravels. Sandy deep weathering mainly formed from the Late Miocene onwards. By the onset of Pleistocene glaciation, differences in mineralogy and fracturing had been etched out by long-term weathering to produce basement terrain that was subtly but pervasively adjusted to rock type and structure. Weathering patterns exerted significant control over glacial and periglacial processes, but the impact of Pleistocene glacial erosion was modest beneath mainly cold-based glaciers. The Buchan palaeosurface shares many characteristics with low-elevation crystalline platforms inside and outside the limits of Pleistocene glaciation elsewhere in Europe.

Keywords

Palaeosurface • Unconformity • Deep-weathering profiles • Kaolin • Flint • Neogene

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21.1 Introduction

The Buchan lowland lies between the Rivers Deveron and Ythan in northeast Scotland (Fig. 21.1). Its landscape has four characteristics that are distinctive in a Scottish context: (i) Buchan has remained as a slowly eroding platform since the Devonian; the platform was buried beneath successive sedimentary covers in the Palaeozoic and Mesozoic and later re-exposed; (ii) the landscape retains Neogene or older kaolinitic quartzite and flint gravels and sands, along with deep-weathering profiles; (iii) the impact of glacial erosion beneath successive Pleistocene ice sheets has been remarkably limited; and (iv) the region includes some of the longest and most detailed stratigraphic records of Middle and Late Pleistocene environments in Scotland. In its long-term geomorphological development, the Buchan palaeosurface resembles other crystalline terrains in Europe both inside and outside the limits of Pleistocene glaciation (Hall 2005).

21.2 The Geological Framework

The metamorphic and igneous rocks of the Eastern Grampian Highlands belong to the Dalradian Supergroup, part of the eroded root zone of the Caledonian mountain belt (Stephenson and Gould 1995; Chap. 2). An original suite of Dalradian sedimentary rocks was much altered by heat and pressure, and intensely folded and faulted, during low-grade ('Buchan-type') metamorphism in the Caledonian Orogeny at ~490–470 Ma. Metamorphism formed quartzites, psammites and pelites, with widely varying mineralogy and generally with high fracture densities. Large volumes of basic magma were emplaced at ~470 Ma to form extensive layered basic and ultrabasic intrusions. Acid magmas were also intruded at this time, including the Strichen Granite, and in a later magmatic phase at 425–395 Ma, including the Peterhead Granite. The so-called Older Granites are mainly grey, muscovite-biotite granites, typically foliated; the

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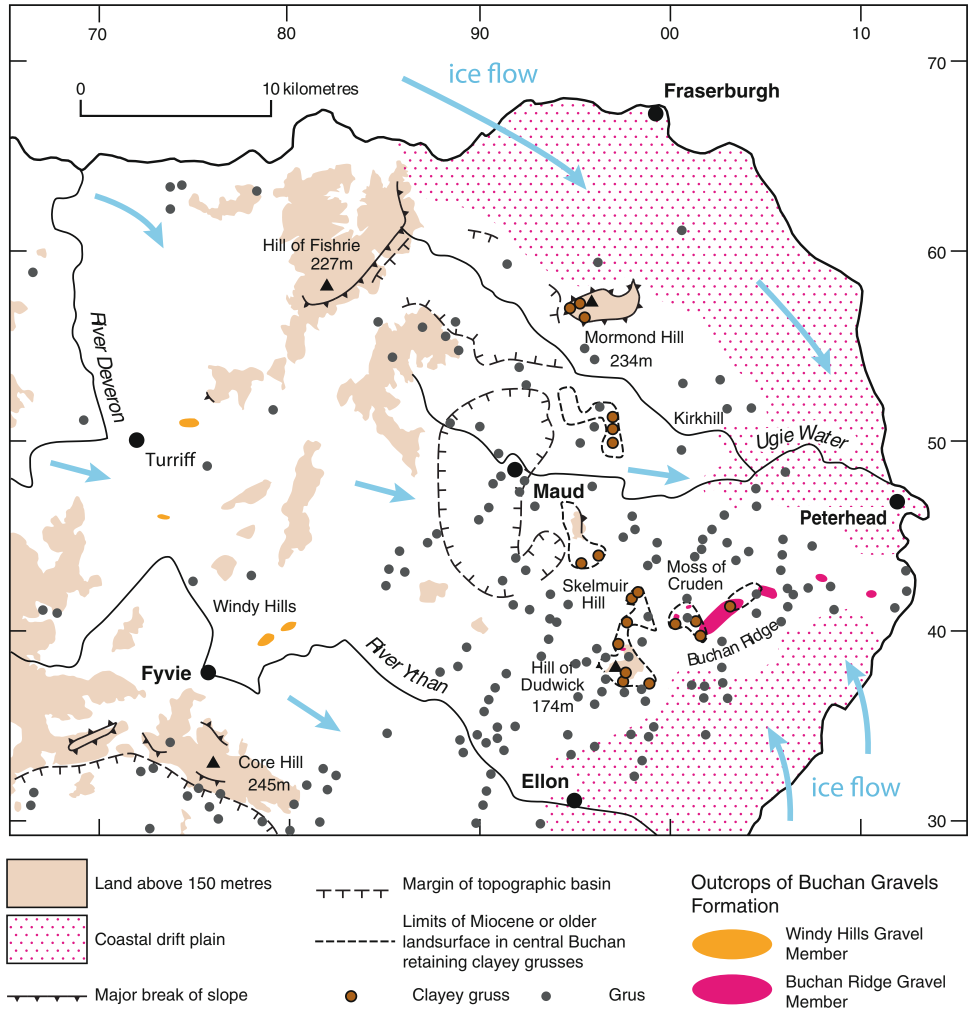


Fig. 21.1 Main features of the Buchan palaeosurface. (Modified after Merritt et al. (2003); adapted from figure P915269 on BGS Earthwise, page version ID 34592 © UKRI)

Younger Granites are pink, coarse-grained biotite granites. These diverse metamorphic and igneous rocks form the basement of the Buchan platform (Fig. 21.2). The principal structures trend mainly NNW–SSE and NNE–SSW in the basement, but north–south faults bound the Turriff Devonian sandstone basin.

The landscape of Buchan bears a strikingly anomalous relationship with its Highland geology, forming one of the largest areas of lowland north of the Highland Boundary Fault. Granites and quartzites occur only as low hills in Buchan, yet these rocks form prominent summits in the Cairngorms and the Western Grampians. When did this levelling of the basement relief occur?

21.3 Erosion and Burial of the Buchan Platform Prior to the Cenozoic

The Caledonian Orogeny ended in the Early Devonian, a period when the Grampian Highlands lay near the Equator. Devonian conglomerates and sandstones were deposited under semi-arid conditions on river floodplains and alluvial fans (Trewin and Thirlwall 2002). Devonian sedimentary rocks lap around the present coastline, and outcrop in a half graben at Turriff (Fig. 21.2). During deposition of these sediments, Buchan was a hilly to mountainous terrain. This exhumed terrain is most clearly seen farther west in

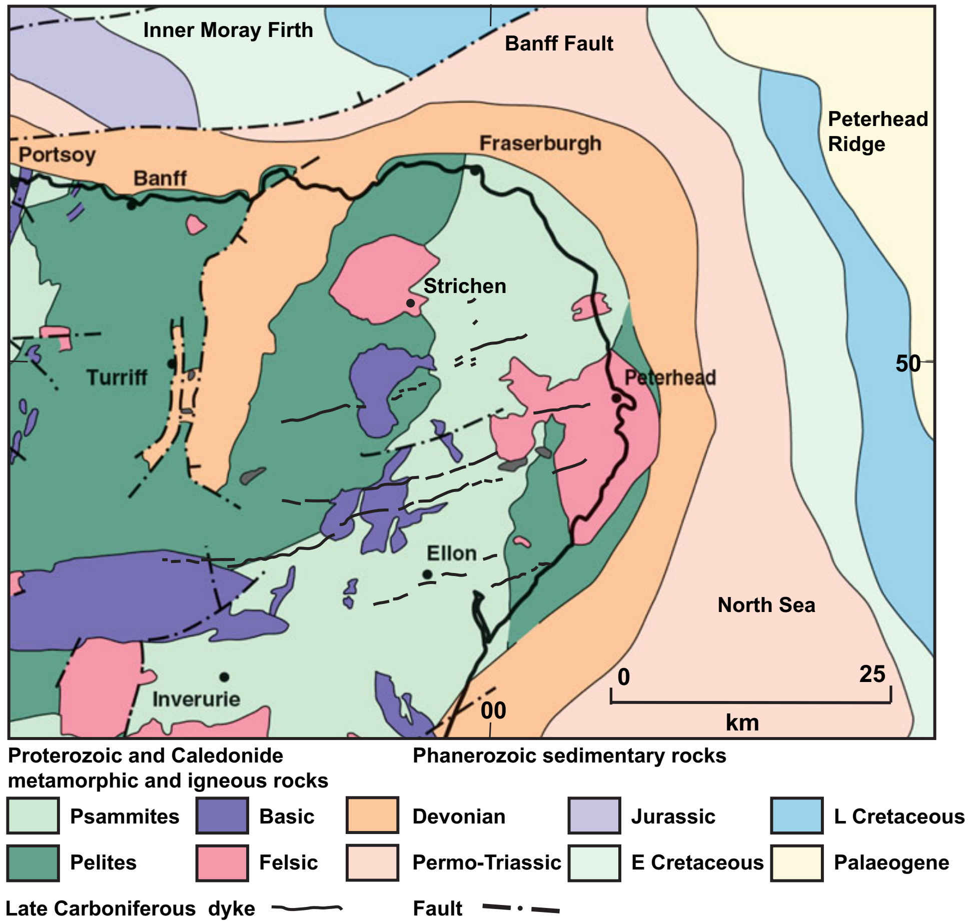


Fig. 21.2 Schematic solid geology of Buchan and the offshore area. (Contains British Geological Survey materials © UKRI 2020)

Banffshire, where small outliers of Devonian rocks occupy basin and valley floors bounded by steep slopes and overlooked by hills representing the stumps of Devonian mountains. The general accordance of the sub-Devonian unconformity and the present landscape in the eastern Highlands requires that the depth of erosion in basement over the last ~400 Ma has been limited (Macdonald et al. 2007). This is strikingly illustrated at Moss of Cruden, west of Peterhead, where three unconformities are superposed. The weathered surface of the Peterhead Granite is overlain by the Devonian Smallburn Sandstone, whilst nearby lies a small outlier of Early Cretaceous greensand which, in turn, is cut into and overlain by the flint-bearing Palaeogene Buchan Ridge gravels; the compressed geological sequence occupies an elevation range of less than 40 m (Merritt et al. 2003).

Middle Devonian uplift and erosion led to formation of an erosional unconformity that extends across the basement beneath the upper Devonian sequence west of Buchan. Late Carboniferous dolerite dykes oriented west–east occur west of Peterhead (Read 1923; Fig. 21.2). A lack of vesicles in these dykes indicates an original emplacement at depths of >0.5 km for rocks currently at the surface and hence suggests the presence of thick Upper Devonian–Lower Carboniferous sedimentary cover in the Late Carboniferous. The Early Permian was a time of desert conditions when Scotland lay ~15° north of the Equator. The Moray Firth area was surrounded by arid uplands. Permo-Triassic sandstones occur close offshore (Fig. 21.2) and likely extended across Buchan. By the end of the Triassic, the outer Moray Firth area formed part of a wide plain (Andrews et al. 1990) which probably extended across Buchan. Jurassic rocks are absent east of Peterhead (Fig. 21.2); the Buchan platform was likely uplifted and eroded in the Early and Middle Jurassic, with further thinning of its Palaeozoic cover and progressive exposure of basement (Hall 1987). The presence of Early Cretaceous chert and Late Cretaceous flint clasts in the Palaeogene Buchan Gravel Formation (Sect. 21.4) indicates former sedimentary covers of these rocks, resting on basement and Devonian sandstone, and thinning westward. The development of low relief on the Buchan platform was progressive and sequential, with initial exposure of basement in the Early Devonian and subsequent thinning of sedimentary cover and re-exposure of the basement to erosion in the Late Devonian, Late Triassic, Jurassic and Early Cretaceous.

The limited depths of erosion since stabilisation of the basement and the multiple cycles of erosion and burial evident in Buchan are features that occur on other European shields and platforms (e.g. Lidmar-Bergström 1993; Bessin et al. 2015; Gunnell 2020; Hall et al. 2021). In Buchan, and on these other platforms, long-term erosion rates in the basement have been low due to long residence times close to base level, cyclic burial and limited uplift.

21.4 Palaeogene and Neogene Weathering and Landform Development

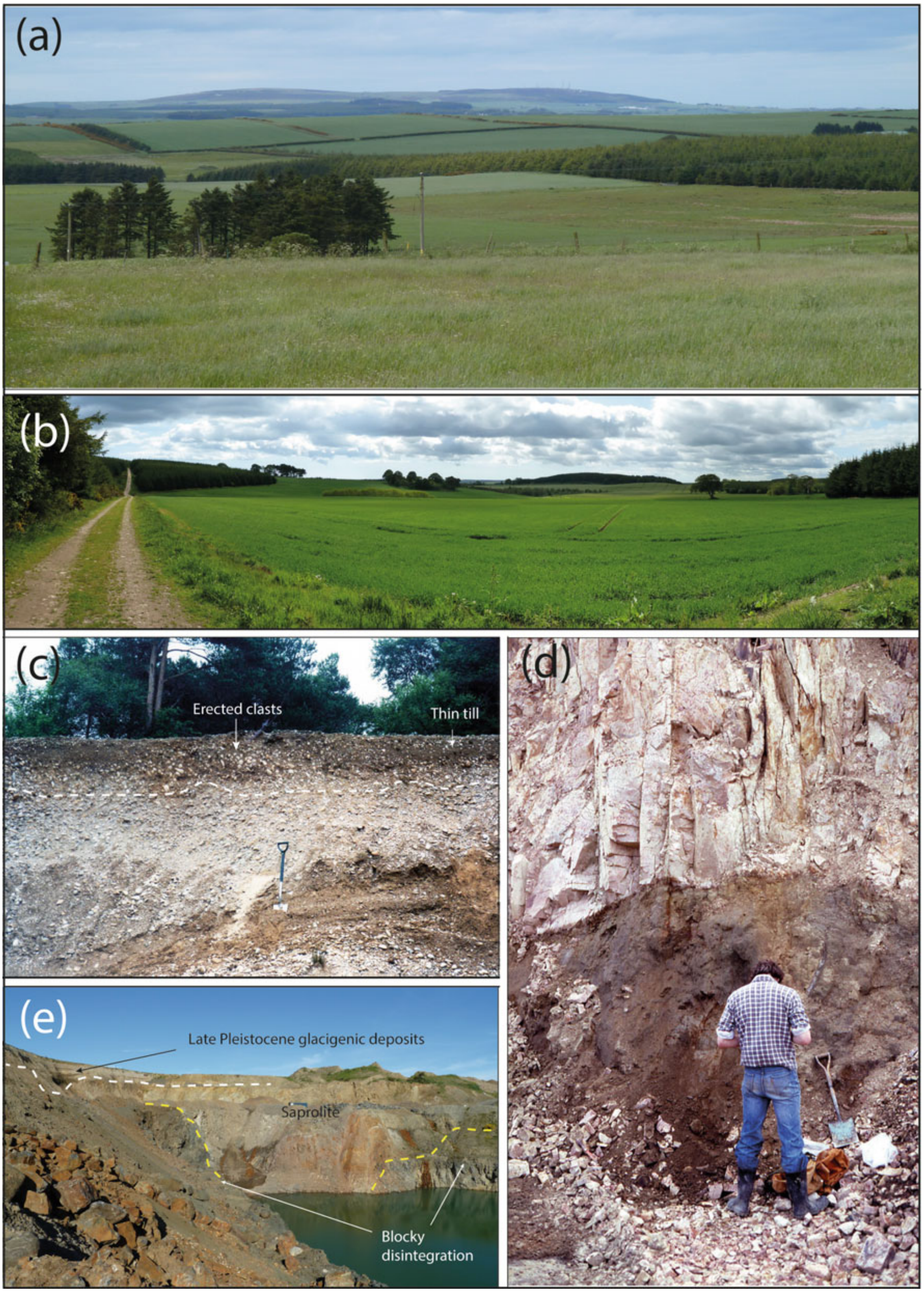
Palaeocene magmatic activity in western Scotland was accompanied by uplift of the Eastern Grampian Highlands (Chap. 3). However, retention of Cretaceous residues indicates that the Buchan area remained close to base level. Sedimentation rates dropped in the North Sea after ~52 Ma, with deposition of mud replacing earlier sands, indicating relief reduction in source areas around the Moray Firth, and with thinning or removal of chalk cover in Buchan. Climates during the Palaeogene were warm and humid throughout Northwest Europe until the global step change from 'greenhouse' to 'icehouse' climates in the Late Oligocene, and a further drop in temperature at ~ 10 Ma. This latter period is marked by an influx of immature sediment into the North Sea Basin. Increasing quantities of feldspar, biotite and chlorite mark a fundamental change in weathering styles in source areas (Huggett and Knox 2006). During the Neogene, the shoreline of northeast Scotland lay well to the east of the present coastline. The Deveron river system remained connected to the Ythan at Turriff, and the eastward-draining Deveron–Ythan system, together with the courses of the Dee and Don, was likely first established in the Palaeogene (Hall 1991).

The Buchan lowland is dominated by a single complex erosion surface, the Buchan palaeosurface, the lowest of a staircase of stepped palaeosurfaces in NE Scotland (Hall 2005). Although relative relief seldom exceeds 60 m, the oldest landscape elements occur at higher elevations, as is typical on platforms found south of Pleistocene glacial limits (Gunnell 2020). In eastern Buchan, hill and ridge tops retain former valley fills, up to 25 m thick, of the flint and quartzite gravels of the Buchan Gravels Formation (Figs. 21.1 and 21.3c). Nodular flint occurs at the base of the gravels, a remanié product of dissolution of the former chalk cover. The flint-dominated deposits of the Buchan Ridge Gravel Member have been interpreted as beach gravels (Bridgland et al. 1997), although the presence of intercalated beds of kaolinitic sand favours a fluvial origin (Merritt et al. 2003). More certainly, the elevated positions of the gravel bodies indicate post-depositional topographic inversion; this can be attributed to the resistance of the siliceous gravel bodies to weathering and erosion. The quartzite-dominated gravels of the western Windy Hills Gravel Member occur below ridge tops but above the present Deveron–Ythan valley. The abundance of detrital rutile in these gravels indicates derivation from the pelites and basic intrusions to the west (Hudson 1980). The presence of clasts up to boulder size on the Buchan Ridge indicates deposition by large ancestral rivers of the Deveron–Ythan system.

The Buchan Ridge gravels rest on kaolinized basement, a geochemically mature weathering type termed *clayey grus*, and the gravels are themselves weathered to a depth of 25 m; oxygen isotopes suggest that the kaolinitic weathering took place in two phases under mean annual temperatures of 23 ± 5 °C and 15 ± 5 °C (Hall et al. 2015a). Comparison with North Sea palaeotemperatures (Buchardt 1978; Chap. 3) shows that the warmest temperatures were matched only in the Palaeogene. Hence, the Buchan Ridge gravels are likely Palaeogene in age; the gravels may correlate with Palaeocene–Eocene sands in the outer Moray Firth. The clayey grus weathering type may include sub-Cretaceous kaolins but appears to have developed mainly in two later phases, first under humid-tropical conditions in the Palaeogene, and later under cooler temperatures in the Miocene. However, there are no absolute ages available for the Buchan kaolins. It is noteworthy that farther south in western Europe, kaolinitic, quartz-rich river gravels span the entire period between the Palaeocene and the Early Pleistocene.

The subdued terrain in Buchan resolves at the 10 km scale into tiered landscapes showing pervasive control by rock type and structure. The upper tier is represented by isolated low hills and broad interfluves, typically developed on pelites and quartzites, and includes the quartzite inselberg of Mormond Hill (Fig. 21.3a). Patches of the Buchan Ridge gravels and kaolinitic weathering occur on this high ground. The extensive middle tier has low-angle slopes developed across closely-fractured and often deeply-weathered rocks (Fig. 21.3b). This terrain includes the Windy Hills gravels (Fig. 21.3c). The lowest tier comprises the floors of valleys, typically developed along fault and shear zones that are locally deeply weathered. Large, shallow topographic basins have formed in rocks of low resistance to chemical weathering. Examples include the basins in the biotite-rich Strichen Granite and the Maud norite (Fig. 21.1). The close links between bedrock mineralogy, fracturing and weathering across the tiered terrain indicate that the landscape developed during lowering by differential weathering and erosion, a form of etchplanation. A Palaeogene age for the Buchan Ridge fluvial gravels implies lowering of the surroundings at average rates of ~2 m Ma^{-1} based on comparison of valley floor elevations.

Remnants of deep weathering are widespread in Buchan (Hall 1985). Weathering is most common in biotite granites, basic igneous rocks and feldspathic psammites. Exposures in quarries and borehole records show that weathering to >5 m depth is common in eastern Buchan (Fig. 21.1), with depths


(a)
(b)
(c)
Erected clasts
Thin till
(d)
(e)
Late Pleistocene glacigenic deposits
Saprolite
Blocky
disintegration

◀ **Fig. 21.3** Characteristic features of the Buchan palaeosurface. **a** View north from near Hill of Dudwick looking towards Mormond Hill. The terrain in the foreground is developed across pelites, semipelites and psammites. Fracture-aligned valleys run WSW–NNE, parallel to Late Carboniferous dolerite dykes. **b** Smooth, rolling topography with ~60 m local relief near Mintlaw. The upper slopes on the left are developed mainly in quartz-rich psammite and small granite masses. The lower slopes are developed across weathered biotite-bearing granite, norite and psammite. **c** Quartzite gravel at Windy Hills above the Ythan valley. Frost-erected pebbles occur in the upper 1.5 m of the gravel, which is also locally reworked into a thin till. **d** Kirkhill Quarry in 1978, showing grussified basic igneous rocks below a fragmented felsite dyke. The weathering underlies the Middle and Late Pleistocene sequence at Kirkhill; at Leys Quarry nearby, weathered rock is incorporated into the basal till. **e** Variably weathered Dalradian psammites and pelites, with mafic masses and younger felsite dykes at Savock Quarry at the northeastern end of the Moss of Cruden ridge. The weathering is >15 m thick, with a present upper surface ~30 m below the base of the flint gravel on Moss of Cruden. The thickness of the original weathering profile may have been >45 m. (Images: **a**, **b**, **e** Adrian Hall; **c**, **d** Rodger Connell)

of 60 m recorded at Hill of Dudwick. Seismic profiles suggest that weathering tens of metres deep exists on the sheared margins of basic igneous masses (Ashcroft and Munro 1978). A common feature is the sharp lateral variation in weathering depths (Hall 1986). The basal surface of weathering is also highly variable in its character and form, in places abrupt and elsewhere diffuse. Saprolite is often interrupted by zones of blocky disintegration, where chemical alteration of the rock remains limited (Fig. 21.3d). The basal surface of weathering is often irregular, with alternating low risers and depressions (Fig. 21.3e), linear zones of deep chemical alteration and scarp-foot weathering zones. Similar local- and regional-scale patterns are reported for shield rocks in Finnish Lapland and on the Kola Peninsula (Hall et al. 2015b).

On the middle and lower tiers of the terrain, the weathering is entirely of a geochemically immature grus type. The saprolites are sandy, with low clay contents, varied clay mineral assemblages and substantial components of little-altered primary minerals. No absolute ages exist for this weathering type, or for the Fe and Mn oxide coatings found on many fracture surfaces. The grus may represent, in places, the basal parts of more evolved clayey-grus weathering profiles (Fig. 21.3e). Similar sandy saprolites developed in western Europe under Neogene humid-temperate climates. Correlation with North Sea sediments indicates that this phase of deep weathering in Scotland lasted from the Late Miocene to the Early Quaternary, prior to the first advances of the Scottish ice sheets. Formation of grusses likely continued into the Pleistocene, as Middle Pleistocene tills and gravels contain grussified clasts (Hall et al. 1989).

21.5 Pleistocene Landscape Modification

During the last glaciation, and probably also earlier in the Pleistocene, ice flowed towards Buchan via three pathways: the inner Moray Firth, the western North Sea and the Eastern Grampians (Chap. 4). Sedimentation from interacting ice masses produced complex stratigraphy: Buchan holds the most complete terrestrial Pleistocene record in Scotland. The oldest-known till units occur at Kirkhill, likely deposited by an ice sheet during MIS 8 (303–245 ka). This Pleistocene sequence is unique in the British Isles' terrestrial record in providing a detailed record of at least three complex stadial–interstadial–interglacial cycles. Deeply buried tills of largely unknown age are also recorded in numerous boreholes in the Ellon and Peterhead areas. Merritt et al. (2017) have provided a recent detailed review of the Pleistocene stratigraphic sequence in NE Scotland.

The geomorphological effects of Pleistocene glacial erosion on the lowlands of NE Scotland have been limited, yet even within Buchan there is significant variation in the impact of former glaciers. Weathering profiles are truncated or erased from areas affected by the relatively vigorous ice streams that moved along the coastal fringes of the Moray Firth and the North Sea (Fig. 21.1). However, in the area formerly covered by ice flowing out of the Eastern Grampians, an inverse relationship exists between the distribution of landforms of glacial erosion, such as streamlined hills and roches moutonnées, and the frequency of weathering (Hall and Sugden 1987). Moving northwards from Inverurie, glacial bedforms become less frequent and then absent, whereas pockets of weathering cover become more common and merge into more continuous regolith cover. In central Buchan, the preservation of the Buchan Gravels Formation implies that Pleistocene glacial erosion has been negligible. Across much of NE Scotland, glacial erosion has been largely restricted to the removal of saprolites, except along the main ice-flow corridors (Glasser and Hall 1997). The limited impact of glacial erosion across Buchan is comparable to some other parts of Scotland, including the Eastern Grampian plateau, along the Caithness-Sutherland border, NW Lewis and the Lammermuir Hills (Chaps. 8, 9, 20 and 27). These areas were remote from ice centres, experienced relatively limited durations of ice cover and remained predominantly under non-erosive, cold-based ice.

Tills deposited by ice moving from the Eastern Grampians typically show a dominance of locally derived debris, with many angular clasts of the bedrock types found in immediately up-ice positions, reworked saprolite and periglacial slope deposits, and clay mineral suites resembling those found in underlying saprolites (Merritt et al. 2003). Such locally derived tills imply short distances of glacial transport that are compatible with limited glacial erosion (Fig. 21.3c).

The smooth slopes that are typical of much of Buchan (Fig. 21.3a, b) are in part products of periglacial slope processes operating across fissile Dalradian rock types, such as pelite, extensive deep-weathering mantles and thin covers of glacial deposits. That such processes were significant during Middle and Late Pleistocene cold stages is shown by the presence of periglacial slope deposits and permafrost indicators, such as ice-wedge casts and erected pebbles (Fig. 21.3c), that occur in at least six separate intervals in the regional stratigraphy (Connell and Hall 1987). Due to its peripheral position, Buchan likely experienced long periods of cool and periglacial environments in the Pleistocene (Hall et al. 2019).

21.6 Conclusions

The Buchan palaeosurface is a diachronous and polygenetic feature. The post-Caledonian basement platform was first exposed in the Devonian, then buried beneath successive thick sedimentary covers and re-exposed at intervals through the late Palaeozoic and Mesozoic. The present erosion level in eastern Buchan remains close to the sub-Cretaceous unconformity. A new landscape developed after removal of chalk cover, with formation of kaolinitic weathering and deposition of flint and quartzite gravel in the Palaeogene on what is today the highest ground. Deep weathering continued under cooler climates in the Neogene and Early Pleistocene and led to development of a tiered relief closely adjusted to geology. Deep-weathering patterns are important for understanding not only the development of the Neogene relief but also its differential modification by glacial erosion in the Pleistocene. The Buchan landscape is of outstanding geomorphological interest as it provides detailed evidence of a long exposure and burial history through the coexistence of tiered unconformities, Palaeogene kaolins and siliceous gravel deposits, Neogene sandy saprolites and various landforms of differential weathering and erosion. Similar slowly eroding platforms are found south of the limits of glaciation in western Europe, but within these limits the only comparable landscapes are in other areas of former cold-based ice cover in northern Finland and the Kola Peninsula.

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22 Beaches and Dunes of the Moray Firth Coast

James D. Hansom

Abstract

The Moray Firth coastline supports a varied assemblage of extensive modern and emerged beach and dune systems that have formed under falling sea levels and plentiful supplies of glacigenic sediment during the Middle and Late Holocene. Waves mainly from the north and east have driven longshore sediment transport to form well-developed modern beaches and spits that have migrated westward on the south shore of the firth and southward on the northwest shore, in most localities continuing a trend displayed by Holocene emerged beaches and spits that lie landward of the modern coast at altitudes of <10 m above present sea level. Within the Dornoch Firth, however, some emerged gravel beaches record fundamental changes in the morphogenetic environment during the Holocene, probably attributable to partial closure of the mouth of the firth by progradation. In some places, notably at Spey Bay and Culbin Sands, the outlets of major rivers were diverted by spit growth, but subsequent breaching of spits has restored the outflow channels to their original locations. Beach and dune development on the Moray Firth coastline can be understood in the context of discrete coastal sediment cells, with boundaries generally formed by prominent headlands, a context likely to have been in place over much of the Holocene. Although the general pattern of progradation and spit extension during the Holocene implies abundant sediment supply, recent research suggests that a reduction in sediment supply, coupled with sea-level rise, has resulted in increased erosion of Moray Firth beaches, a pattern common to many beaches in Scotland.

Keywords

Sediment supply • Sea level • Beach • Dune • Emerged beach • Longshore transport • Spit

22.1 Introduction

The Moray Firth forms a broad embayment on Scotland's northeast coast, between Cairnbulg Point and Duncansby Head (Fig. 22.1). Open to waves from the north and east, the main beach and dune areas lie in the inner reaches of the firth, south of rocky outcrops at Brora in the north and west of Portnockie in the east, and include beaches in the smaller inlets of Loch Fleet and the Dornoch Firth. The main beach and dune complexes are located seaward of emerged features deposited when relative sea level was higher than at present and often form part of large composite shorelines including forelands, spits, and bars. Both wave erosion of the underlying sandstone bedrock and supplies of glacigenic sediment from offshore and terrestrial sources have locally contributed sediment for beach building. Large beaches, dunes, and spits occur in the inner Moray Firth at Spey Bay, Culbin Sands and Whiteness Head, in the Dornoch Firth at Morrich More, Dornoch Point and Ness of Portnaculter, and at the mouth of Loch Fleet (Fig. 22.1).

West of Spey Bay and south of Brora, the Moray Firth coastlands are almost entirely underlain by the Devonian sedimentary rocks of the Old Red Sandstone Supergroup; east of Spey Bay the coastline consists of Dalradian metapsammites and metapelites. Structurally, the northwest coast of the Moray Firth is closely associated with two NE-trending faults, the Great Glen Fault and Helmsdale Fault, which, respectively, control coastal configuration south and north of the Dornoch Firth, but the southern coastline is only weakly related to a series of W–E trending faults. Offshore, the Moray Firth forms part of a large Mesozoic basin that dips uniformly toward the center of the basin (Andrews et al. 1990).

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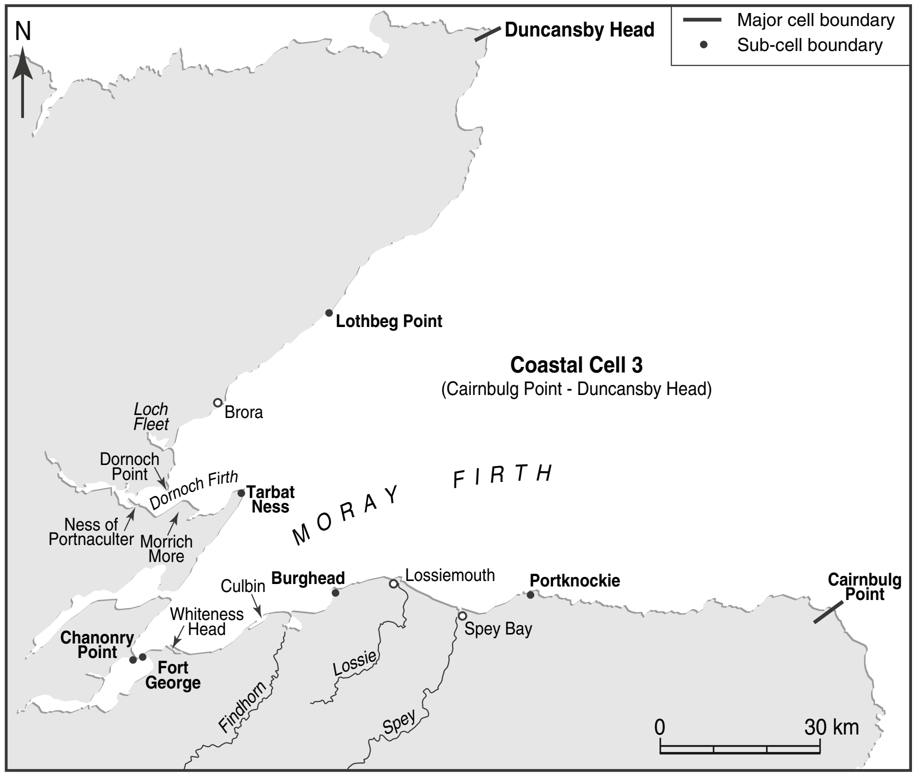

Fig. 22.1 Location of the main beach and dune areas of the Moray and Dornoch Firths, showing sediment cell and sub-cell boundaries

Beach and dune development on the Moray Firth coastline commenced soon after deglaciation, but most of the main constructional features below about 10 m OD (Ordnance Datum) accumulated over the Middle and Late Holocene in response to variations in sediment supply and gradually falling relative sea level (RSL). The impact of these factors has produced a wide range of beach and dune responses over time. For example, sediment accretion over the past ~7 ka has produced a massive constructional foreland at Morrich More in the outer Dornoch Firth, and another at Culbin Sands in the Moray Firth. In contrast, the large gravel spit complexes close to the Ness of Portnaculter are now largely abandoned following a reduction in wave energy in the central Dornoch Firth, and the decline of gravel supply to Spey Bay has resulted in extensive erosion in the east that now sustains beach extension farther west.

22.2 Controls on Coastal Development

22.2.1 The Impact of Glaciation

During successive episodes of Pleistocene ice-sheet glaciation, the Moray Firth was occupied by a major ice stream flowing northeastward from the Great Glen and the Spey valley. During the last (Late Devensian) ice-sheet glaciation of ~35–14 ka, ice-flow directions varied relative to the present coast, with ice at different times flowing parallel with or across the coastline (Hughes et al. 2014; Merritt et al. 2017, 2019), and it is likely that similar changes in ice-flow directions occurred during earlier glaciations. Successive glaciations have profoundly affected coastal development, not only by eroding the embayments now occupied by the Dornoch and inner Moray Firths (Fig. 22.1), but also through deposition of copious glacigenic sediments, both onshore and offshore, that have provided much of the material subsequently available for beach building.

Deglaciation of the southern Moray Firth coast may have begun as early as ~17.7 ka, with the northwest coastline being ice-free by ~17.5–17.0 ka, though ice cover may have persisted in the inner Moray Firth until ~15.6 ka or later (Ballantyne and Small 2019). Retreat of the ice margin exposed vast amounts of glacigenic sediment in the nearshore and offshore zones, and along the present coastline: offshore and nearshore sediments are dominated by till and glacimarine deposits, while onshore deposits include prominent readvance moraines, glaciolacustrine sediments and outwash deposits (Merritt et al. 2017; Chap. 15). Immediately inland of the beaches of Spey Bay, Culbin and Whiteness Head, for example, are thick deposits of till and glacifluvial sand and gravel that have been trimmed by wave activity. Farther north, ice recession in the Dornoch Firth exposed thick sequences of till and glacifluvial sediments that subsequently provided material for beach building in coastal locations (Firth et al. 1995).

22.2.2 The Legacy of Sea-Level Change

Lateglacial and Holocene RSL changes have fundamentally affected the configuration of the Moray Firth coast. Such changes have reflected the interplay between local glacial isostatic rebound, which increased southwestward across the area, and rising eustatic sea level. The Moray and Dornoch Firths share similar RSL trends (Shennan et al. 2018). As the margin of the last ice sheet moved inland, the Lateglacial sea flooded low-lying coastal areas to altitudes of up to ~20 m OD, as recorded by widespread deposition of glacifluvial and glacimarine terraces graded to the Lateglacial marine limit. Thereafter, RSL fell to a minimum at ~11 ka, when it lay close to present sea level (0 ± 2 m OD) in the Dornoch Firth and inner Moray Firth, then gradually rose to the Holocene marine limit (~6 ± 2 m OD) at ~7 ka or later (Smith et al. 2019). Almost everywhere along the Moray and Dornoch Firth coasts the Holocene marine limit is marked by an abandoned cliff, generally cut in glacigenic sediments and fronted by emerged shorelines.

Subsequent gradual fall in RSL led to abandonment of the highest Holocene shoreline and formation of staircases of progressively lower gravel and sand beaches. During this period, beach building was nourished from several sources: from fluvial erosion of inland glacigenic deposits, marine trimming of cliffs composed of glacigenic sediments and, crucially, the arrival of nearshore and offshore sediments, originally of glacigenic origin, driven onshore by waves as sea level fell. Many of the earliest emerged Holocene beaches that formed in the central Dornoch Firth, for example, comprise mainly gravels sourced from local glacifluvial sediments, whereas in the outer firth the mid-Holocene emerged beaches are dominated by sands sourced from offshore (Firth et al. 1995). The same general pattern occurs in the composition of emerged beaches in the Moray Firth, although modulated by the input of gravels of glacigenic origin that reached the coast via the arterial Spey and Findhorn Rivers (Figs. 22.1 and 22.2a). The status of current sea-level change is uncertain, but tide gauge records for the past few decades suggest that RSL may be rising, a trend possibly linked to recent beach erosion (Rennie and Hansom 2011). Evidence of degrading saltmarshes in the region (Merritt et al. 2017) also suggests recent local RSL rise.

22.2.3 Present-Day Morphogenetic Controls

Tidal currents in the Moray Firth area are generally weak, except at the outlets of major rivers and in narrow inlets such as Loch Fleet (Ramsey and Brampton 2000). In consequence, wave-induced currents exert the most significant impact on sediment movement along the coast. Scotland is strongly influenced by prevailing southwesterly winds, making the Moray Firth coast a leeward shore with mainly offshore winds that limit waves from this direction. The total wave and swell climate offshore of the Moray Firth is dominated by wave approach from the north and northeast, with an important secondary component from the east and southeast (Ramsey and Brampton 2000). Modulated locally by coastal orientation, a westerly wave-produced net drift direction occurs along the Moray Firth south coast, for example at Spey Bay and Culbin, and a net southerly drift takes place along the north coast, as at Loch Fleet and the Dornoch Firth.

The general context of beach and dune development in Scotland is provided by a series of self-contained sediment transport cells, within which accumulated sediment can be exchanged but not exported to adjacent cells (Ramsey and Brampton 2000). Since prominent headlands generally form the cell boundaries, most cells are likely to have been in place at least since the mid-Holocene. Within each cell, the boundaries of smaller sub-cells are not totally sediment-tight, allowing only very limited export or import of sediments; all the Moray Firth lies within coastal cell 3, from Cairnbulg Point to Duncansby Head (Fig. 22.1).

22.3 Key Sites

22.3.1 Spey Bay

Spey Bay, on the southern coast of the Moray Firth, extends for ~15.5 km west of Portknockie (Fig. 22.1), enclosing a spectacular gravel beach on either side of the braided mouth of the River Spey, which drains a catchment dominated by glacigenic deposits (Fig. 22.2a; Chap. 19). To the rear of the modern gravel beach an emerged cliff cut mainly in glacifluvial and glacimarine deposits is fronted by a staircase of coast-parallel emerged gravel ridges at ~5–10 m OD (Firth and Haggart 1989; Hansom 2003a; Fig. 22.2b). The mouth of the River Spey is marked by a complex series of gravel ridges and bars that exhibit frequent changes in response to high river discharge events and wave processes. Dominated by a strong, wave-driven longshore current that periodically builds a gravel spit that deflects the mouth of the River Spey westward, recurrent spit breaching during flood events returns the river outlet close to its present position. Recent breaches, however, have been engineered to reduce the risk of flooding and erosion west of the river mouth, with the beach in front of an adjacent settlement being artificially nourished to maintain its position and stability (Hansom 2003a).

The modern beach complex is gravel-dominated, with older, inland-trending ridges being truncated and buried by more recently deposited ridges (Fig. 22.2c). The eastern part of Spey Bay appears to have undergone (and continues to

Fig. 22.2 **a** Gravel ridges and overwash lobes at the mouth of the River Spey (image width 2.7 km). The modern ridge seaward of the houses was artificially recharged with gravels excavated from the river mouth in 1995, with no recharge since. **b** Emerged gravel ridges descend to the modern beach from the abandoned Holocene cliff at ~10 m OD. **c** Emerged gravel ridges at ~4 m OD recurve inland and are buried and truncated by the modern beach ridge. The line of concrete blocks marks the 1940s shore position and orientation. (Images: **a** Google Earth™; **b**, **c** © Anne Burgess CC BY-SA 2.0)

experience) net erosion, releasing sediment to feed gravel ridge accretion farther west (Hansom 2003a). At present, despite a healthy sand and gravel supply from the River Spey (Fig. 22.2a), erosion extends for several kilometers west of the river mouth, placing houses at risk. A strong, wave-driven, longshore current transports sediment westward toward Lossiemouth, and gravel deposition has progressively replaced the sand beach in the west of the bay. Historical records show that between 1870 and 1995, gravel has extended 4.2 km along the coast at an average rate of 34 m a^{-1} (Gemmell et al. 2001a), and that over the same period the sand beach and backing dunes in the west near Lossiemouth have experienced net erosion, which continues at present. An indicative sediment budget for Spey Bay, calculated by Gemmell et al. (2001a; Fig. 22.3), incorporates the input of sand and gravel from the River Spey and suggests that >10,000 m^3 a^{-1} is currently being removed from the eastern part of Spey Bay, but that only about half of this contributes to sediment accretion farther west. There is also evidence of a reduction in the amount of sediment delivered to the coast by the River Spey because of flood-bank construction upstream (Gemmell et al. 2001b).

22.3.2 Culbin Sands

Culbin Sands, between Nairn and the tidal basin at the mouth of the River Findhorn (Fig. 22.1), is one of the most spectacular coastal sites in Scotland due to its complex

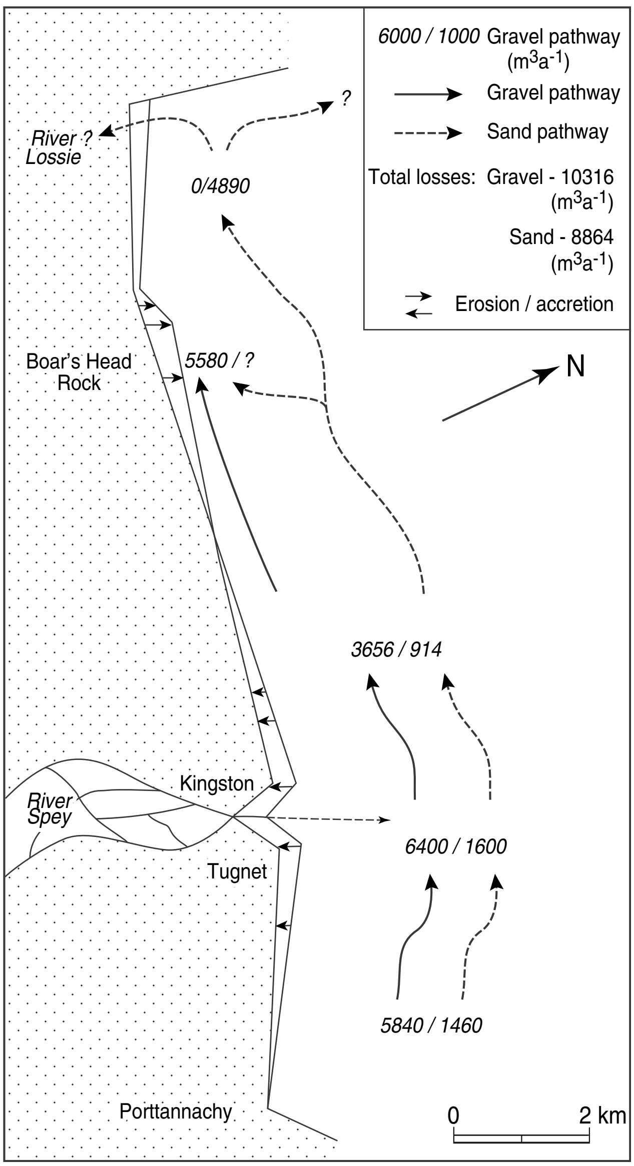

Fig. 22.3 Diagrammatic representation of an indicative annual sediment budget for Spey Bay. (From Gemmell et al. (2001a). Licensed under Open Government Licence v3.0)

history of spit development and, inland from the coast, the most extensive dunefield in the country. The site consists of ~30 km^2 of emerged sand and gravel strandplain backed by a prominent ~7 m high emerged Holocene cliff whose base lies at ~9 m OD (Hansom 2003b). The strandplain is crossed by gravel ridges that represent forerunners of the present active spits and is extensively buried by a dunefield that was formerly active but has now been stabilized by afforestation (Comber 1995); the gravel ridges closest to the present shore also protect areas of active saltmarsh. At the coastline, two substantial spits extend westward. The larger of the two, known as The Bar, is a 3.5 km long gravel barrier that is almost detached from the mainland and comprises several generations of recurved gravel ridges; the older ridges are capped by high sand dunes, whereas the younger ridges support heath vegetation, and are truncated at their seaward ends by modern beach ridges (Fig. 22.4). Erosion and overwashing occur at the neck along the eastern section of The Bar, with sand and gravel encroaching onto the saltmarsh behind (Comber et al. 1994). The Bar has extended westward at an average rate of ~15 m a^{-1} from its earliest plotted position in 1685, its coarse gravel recurves being produced by waves refracting around the tip into deeper water to the west (Fig. 22.5a).

Partly concealed beneath the main dunefield at Culbin are at least three sequences of emerged gravel ridges that splay out to the southwest away from the present east–west coastal orientation (Fig. 22.5a). The landward ridges are truncated by younger ridges to seaward, indicating erosion of the earlier forms (Comber 1995). These ridges represent remnants of shorelines (spits) that formerly extended seaward at high angles to the present coastline. The earliest surviving set of ridges (group 1 in Fig. 22.5a) was part of a former shoreline lying oblique to the present coast and extending southwest that was progressively truncated by updrift erosion in the east. The eroded sediments formed the ridges of group 2, which were then eroded to form group 3. This sequence of spit development diverted the mouth of the River Findhorn westward, its former course now being marked by a line of peat bogs and small lochs along the foot of the abandoned Holocene cliff (Fig. 22.5a). Radiocarbon dating of peats along the foot of the cliff suggests that the abandonment of this route occurred sometime between 5.4 and 3.6 ka when the River Findhorn breached northward through the narrowing updrift spit ridges. A similar pattern of westward spit extension is suggested for the historically dated spit positions (groups 4–7 in Fig. 22.5a) between Buckie Loch and The Bar (Hansom 2001).

Culbin is also noted for its extensive dunefield, which includes: parabolic dunes up to 15 m high and 400 m in width and length, the largest of their kind in Europe; butte dunes representing the eroded remnants of former vegetated dunes; and extensive tracts of formerly mobile transgressive

Fig. 22.4 The Bar at Culbin has extended westward at ~15 m a^{-1} since the late seventeenth century. The older ridges are now truncated by modern ridges at their seaward ends; sedimentation behind the ridges supports sandflat and saltmarsh. (Image: © P. and A. MacDonald/Aerographica/NatureScot)

dunes up to 30 m high. Before stabilization by forestry, the transgressive dunes were smooth, rounded, unvegetated features that migrated downwind over the underlying gravel ridges at up to 6.5 cm per day (Fig. 22.5b), periodically burying and re-exposing abandoned buildings as they migrated eastward. Culbin is noted for a record of sand inundation events that finally buried the fertile lands and buildings of the Culbin Estate in 1694; this event led to a Scottish Parliament Act of 1695, prohibiting the harvesting of marram grass, a practice that had likely contributed to dune surface instability (Ross 1992). As recently as 1922, Culbin remained the largest area of open, mobile and unvegetated sand dune in Britain, but by 1963 had been largely stabilized by afforestation.

22.3.3 Whiteness Head

Whiteness Head, to the west of Culbin, is a 3.5 km long active spit that has prograded westward along the south shore of the Moray Firth (Fig. 22.1). Composed of gravel ridges capped by small amounts of wind-blown sand, its distal end recurves southward, impounding an inlet with areas of saltmarsh and sandflat (Fig. 22.6a). Historical records show that since 1880 the spit extended westward at an average rate of 15.2 m a^{-1} (Stapleton and Pethick 1996; Fig. 22.6b), its growth fuelled by erosion updrift (Hansom 2003c). Land reclamation has reduced the formerly more extensive saltmarsh to an area of ~1 km^2, and western spit extension has been curtailed by maintenance dredging. Landward of the spit and inlet, a complex of emerged Holocene sand and gravel beach deposits is backed by a 10–12 m high, emerged Holocene cliff (Ritchie et al. 1978). Whiteness Head spit supports several recurved gravel ridge complexes at 3–5 m OD, each truncated downdrift by the succeeding one and separated by small bays of saltmarsh and dry gravel slacks (Ritchie et al. 1978). Extending along the entire length of the spit, the present-day active gravel ridge is almost 2 m higher than the older recurves behind and is periodically overtopped during storms. This fronting ridge declines in height eastward to form low gravel overwash lobes that encroach over the saltmarsh behind.

22.3.4 Ness of Portnaculter

The Dornoch Firth is flanked by both emerged and modern spits and beaches. Within the central firth, mid-Holocene high sea levels resulted in the formation of spit complexes at Ness of Portnaculter and nearby Ardjachie Point (Fig. 22.1).

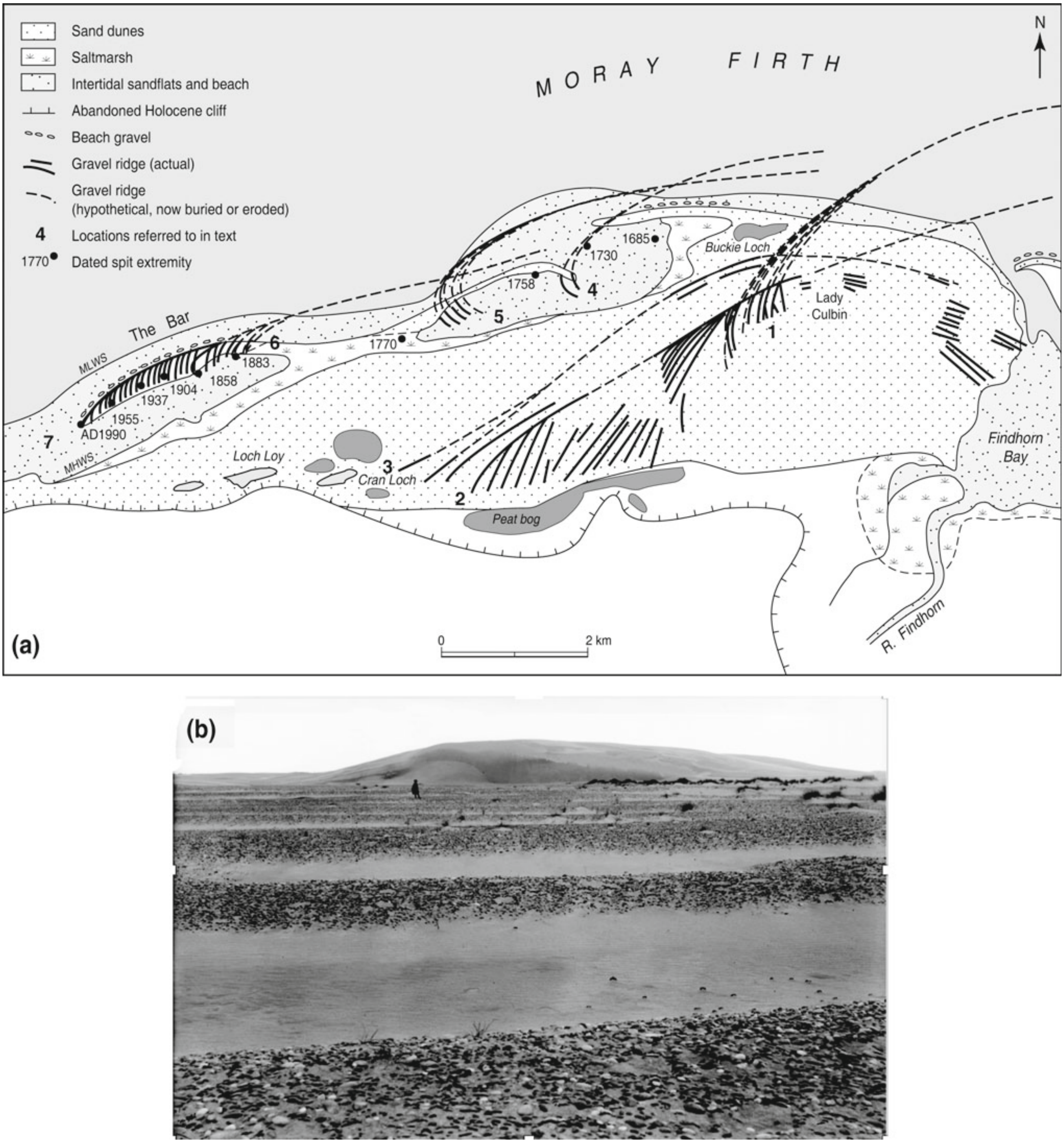

Fig. 22.5 **a** Culbin Sands is a largely forested and dune-covered gravel strandplain composed of emerged gravel spits that extend westward. These formerly deflected the outlet of the River Findhorn westward; northward flow into the Moray Firth was resumed by breaching of the spit neck, and westward spit extension continued by erosion updrift and deposition downdrift. Reconstructed spit sequences are numbered 1–7. **b** Culbin in 1912 was a rapidly changing dunescape. This view toward Lady Culbin, the unvegetated dune in the distance, shows dune migration over emerged gravel ridges. (Reproduced by permission of the British Geological Survey, Licence CP20/088)

The Ness of Portnaculter is an emerged spit composed of several ridges of well-rounded and sorted coarse gravels at up to 9.3 m OD extending northwest. The ridges are truncated along the coastline to the east but in the west a prominent gravel ridge extends 2 km northwestward, with well-developed westward recurves that mark progressive spit extension into the firth (Fig. 22.7). The orientations of the recurves suggest easterly or northeasterly waves recurving

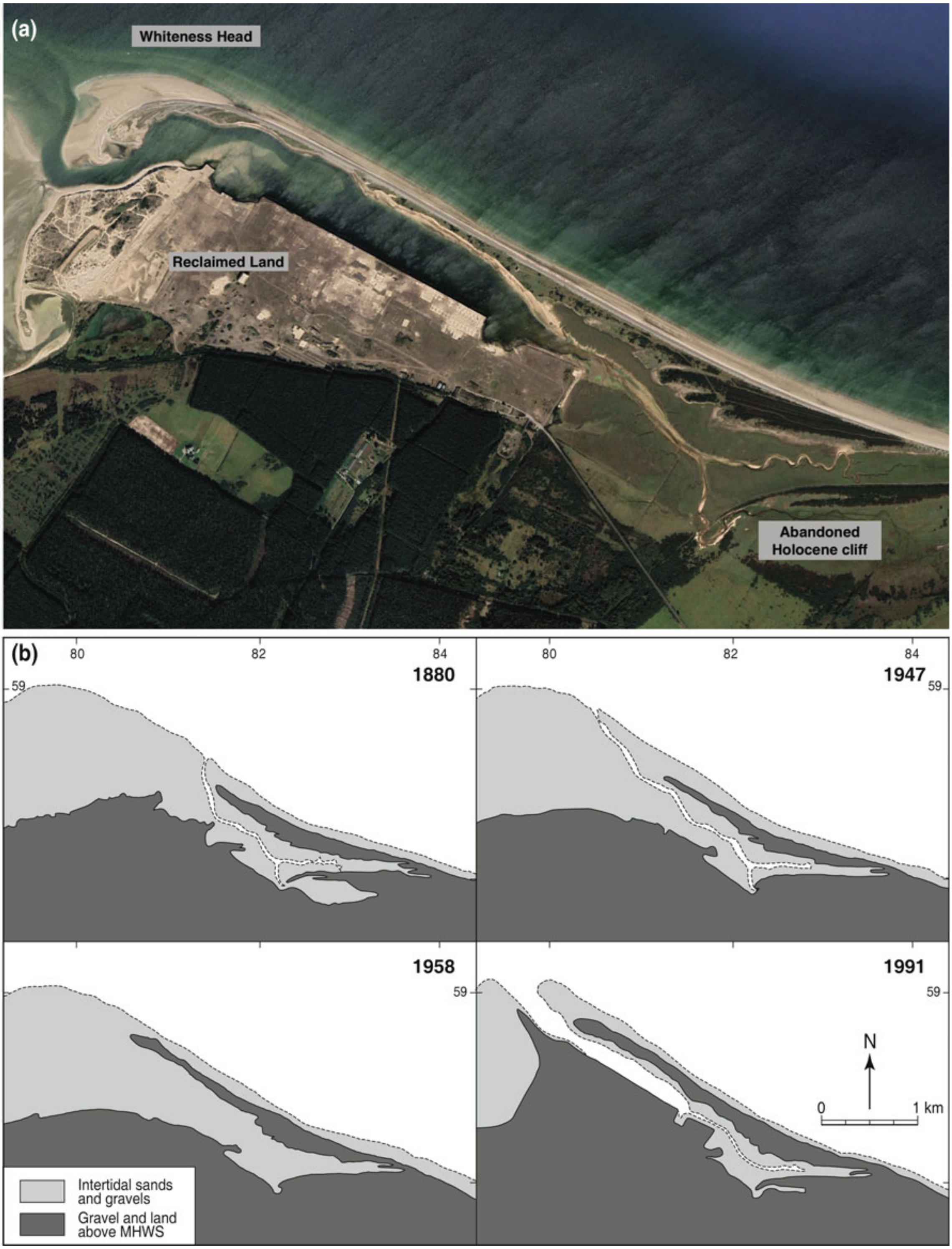

Fig. 22.6 **a** Gravel spit of Whiteness Head encloses an inlet and former areas of saltmarsh. (Image: Google Earth™; image width 3.9 km). **b** Spit development at Whitehead Ness reconstructed from historical sources; landward changes by 1991 reflect industrial development and channel dredging. (From Stapleton and Pethick (1996). Licensed under Open Government Licence 3.0)

into deep water to the north, and their altitude implies emplacement during the period of highest Holocene sea level (Firth et al. 1995). The source of the sediments appears to have been erosion of adjacent glacigenic terraces and the Holocene cliff to the east. Similar emerged gravel structures occur on the north shore at Meikle Ferry and Ard na Cailc

(Fig. 22.7). At Ardjachie Point, a lower set of ridges (up to 3.8 m OD) lies at right angles to the orientation of the Ness of Portnaculter ridges and extends northeast for 1 km. These ridges are composed of finer, less well-sorted gravels and, with southeast-trending recurves, their lower altitude and alignment suggest more recent emplacement by westerly waves rather than easterly waves. The sediment source appears to have been erosion of the emerged spit at Portnaculter.

The contrasting alignment, altitude, and composition of the two spits suggest marked changes in wave climate following the Holocene sea-level highstand. The formation of the substantial coarse-gravel spit at Ness of Portnaculter implies a regime of high-energy waves from the east or northeast, but the present wave climate of the central Dornoch Firth is low energy, with wave heights rarely exceeding 0.4 m (Firth et al. 1995). This apparent anomaly is explicable by partial closure of the mouth of the Dornoch Firth, after the formation of the Ness of Portnaculter spit, by progradation at Morrich More and Dornoch Point (Sect. 22.3.5). This changed the wave regime in the central firth and led to the formation of the later east-trending spit at Ardjachie, which formed, at least in part, through cannibalization of sediment from the Ness of Portnaculter spit. Corroboration of this change in wave climate comes from erosion of the west side of Morrich More (see below) which is now characterized by sediment transported eastward under waves from the west.

22.3.5 Morrich More and Dornoch Point

Morrich More (Figs. 22.1 and 22.8) is a spectacular ~20 km^2 emerged foreland in the outer Dornoch Firth. The foreland has been built eastward by the progressive accretion of ~50 NW–SE trending emerged sandy ridges backed by a prominent Holocene cliff cut into glacigenic sediments and whose base lies at ~10 m OD. The ridges are everywhere capped by low dunes, except in the western corner where substantial, 14 m high parabolic dunes occur. The youngest ridges are located on the outer (eastern) coast where they are represented by the prominent dune-capped, sand-cored tidal islands of Innis Mhòr and Paterson Island (Figs. 22.8 and 22.9a). Extensive sand flats and saltmarsh occur in the sheltered area between the seaward-most emerged beach ridges and the tidal islands. In the center

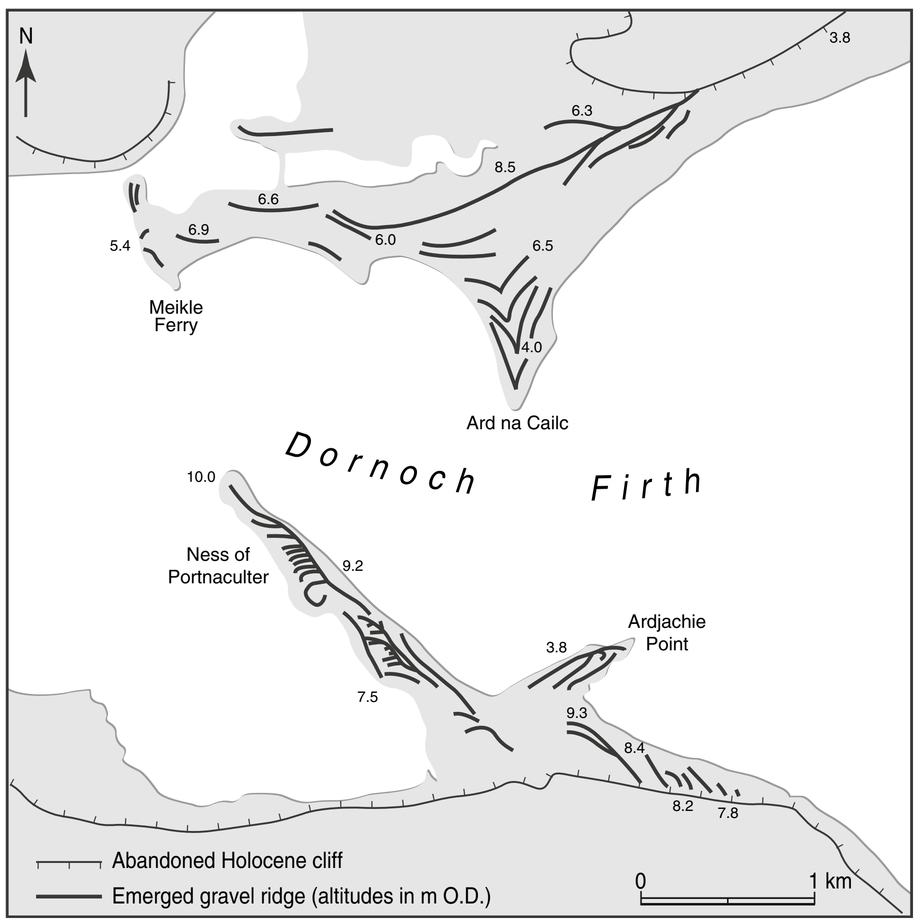

Fig. 22.7 Ness of Portnaculter and Ardjachie spits on the south shore of the Dornoch Firth and Meikle Ferry and Ard na Cailc spits on the north shore. Emerged gravel ridge orientation shows westward migration of Ness of Portnaculter, with the eastward orientation of the lower emerged ridges at Ardjachie indicating a change in wave direction as sea level fell and the influence of waves from the east diminished. (From Hansom (2001) © 2001 Elsevier; reproduced with permission)

and west of the foreland, the emerged ridges are separated by areas of damp slack, but in the east they are interdigitated with elongate depressions occupied by well-developed saltmarsh (Hansom 2003d; Fig. 22.9b). On the outer coast, the intertidal zone displays active, onshore-moving sand bars, whereas on the inner west coast, tidal sandflats extend over 1 km into the firth. Along the inner shore, dune-capped emerged ridges are being truncated by wave erosion, with eroded sediment transported northeastward to accumulate in tidal flats and saltmarshes behind Innis Mhòr (Hansom and Leafe 1990; Fig. 22.9a).

The Morrich More emerged beach ridges are composed of medium to fine sand, brought onshore by waves moving over a shallow and extensive offshore zone covered by glacigenic sediments. Grouped by altitude, the highest ridges lie at 6.0–8.6 m OD and are ~100 m apart (Hansom 2003d). Peat at 6.4 m OD beneath a ridge at the foot of the Holocene cliff yielded a calibrated radiocarbon age of ~7.4 ka, suggesting that ridge formation occurred at the peak of Holocene sea-level rise in the area (7.8–6.2 ka). The subsequent fall in RSL led to a second group of ridges forming at 4.4–5.5 m OD (Fig. 22.10), and the lowest sand ridges occur at 2.5–4.0 m OD and 1.4–2.4 m OD, the latter being the same altitude as modern features (Hansom and Leafe 1990). The younger ridges are higher than the older ones, suggesting that sediment supply may be progressively declining, permitting more prolonged building of individual ridges rather than the formation of additional ones.

Both geomorphological and historical evidence indicate progressive eastward migration of the Morrich More foreland, with erosion of the western shore being counterbalanced by northeastward progradation of the eastern shore (Hansom 1999, 2003d). The alignment of the oldest sand ridges, which are now truncated at the western shoreface, suggests that, at the peak of the Holocene sea-level rise, the apex of the foreland lay several kilometres west of its present location and has subsequently migrated eastward (stages 1–3, Fig. 22.10). Recession of the western shore and accretion of the northeast shore are confirmed by maps from 1730 and 1887 (stages 4 and 5, Fig. 22.10). Ordnance Survey maps from the early twentieth century depict Innis Mhòr and Paterson Island as unvegetated sand bars, but both are now substantial, 4–5 m high dune islands, indicating net accretion over the past century (stage 6, Fig. 22.10). However, recent erosion, overwash, and breaching of the seaward face of Innis Mhòr and Paterson Island suggest that sediment delivery from offshore may now be declining, a trend that is likely to be exacerbated by rising relative sea level (Hansom et al. 2017).

Dornoch Point is a 3 km long recurved sand spit on the north shore of the outer Dornoch Firth, and extends southward across the mouth of the firth toward Morrich More (Fig. 22.8). A prominent east–west trending abandoned Holocene cliff composed of glacigenic sand and gravel is fronted by two gravel barriers at 6.5–7.0 m OD that extend westward and probably represent the culmination of the mid-Holocene sea-level rise. Seaward of these ridges, a broad sandflat

Fig. 22.8 Innis Mhòr on Morrich More and Dornoch Point both recurve west into the Dornoch Firth under the influence of waves from the east, while extensive ebb tidal sand banks extend seaward of Dornoch Point. Wide tidal sand banks occur on either side of the firth. The emerged beach and dune ridges of Morrich More are interdigitated with saltmarsh landward of the tidal islands of Paterson Island and Innis Mhòr. (Image: Google Earth™; image width 10.5 km)

Fig. 22.9 **a** Morrich More: view SE over dune-capped Innis Mhòr at low tide, with tidal sand flat and developing saltmarsh to landward. The seaward edge of Innis Mhòr is now erosional. **b** Morrich More: view NW at high tide with Paterson Island on the right. Emerged dune-capped beach ridges, rising in altitude to the southwest, are interdigitated with saltmarsh creeks. (Images: **a** © P. and A. MacDonald/Aerographica/NatureScot; **b** Jim Hansom)

formed as RSL subsequently fell, but no sand or gravel ridges analogous to those at Morrich More or Culbin are evident under the blown sand veneer. The sandflat merges eastward with a prominent sand ridge at 2.4 m OD that extends southward to Dornoch Point. This ridge is capped by low sand dunes in the north but overlain by a prominent active dune cordon in the south that is locally breached and overwashed during storms (Hansom and Dunlop 2011). Seaward of the ridge, a wide sandy beach appears to be fed by sand deposits from offshore (Reid and McManus 1987) and by southward longshore drift. Dune-capped, west-trending recurves occur at the point, providing shelter for saltmarsh development. The intertidal zone west of Dornoch Point is dominated by a low-gradient, ~1 km wide sandflat, while to the east lie extensive tidal sand banks produced by ebb tidal flow out of firth (Fig. 22.8).

Unlike Morrich More, establishing the time and mode of emplacement of Dornoch Point is problematic since few well-defined landforms exist landward of the dunes. Both features lie within a coastal sub-cell, and so sediment exchange might be expected between them, but the dominant sediment drift direction is southward, toward Dornoch Point and its extensive intertidal and nearshore bars (Ramsey and Brampton 2000). Historical map evidence can be used to establish the longevity of sediment drift direction and its impact on the development of Dornoch Point. The earliest (1747–1755) reliable map depicts Dornoch Point and Innis Mhòr as almost-connected low-tide sand bars (Fig. 22.11a), and Admiralty charts of 1845 depict a short and stubby Dornoch Point with nearshore bars extending eastward, separated from Morrich More by a narrow channel only ~3.4 m deep at low tide. (Fig. 22.11b). This shallow channel may have been a sediment transit route from Dornoch Point to Morrich More, potentially contributing to the rapid accretion of Morrich More. Formed well before Dornoch Point reached its present position, the narrowing of both north and south shores of the firth appear to be longstanding, influencing the wave regime and associated coastal landform development in the central Dornoch Firth (Firth et al. 1995; Sect. 22.3.4).

22.3.6 Loch Fleet and Coul Links

Loch Fleet is a 6 km^2 inlet in the outer Dornoch Firth, partly enclosed by emerged and modern beach and dune systems backed by an abandoned Holocene cliff whose base lies at ~10 m OD (Figs. 22.1 and 22.12). Seaward of the abandoned cliff lies an emerged estuarine terrace at ~6 m OD, and peat buried below sand at 5.2 m OD has yielded a calibrated radiocarbon age of ~7.8 ka (Smith et al. 1992), consistent with deposition of the terrace (and cutting of the backing cliff) during the Holocene sea-level highstand. The emerged terrace is underlain by carse (estuarine silt and clay) deposits that are truncated seaward by at least two emerged spits at 10–11 m OD that extend southward and recurve into the mouth of Loch Fleet. Fronting these, the present sand and gravel spit extends ~4 km south from Golspie, its southward extension being arrested by outflow from Loch Fleet (Fig. 22.12). The altitude and configuration of the emerged spits indicate southward longshore sediment transport during the Holocene (Ramsey and Brampton 2000), with updrift erosion fuelling downdrift accretion (Smith and Mather 1973). Recent artificial protection of the northernmost section of the modern ridge has accelerated this trend (Hansom et al. 2013). Blown sand thinly veneers the emerged and present beach ridges, but sand dunes are only fully developed close to the ebb-tidal delta at the mouth of Loch Fleet where the sand supply remains locally plentiful.

To the south, beyond the intertidal sand bars produced by the outflow of Loch Fleet, a wide sandy beach fronts the extensive dune system of Coul Links (Smith and Mather 1973). Coul Links is a large foreland composed of emerged marine terraces that descend toward north–south aligned

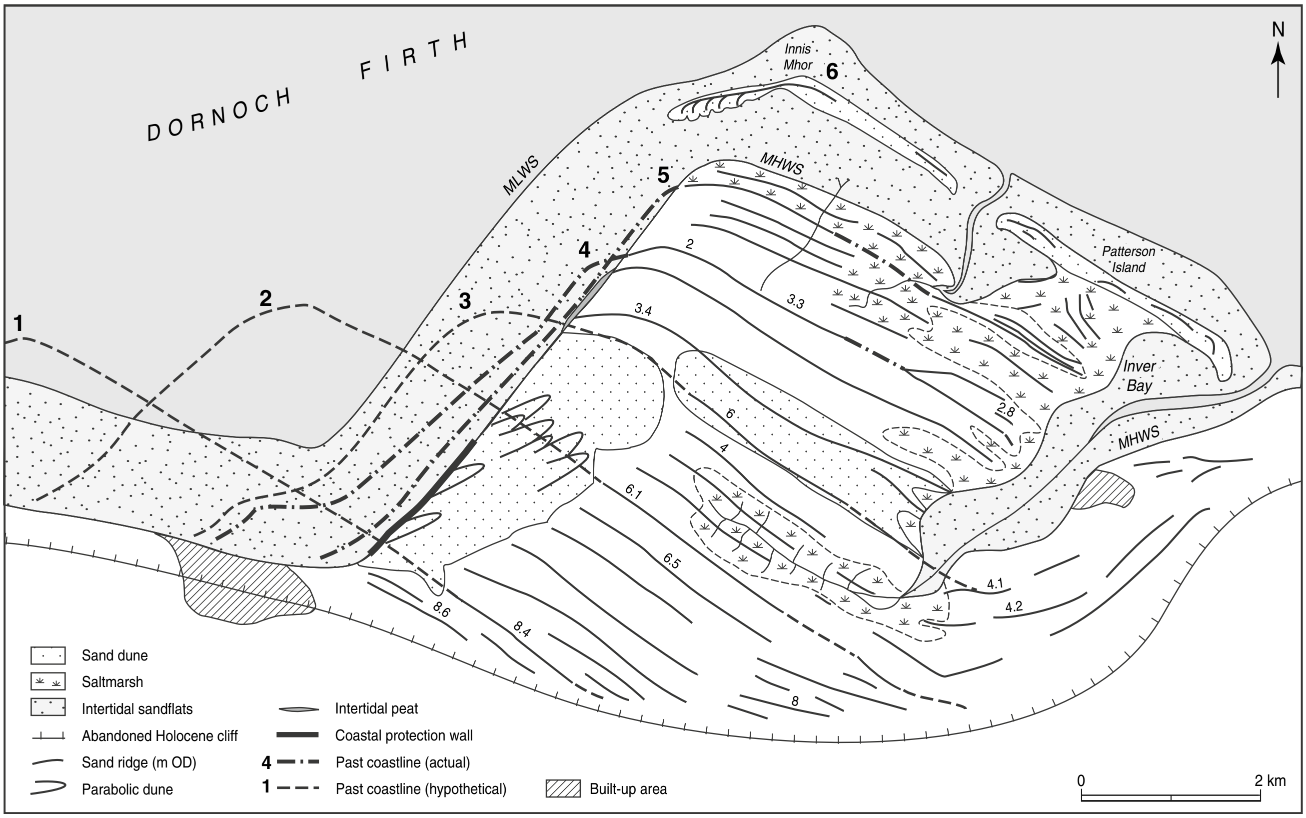

Fig. 22.10 Stages in the migration of the emerged foreland of Morrich More after ~7.4 ka (stages 1–6), reconstructed from the alignment of older emerged sand ridges, radiocarbon dating and historical evidence. Long-term erosion of the western shore has been accompanied by sediment accretion and shoreline progradation to the northeast. (From Hansom (1999). Licensed under Open Government Licence v3.0)

gravel ridges whose northern ends recurve westward into the mouth of Loch Fleet. The ridges are capped by thick deposits of blown sand with dune crests reaching heights of 7 m OD between low, inter-ridge, seasonally flooded dune slacks (Fig. 22.13). The sand supply to the Coul Links beach and dunes has been augmented by sand eroded from the beaches north of Loch Fleet that lie within the same sediment sub-cell, initially accumulating on the ebb-tidal bars at the loch mouth and subsequently being blown landward. However, recent shoreface erosion of the dune cordon (Fig. 22.13) suggests that Coul Links and its dunefield are now experiencing losses that are likely to be exacerbated by future sea-level rise.

22.4 Conclusion

The emerged and present-day beaches and dunes of the Moray and Dornoch Firths comprise an assemblage of coastal landforms unparalleled in Scotland. Despite the variety and range of coastal landforms in these areas, several factors are common to all. First, all have developed as relative sea level declined from a mid-Holocene highstand that in most areas is marked by a backing cliff with its base at ~10 m OD; subsequent sea-level fall can be traced by a progressive reduction in the height of emerged beach ridges, though not all parallel the present coastline. Second, throughout the period of falling relative sea level, spit extension and shoreline migration and progradation have been fed by an abundant supply of glacigenic sediment derived from both onshore and offshore sources. Third, longshore sediment transport driven by waves from the east and north has dominated the outer coasts throughout most or all the Middle and Late Holocene. As a result, sediment transport (and spit extension) has been westward along the southern shore of the Moray Firth and southward along its northwestern shore; only within the Dornoch Firth is there evidence of a switch in the direction of longshore drift. Collectively, these three factors have favoured the development of extensive suites of emerged and modern beaches and spits that have supported extensive sand dune development. Geomorphological, chronological, and historical evidence demonstrates that the Moray Firth coastline has experienced continuous change, in some cases involving simple down-drift migration of forelands or extension of spits, as at Whiteness Head, Morrich More, Dornoch Point and Loch Fleet, in others reflecting changing morphogenetic conditions, as at Ness of Portnaculter and Ardjachie Point,

Fig. 22.11 **a** Roy Map of 1747–1755 depicts the proximity of the twin sand banks of Dornoch Point and Morrich More. **b** Admiralty chart of 1845 (depths in feet) showing the deepest part of the channel at the mouth of the Dornoch Firth to be only ~3.4 m below low tide, enabling southward sediment transfer to Morrich More. (Images: **a** Crown copyright; **b** Courtesy of the National Library of Scotland)

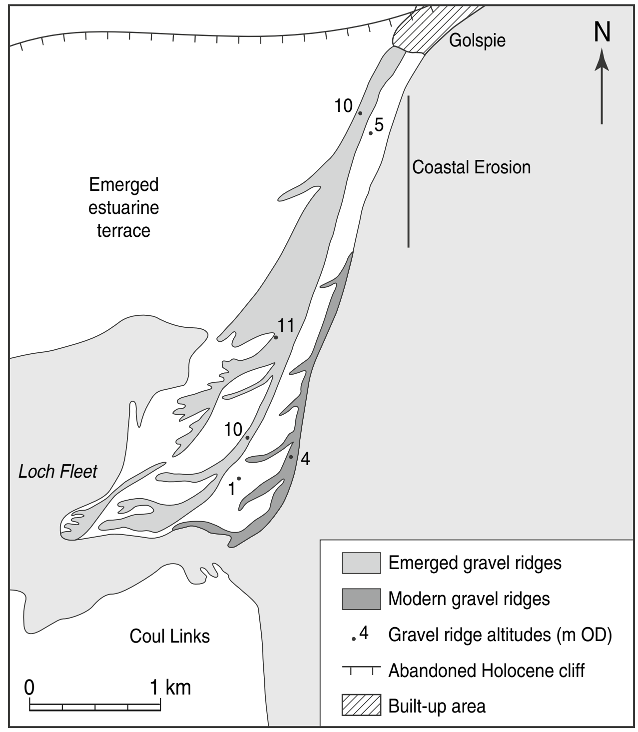

Fig. 22.12 Southward-extending emerged gravel spits enclosing Loch Fleet and its emerged estuarine silts and clays. Recent erosion in the north and associated southward sediment transfer continues a long-standing pattern. (From Smith and Mather (1973). Licensed under Open Government Licence v3.0)

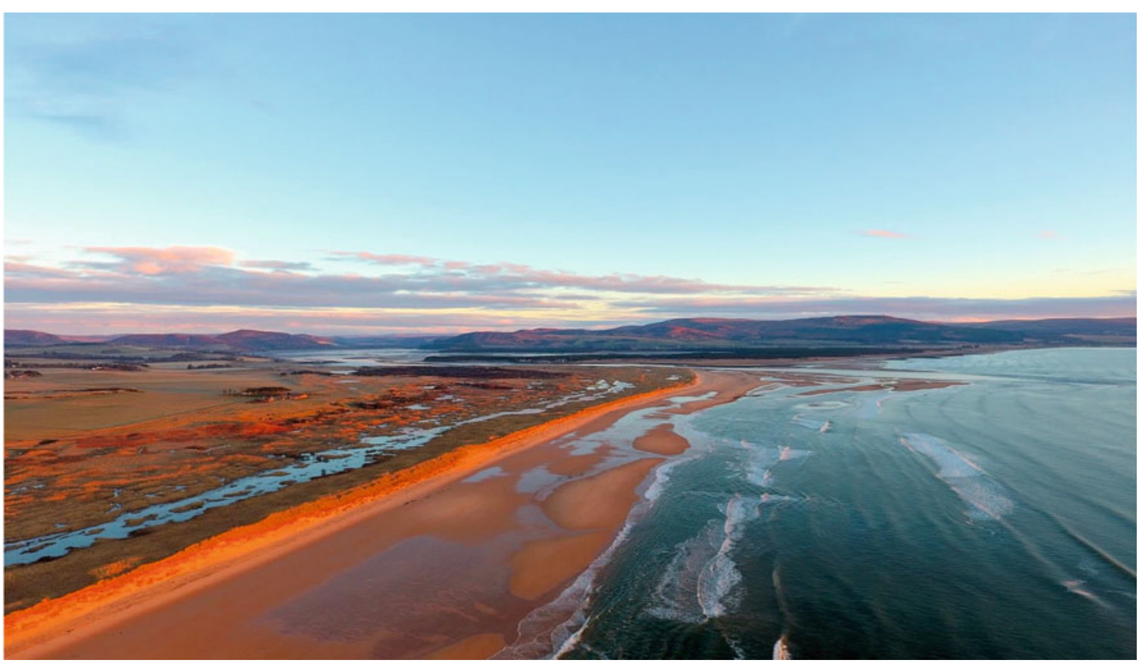

Fig. 22.13 View north over Coul Links toward the mouth of Loch Fleet. The frontal dune cordon is erosional, despite a wide intertidal beach and accretion on the ebb-tidal sand banks of Loch Fleet. Flooded dune slacks occur landward of the frontal dune ridge. (Image: Craig Allardyce)

and at some involving diversion of river outlets and subsequent breaching of spits, as at Spey Bay and Culbin Sands. Some formerly dynamic dune systems have been artificially stabilised, while others have evolved with less human interference. Despite a long history of accretion, increased beach erosion suggests that sediment supply may now be diminishing at a time of rising sea level, a scenario that affects many Scottish beaches and threatens the long-term stability of the coastal landforms.

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Beaches and Dunes of Eastern Scotland

23

James D. Hansom

Abstract

Substantial beach and dune complexes have formed along the east coast of Scotland, particularly, within St Andrews Bay, Montrose Bay, Aberdeen Bay and Belhaven Bay, where the interplay of relative sea-level change and availability of glacial and paraglacial sediment sources has locally favoured sand and gravel accumulation since the mid-Holocene. Most of these complexes began to develop when sea level was higher and sediment supply abundant, but have since merged, often imperceptibly, into younger and modern features at present sea level. Some substantial dune systems have now stabilised into fixed parabolic dune forms, whilst others remain highly active and mobile. In locations, where local factors have offset an overall decline in the availability of beach-building sediments, coastal accretion has continued, albeit supplied from the ongoing erosion of beaches and dunes elsewhere. Beach and dune development in eastern Scotland can be understood in the context of discrete coastal sediment cells, with boundaries generally formed by prominent headlands, a context likely to have been in place over much of the Holocene.

Keywords

Coastal erosion and accretion • Beaches • Spits • Sand dome • Parabolic dunes • Emerged beaches • Sea-level change • Sediment cells

23.1 Introduction

The east coast of Scotland stretches from Cairnbulg Point in the north to St Abbs Head in the south and is punctuated by the major estuaries of the Firth of Forth and Firth of Tay (Fig. 23.1). Beach and dune landforms are widely developed and form the most prominent depositional landforms of this coast, particularly, north of the Forth estuary (within coastal cell 2, Sect. 23.2.3) where they are contained within the large embayments of St Andrews Bay (Tentsmuir and Barry Links), Montrose Bay (Montrose and St Cyrus beaches) and Aberdeen Bay (Sands of Forvie). South of the Forth estuary (within coastal cell 1), the development of large beach and dune systems is more limited, although the spit, dune and saltmarsh complex of Belhaven Bay is a notable exception. The bays and beach complexes are typically pinned between rock headlands and stretches of rocky coast that include cliffs, arches, stacks, caves and shore platforms, many of which developed under formerly high sea levels.

The beaches and dunes of the east coast have mainly accumulated over the last ~6–7 ka in response to the interplay of eustatic sea-level change, glacio-isostatic land uplift and variations in paraglacial sediment supply. The local impact of these factors is manifest in a range of contrasting beach and dune responses. For example, long-term sediment accretion continues today at Tentsmuir, where there has been almost continuous seaward advance of the shoreline over at least the last ~5 ka, some of it sourced by longshore transport (Fig. 23.2), and south of the Forth estuary at Belhaven Bay. By contrast, the diminution of sediment supply to the beaches and dunes of Montrose and Aberdeen Bays has resulted in recession in the south and sand transfer to the north of both bays, producing accreting beach and dune systems at St Cyrus in Montrose Bay and at Sands of Forvie in Aberdeen Bay.

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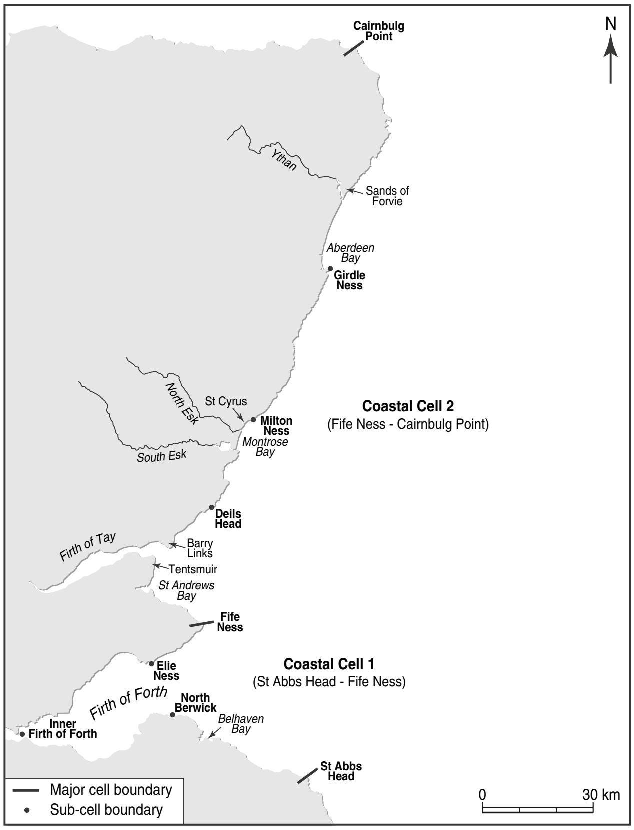

Fig. 23.1 East coast of Scotland, showing the location of key sites and coastal sediment cell and sub-cell boundaries

Many of the modern coastal systems on the east coast of Scotland are adjacent to emerged or buried beaches emplaced under past Holocene sea levels and now adorned with dune systems that have been modified by processes similar to those that are active at present. All of the beaches and dunes in this region owe their development to the reworking of glacial and paraglacial sediments that veneer the underlying onshore and offshore geological structure and limit its local control over beach development. Sea-level change and the local morphogenetic environment are also key controls. Several cities and towns are located on the east coast, and human activities have increasingly impacted the development of coastal landforms, particularly, where the coast has been modified by artificial defences.

Some of the beach and dune systems on this coast lie within the Midland Valley, a terrane dominated by Devonian sedimentary rocks north of St Andrews and Carboniferous sedimentary rocks farther south, with very limited outcrop of more resistant igneous rocks at the coast. North of the Highland Boundary Fault, which crosses the coast south of Aberdeen, the underlying rocks are mainly Dalradian metasedimentary rocks, though Aberdeen Bay itself is partly underlain by Devonian sedimentary inliers and enclosed by headlands of resistant metapelites. Rock structure and lithology, however, represent only passive controls on beach development on this coastline, as sediment supply has been dominated by reworking of terrestrial and offshore glacigenic deposits.

Fig. 23.2 Low-tide view east over the mouth of the Tay estuary with Barry Links and Buddon Ness on the north bank and Tentsmuir Point in the foreground on the south bank. The main channel of the Tay is marked by linear tidal and subtidal banks of Gaa Sands in the north and Abertay Sands in the south. (Image: © P. and A. Macdonald/ Aerographica/NatureScot)

23.2 Controls on Coastal Development

23.2.1 The Impact of Glaciation

The legacy of successive Pleistocene glaciations has influenced coastal development in two ways: erosion of pre-existing valleys within which beaches have later developed and provision of sediment for beach development. The last Scottish Ice Sheet extended offshore from the present coastline of eastern Scotland across the floor of the North Sea Basin (Chap. 4). A large moraine complex, the Wee Bankie Moraine, marks a significant north–south aligned recessional stage that interrupted overall retreat of the ice margin (Bradwell et al. 2008). It lies ~35–50 km east of the present coastline and forms part of the Wee Bankie Formation of glacigenic sediments, which provided a source of abundant sediment available to move onshore as relative sea levels fell following deglaciation. Shorewards of the Wee Bankie Formation, glacimarine and estuarine sediments were deposited between ~17.8 and 16.0 ka as the ice margin retreated into the Firth of Forth and Firth of Tay; these correspond to the Errol Clay Formation, which occurs onshore along the Tay estuary and extends into St Andrews Bay, Montrose Bay and Aberdeen Bay (Peacock 1999; Roberts et al. 2019). By ~16.0 ka, the last Scottish Ice Sheet had retreated from the Firth of Tay and probably much or all of the east coast (Ballantyne and Small 2019; Chap. 4), depositing large amounts of glacifluvial sand and gravel in the nearshore zone and low-lying coastal areas.

Fluvial reworking and transportation of onshore glacigenic sources during the Lateglacial and Holocene have provided an abundant supply of sediment to beaches on the outer coast, such as the emerged beaches behind Barry Links, Tentsmuir and Belhaven (Whitbread et al. 2014). Where the Tay estuary meets the North Sea, the process continues today with substantial amounts of sand deposited and reworked locally (Fig. 23.2). However, modern erosion in Montrose and Aberdeen Bays suggests that the supply of sediment from offshore and onshore glacigenic sources has diminished substantially. Similarly, Ritchie (2000) considered the once plentiful supply of beach and dune-building sand at Forvie to have been a one-off paraglacial legacy based on non-renewable glacigenic sediment sources.

23.2.2 The Legacy of Sea-Level Change

Sea-level change during the Quaternary has influenced both coastal configuration and position. On the east coast of Scotland, Lateglacial and Holocene relative sea level (RSL) has fluctuated in response to changes in global sea level and local patterns of glacio-isostatic adjustment. In the Tentsmuir, Barry Links and Montrose areas this has resulted in buried as well as emerged shorelines. More nuanced impacts on the development of Scottish coasts relate to the way postglacial sea-level change affected coastal sediment budgets and the timing, volume and type of sediment arriving from offshore (Hansom 2001).

At ~16–14 ka, emerged beaches formed along most of the east coast (Cullingford and Smith 1980), but the depositional forelands of Tentsmuir and Barry Links at the mouth of the Tay estuary did not exist. As the margin of the last ice sheet retreated, marine and estuarine clays of the Errol Clay Formation were deposited in sheltered locations around the east coast (Peacock 1999). RSL continued to fall until ~11 ka, but beaches formed during this lowstand were subsequently buried by later estuarine and beach sediments as rising eustatic sea level temporarily outpaced glacio-isostatic uplift, leading to widespread Holocene marine transgression after ~7.8 ka. In estuaries and other sheltered locations, this transgression resulted in the deposition of the estuarine clays, silts and sands of the Carse Clay Formation. Close to the centre of isostatic uplift, in the Forth and Tay estuaries, the Holocene marine limit (Main Postglacial Shoreline) was reached within the period ~7.8–6.2 ka, but farther from the uplift axis the Holocene marine limit was diachronous, culminating at ~5.7–3.7 ka, and forming what is now termed the Blairdrummond Shoreline (Shennan et al. 2018; Smith et al. 2019).

Along most of the east coast of Scotland, RSL fall from the Holocene marine limit was accompanied by ongoing deposition of carse sediments (silty-clay estuarine deposits), some of which are now terraced by later incision, and the formation of suites of beach ridges that represent the development of progressively lower shorelines as RSL fell from its Holocene highstand (Armstrong et al. 1985). Many of these ridges merge imperceptibly with younger and modern shorelines, especially where veneered by blown sand. RSL continued to fall throughout the Late Holocene, but tide-gauge trends over recent years suggest that RSL rise may now be occurring and causing localised increases in the extent and rate of beach erosion (Hansom et al. 2017).

23.2.3 Present-Day Morphogenetic Controls

The general context of beach and dune development in Scotland is provided by a series of sediment-transport units called coastal cells; accumulated sediment can be exchanged within each cell, but not exported to adjacent cells (Ramsey and Brampton 2000a). Since prominent headlands generally form the cell boundaries, most cells are likely to have been in place at least since the mid-Holocene. Within each coastal cell, smaller sub-cell boundaries are not totally sediment-tight and can allow limited exchange of sediments. The east coast beaches from St Abbs Head northwards to Fife Ness lie within coastal sediment cell 1 (Ramsey and Brampton 2000a), and those from Fife Ness northwards to Cairnbulg Point fall within cell 2 (Ramsey and Brampton 2000b; Fig. 23.1).

The total wave and swell wave climate offshore of the Firth of Forth and Firth of Tay is dominated by wave approach from the north and northeast, with an important secondary vector from east and southeast (Ramsey and Brampton 2000a, b). Modulated locally by coastal orientation, a mainly westward wave-produced net drift direction occurs within the Firth of Forth and a generally northward net drift direction occurs on the coast from St Andrews Bay to Aberdeen Bay (Fig. 23.1). Within St Andrews Bay, however, local wave refraction produces a net southward drift in the north but a northward drift in the south.

23.3 Key Sites

23.3.1 St Andrews Bay: Tentsmuir and Barry Links

The extensive emerged beach and dune complex of Tentsmuir lies within St Andrews Bay, a shallow bay bounded to the north by the sand banks of the Tay estuary and to the south by a shore platform and rock headland (Fig. 23.1). Tentsmuir shares its development and sedimentary context with the large beach and dune foreland of Barry Links, on the north shore of the Tay estuary. Both are sites of long-term sediment accretion, with Tentsmuir having prograded seaward by 3.8 km over at least the last ~5 ka (Ferentinos and McManus 1981), and its sand dunes now comprise one of the largest areas of blown sand in Great Britain (Dargie 2000). The beaches at Tentsmuir Point include extensive areas of intertidal sand spits, banks and bars (Abertay Sands), these being matched by intertidal and subtidal sand bars (Gaa Sands) at Barry Links (Hansom 2003a, b; Figs. 23.2 and 23.3a).

Rockhead occurs at depth below Barry Links, overlain by till and glacifluvial sediments, themselves overlain by up to 20 m of estuarine clay of the Errol Clay Formation and 10 m of beach sand (Buddon Sand), the top of which is capped by a series of emerged shorelines (Paterson et al. 1981). The same general stratigraphy appears to continue beneath much of Tentsmuir (Ferentinos and McManus 1981). Coastal development at Tentsmuir and Barry Links has been dominated by sedimentation, both sites having inherited a sedimentary regime fuelled from onshore and offshore glacigenic sources. Sediment fingerprinting shows the present sediment regime of the Tay estuary to be marine-dominated, with 51% of sediment sourced from the south and 29% from the north, and only 20% of fluvial origin (Jenkins et al. 2002). Open-coast wave and tide interaction with the estuarine currents of the Tay have also impacted on the Holocene and recent coastal development at both sites.

Wave refraction modelling shows northeast waves driving a southeastward current at Barry Links and a weak northward current at Tentsmuir (Ferentinos and McManus 1981; Sarrikostis and McManus 1987). Waves refracted from the southeast produce a similar pattern. These wave-driven longshore currents are matched by tidal flows: the flood tide moves south along Barry Links, and northward along Tentsmuir due to a flood tidal gyre in St Andrews Bay, with ebb tides sweeping south over Gaa Sands and Abertay Sands aided by outflow from the Tay estuary (Ferentinos and McManus 1981; Fig. 23.3b). St Andrews Bay is shallow (Fig. 23.3a) and covered with glacigenic and fluvial sediments, allowing wave and tidal processes to transport large amounts of sediment to the beaches of Tentsmuir and Barry Links (Hansom 2003a, b). Since the overall orientation of the coastline has remained relatively unchanged since the mid-Holocene, the modern wave and tidal depositional context is likely to have been similar over the past ~6 ka, resulting in the long-term construction of both forelands within the context of falling RSL.

Nevertheless, the development of Tentsmuir and Barry Links over the Middle and Late Holocene likely produced changes in the hydraulic efficiency of the estuary, forcing spatial and temporal changes to the ebb-dominant Tay estuary mouth and its tidal and nearshore sediment banks. Early Holocene RSL rise almost certainly produced an increased transfer of marine sediments into the estuary, followed by a decline as RSL subsequently fell. More recently, the balance between marine and fluvial sediment input may have changed as a consequence of falling fluvial sediment delivery due to dam construction within the Tay catchment, leading to enhanced marine sediment dominance (Jenkins et al. 2002). The building of rail and road bridges across the Tay estuary at Dundee in the nineteenth and twentieth centuries and extensive land reclamation on the north shore may have further influenced estuarine hydraulics and sand-bank response. Despite this, hydrographic charts from 1795 show that only minor changes have subsequently occurred in the configuration and extent of Gaa Sands, Abertay Sands, and the intertidal bars on the north side of the Tay estuary. Ongoing vigorous longshore transport within St Andrews Bay and a reduction in sediment from the Barry Links coast due to twentieth-century defence works may force additional changes to estuary-mouth dynamics.

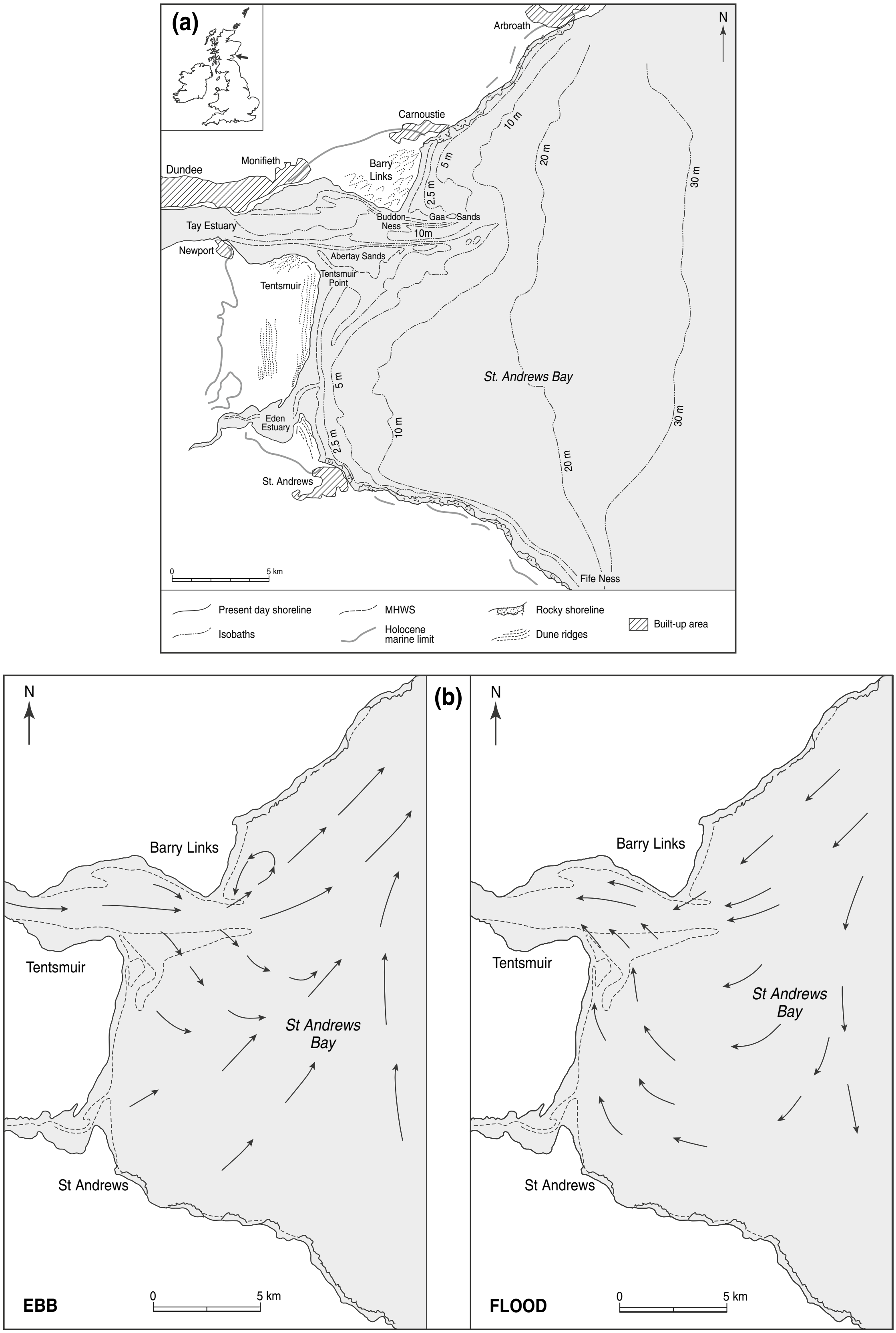

Fig. 23.3 **a** Forelands of Tentsmuir and Barry Links have built seaward from the Holocene marine limit at the mouth of the Tay estuary. Wave-driven sediment transport and interaction with the Tay outflow have produced extensive tidal and subtidal sand banks at Abertay Sands and Gaa Sands. MHWS: mean high water springs. **b** Mid-ebb and mid-flood tidal streams in St Andrews Bay, showing the general trend of tide-induced sediment movement towards the points at Tentsmuir and Barry Links. (From Ferentinos and McManus (1981) © 1981, reproduced by permission of John Wiley and Sons)

23.3.2 Tentsmuir

The Holocene marine limit at Tentsmuir lies at 7–9 m OD (Ordnance Datum), seaward of which lie more than 3.5 km of low sandy beach ridges trending parallel to the present coastline, with intervening lochs and damp slacks. Although a cover of blown sand makes identification of the emerged beach-top surfaces imprecise, two major sequences can be identified, one set trending east–west along the Tay estuary and the other north–south. Dunes in the south reach 6 m OD but drop to 2–4 m OD towards Tentsmuir Point (Hansom 2003a; Fig. 23.4). Continuing accretion at Tentsmuir Point is linked to sand delivered via the intertidal Abertay Sands, which reach ~1 km in width at the southern entrance to the Tay estuary and stretch eastwards for ~7 km beyond the point (Fig. 23.2). Rapid dune development at Tentsmuir Point is also fed from erosion farther south: between 1978 and 1990, an estimated 46,000 m^3 of sand was eroded from the dunes to the south with 33,000 m^3 arriving at Tentsmuir Point (McManus and Wal 1996); the remainder probably accumulated in the Abertay Sands and was subsequently recycled into the beach system. Since the 1890s, long-term northward sediment transport has contributed to northeastward extension of Tentsmuir Point by 870 m, at an average rate of almost 5 m a^{-1}. By 2018 much of the Tentsmuir shore had prograded, except in the south, with local variability at Tentsmuir Point, likely forced by dynamism within Abertay Sands (Fig. 23.5).

23.3.3 Barry Links

Inland of Barry Links, glacifluvial terraces of sand and gravel merge shoreward with emerged sandy beaches aligned SE–NW at the Holocene marine limit. However, the overall pattern of these beach ridges is masked by a blanket of dune sand, much of which has been blown into well-developed parabolic dunes (Hansom 2003b; Fig. 23.6 a), a dune form that is rare in Scottish coastal dunes. The Barry Links parabolic dunes have a pronounced V-shaped planform and elongate hairpin shapes that are unique in Britain. Landsberg (1956) used the resultant vector of winds (238°) to explain the orientation of the Barry parabolic dunes (Fig. 23.6b), with open exposure to the southwest permitting development of a regular parabolic form and northeastward dune migration. However, at some stage, the parabolic dunes became stabilised by vegetation colonisation and sand interception upwind by younger dunes. It is also possible that the inception and movement of the parabolic dunes may be related to windier conditions, for example, during the Little Ice Age (Jackson et al. 2019), with revegetation and stability being favoured by subsequent reduction in wind strength. Ongoing frontal erosion and truncation of the easternmost parabolic dunes (Fig. 23.6) may be related to recent sea-level rise, itself climate-related.

Buddon Ness at the southern tip of the foreland (Fig. 23.6) has undergone considerable historical change: by the early nineteenth century, the sixteenth-century lighthouse

Fig. 23.4 View north over Tentsmuir Point, showing tidal sand bars migrating northwards to recurve into the estuary or merge with Abertay Sands. Older, recurved, dune-capped beaches lie behind, backed by forested higher dunes. (Image: © P. and A.

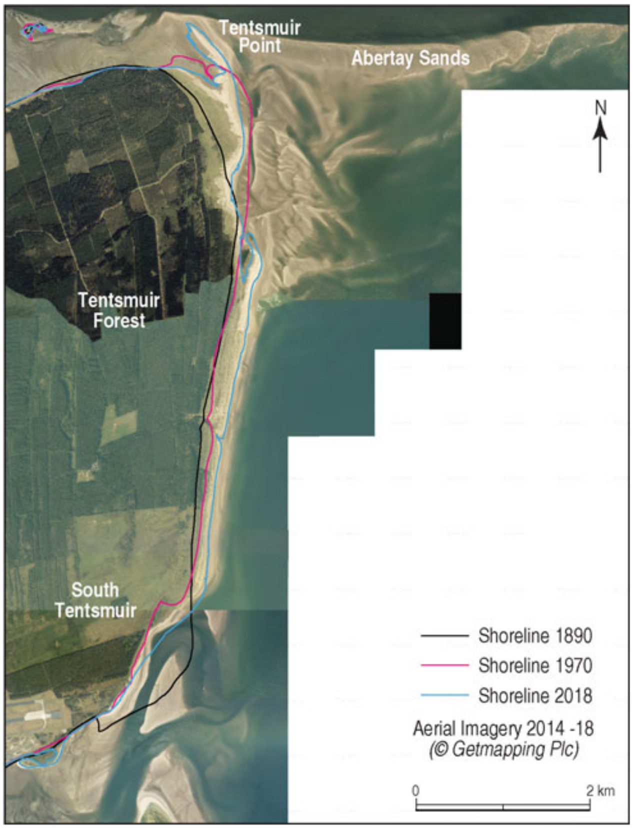

Fig. 23.5 Successive shoreline positions at South Tentsmuir and Tentsmuir Point in 1890, 1970 and 2018, showing erosion in the south between 1890 and 1970, but accretion along much of the coast to the north, albeit with more variability close to Tentsmuir Point itself. (Based on shoreline data from Hansom et al. 2017)

site was 6 m under water, and lay 2 km southeast of the present lighthouse (Wright 1981). Buddon Ness experienced 350 m of erosion between 1970 and 2014, and the east and west beaches have both retreated since the 1890s (Hansom et al. 2017). At a time when the overall sediment supplies to Scottish coasts are declining (Clayton 2003), the sediment supply to the Barry Links beaches has been further impacted by erosion protection structures. Between 1967 and 1996, 55 m of erosion occurred along 3.5 km of the northern part of the east beach; rock armour inserted to combat this has resulted in increased erosion rates towards Buddon Ness (Hansom et al. 2011).

23.3.4 Montrose Bay

Montrose Bay extends for 9 km between rock headlands that enclose broad sandy beaches in the south at Montrose and in the north at St Cyrus, backed by sand dunes and punctuated by two rivers (Fig. 23.1). The southern river (the South Esk) flows through a large (~7 km^2), enclosed, sediment-filled tidal basin that limits the delivery of sediment to the coast, and is flanked by emerged shoreline terraces and gently rising, glacially moulded topography (Smith and Cullingford 1985). The northern river (the North Esk) currently supplies sand and gravel to the coast at St Cyrus. The east coast wave

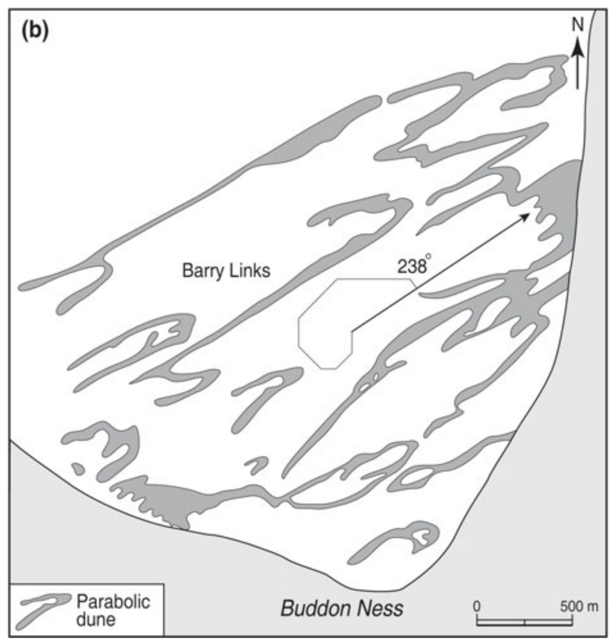

Fig. 23.6 **a** View northeast over Barry Links and Buddon Ness in the 1980s showing the vegetated parabolic dunes (now truncated in the east) and newly formed dune ridges in the foreground along the western shore. **b** Parabolic dunes at Barry Links in 1956 and the resultant vector of wind direction; by the 1980s some parabolic dunes had been truncated by frontal wave erosion. (Images: **a** © P. and A. Macdonald/Aerographica/NatureScot; **b** From Landsberg (1956), reproduced by permission of John Wiley and Sons)

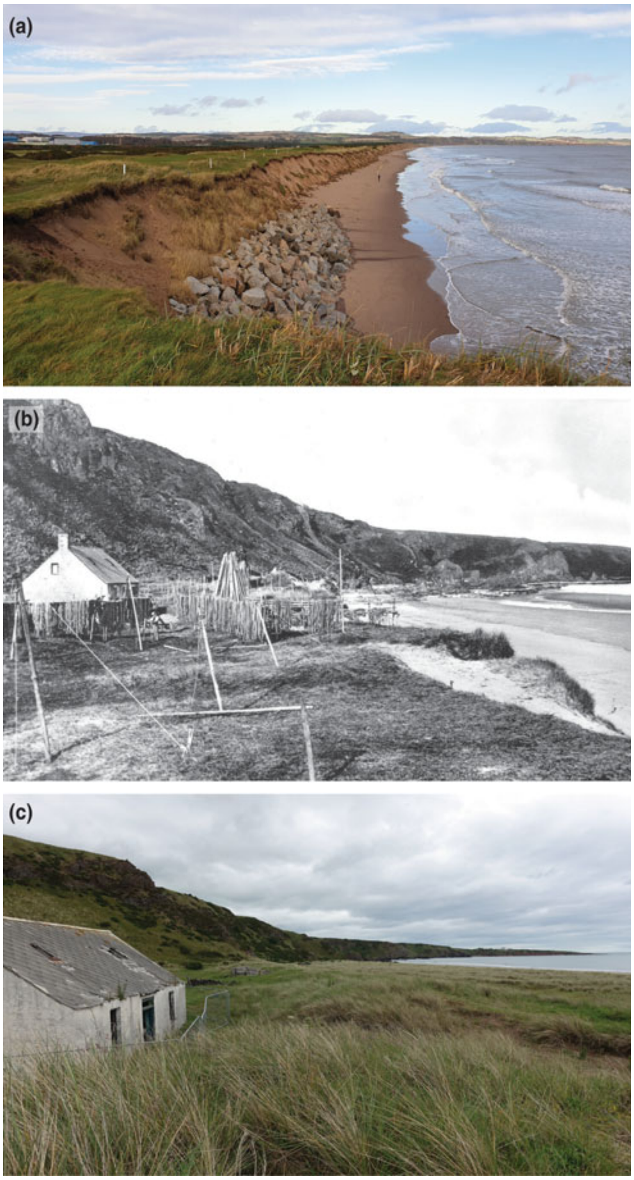

Fig. 23.7 **a** View north over Montrose Bay south, showing ongoing frontal dune erosion despite a wide intertidal zone. **b** A fishing station at St Cyrus around 1940, showing the dune coastal edge close to the backing cliff. **c** The same fishing station in 2020, showing up to ∼50 m of seaward migration of the dune edge since the 1890s, much of this during recent decades. (Images: courtesy of John Adams)

climate is generally dominated by north and northeast waves, but east and southeast waves are locally more important due to the orientation of the coastline and are responsible for a net northward longshore sediment transport (Ramsey and Brampton 2000b). As a result, the southern beach at Montrose Bay has a long history of coastal erosion (Fig. 23.7a) complemented by substantial beach accretion in the north at St Cyrus (Fig. 23.7b, c).

Prolonged twentieth-century beach erosion at Montrose suggests a decline in sediment supply. Since the 1890s, longshore and northward removal of sediment has resulted in beach lowering along the southern part of Montrose Bay, where wave erosion of the frontal dunes has narrowed the dune cordon (Fig. 23.7a). There is now a risk of overtopping at low points and flooding of the lower interior during storm conditions. Over the last 50 years or so, ∼90 m of beach and dune erosion in south Montrose Bay has contributed to ∼50 m of beach and dune progradation at St Cyrus (Hansom et al. 2017). This northward sediment shift has produced a slight clockwise beach rotation and appears to be a nineteenth- and twentieth-century effect driven by reduction in sediment delivery to Montrose Bay, with no evidence of any return movement of sediment southward.

23.3.5 Aberdeen Bay: Sands of Forvie

Aberdeen Bay extends for 24 km north from Girdle Ness to Collieston and is bounded by rock headlands that enclose a wide bay into which three rivers flow, feeding sediment to beaches that front dune systems (Fig. 23.1). The southernmost 3.5 km of the bay at Aberdeen is now artificially stabilised. To the north, sand and gravel beaches are backed by blown sand that veneers glacigenic deposits and emerged postglacial beaches. Much of the northern part of Aberdeen Bay is characterised by dune systems that have migrated downwind and sub-parallel to the coast. Of these, the only remaining undisturbed dune system is at Sands of Forvie, where a massive, free-standing, unvegetated dune is flanked by several arcs of mobile dunes in the south and large migrating parabolic dunes in the north (Hansom 2003c; Fig. 23.8a).

Although waves from the north and northeast are important, especially during storms, the northeastward trend of Aberdeen Bay coast favours waves from the east and southeast driving net northward longshore sediment transport. As a result, sediment eroded from the south is transported to sandy beaches in the north that are backed by spectacular dune systems. Best developed at the mouth of the River Ythan, the Sands of Forvie represent the fifth largest and least disturbed sand-dune system in Britain (Dargie 2000). The Sands of Forvie consist of two components: a southern peninsula (South Forvie) that contains a remarkable assemblage of active blown-sand landforms sitting on emerged gravel spreads, estuarine terraces and glacigenic ridges that reach 12 m OD (Fig. 23.8a); and a northern bedrock step (North Forvie) that reaches 57 m OD and is covered in glacial deposits capped by large parabolic dunes.

The dunes at Forvie likely originated prior to 5 ka (Ritchie 2000) and were interpreted by Landsberg (1956) as a series of large northward-migrating sand waves. South Forvie is characterised by an outer zone of active coastal

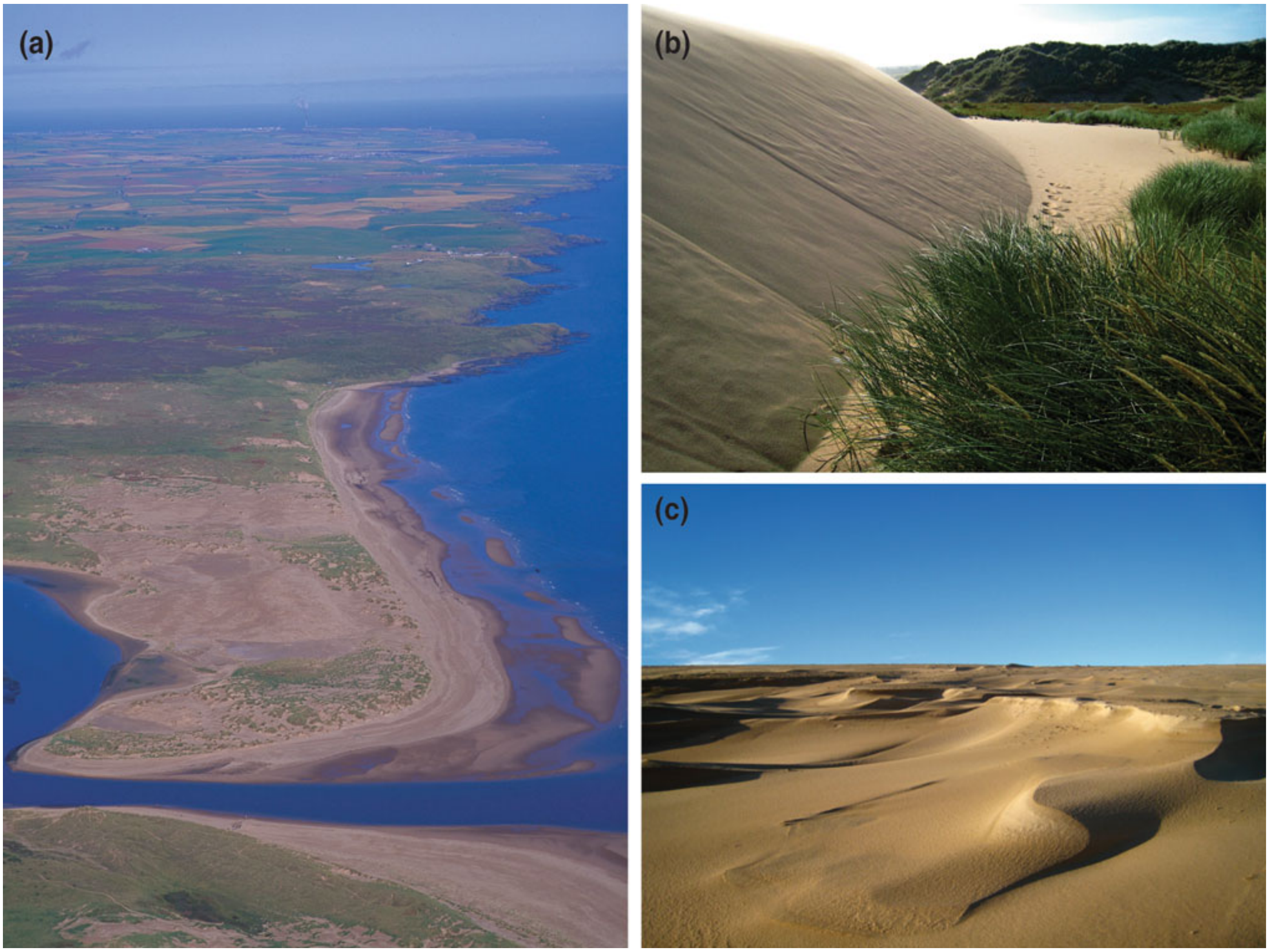

Fig. 23.8 **a** View north over the Ythan estuary to the South Forvie sand dunes, with the North Forvie heathland and dune plateau in the middle distance; South Forvie is dominated by a massive, unvegetated sand dome. **b** Active dunes at South Forvie encroaching northwards over earlier surfaces. **c** The windblown mobile surface of the unvegetated sand dome at South Forvie. (Images: **a** © P. and A. Macdonald/Aerographica/NatureScot; **b**, **c** © Martyn Gorman, CC BY-SA 2.0)

dune ridges fronted by the present beach, and an inner zone of great sand arcs limited in the west by the Ythan estuary. Parts of the sand arcs are unvegetated and all extend across the entire width of the peninsula, reaching a height of 35 m OD in the north where sand migrates over vegetated dune surfaces underlain by palaeosols and till (Fig. 23.8b). The most distinctive landform of South Forvie is an unvegetated dome of sand more than 1 km long, 200 m wide and over 25 m high (Fig. 23.8a). The dome surface is ornamented with sand waves and ripples (Fig. 23.8c); sand on the dome migrates both north across vegetation and deflation surfaces and west into the Ythan estuary.

Ritchie (1992) suggested that the South Forvie dunes have spread northward from the Ythan estuary at a progressively decreasing rate, but that North Forvie evolved differently, with sand blown upslope from the southern dune arcs becoming isolated and scattered on the open topography of the northern bedrock step. Episodes of sand influx from the south inundated farmland and two lochs in the north during the fifteenth, eighteenth and nineteenth centuries (Ritchie 1992; Fig. 23.9). Some of the parabolic dunes of North Forvie are now partly stabilised, but some active dunes reach heights of 20 m above the adjacent surface, their development being linked to vortices within crestal wind-flow patterns (Robertson-Rintoul 1990).

Cartographic evidence suggests the sediment budget of the Ythan estuary mouth and the Forvie sand dome has remained in equilibrium for at least 150 years, ensuring relative stability to the semi-closed sand-transport cell that feeds the bare sand dome: Aeolian sand transport westward over the dome feeds sediment to the estuary behind, whence it is carried by river flow and ebb tides back onto the beach to form the sand feed to the dome (Weatherill 1980). Given the general decline in sediment availability over the

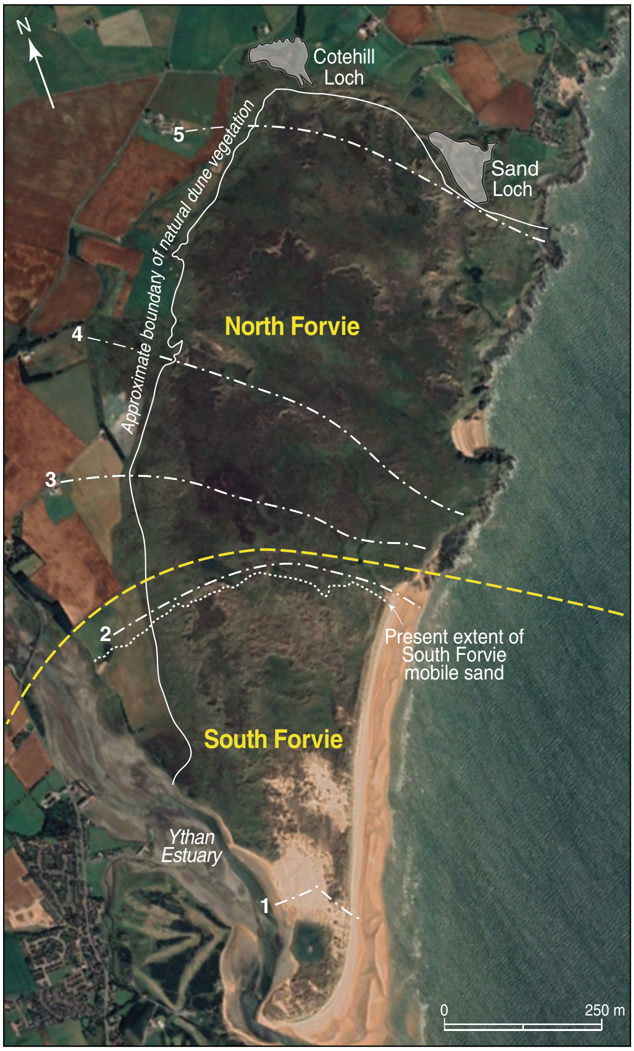

Fig. 23.9 Limits of sand movement at South and North Forvie showing the approximate limits of drifting sand. 1: at ~2 ka; 2: at 1.9 ka; 3: at start of the fifteenth century; 4: at end of the fifteenth century; 5: at end of the eighteenth century. Sand also intermittently inundated Cotehill Loch and Sand Loch until 1866. (Image: Google Earth™; data from Ritchie 1992)

Holocene (Hansom 1999), the Forvie dune system almost certainly originated at a time when substantially more sand was available in the nearshore sedimentary environment (Ritchie 2000).

23.3.6 Beaches South of the Forth: Belhaven Bay

Few substantial beach and dune formations occur south of the Forth estuary, but a notable exception is the large beach and spit complex at sheltered Belhaven Bay. The outer Firth of Forth is dominated by waves from the north to southeast (Ramsey and Brampton 2000a) that drive a net westward long shore sediment transport estimated to approach 300,000 m^3 a^{-1} (Pontee et al. 2004). The Firth of Forth is therefore a net sediment sink, but its crenulate southern coastline interrupts sediment transport, creating relatively closed local sediment systems isolated by rock headlands and deeper water, such as at Belhaven Bay (Fig. 23.1).

The River Tyne enters Belhaven Bay between opposing spits capped by low dunes, with an intertidal shore platform in the north and a wide sandy beach in the south (Fig. 23.10a). Extensive areas of salt marsh sit behind the spits, with ~5 km of emerged carse and estuarine flats rising landward from ~2 m OD to merge with the Holocene marine limit at ~6 m OD. The twin spits are slightly offset from each other, with maps from 1822 and 1890 showing the outer spit (Spike Island) to be absent and the inner spit (Sandy Hirst) in approximately its modern position and form, an apparent longevity supported by mature vegetation growth. Atypically for a spit, the narrow northern neck of Sandy Hirst sits on a 400 m wide intertidal shore platform strewn with stranded glacial erratics (Whitbread et al. 2014). At over 1.2 km long and 70 m wide at its wider distal end, the spit is intermittently active with wave refraction moving sediment southward into beach ridges and low dunes. The backing saltmarsh contains a well-developed dendritic pattern of creeks and linear salt pans. Large erratic boulders within the saltmarsh mud suggest that thin saltmarsh covers a landward extension of the intertidal shore platform. The 1890 shoreline shows the recurves of a now wooded former spit lying landward of the present-day spit of Spike Island (Fig. 23.10a).

Spike Island is a more recent feature. Prior to the 1940s, maps show only tidal sand banks in the spit's present position, and the beach ridges appear to have rapidly accreted since then (Whitbread et al. 2014; Fig. 23.10a). The source of this rapid sediment influx is unknown but substantial enough to allow low sand dunes to develop on top of beach ridges and provide protection for saltmarsh to develop in the lee of the spit. The present coastal margin of Spike Island is low and accreting at either end, though its northward extension is restricted by the River Tyne channel which runs along the southern edge of the shore platform. Nevertheless, Spike Island experiences phases of erosion under storm conditions and is subsequently rebuilt by landward migrating tidal sand bars (Fig. 23.10b). The influx of sediment sufficient to allow the recent construction of Spike Island makes Belhaven Bay unusual in a modern

Fig. 23.10 **a** Belhaven Bay in 2020, with Sandy Hirst and Spike Island spits lying seaward of intertidal saltmarsh. The time-series shoreline positions show the 1890s abandoned spit of West Barns Links (now forested) landward of the more recently formed Spike Island (from Hansom et al. 2017). **b** Spike Island foreshore erosion in November 2013, showing a lowered post-storm beach surface exposing gravels and 1940s defensive pillars. A new onshore-migrating sand bar on the seaward side will restore the pre-storm profile. (Images: **a** based on data from Hansom et al. (2017); **b** © Richard West, CC BY-SA 2.0)

Scottish context, where a reduced sediment supply is the norm.

23.4 Conclusion

The highlights of the beach and dune complexes of the east coast of Scotland are those locations where plentiful sediment supply in the past has been maintained by local factors to allow ongoing accretion to occur, despite coastal erosion elsewhere. This has produced several extensive beach and dune complexes that contain both older landforms that have now stabilised, such as the parabolic dunes at Barry Links and North Forvie, as well as highly dynamic landforms that continue to be subject to ongoing change. At Tentsmuir, the large foreland that has developed over the Middle and Late Holocene continues to receive sediment that drives accretion at Tentsmuir Point. At South Forvie, sediment transport over the bare sand dome to the Ythan estuary behind continues to be locally recycled back to the fronting beach, and at Belhaven Bay, substantial spit development has occurred since the 1940s. Elsewhere, beach and dune development is largely sustained by sediments eroded from updrift parts of the coast, such as occurs within Montrose Bay. Collectively, the beach and dune systems of the east coast of Scotland are representative of those where gradually falling Late Holocene sea levels and a diminishing supply of sediment have resulted in a present-day pattern of updrift erosion and coastline recession counterbalanced in places by downdrift accretion and coastal progradation, mitigated by local circumstances and modern sea level rise.

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24 Loch Lomond, Menteith and the Forth Valley

David J. A. Evans

Abstract

The physiography of south Loch Lomond and the western Forth and upper Teith valleys is strongly controlled by the lithology and structure of three dominant bedrock zones: the Dalradian metamorphic rocks of the Highlands, the steeply dipping strata along the Highland Boundary Fault and the Carboniferous lava plateaux of the Campsie Fells and Kilpatrick and Gargunnock Hills. A regional pre-Quaternary drainage network of rivers flowing east or west from an initial watershed that coincided with present-day highest ground was terminated by watershed breaching and the overdeepening and oversteepening of topography during numerous Quaternary glaciations. Impacts of the last (Devensian) ice-sheet glaciation are evident in overprinted drumlins that record Highland glacier ice flow down the Firth of Clyde, and from Loch Lomond and the upper Forth and Teith valleys to feed the Forth Ice Stream. Superimposed on these glacial landforms are those of the Loch Lomond Stade, including the end moraines of the Lomond, Menteith and Teith piedmont glacier lobes, which are associated with a chronostratigraphic record that represents the stratotype for the Younger Dryas in Britain. The glacitectonically thrust moraines of the Menteith ice lobe indicate possible surging behaviour. Glacioisostatically driven relative sea-level changes since the Late Devensian ice-sheet deglaciation are recorded by raised marine landforms and deposits within an altitudinal range of 10–15 m OD. Postglacial geomorphic processes are dominated by pervasive river incision and basin infill, and rock-slope failures that occur in a dense cluster in the Highland terrain to the west of Loch Lomond and around the margins of the basalt lava plateaux.

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Keywords

Loch Lomond Stade • End moraine • Hill-hole pair • Glacitectonic composite ridge • Ice-dammed lake • Spillway • Carse • Main Lateglacial Shoreline • Paraglacial rock-slope failures

24.1 Introduction

The Loch Lomond, upper Forth and upper Teith valley systems are typical of the large number of glacially overdeepened troughs created by the drainage of outlet glaciers from the Highland ice dispersal centres during multiple glacial cycles (Fig. 24.1). Such troughs, including the Forth and Teith valleys, partially follow the courses of former fluvial valley networks, as indicated by the sinuous planforms of their chains of elongate rock basins. Others reflect greater glacial modification and contain evidence of large-scale watershed breaches, best illustrated by Loch Lomond. In all three valley systems, glacier ice has crossed and dissected not only pre-glacial watersheds but also major geological structures (Fig. 24.2). Depositional landforms in the Loch Lomond, Forth and Teith drainage basins are dominated by those relating to the Loch Lomond Readvance, the final episode of glaciation in this area, which occurred during the Loch Lomond Stade of ~12.9–11.7 ka (Golledge 2010; Bickerdike et al. 2016, 2018a, b; Chap. 4). Older glacial landforms include west–east-orientated drumlins, created by subglacial streamlining when Highland glacier ice flowed from the dispersal area centred over Rannoch Moor, down the upper Firth of Clyde, Loch Lomond and the upper Forth and Teith drainage networks to feed into the eastward-draining Forth Ice Stream. The alignment of streamlined bedforms indicates former ice flow WSW–ENE between the Gargunnock and Ochil Hills and NW–SE over the Gargunnock Hills (Rose 1981; Hughes et al. 2014; Chap. 26). Overprinted drumlins reveal ice-flow directional changes during the Late Devensian ice-sheet

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glaciation (Dimlington Stade; ~31–14.7 ka). Loch Lomond Stadial subglacial bedforms are also superimposed over those of the last ice sheet in the area (Rose 1987; Fig. 24.3a). Glacioisostatically driven relative sea-level changes since Late Devensian ice-sheet deglaciation (Smith 1993; Peacock 2003; Smith et al. 2019) are well recorded by raised marine landforms and deposits, including eroded cliff lines around south Loch Lomond and the former expansive Flanders Moss estuarine flats (the Carse of Stirling) in the western Forth valley (Rose 1980a, b, c, 1981, 2003a, b; Hansom and Evans 2000).

24.2 Geology and Landscape Evolution

The landscape of the region is strongly influenced by the lithological controls of the Highland Boundary Fault, the adjoining Devonian sandstones and marls, and the Lower Carboniferous lava plateaux of the Campsie Fells and Kilpatrick and Gargunnock Hills (Hall et al. 1998; Figs. 24.1 and 24.2). The Highland Boundary Fault forms the boundary between the Midland Valley terrane to the south and the

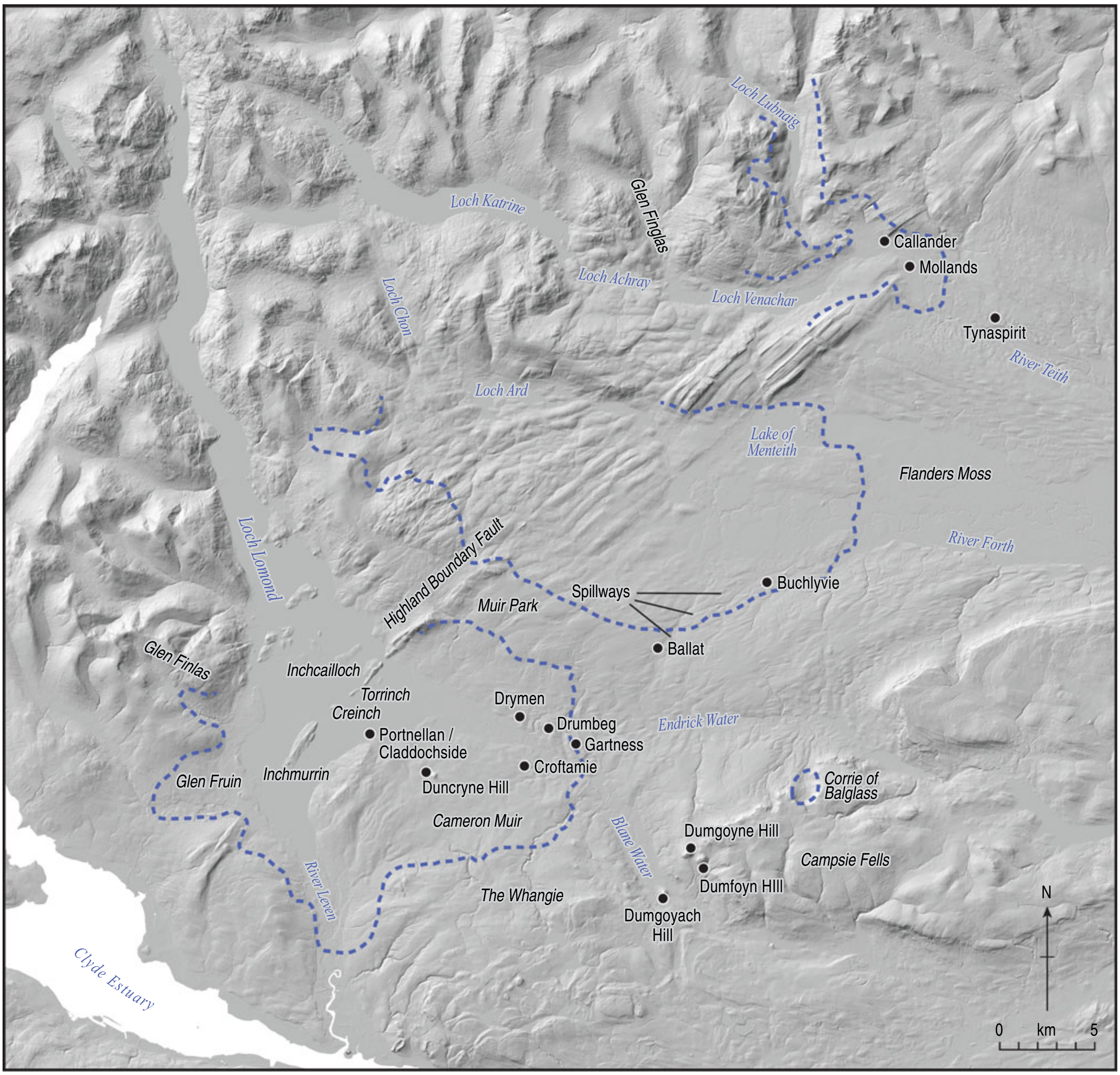


Fig. 24.1 Digital elevation model showing the general physiography of the western Forth valley and south Loch Lomond area; dashed blue lines mark the maximum extents of Loch Lomond Stadial glaciers. (NEXTMap Britain data from Intermap Technologies Inc., provided courtesy of NERC via the NERC Earth Observation Centre)

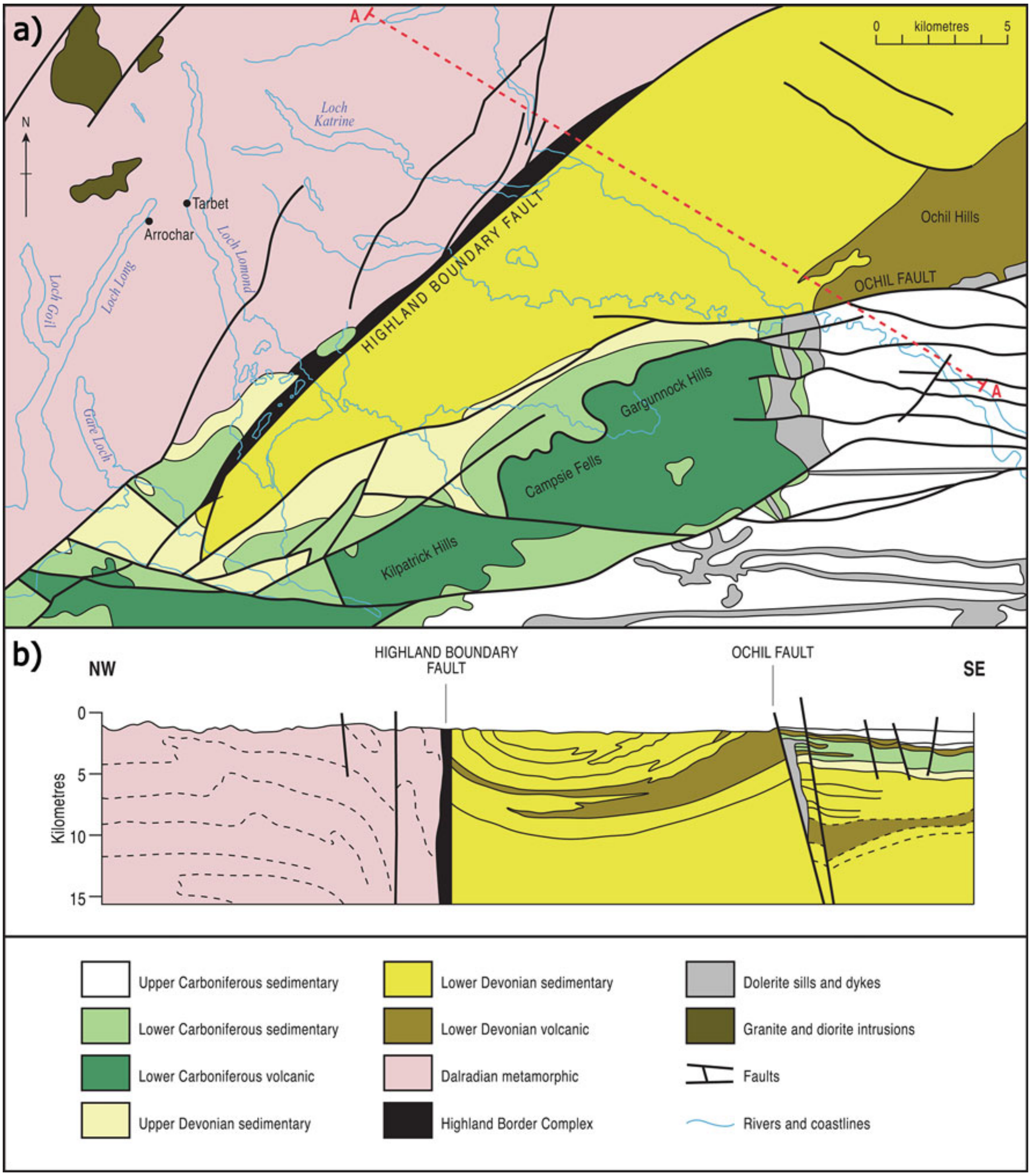


Fig. 24.2 **a** Generalized geology of the western Forth valley and south Loch Lomond area. **b** Cross section through the geological structures along the transect line A–A. (From Evans and Rose 2003a, after Browne and Mendum 1995)

Grampian Highlands terrane to the north. This part of the Midland Valley mainly consists of lowland areas underlain by Lower Devonian sedimentary rocks and upland plateaux of basaltic lavas and tuffs of the Clyde Plateau Volcanic Formation. The plateaux display stepped marginal cliffs reflecting stacked lava sheets, through which protrude cone-shaped hills (e.g. Dumgoyne, Dumfoyn and Dumgoyach) representing the volcanic necks that fed the lava sheets during the Lower Carboniferous (Figs. 24.1 and 24.4a). The SW Highlands are composed of the more resistant rocks of the Neoproterozoic Dalradian Supergroup, which were initially uplifted during the Caledonian Orogeny (Fig. 24.4b; Chap. 2). The Highland Boundary Fault is represented by a SW–NE aligned, prominent multicrested ridge that represents several fault-bounded slices of Cambrian and Ordovician strata that have been heavily deformed, metamorphosed and hardened, known as the Highland Border Complex. The regional morphology of this strong linear feature is best exemplified by Conic Hill on the east shores of Loch Lomond and the aligned islands of

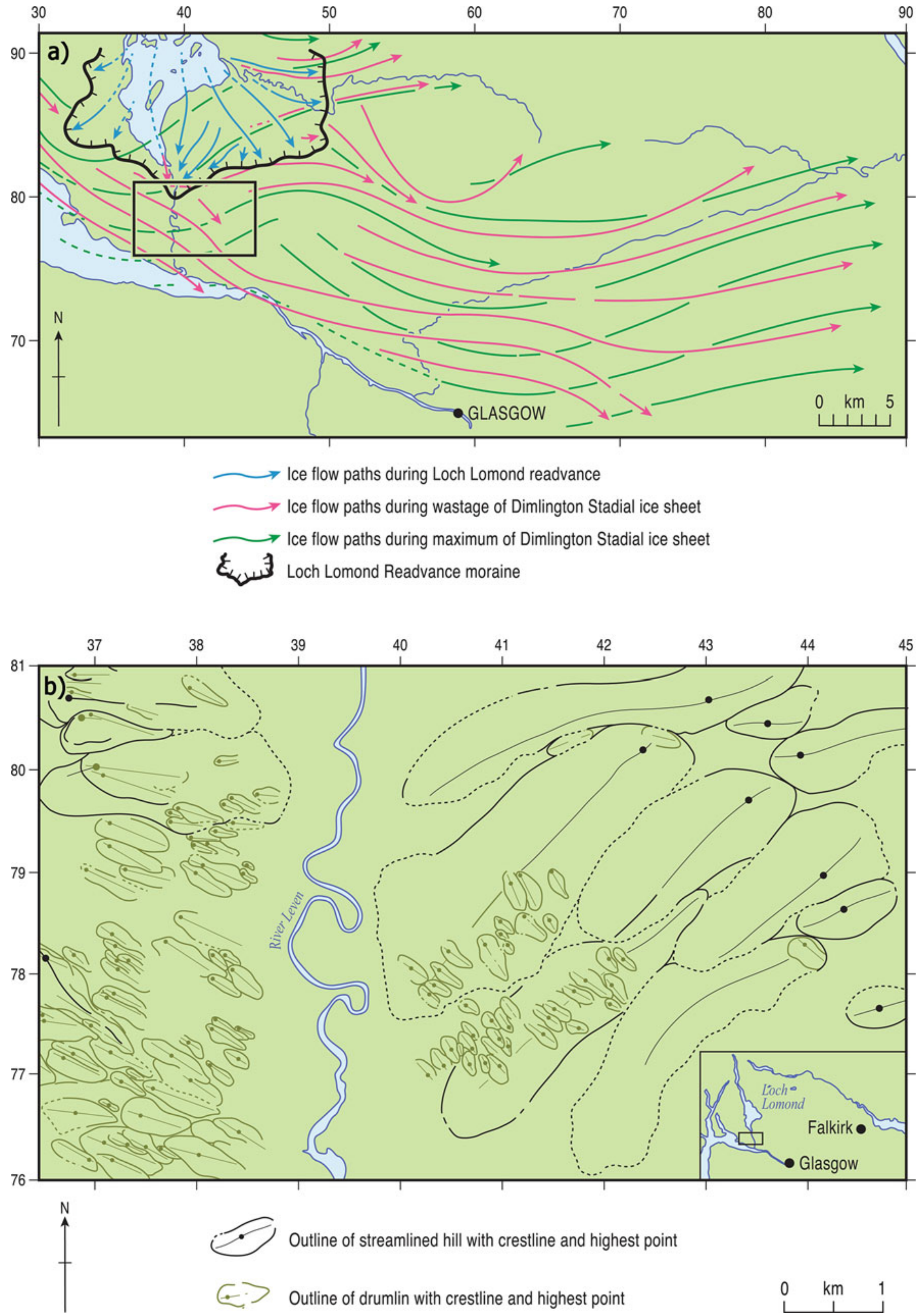


Fig. 24.3 Patterns of ice flow and subglacial landform superimposition. **a** Reconstructed ice-flow pathways, showing changes in ice-flow direction during the last ice-sheet glaciation and the radial flow of the Lomond glacier lobe during the Loch Lomond Stade. **b** Local details of drumlin superimposition in the boxed area in (**a**) as an example of the evidence used to infer changing ice-flow directions of the last ice sheet. (From Rose (1987) © 1987, Reproduced with permission from Taylor and Francis Group)

Inchcailloch, Torrinch, Creinch and Inchmurrin (Fig. 24.1). Conic Hill (Fig. 24.4c, d) is a classic hogback ridge with multiple crests representing the surface manifestation of steeply dipping, resistant strata (Evans and Rose 2003a).

24.3 Pre-glacial Fluvial Systems and Their Modification by Glacial Erosion

The pattern of pre-glacial drainage in the Loch Lomond area and the upper valleys of the Forth and Teith has been reconstructed by Linton and Moisley (1960) as four major drainage basins, across which the Loch Lomond trough was later excavated by glacial erosion (Fig. 24.5). According to their reconstruction, the Glen Falloch system in the north drained via Glen Dochart into the Tay. South of this, they suggested that the Teith was fed by drainage from the slopes of Ben Ime in the west and then eastwards through Loch Arklet and Loch Katrine. They envisaged that the pre-glacial River Forth drainage was fed by two southern drainage basins, one draining the eastern slopes of Ben Lomond and through the Duchray Water valley and Loch Ard, the other draining from Glen Douglas and Glen Luss; westwards drainage from the watershed of these two southern drainage basins was dictated by the fault-controlled Loch Long basin.

The termination of this pre-glacial fluvial drainage pattern occurred once the divides between the four major drainage basins were breached (Fig. 24.5) by successive episodes of southwards glacier flow from the West Highland ice dispersal centre on Rannoch Moor (Payne and Sugden 1990). The largest glacial breaches initiated the incision of the Loch Lomond trough and coincide with the overdeepened floor of

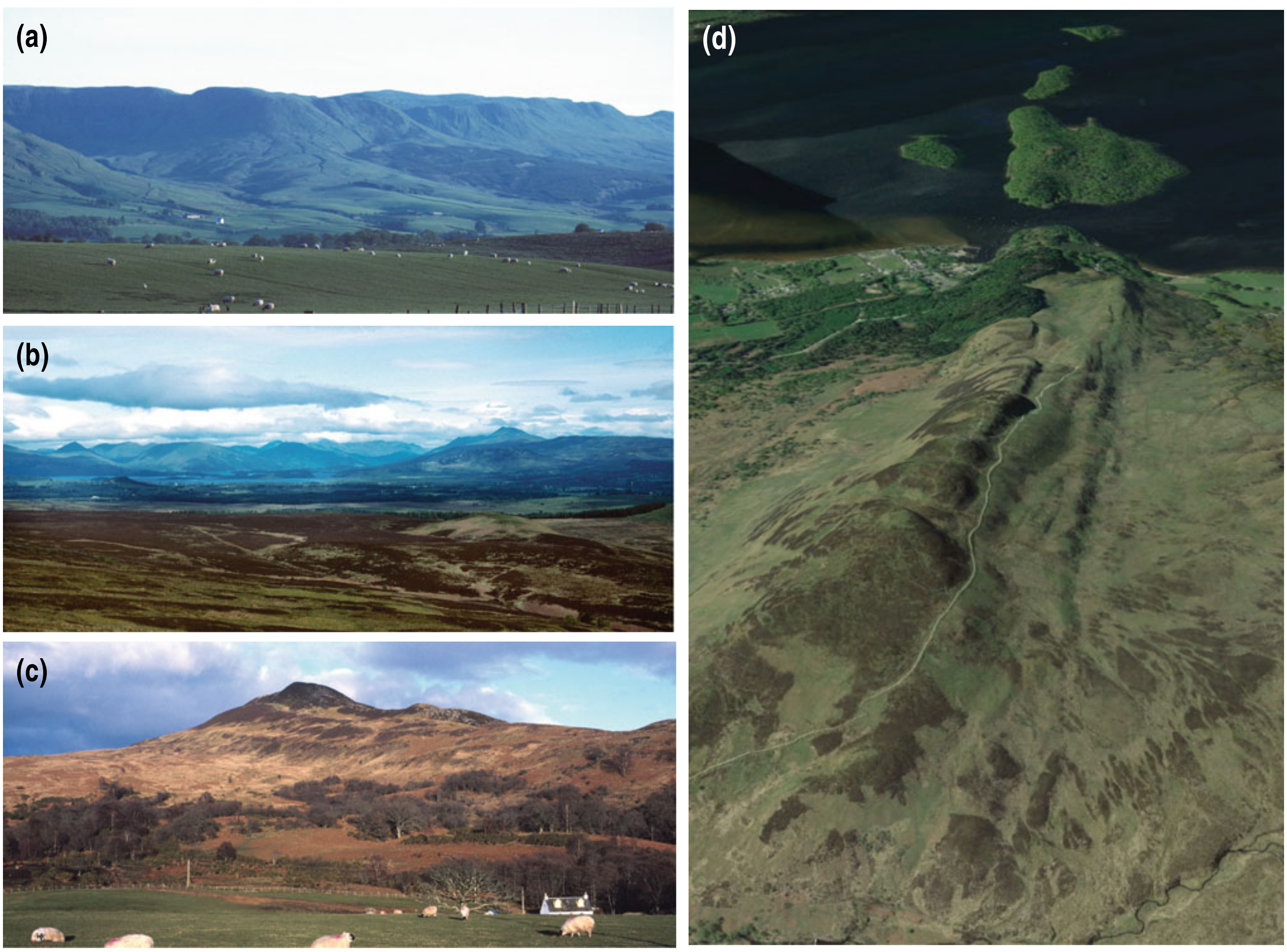


Fig. 24.4 Geologically controlled physiography. **a** The north side of the Campsie Fells, a plateau formed by the Clyde Plateau Volcanic Formation; the Corrie of Balglass cuts back into the plateau edge on the left. In the foreground are drumlins formed during Late Devensian ice sheet glaciation. **b** View from the southeast towards south Loch Lomond and into the Highlands, a mountainous region of resistant Neoproterozoic rocks. Ben Lomond (974 m) is the summit on the right. **c** and **d** Ground and aerial views of the hogback ridge of Conic Hill, showing the multiple summit ridges indicative of steeply inclined Cambrian and Ordovician strata. (Images: **a–c** David Evans; **d** Google Earth™)

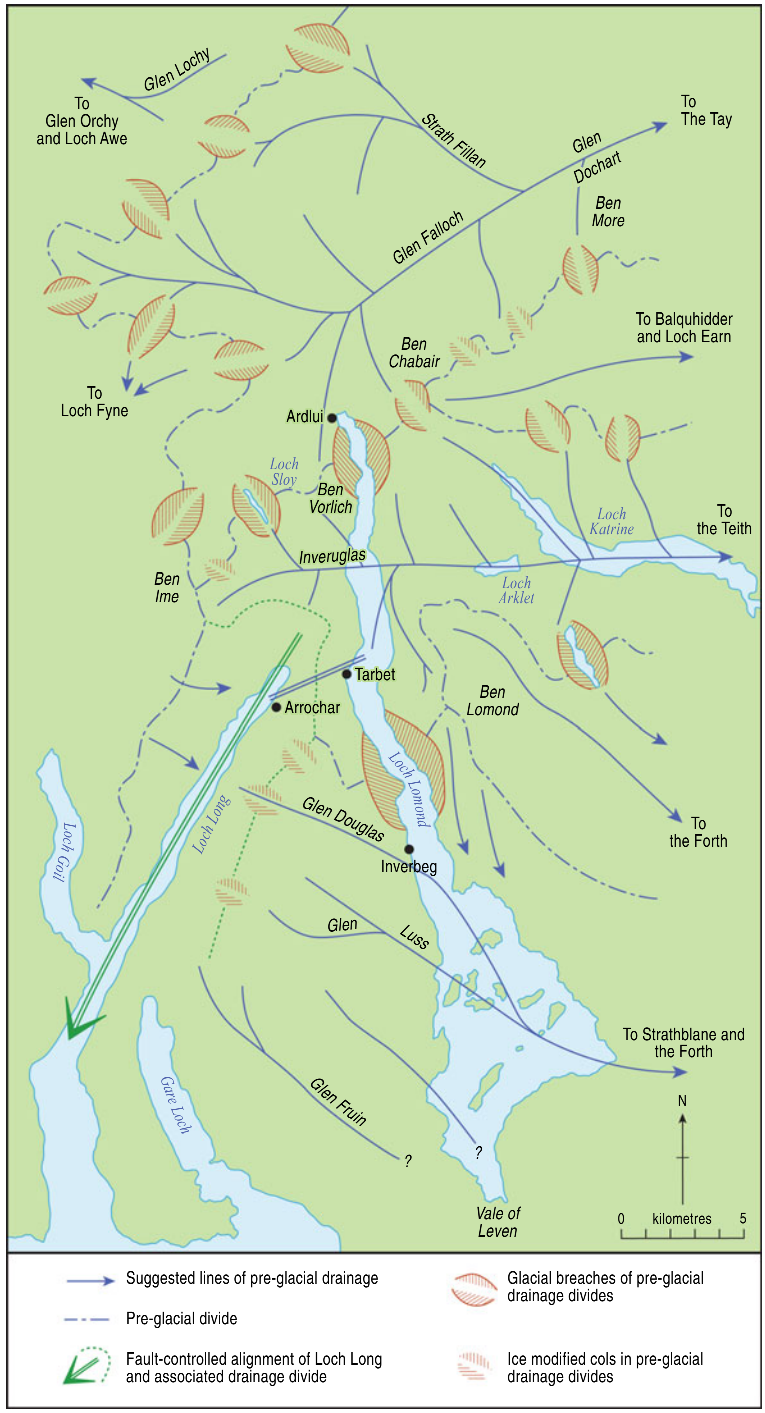


Fig. 24.5 Map of the pre-glacial drainage patterns of the Loch Lomond area and features related to the subsequent modifications by glacial erosion. (After Linton and Moisley 1960)

the loch between Ardlui in the north and Inverbeg in the south (Fig. 24.5). Linton and Moisley (1960) used the evidence provided by truncated spurs along the loch margins to reconstruct the pre-glacial land surface, which together with the depth of the overdeepened loch basin indicates a maximum of ~260 m of vertical erosion. As overdeepenings are typically located where ice discharges are relatively high and therefore likely around the zone of the long-term equilibrium line of glaciers (Løken and Hodgson 1971; Boulton 1996; Hallet et al. 1996; Anderson et al. 2006; Cook and Swift 2012), this depth of erosion is most likely associated with average Quaternary glacier coverage, the style of ice cover that predominated for the longest cumulative time during the Quaternary (Porter 1989; McCarroll 2006) and hence a mountain icefield, valley and piedmont lobe style of glacierization. Porter (1989) regarded the Younger Dryas style of glacierization as being indicative of the average glaciation style for the northern hemisphere and it is notable that the Loch Lomond Stadial ice cover in this region was one of mountain icefield, valley outlet glaciers and piedmont lobes (Bickerdike et al. 2016, 2018a, b). The islands of southern Loch Lomond indicate a reduction in vertical erosion in the area of, and south of, the Highland boundary, where former ice flow escaped the topographic constriction of the adjacent mountains. Hence, the bottom topography of Loch Lomond resembles that of fjords, where overdeepenings formed at the zones of greatest erosion give way down-ice to sills or thresholds where ice flow is less constrained and hence ice velocities decrease (Løken and Hodgson 1971). Indeed, due to glacioisostatic depression of the crust, Loch Lomond was a seawater loch during glaciations, as indicated by the raised marine deposits around its southern shorelines, which were laid down during recession of the Late Devensian ice-sheet margin (Rose 1981; Browne and McMillan 1989; Hall et al. 1998; Peacock 2003). The locations of substantial lochs along the upper Forth valley (Lochs Chon and Ard) and upper Teith valley (Lochs Katrine, Achray and Venachar; Fig. 24.1) are representative of overdeepening of these valley systems by the erosive outlet glaciers flowing southeastwards from the Highland ice dispersal centre.

24.4 Glacial Geomorphology of the Loch Lomond Stade

Since their identification by Simpson (1933), the extensive end moraine complexes of south Loch Lomond and the western Forth and upper Teith valleys (Fig. 24.6) have been regarded as a special case in the geomorphology of the Loch Lomond Stade, because they record the development of large terrestrial piedmont lobes (Lomond, Menteith and Teith glaciers, respectively), a relatively unusual palaeoglaciological configuration for the stade, compared to the mountain icefield, cirque and valley glaciation styles that predominated elsewhere in Scotland (Bickerdike et al. 2018a). Also, unusual is the large volume of glacitectonically deformed sediment, particularly in the end moraines of the Lomond and Menteith palaeo-glaciers.

24.4.1 Glacial Landsystems of the Lomond Piedmont Lobe

The glacial geomorphology of the Lomond glacier lobe is associated with an excellent stratigraphic record of Dimlington Stadial (last ice sheet) and Loch Lomond Stadial tills separated by organic deposits dated to the Lateglacial Interstadial ($\sim$14.7–12.9 ka), thereby justifying this as the type area for the Younger Dryas in Britain (Hall et al. 1998; Evans and Rose 2003b; MacLeod et al. 2011). At the southern margin of the Lomond lobe, the occurrence of stratigraphic sequences of tills, glacilacustrine and glacimarine deposits has enabled the reconstruction of a former ice-dammed lake (Glacial Lake Blane) that formed during the retreat of the last ice sheet and subsequent Loch Lomond Readvance, as well as the incursion of the sea due to glacioisostatic depression at the end of the Dimlington Stade (Fig. 24.7).

The outer limit of the Lomond glacier is marked by a continuous belt of prominent moraine ridges. The landscape within this moraine limit is characterized by widespread till hummocks interspersed with eskers and kamiform features as well as drumlins that record radial ice flow beneath the piedmont ice lobe. These drumlins are superimposed on an older set that formed under the last ice sheet. The older drumlins dominate the terrain beyond the Loch Lomond Readvance limit and record NW–SE or west–east regional ice flow (Rose 1980d, 1981; Figs. 24.3 and 24.8; Chap. 26). The exhaustion of subglacial sediment up-flow from the Lomond ice margin is evident at Rowardennan, where whalebacks or rock drumlins record subglacial erosion. The end moraine can be traced from Glen Finlas and Glen Fruin in the west, into the Vale of Leven, across Cameron Muir and the valleys of the Blane and Endrick Waters and then along the south slopes of Conic Hill, where it merges with the upper limit of gullied drift (Fig. 24.8a, b). Along much of this length the moraine appears to be a single or complex terrestrial push moraine but in the Endrick–Blane lowlands there is a series of up to six inset ridges that records active ice-marginal recession in contact with Glacial Lake Blane. The outermost moraine forms a sharp-crested ridge in the north, but a flat-topped feature comprising laminated silts and glacifluvial sands and gravels in the south, indicating that the southern part of the moraine formed subaqueously in Lake Blane. A quarry section at Drumbeg, near Drymen, reveals that the ice margin fed subaqueous fans and then an ice contact, Gilbert-type delta into the lake before advancing into the stratified sediments, deforming them and creating a glacitectonite and till carapace (Rose 1981; Benn and Evans 1996; Phillips et al. 2002; Evans and Rose 2003a; Fig. 24.9). The water level in Lake Blane was dictated by a spillway located at $\sim$65 m OD at Ballat (Figs. 24.1 and 24.7e), which probably also formed the lake outlet during the earlier damming event at the close of the Dimlington Stade (Fig. 24.7b). Hence, the drainage in the Blane and Endrick basins was forced to flow eastwards into the Forth basin, opposite to its normal north-westward course into Loch Lomond. The former floor of Glacial Lake Blane is represented by a blanket of glacilacustrine deposits that form a locally gullied plain at $\sim$65 m OD along the valley of the Blane Water (Fig. 24.1). At the northern end of the Ballat

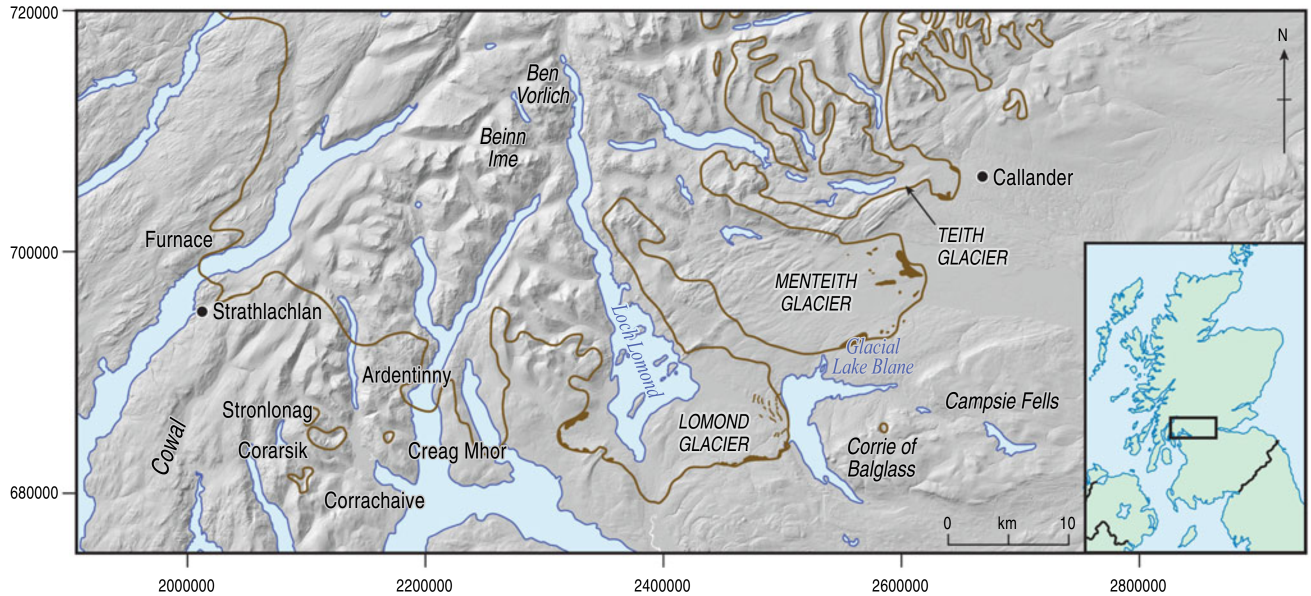


Fig. 24.6 Loch Lomond Readvance major moraines (brown) and associated ice-margin reconstructions (brown line) for the piedmont glacier lobes of the southern Highlands. (Modified from Bickerdike et al. 2018a)

spillway, the drainage routeway is marked by a series of inset channels aligned with the right lateral moraines of the Menteith glacier lobe (Rose 1980d, 1981; Figs. 24.1 and 24.7e).

24.4.2 Glacial Landsystems of the Menteith Piedmont Lobe

The moraines that directed the spillway waters at the end of the Ballat channel form part of an extensive latero-frontal moraine system that continues 8.5 km northwestwards towards the Highland Boundary Fault and eastwards for 12.5 km in an arc across the upper Forth valley around the Lake of Menteith, thereby demarcating the extent of the former Menteith glacier lobe (Figs. 24.1 and 24.6; Simpson 1933; Gray and Brooks 1972; Smith 1993). The varied characteristics of the moraine belt between the Lake of Menteith and Buchlyvie have enabled a detailed reconstruction of the former glacier dynamics (Evans and Wilson 2006). A subdivision of the moraine belt into four segments (A–D, Fig. 24.10) by Evans and Rose (2003a) was based on distinctive morphological signatures, partly verifying a similar approach by Smith (1993). In the south, segment D comprises two closely spaced moraine ridges, which continue in a broad arc from Buchlyvie westwards and then northwestwards, becoming the right lateral moraine of the Menteith glacier lobe on the Muir Park uplands. It is these moraines that are dissected by channels created by the water emerging from the end of the Ballat spillway from Glacial Lake Blane. Segment C is composed of an arcuate assemblage of shorter moraine ridges and intervening linear depressions that are subparallel in plan form. Segment B is an area of low-amplitude sand and gravel hummocks, kettle holes and linear ridges, aligned both parallel and transverse to the former direction of ice flow on a bedrock high north of the upper River Forth. Smith (1993) interpreted the hummocks and ridges as crevasse-fill features, created when sediment was deposited into crevasses in the Menteith glacier lobe as it advanced over the bedrock high. Finally, segment A comprises a relatively narrow and elongate, 30 m high ridge with a surface composed of numerous parallel and straight to slightly sinuous subsidiary ridges and dissected in places by sinuous channels (Fig. 24.11). This segment forms the eastern margin of the Lake of Mentieth, and its steep west-facing slope forms the linear eastern lake shoreline; the northern and southern shorelines are also largely linear and align with the northern and southern margins of moraine segment A.

The moraines in segment D are typical of the Loch Lomond Stadial terminal moraines identified in a variety of settings in the British Isles in that they form a dual ridge assemblage, probably created by ice-marginal pushing. Dating and numerical modelling of Loch Lomond Stadial ice margins in the English Lake District (Hughes et al. 2012; Brown et al. 2013) indicates that this dual moraine production likely relates to a strong climatic control, whereby the maximum ice extent was attained in the early, coldest part of the stade, followed by a more restricted but longer stillstand (cf. Baroni et al. 2017). In contrast, the landforms in moraine segments A, B and C indicate a quite different glacier dynamic scenario. The landform assemblage at

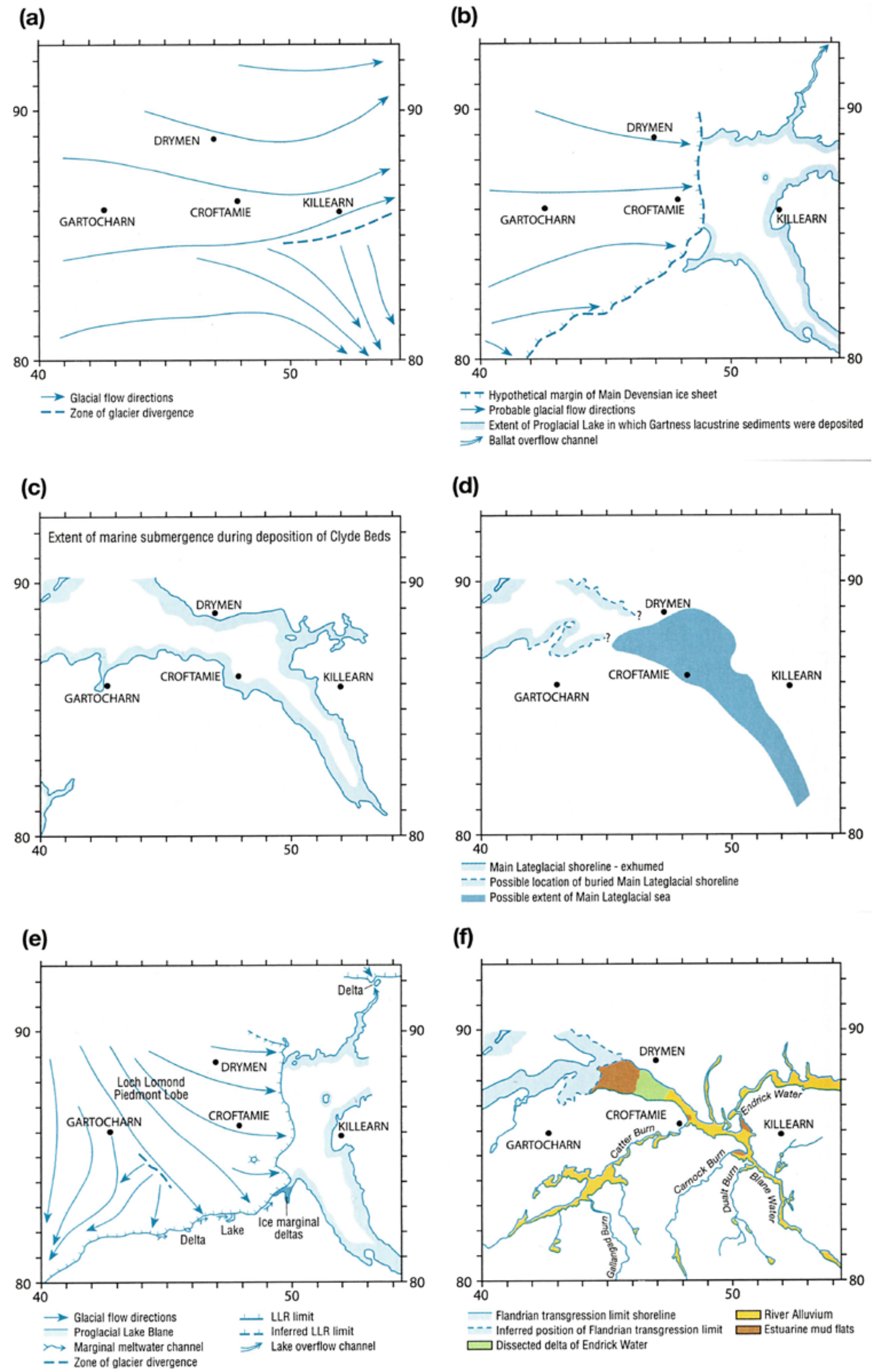


Fig. 24.7 Palaeogeographic reconstructions of events in the south Loch Lomond area. **a** Last glaciation (Dimlington Stadial) ice-flow directions. **b** Development of Glacial Lake Blane during ice-sheet recession. **c** Lateglacial Interstadial marine submergence and deposition of the Clyde Beds (marine clays). **d** Formation of the Main Lateglacial Shoreline. **e** Loch Lomond Readvance and damming of Glacial Lake Blane, which drained northwards towards the margin of the Menteith glacier lobe. **f** Holocene marine submergence and the construction of a delta at the mouth of the Endrick Water in Loch Lomond. (After Rose 1980d)

moraine segment A, together with the Lake of Menteith and the straight shorelines aligned with the boundaries of the moraine, display the characteristics of a glacitectonic hill-hole pair (Evans and Wilson 2006; Fig. 24.11b). Similarly, the multiple parallel ridges and elongate depressions of moraine segment C closely resemble the diagnostic form of a glacitectonic composite ridge. In both moraine segments, each minor ridge represents an individual thrust slice or the nose of an individual fold in an overthrust folded sequence of strata. Evans and Wilson (2006) also reported asymmetric

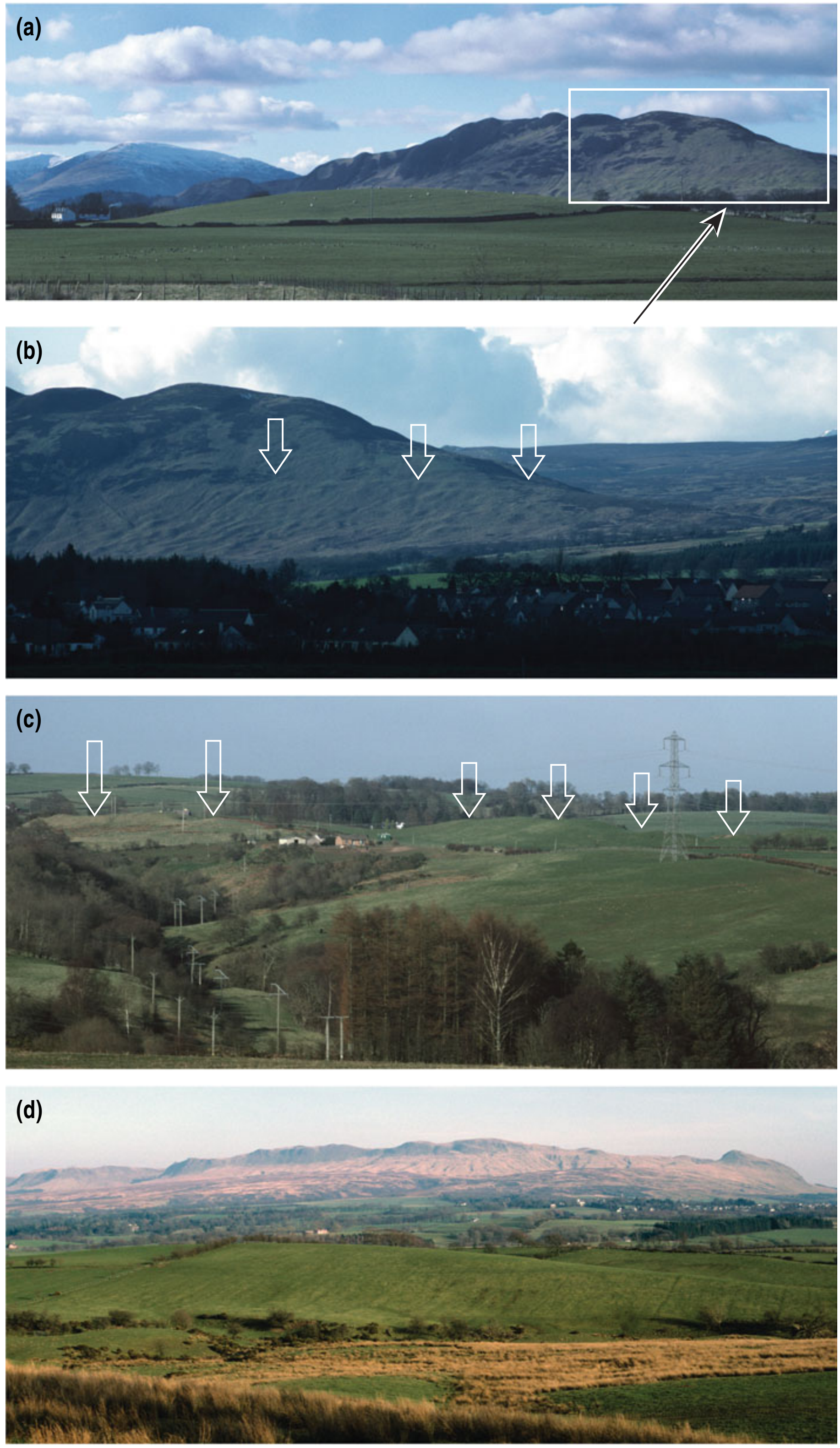
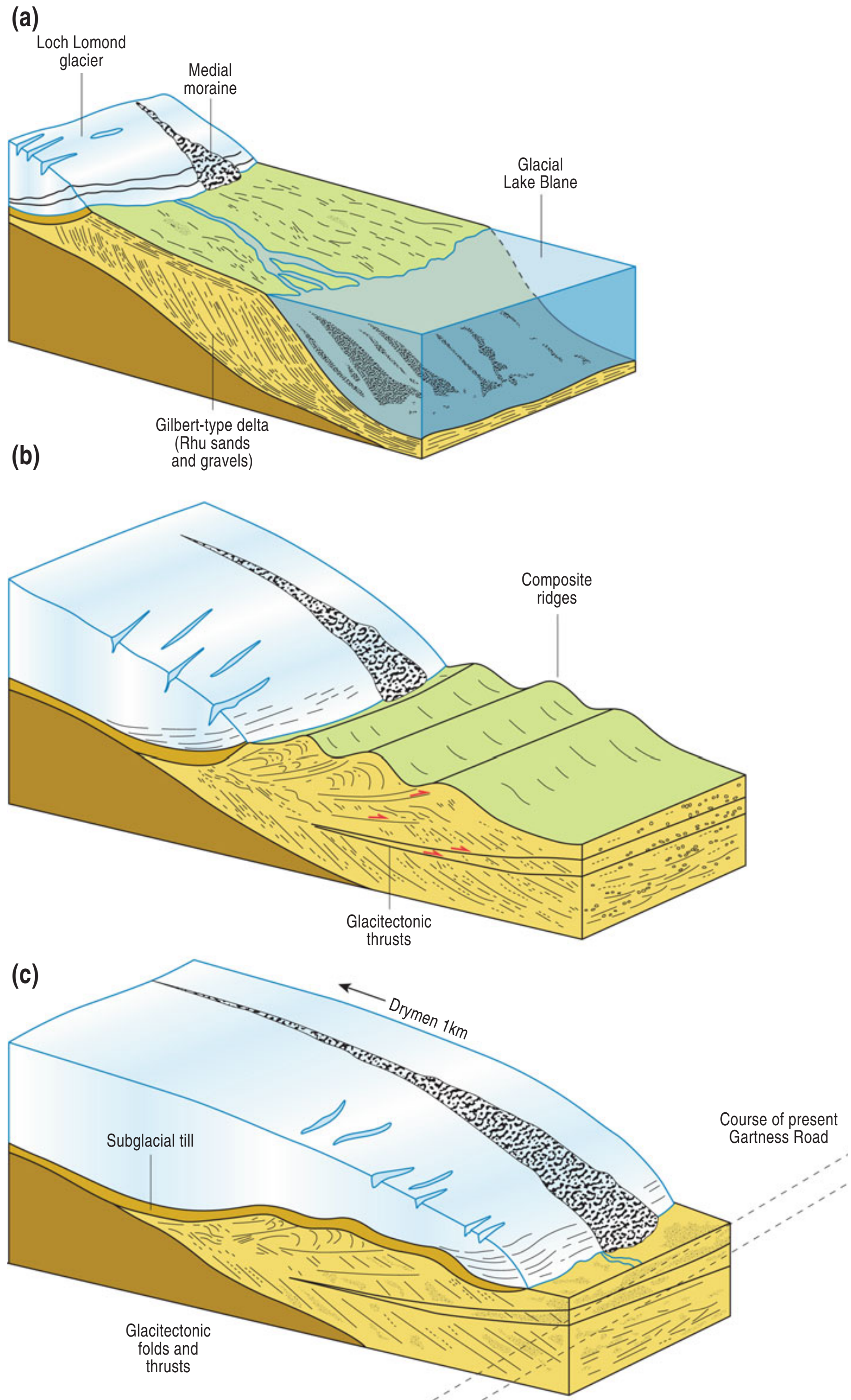


Fig. 24.8 Glacial landforms in the south Loch Lomond area. **a** Loch Lomond Stadial drumlin near Drymen looking across to the drift limit on Conic Hill. **b** The Conic Hill drift limit (arrowed). **c** The outermost Loch Lomond Stadial moraine (arrowed) near Gartness. **d** West–east aligned drumlin near Balfron produced by Dimlington Stadial ice-sheet flow, looking across to the Campsie Fells. (Images: David Evans)

cross profiles for the ridges, comprising a steep ice-proximal slope and an undulatory, gently dipping distal slope (Fig. 24.11b), typical of proglacially thrust masses in which individual folds or thrust slices diminish in size with distance through the proglacial stress field. The few exposures through these landforms indicate that they contain a complex

Fig. 24.9 Reconstruction of events relating to the deposition of the Glacial Lake Blane delta at Drumbeg near Drymen and its subsequent glacitectonic deformation by the Lomond glacier lobe. **a** Ice recession and delta deposition. **b** Glacier readvance and glacitectonic deformation. **c** Glacier overriding. (From Evans and Rose 2003a)



stratigraphy that includes shelly marine clay, till, sand and gravel (Smith 1993), supporting their interpretation as glacitectonically constructed moraines. Moreover, the straight margins of the Lake of Menteith are diagnostic of the strike-slip tear faults that mark the lateral boundaries of sediment masses displaced from the source depressions of hill-hole pairs.

Identifying a likely glacitectonic origin for moraine segment A, Thorp (1991) proposed that the Menteith glacier lobe had a very low-angled surface profile and flowed

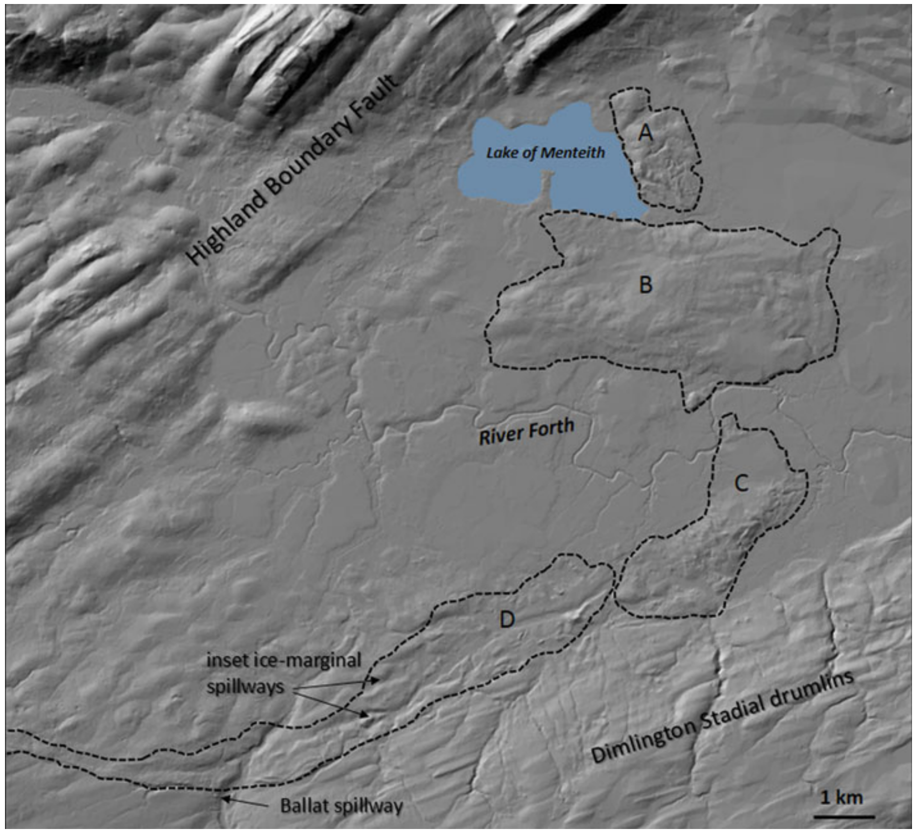


Fig. 24.10 Digital elevation model of the upper Forth valley and Lake of Menteith, showing the four segments (A–D) of the Menteith glacier lobe end moraine. (NEXTMap Britain data from Intermap Technologies Inc., provided courtesy of NERC via the NERC Earth Observation Centre)

rapidly over unconsolidated marine sediments with relatively high pore-water pressures. These marine sediments were deposited in the upper Forth valley when this area formed an estuary under the relatively higher sea levels of the Lateglacial Interstade (Sect. 24.4.1). He also suggested that the basal shear stress of the Menteith glacier lobe was abnormally low compared to most other outlets of the West Highland glacier complex and that consequently it may have experienced surging behaviour. This scenario was supported by Evans and Wilson (2006), who alluded to a surging glacial landsystem signature for the area (Evans and Rea 2003) based on the presence of a hill-hole pair, a composite moraine ridge and possible crevasse-fill landforms along the former glacier margin.

24.4.3 Other Loch Lomond Readvance Limits

A less extensive and less lobate outlet glacier emerged from the Loch Venachar trough into the upper Teith valley near Callander, where it deposited an arcuate end moraine (the Callander moraine) amongst older (ice-sheet deglacial) kame and kettle topography (Thompson 1972; Merritt and Laxton 1982; Merritt et al. 2003; Fig. 24.1). This location is a key site for dating the timing of both ice-sheet retreat (~15.5–15.0 ka) and readvance of ice during the Loch Lomond Stade. Cores obtained from an infilled kettle hole inside the Callander moraine record lacustrine deposits underlying a Loch Lomond Readvance till; those obtained from a similar site outside the moraine record inwashed minerogenic sediments overlying organic deposits of Lateglacial Interstadial age, thus demonstrating that the 'outside' site was not over-ridden by ice during the Loch Lomond Stade (Merritt et al. 1990, 2003; Lowe 1993a, b).

In addition to the glacial landform assemblages indicative of lowland piedmont lobes, the area also contains evidence of a small niche glacier in the Corrie of Balglass on the northern flanks of the Campsie Fells (Brown 2003; Figs. 24.1 and 24.4a). The limits of this glacier are marked by a terminal moraine, inset within which are up to six recessional moraines. Glaciological modelling indicates that the former glacier at this site could have existed under Loch Lomond Stadial climatic conditions, but was marginally viable if interpreted as a cirque-based ice mass fed by snowblow and avalanching rather than an outlet of a possible plateau icefield (Brown 2003; Carr and Coleman 2007).

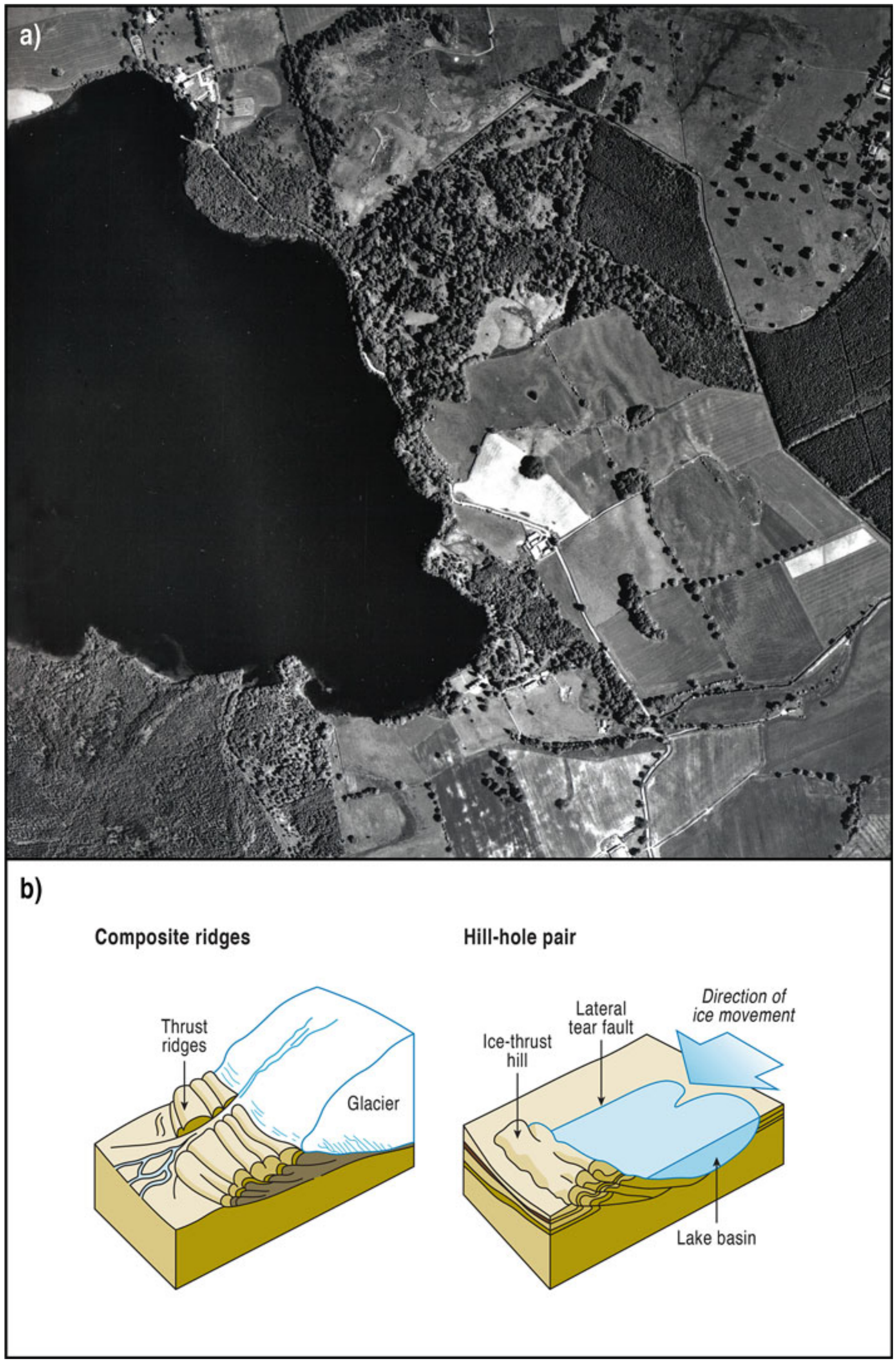


Fig. 24.11 Lake of Menteith hill-hole pair. **a** Aerial photograph extract, showing the surface morphology of moraine segment A, characterized by parallel, low-sinuosity ridges (Image 51288/128, All Scotland Survey 1988, Crown Copyright, HES). **b** (Conceptual diagrams of glacitectonic composite ridges and hill-hole pair. From Evans and Rose 2003a)

24.5 Raised Marine Landforms

Complex relative sea-level changes occurred in the Forth valley after the retreat of the last ice sheet, resulting in the development of superimposed erosional and depositional coastal landforms and sediments (Smith 1993; Peacock 2003; Shennan et al. 2018; Smith et al. 2019). From ~16.0 to ~14.5 ka, glacioisostatic rebound resulted in a drop in relative sea level in the region from a highstand of up to ~40 to ~12 m OD. Following a rise to ~19 m OD at ~13.5 ka, it fell to ~2.5 m OD at ~11.0 ka. A subsequent mid-Holocene transgression up to ~10 m OD was achieved by ~7.5 ka, after which continuing glacioisostatic uplift resulted in a gradual fall in sea level to its present position (Fig. 24.12a).

On the south shore of Loch Lomond, erosional marine landforms of probable Lateglacial Interstadial age are partly

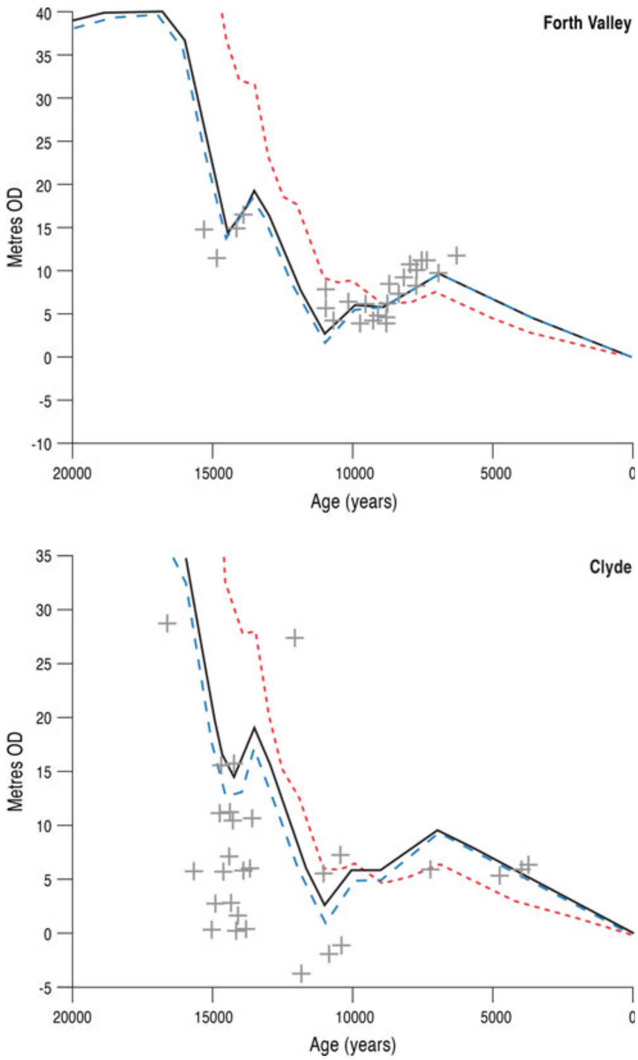


Fig. 24.12 Relative sea-level curves for the Forth valley and the Clyde estuary. Crosses are dated sea-level index points and the three separate curves are different reconstructions compiled from various sources. (Reprinted from Shennan et al. (2018) with permission from Elsevier)

buried by Loch Lomond Stadial glacial deposits. These erosional landforms include a shore platform that cuts across both bedrock and older Quaternary deposits, extends up to 500 m offshore, and rises inland from ~1 m below the present level of the loch to ~12 m OD, where it meets a backing cliff up to 15 m high (Rose 1980a, b, c, 1981, 2003a, b). This cliff is also exposed along the northern shores of the Clyde estuary, for example at Ardmore (Fig. 24.13) and forms part of the Main Lateglacial Shoreline (Chap. 4). Regionally this platform and cliff have been related to the Main Rock Platform of western Scotland (Sissons 1974; Gray 1978), which is thought to have been formed through frost-wedging of bedrock and removal of debris by waves and sea ice under the severe periglacial conditions of the Loch Lomond Stade. The later phase of marine transgression at ~7 ka (Fig. 24.12b) is recorded in south Loch Lomond by shorelines up to 12 m OD.

More extensive raised marine depositional features such as estuarine flats (carse) are well represented in the area, notably the Carse of Stirling (Hansom and Evans 2000) which extends over 50 km from Grangemouth to the Lake of Menteith, where raised estuarine deposits onlap the slopes of the glacitectonic hill-hole pair and penetrate inside the end moraine of the Menteith glacier lobe (Smith 1993; Fig. 24.14). At the time of their deposition (~11.0–7.5 ka), due to glacioisostatic depression, sea level in the region was at least 10–14 m higher than present (Fig. 24.12a) and the carse surface forms part of the Main Postglacial Shoreline that represents the Holocene marine limit in areas close to the centre of glacioisostatic uplift (Smith et al. 2019). This extensive surface is underlain by estuarine clays and silts that represent suspended sediment settling in a low-energy and sheltered depositional environment with limited wave activity, so that sediment suspended in the water column was carried up-estuary by the flood tide.

24.6 Landforms Related to Paraglacial Landscape Change

Significant paraglacial rock-slope failures (RSFs) characterize the mountain slopes of the Highland terrain west of Loch Lomond, where they are developed in Dalradian metamorphic rocks, and the margins of the basalt lava plateaux of the Kilpatrick Hills, Campsie Fells and Gargunnock Hills (Jarman 2003a). Glacial loading, erosion and unloading of terrain are generally acknowledged to be prime drivers of RSFs. These processes, together with the effects of thermomechanical and hydromechanical fracturing during deglaciation, result in rock-mass weakening through internal rock damage and alter the stress field under slopes. As a result, failure may occur shortly after deglaciation or lag deglaciation by centuries or millennia due to progressive shear of internal rock bridges and reduction of friction angles to residual levels. It is also possible that seismic activity, induced by fault reactivation due to glacioisostatic rebound, contributed to reduction in rock-mass strength (seismic fatigue) or the triggering of failure in some cases (Ballantyne et al. 2014; Cave and Ballantyne 2016; Chap. 14).

A particularly large cluster of ~100 RSFs occurs in the mountains between Loch Lomond and Loch Fyne. This cluster was explained by Jarman (2003a) as a result of the area having some of the most intensively dissected mountain terrain in Scotland. The RSFs in this area are dominated by arrested rockslides and rock-slope deformations rather than catastrophic failures. The most spectacular RSFs are those

Fig. 24.13 Main Lateglacial Shoreline cliff eroded into Devonian conglomerate bedrock at Ardmore on the Clyde estuary. (Image: James Rose)

Fig. 24.14 Carse of Stirling viewed from Stirling Castle. (Image: James Hansom)

on The Cobbler in the Arrochar Hills (Fig. 24.15a). This mountain has a distinctive and locally unusual summit morphology, comprising a small but precipitous east-facing corrie surrounded by typically alpine terrain comprising arêtes and three distinct peaks (Jarman 2004). The complex RSFs on this mountain include rock-slope deformations, arrested translational slides and collapse features, all of which have contributed to the unusual summit topography, giving credence to the notion of Turnbull and Davies (2006) that cirque forms may be created by non-glacial processes in certain circumstances. More typical of the RSFs in this part of the Highlands are the arrested rockslides on Tullich Hill (Fig. 24.15b), with upper-slope headscarps and rupture surfaces and slipped masses or debris lobes farther downslope (Jarman 2003b).

Around the margins of the Kilpatrick Hills and Campsie Fells a number of RSFs represent different styles of mass failure in the basalt lavas where they overlie weaker sandstones (Evans and Hansom 1998). For example, located on the north face of the Kilpatrick Hills, The Whangie is a spectacular 10 m high, 3 m wide slab of basalt separated from the plateau edge cliff by a narrow chasm (Fig. 24.15c); downslope of this slab are older blocks in various stages of collapse, indicating that the assemblage of features represents an arrested translational landslide or block glide. On the western cliffs of the Campsie Fells, above the village of Strathblane, collapse of stacked lava flows over the weaker underlying sedimentary rocks has exposed a steep headwall, with flow-like runout in the form of a lobe of hummocky debris at the slope foot (Fig. 24.15d). Similar rockwall

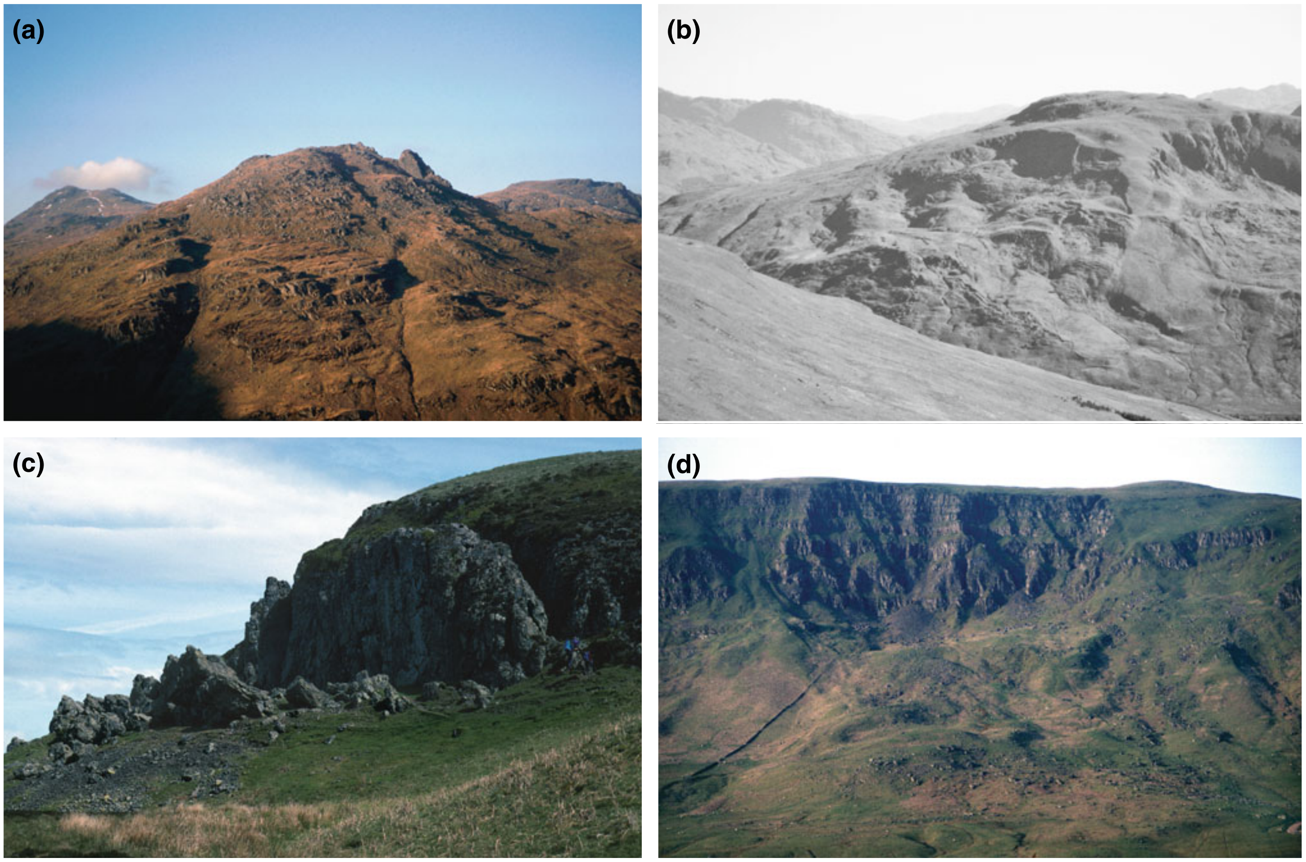


Fig. 24.15 Rock-slope failures. **a** The Cobbler, Arrochar. **b** Tullich Hill, Arrochar. **c** The Whangie, Kilpatrick Hills. **d** Strathblane rock-slope failure, Campsie Fells. (Images: **a**, **b** David Jarman; **c**, **d** David Evans)

collapse is evident at various sites on the north slopes of the Gargunnock Hills.

24.7 Conclusion

A wide range of landforms relating to a both long-term landscape evolution and the last glacial–deglacial cycle occurs in the area of Loch Lomond and in the western Forth and upper Teith valleys. Clear physiographic zones are strongly controlled by bedrock lithology and structure, with the Highland Boundary Fault separating the Dalradian metamorphic rocks of the Grampian Highlands terrane from the sedimentary rocks and Lower Carboniferous lava plateaux of the Midland Valley terrane. Glacially overdeepened troughs were created in the Highland terrane over the course of numerous Quaternary glaciations by outlet glaciers draining from the Highland ice dispersal centres, with some troughs being developed partly along pre-glacial fluvial valley networks and others breaching ancient watersheds. This overdeepened and oversteepened topography is a product of average Quaternary glacier coverage, namely repeated episodes of limited alpine-style glaciation, a style of ice cover that predominated for the longest cumulative time over the Quaternary Period.

The oldest glacial depositional landforms of the area are the overprinted drumlins of the Devensian ice-sheet glaciation, which record glacier flow down Loch Lomond and the upper Firth of Clyde and eastwards along the upper Forth and Teith valley networks to feed into the Forth Ice Stream. Superimposed on this older glacial depositional landsystem are the landforms and sediments of the Loch Lomond Stade. These provide evidence of the development of the large Lomond, Menteith and Teith piedmont glacier lobes, and are regarded as characteristic of the piedmont lobe landsystem signature for the Younger Dryas in Britain. Moreover, the excellent chronostratigraphic record associated with this geomorphology qualifies the area as the type locality for the Younger Dryas in Britain. Notable glacial landforms include the continuous end moraine belts, inside of which are widespread till hummocks, eskers and kamiform features and drumlins indicating former radial ice flow. Also, significant are the landforms and sediments that represent the former existence of Glacial Lake Blane, dammed by the

Lomond lobe in the Endrick and Blane valleys and spilling eastwards to the margin of the Menteith ice lobe. The distinctive landforms of the Menteith ice lobe, particularly the glacitectonically thrust moraines, suggest that it underwent surging during the Loch Lomond Stade. The glacioisostatically driven relative sea-level changes that took place during and after Late Devensian ice-sheet deglaciation are represented by raised marine landforms and deposits that include the Main Lateglacial Shoreline, mid-Holocene raised shorelines and the former estuarine flats of the Carse of Stirling, all of which lie within an altitudinal range of 10–15 m OD. The most prominent postglacial landforms are paraglacial rock-slope failures that occur in a dense cluster in the glacially dissected terrain west of Loch Lomond and at a number of locations around the margins of the basalt lava plateaux of the Kilpatrick Hills, Campsie Fells and Gargunnock Hills.

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David J. A. Evans is a Professor of Physical Geography at Durham University, England, specializing in glacial geomorphology. He specifically works on glacial landsystems (especially modern process-form models) and their application to reconstructing past glaciations (palaeoglaciology) and has undertaken research in a range of modern and ancient glacial environments, including Iceland, the Canadian High Arctic, Svalbard, Norway, the Canadian Prairies, New Zealand, South Georgia and the British Isles. He has authored and co-authored a number of books in this research arena, including *Glaciers and Glaciation* (1998, 2010), *Glacial Landsystems* (2003), *A Practical Guide to the Study of Glacial Sediments* (2004), *Vatnajökull National Park (South Region)—Guide to a Glacial Landscape Legacy* (2016) and *Till—A Glacial Process Sedimentology* (2018). He was awarded the Busk Medal by the Royal Geographical Society in 2017 in recognition of his contributions to glacial geomorphology and research-led teaching in glaciated environments.

25 Glacifluvial and Glacilacustrine Landforms of the Midland Valley

David J. A. Evans

Abstract

Widespread meltwater erosional landforms and accumulations of glacifluvial sands and gravels have long been recognised in the Midland Valley and the adjacent upland areas. Erosional channels that record the drainage of subglacial and marginal meltwater and relate to the regional drainage of receding ice lobes, together with former ice-dammed lake spillways, are now represented by dry valleys and channels occupied by underfit streams. Depositional assemblages record the spatial and temporal evolution of glacier meltwater routeways and associated glacier karst systems in the marginal zones of the last ice sheet as it retreated from the area. These are represented by the kamiform and esker landforms created in complex and interlinked supraglacial, englacial and subglacial drainage networks and ephemeral depocentres. The glacifluvial and glacilacustrine landforms of the region are illustrated by four landsystems characteristic of general styles of glacifluvial landform-sediment production: (1) ice-sheet downwasting and marginal uncoupling in lowland terrain, exemplified by the Carstairs Kames; (2) subglacial meltwater erosion and esker sedimentation within a former palaeo-ice stream, illustrated by the subglacial meltwater channels and Alyth-Forfar esker belt of the Strathmore area; (3) ice recession into upland topography and concomitant thermal regime change, demonstrated by the lateral, sub-marginal and subglacial meltwater channels, eskers and kames formed during glacier retreat in Strathallan; and (4) ice-dammed lake formation and spillway incision, illustrated by the Ochil Hills glacilacustrine landforms and watershed overspill channels.

Keywords

Meltwater channels • Eskers • Kames • Glacier karst • Kame and kettle topography • Ice-dammed lake • Spillways

25.1 Introduction

The Midland Valley lies between the Highland Boundary Fault to the north and Southern Upland Fault to the south. Most of the area consists of low ground below 200 m ordnance datum (OD) underlain by a range of Devonian and Carboniferous sedimentary rocks, mantled by tills of variable thickness (Kearsey et al. 2019) and interrupted by uplands of stacked lava flows and sills (Fig. 25.1; Chap. 2). The region contains outstanding examples of glacifluvial landforms that developed in a range of topographic settings as the last ice sheet retreated. These include both erosional features in the form of proglacial, ice-marginal, sub-marginal and subglacial meltwater channels and a wide range of depositional landforms composed of stratified glacifluvial sands and gravels. In some areas, active retreat of the ice margin resulted in deposition of proglacial outwash plains (sandar), but in others, downwastage and stagnation of the retreating ice margin formed systems of meltwater channels, glacier karst features and ice-dammed lakes.

Meltwater channels occur extensively in the Midland Valley and include both isolated examples as well as densely spaced networks. The latter include lateral (ice-marginal or sub-marginal) meltwater channel assemblages that record ice recession into, or along the margins of, upland areas. Notable concentrations occur, for example, in Strathallan northeast of Stirling (Evans et al. 2017) and along the south flank of the Gargunnock Hills (Sissons 1963). The density of such valley-side channel networks provides important evidence for a change in glacier thermal regimes during overall ice-sheet recession, particularly the development of

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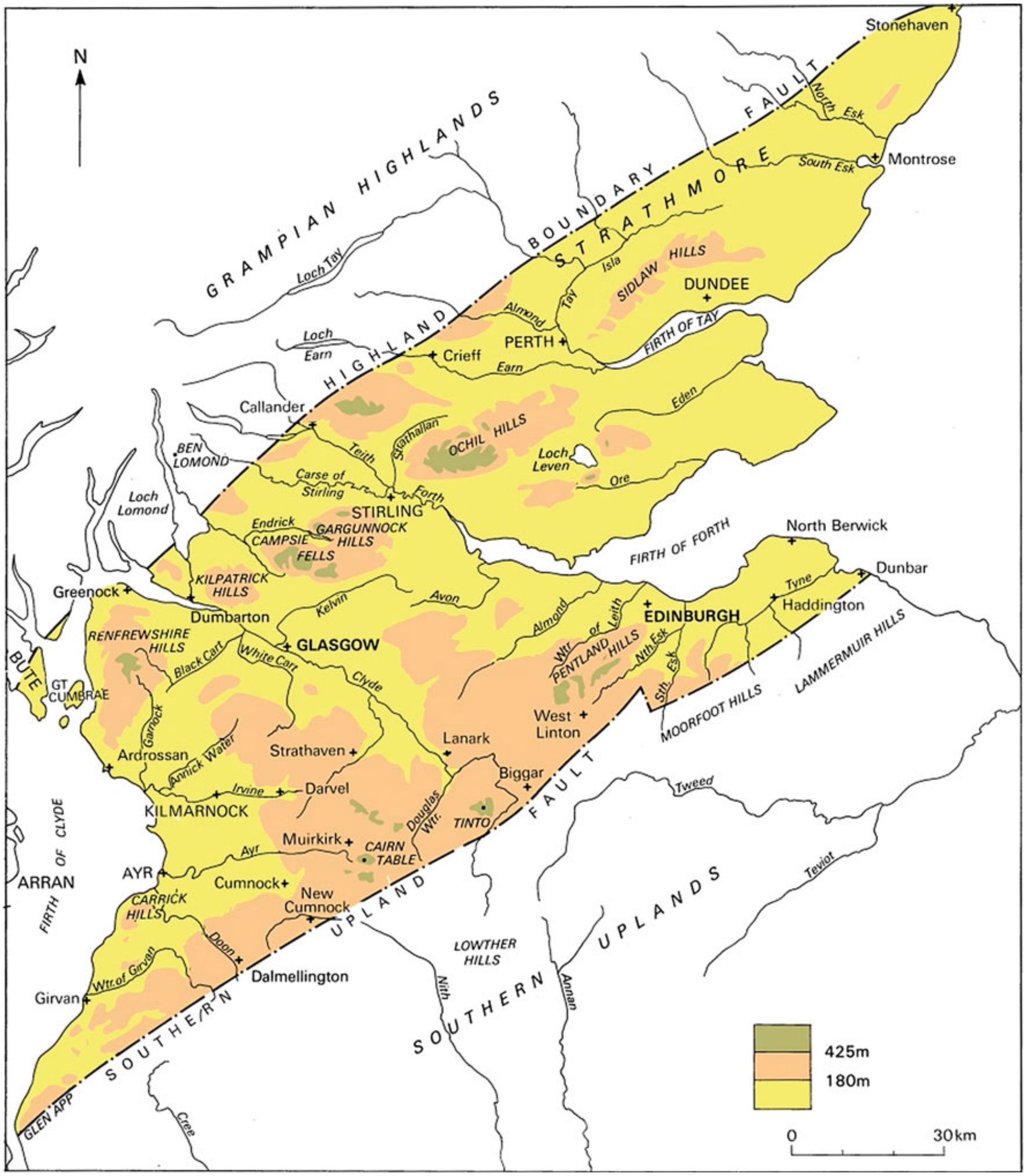

Fig. 25.1 The Midland Valley: relief and key locations cited in the text. (Reproduced with permission of the British Geological Survey. Permit Number CP20/087 British Geological Survey Materials © UKRI 2020. All rights reserved. Sourced from: https://earthwise.bgs.ac.uk/index.php/Physiography, Midland Valley of Scotland)

cold-based ice margins at the more advanced and topographically confined stages of ice-sheet recession (cf. Dyke 1993; Livingstone et al. 2010).

The spatial and temporal evolution of a glacier meltwater and associated glacier karst system in a retreating ice sheet results in recurrent glacifluvial reworking of sediment in complex and interlinked supraglacial, englacial and subglacial drainage networks and ephemeral depocentres (Fig. 25.2). Tunnel collapse and widening of ice-walled channels and their infilling with sand and gravel can lead to the burial of adjacent ice, thereby retarding ablation and isolating stagnant ice blocks that later melt to create water-filled kettle holes and irregular kame and kettle topography. Sediment-choked subglacial and englacial conduits may emerge through downwasting ice as eskers that wind through the more chaotic kamiform topography (Evans and Twigg 2002). Localised ponding can occur as ice-walled lakes that develop due to the collapsing glacier

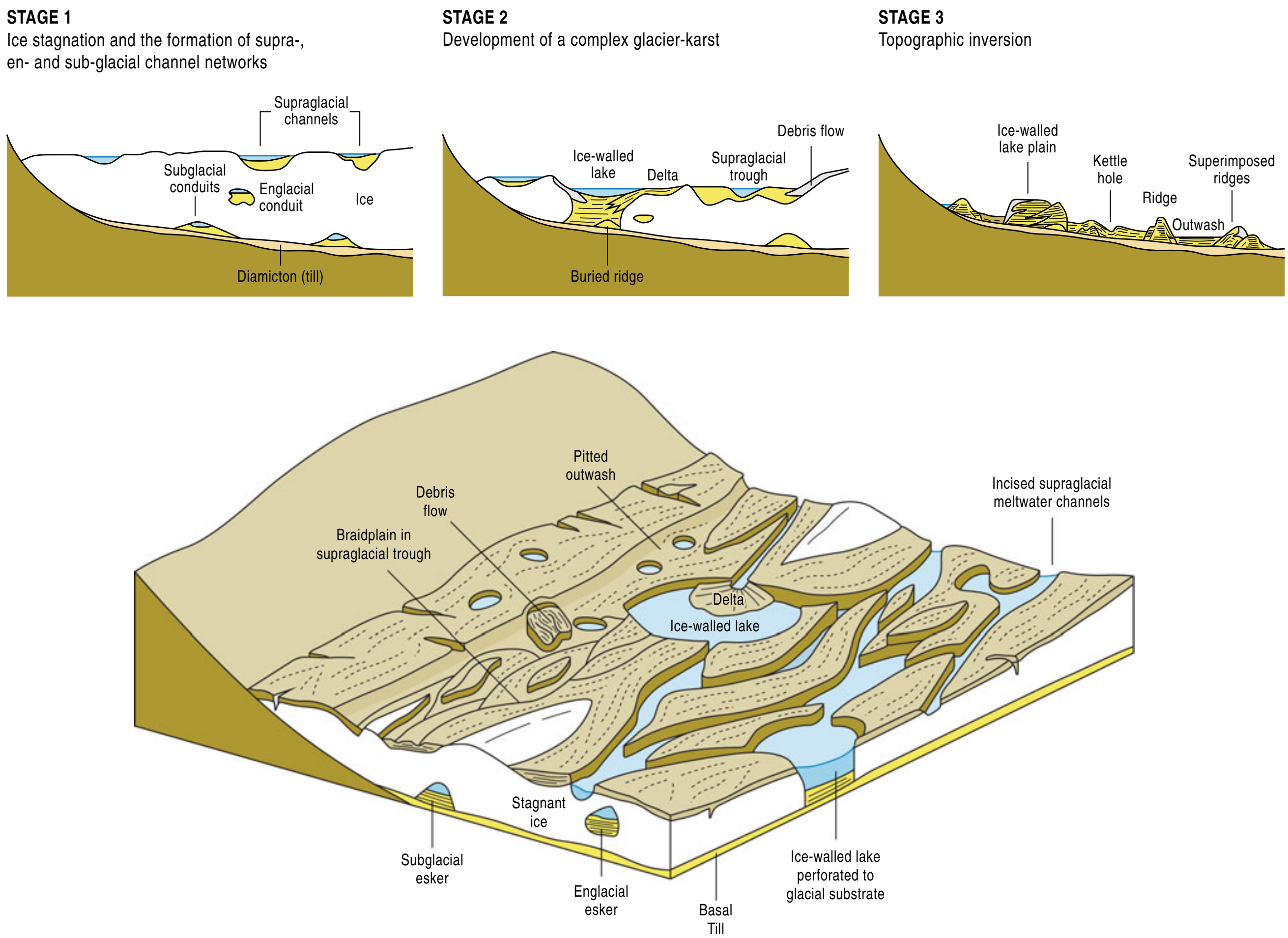

Fig. 25.2 Conceptual model of the development of glacifluvial landforms. (Modified after Livingstone et al. 2010)

karst, giving rise to ice-walled lake plains that also stand proud of the surrounding glacifluvial landforms after melt-out of the surrounding ice. At a downwasting ice margin, increasing volumes of supraglacial meltwater deliver sediment to form kame terraces, which may eventually undergo collapse and terrace fragmentation due to ice melt-out (Fig. 25.2).

Sources of information concerning the distribution of glacifluvial deposits in the Midland Valley include several British Geological Survey Mineral Assessment Reports, notably those by Goodlet (1970), Browne (1977a, b), Cameron et al. (1977), Paterson (1977) and Smith et al. (2008). The BRITICE 2 database and compilation map of Clark et al. (2018) also record a wide range of glacifluvial features in the region, including meltwater channels and glacifluvial landform-sediment assemblages. These sources suggest that four general glacifluvial landsystems can be recognised in the region; these are illustrated below with reference to particular classic locations. First, ice-sheet marginal uncoupling in lowland terrain is demonstrated by the landforms and sediments of the Clyde-Forth corridor. Second, the development of subglacial meltwater erosion and esker sedimentation in a probable ice stream suture zone is illustrated by the Strathmore area. Third, recession into upland topography and concomitant development of lateral, sub-marginal and subglacial meltwater channels and deposits due to thermal regime change is represented by the landform assemblages in Strathallan. Finally, ice-dammed lake formation and associated spillway incision are illustrated by the Ochil Hills glacilacustrine landforms and watershed overspill channels that developed as this upland area became exposed by downwastage of ice occupying the Forth valley.

25.2 Ice-Sheet Marginal Uncoupling in Lowland Terrain: The Carstairs Kames

The Carstairs Kames originally comprised a 7 km long series of anastomosing sub-parallel ridges of sand and gravel interspersed with kame and kettle topography (Fig. 25.3) but

Fig. 25.3 Aerial view of the Carstairs Kames, showing the multiple esker ridges. (BGS © UKRI, image P002940, reproduced with permission of the British Geological Survey and available at https://geoscenic.bgs.ac.uk)

have been extensively quarried for aggregate. Regionally, they form part of a WSW–ENE network of glacifluvial deposits recording downwasting and uncoupling of the Southern Upland and Highland ice masses over the Midland Valley (Fig. 25.4a, b). They have long been recognised as an important glacial depositional landform in Scotland and have been at the centre of debates on the precise origins of eskers and kames more generally. Together with the Polmont Kames, which mark a subglacial and englacial drainage corridor stretching from Falkirk to Linlithgow, the Carstairs Kames were widely considered in early research to represent ice-contact glacifluvial deposition (Gordon 1993). It now recognised, however, that the complex assemblage of features within the Carstairs Kames represents former sediment deposition in an expanding network of eskers at the suture zone or interlobate sediment sink between the margins of the retreating Highland and Southern Upland ice masses at the close of the last glaciation (Huddart and Bennett 1997; Thomas and Montague 1997). It is common to find major assemblages of glacifluvial landforms at the former suture zones between ice flow units, where meltwater drainage becomes naturally concentrated, and indeed some of the world's largest 'interlobate moraines' have a glacifluvial origin (e.g. Punkari 1997; Russell et al. 2003; Sharpe et al. 2007). Thomas and Montague (1997) used the evidence of substantial quantities of glacilacustrine deposits to propose a genetic model in which the esker network evolved into an ice-contact subaqueous depocentre on the floor of a proglacial lake (Fig. 25.4c). This was followed by continued ice-surface lowering and the emergence of the subglacial conduit to eventually form a supraglacial ice-walled channel and ultimately a supraglacial outwash fan and delta. Ice-walled channels filled with coarse gravel were topographically inverted after deglaciation to produce a series of sub-parallel (braided) and slightly sinuous ridges (cf. Evans and Twigg 2002; Storrar et al. 2015).

25.3 Subglacial Meltwater Landforms in an Ice Stream Suture Zone: Strathmore

The prominent subglacial streamlining associated with the Strathmore Ice Stream and its tributary ice flow units emerging from the Highland Boundary terrain to the north (Golledge and Stoker 2006; Chap. 26) is superimposed by an elongate WNW–ESE aligned and 3 km wide chain of hummocky, sand- and gravel-cored topography extending for ~32 km between Alyth and the head of Lunan Water, east of Forfar. This incorporates a series of elongate lochs (e.g. Forfar Loch and Rescobie Loch around Forfar) aligned along the floor of a sinuous drainage channel that extends from the hummocky topography around Forfar and then ESE across the watershed of the east Sidlaw Hills into the valley

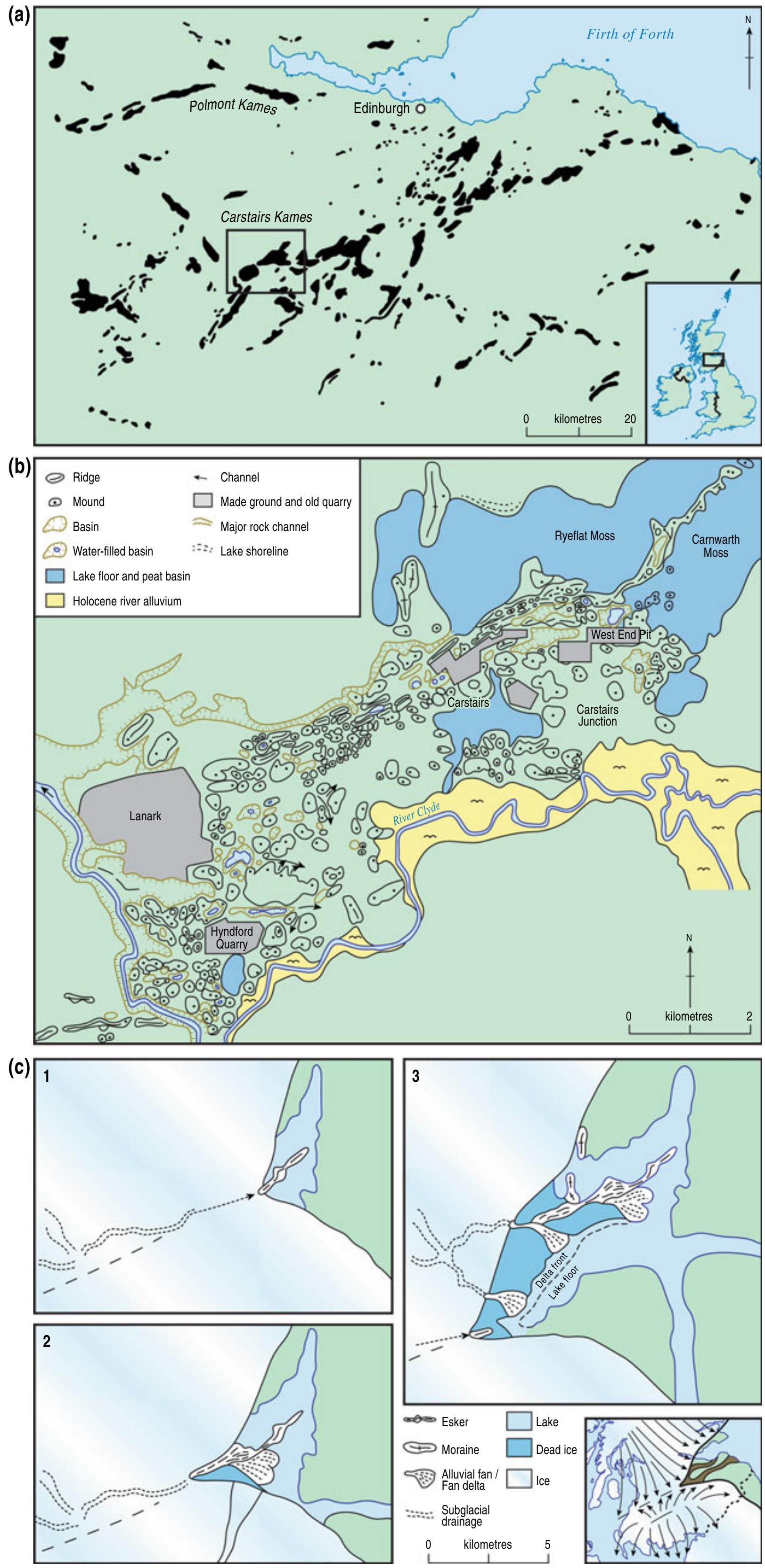

Fig. 25.4 Glacifluvial landforms of the southern Midland Valley. **a** The regional network of depositional glacifluvial landforms (black) that record the uncoupling of Southern Upland and Highland ice masses, including the Carstairs Kames. **b** The landforms of the Carstairs Kames. **c** Reconstruction of the events responsible for the construction of the Carstairs Kames, involving the deposition of eskers and deltas in a re-entrant created in the ice front during its downwasting. The regional map shows the interlobate nature of the network of glacifluvial landforms. (Reprinted from Thomas and Montague (1997); with permission from Elsevier)

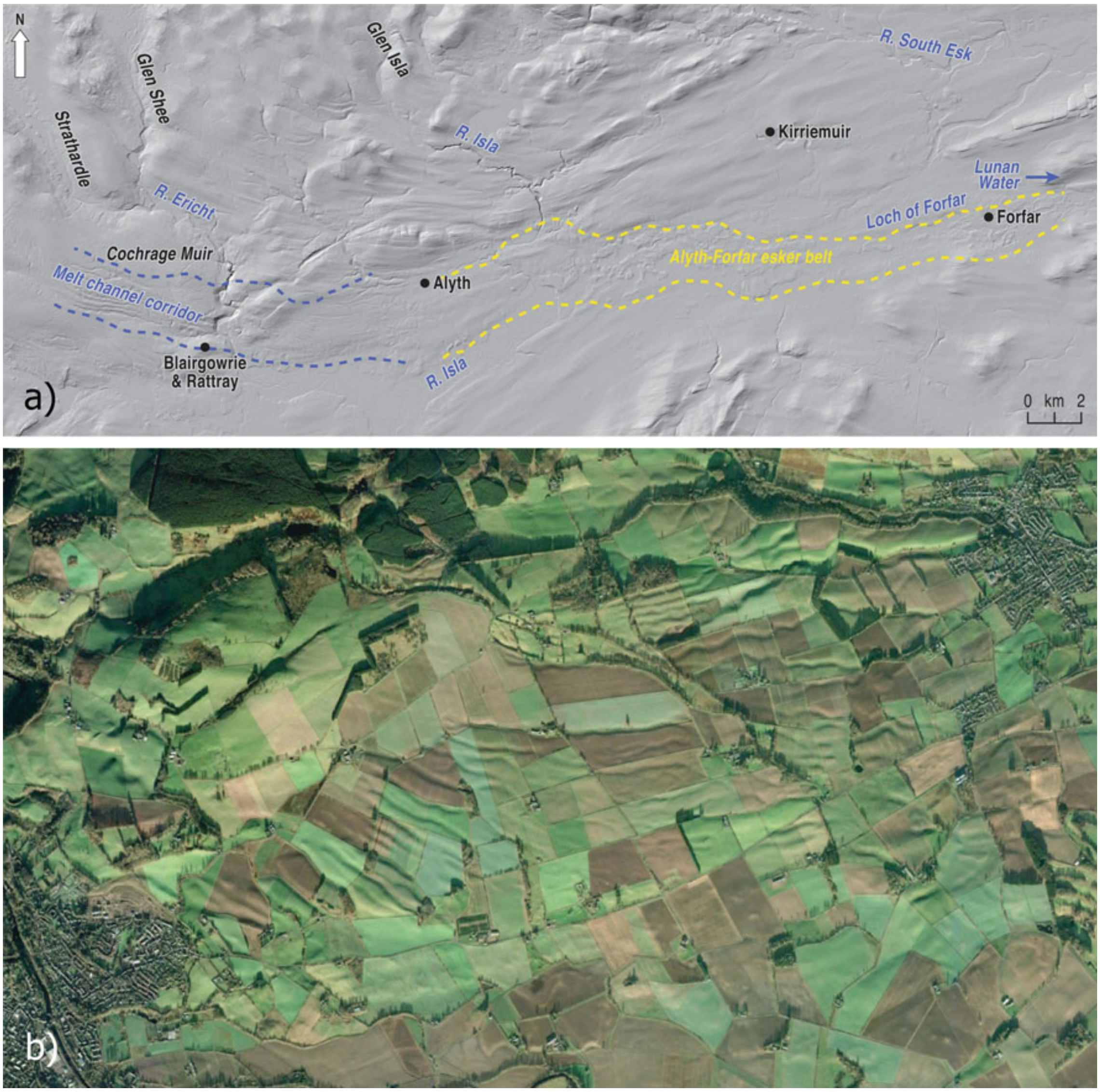

Fig. 25.5 Geomorphology of Strathmore. **a** Digital elevation model of the watersheds separating the Rivers Ericht, Isla and Lunan Water, showing the subglacially streamlined bed of the former Strathmore Ice Stream and its tributaries emerging from the Highland Boundary, superimposed by the chain of hummocky, sand-and-gravel-cored topography of the Alyth-Forfar esker belt, which extends westwards into closely spaced sub-parallel bedrock meltwater channels on Cochrage Muir. (NEXTMap Britain data from Intermap Technologies Inc, provided courtesy of NERC via the NERC Earth Observation Centre). **b** The area northeast of Blairgowrie and Rattray, showing the prominent meltwater channels on the higher ground immediately west of the Alyth-Forfar esker belt. The towns of Rattray and Alyth are visible at bottom left and top right, respectively, and the image is 7.10 km across, with north to the top. (Google Earth™ image)

that forms the head of the Lunan Water (Figs. 25.1 and 25.5a). Although the sand and gravel deposits in this belt were mapped by Paterson (1977) and are included in the BRITICE 2 compilation of Clark et al. (2018), no systematic research has been undertaken on the landform assemblage in this area. Digital elevation models (Fig. 25.5a) clearly indicate that the hummocky topography comprises an assemblage of discontinuous sinuous ridges, chaotic hummocks and depressions and numerous flat-topped mounds and ridges, an assemblage that is characteristic of features created by glacier karst (Fig. 25.2). Especially prominent in this assemblage are the flat-topped ridges, which almost certainly represent former subaerial ice-walled channels, created when sediment-filled subglacial channel and englacial tunnel networks expanded and underwent roof collapse. These features become wider and more prominent within the landform assemblage in an easterly (down-ice) direction, but they have been extensively quarried immediately east of Forfar. As this landform assemblage records the existence of a former ice-contact meltwater drainage pathway, it is hereby designated as the 'Alyth-Forfar esker belt' (Fig. 25.5a).

Extending a further 17 km westwards and northwestwards in a broad arc from Alyth to the slopes above Blairgowrie and Rattray and on to Cochrage Muir is a 3 km wide corridor of closely spaced, sub-parallel bedrock channels, whose floors display undulating profiles and which often cut across topographic high points; indeed the channels continue on either side of the River Ericht valley (Fig. 25.5). Such characteristics are typical of subglacially incised meltwater channels representative of a relatively stable drainage axis or pathway. Charlesworth (1956) interpreted some of these channels as representing the uncoupling zone between the westward-retreating Strathmore ice lobe and glaciers retreating northwards towards their sources in the SE Grampians. He also proposed the former existence of a large proglacial lake (Lake Isla), stretching from Forfar to Rattray, that developed in front of the receding Strathmore ice margin between the South Esk and Isla valleys. Evidence for such a lake is patchy, however, amounting to small areas of glacilacustrine deposits among the extensive chain of hummocky, sand-and-gravel-cored topography of the Alyth-Forfar esker belt. The limited distribution of lacustrine deposits suggests that any lake(s) that did develop in the area during deglaciation took the form of numerous sub-basins created by the partial flooding of a downwasting ice margin (cf. Bennett and Evans 2012).

The subglacial meltwater channel corridor and the Alyth-Forfar esker belt are both aligned sub-parallel to the streamlined subglacial bedforms of the former Strathmore Ice Stream and are interpreted as representing a relatively stable meltwater drainage pathway that was initiated during and immediately after ice stream cessation. The formation of this major drainage axis may also have effectively shut down the ice stream (cf. Lelandais et al. 2018). Although the subglacial channels could have been occupied by meltwater during ice stream operation, the depositional features of the Alyth-Forfar esker belt are superimposed on subglacial bedforms and therefore must have formed after ice streaming ceased. Meltwater flow along this system of channels was dictated by ice stream hydrology and cuts across the present drainage network, crossing the Ericht and Isla valleys and into the Lunan Water valley east of the Sidlaw Hills. The positioning of this 49 km long glacial meltwater corridor is significant in terms of drainage pathways through a former ice stream corridor and may represent the location of the former suture zone between ice flowing eastwards from the Tay valley and ice flowing SE from Strathardle, Glen Isla and Glen Shee to feed the Strathmore Ice Stream.

25.4 Lateral, Sub-marginal and Subglacial Meltwater Channels and Deposits: Thermal Regime Change in Strathallan

Persistent ice-marginal or sub-marginal meltwater drainage is a characteristic of cold-based glacier termini, because meltwater cannot penetrate to the glacier bed. In the landform record, this produces inset sequences of lateral meltwater channels incised into valley sides, from which the former positions of the receding glacier can be mapped (Dyke 1993). The lower ends of such channels often grade into kame terraces, engorged eskers or kame and kettle topography at locations where the erosional capacity of meltwater streams declines due to shallowing gradients. Glacier thermal regime change can be manifested in a spatial and temporal change from lateral, to sub-marginal to subglacial drainage. This culminates in the eventual flooding of the downwasting glacier snout due to higher ice-melt rates and results in the development of an increasingly complex and expanding glacier karst (Fig. 25.2).

An outstanding example of the spatial and temporal evolution of erosional meltwater channels and their association with glacifluvial depositional landforms is the glacial landsystem of Strathallan and the adjacent River Knaik valley between Dunblane and Auchterarder (Fig. 25.6). During deglaciation, the recession of the last ice sheet into the valleys along the Highland Boundary was characterised by the development of topographically confined glacier lobes (Clark et al. 2012). The Strathallan lobe retreated westward between the Ochil Hills to the south and the Uamh Bheag uplands to the north and was associated with eastward drainage of meltwater, contrary to the present southwestward flow of the Allan Water (Forsyth 1970).

A substantial assemblage of glacifluvial depositional landforms lies on the Strathearn–Strathallan watershed (Figs. 25.6 and 25.7a, b), the most renowned of which are the multiple esker ridges of the Gleneagles golf course near Auchterarder (Fig. 25.7b), used by Price (2002) to classify the course as an ‘esker and kame, parkland type’. The Gleneagles esker complex is part of a linear assemblage of esker ridges and elongate kame and kettle topography that extends eastwards for 13 km from Kinbuck to the mouth of Kincardine Glen (Fig. 25.6a), a remarkable glacial meltwater channel and spillway (Fig. 25.7a). This linear assemblage is almost certainly also related to the former position of the meltwater drainage axis responsible for the deposition of the eskers farther east and comprises a complex of esker

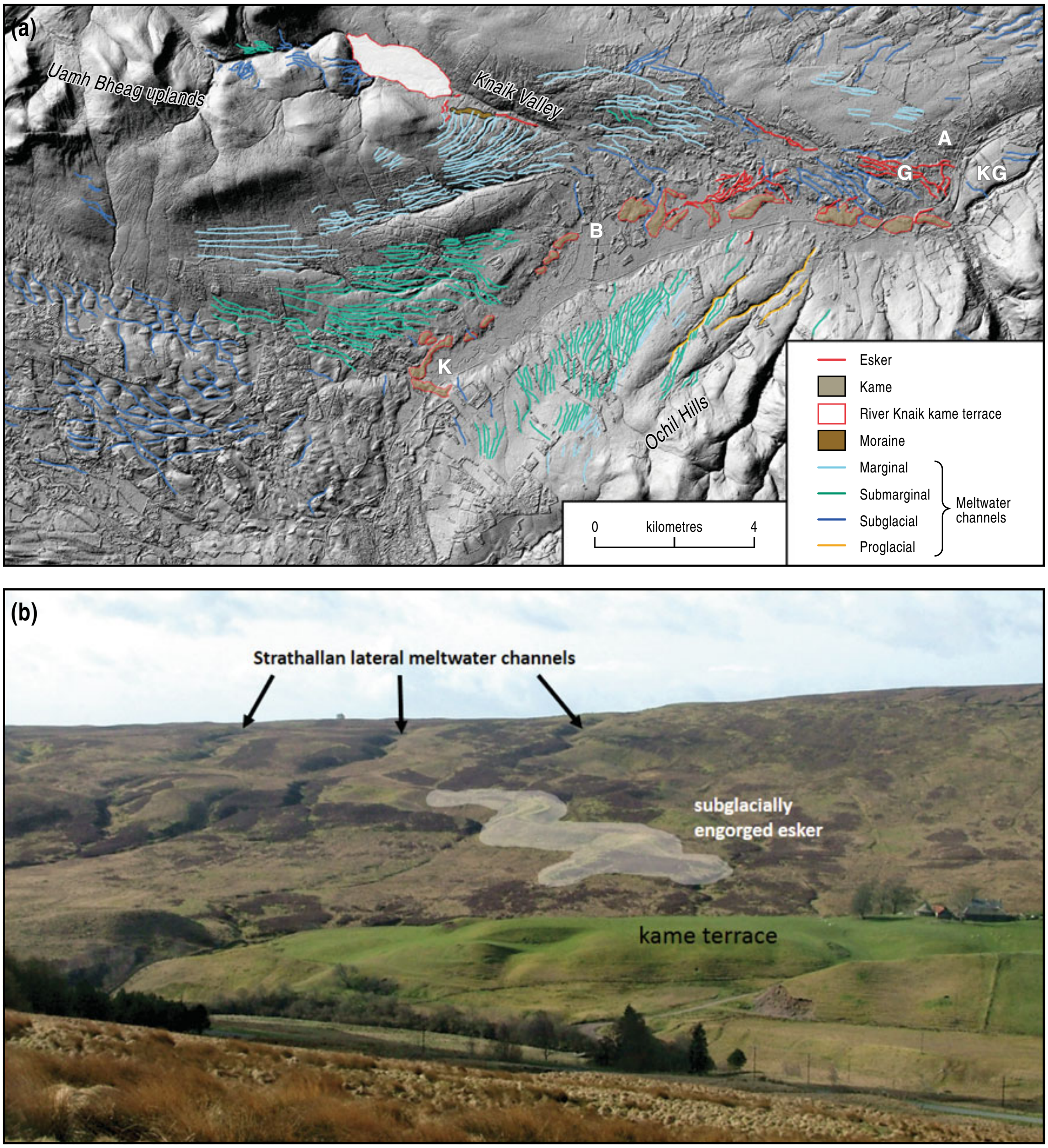

Fig. 25.6 Glacifluvial landforms of the River Knaik valley and Strathallan. **a** Digital elevation model showing the glacifluvial landforms in Strathallan and the valley of the River Knaik. (From Evans et al. (2017); reprinted by permission of the publisher, Taylor & Francis Ltd., http://.www.tandfonline.com). K: Kinbuck; B: Braco; G: Gleneagles; A: Auchterarder; KG: Kincardine Glen. **b** Ground view looking west of the River Knaik kame terrace, showing lateral meltwater channels on the north side of Strathallan, with a subglacially engorged esker emerging from the base of the highest meltwater channel. (Image: David Evans)

segments and kame mounds indicative of glacier karst development. The large volume of glacifluvial deposits in this area justified their inclusion in the British Geological Survey Mineral Assessment survey (Aitken and Shaw 1983). The assemblage of glacifluvial sands and gravels is attributed in this report to incision and removal of valley-side drift by densely spaced and extensive meltwater channels on the mid-slopes of the surrounding uplands. Also evident is the role of the Kincardine Glen meltwater channel in controlling the water levels in the downwasting ice on the Strathearn–Strathallan watershed, manifest in the eastward decline in the surface altitudes of incised flat-topped kames west of the glen. Drainage and sedimentation controlled by downwasting ice occupying the watershed are also evident in the flat tops of the kames, as these are indicative of subaerial sedimentation in ice-walled channels at altitudes higher than that

Fig. 25.7 Glacifluvial and glacilacustrine landforms of Strathallan, Gleneagles and Glen Devon. **a** Part of the Kincardine Glen meltwater gorge. **b** Esker ridges at Kirkton on the Gleneagles golf course. **c** The delta of Glacial Lake Devon at the junction of Glenquey and Glen Devon. Its upper surface records a lake level of ~269 m, and lower benches record later stages of lake-level lowering. (BGS © UKRI images P218075, P218067 and P002691, reproduced with permission of the British Geological Survey and available at https://geoscenic.bgs.ac.uk)

(126 m) of the present-day watershed. The final stages of westward ice recession out of Strathallan led to the damming of a proglacial lake on the valley floor, at around the time when the glacier margin terminated at a gravel ridge near Kinbuck, north of Dunblane (Aitken and Shaw 1983). This is recorded by 30 m thick deposits of glacilacustrine clays lying below modern floodplain alluvium to the south of Braco.

These prominent valley-floor depositional landforms lie below extensively incised hillslopes on the flanks of the Ochil Hills to the south and the Uamh Bheag uplands and Knaik valley to the north (Fig. 25.6). The incisions are series of inset, relict, meltwater channels that descend the slopes obliquely. Their inset, or *en echelon*, pattern indicates sequential cutting of the channels in a downslope direction, guided by the former downwasting and westwards-retreating valley glacier margin. The channels have been classified according to their genesis in relation to the changing nature of the meltwater pathways over time (Evans et al. 2017). This classification identifies a downslope change from inset sequences of marginal channels first to sub-marginal channels and then to subglacial channels. This progression reflects the changing nature of the glacial hydrology and glacier thermal regime of the Strathallan ice lobe once it had retreated to the Strathearn–Strathallan watershed. The occurrence of a dense network of marginal or lateral meltwater channels on the upper slopes of the uplands above the watershed indicates the establishment of a cold-based thermal regime, at least at the glacier margin, concomitant with the development of a topographically confined ice lobe. Such channels are genetically linked to cold-based, receding glacier margins because they are orientated almost contour-parallel and rarely coalesce (Dyke 1993). Inset downslope of the marginal channels is a series of sub-marginal channels, whose diagnostic characteristics are that they trend downslope obliquely, hence display steeper gradients and often end in subglacial chutes or contain chutes along their lengths, such chutes being indicative of marginal water plunging down sub-marginal tunnels. Farther downslope, and linking to the glacifluvial depositional landforms described above, is a series of subglacial channels, identifiable by their undulatory long profiles, created by confined water flow under high hydraulic pressure. These subglacial channels, together with the eskers and associated kame and kettle features, record the final stages of glacier snout downwasting, glacier karst enlargement and infilling of channels with sands and gravels.

In summary, the establishment of a valley-confined glacier lobe during overall ice-sheet recession was associated with initiation of lateral meltwater channels and hence a cold-based thermal regime. As downwasting progressed, the meltwater became focused first in sub-marginal and then subglacial channels indicative of a progressively more polythermal and then warm-based thermal regime. A similar spatial and temporal evolution of drainage styles has been documented in other parts of the British Isles in association with the advanced stages of ice-sheet recession (e.g. Livingstone et al. 2010; Evans et al. 2018a).

In addition to the belt of glacifluvial landforms on the Strathearn–Strathallan watershed, curvilinear chains of glaciofluvial mounds extend from the downslope ends of the sub-marginal meltwater channels and are thought to indicate deposition of sediment in ice-walled channels and tunnels that developed along the lower margins of the receding glacier snout. Hence, the expansion of these channels and tunnels and the progressive burial under sediment of the lower glacier snout on the watershed led to the widespread development of glacier karst involving deposition of a series of arcuate bands of kame and kettle topography. Such features are essentially subglacially engorged or valley eskers that have formed due to meltwater re-entering the glacier after having exited from it farther up-ice (Evans et al. 2018b).

25.5 Ice-Dammed Lakes and Spillways: Glen Devon and the Ochil Hills

Although many Scottish meltwater channels that were once attributed to drainage of glacial lakes dammed along the margins of the retreating ice sheet have been shown to have an ice-marginal, sub-marginal or subglacial origin (Evans 2004), some landform-sediment assemblages in the Midland Valley are clearly the product of damming and reorganisation of drainage by ice blockage. An excellent case study is provided by the landform assemblage in Glen Devon and the adjacent Ochil Hills, where a range of meltwater channel types has been explained as the erosional products of ice-marginal, sub-marginal and subglacial drainage related to downwasting of the Forth valley ice lobe against the southern slopes of the Ochil Hills (Russell 1995). Later stages of ice downwasting in this area resulted in the accumulation of glacilacustrine deposits contained within terrace and delta landforms that record the former existence of an ice-dammed lake, Glacial Lake Devon (Fig. 25.7c). This lake was dammed by the Forth glacier and drained through a channel in the 269 m col through Glen Eagles into the Strathearn drainage basin, probably joining meltwater sourced from the downwasting Strathallan ice lobe and possibly contributing to the excavation of the Kincardine Glen meltwater channel (Francis et al. 1970; Russell 1995). Continued downwasting and recession of the Forth glacier exposed a col at 233 m, thereby allowing lake drainage through Glen Queich and a fall in lake level to that altitude (Fig. 25.8). In Glen Devon, the collective evidence provided by exposures of lacustrine deposits in terraces and deltas and the corresponding altitudes of the spillway channels provides conclusive evidence for the former existence of a glacial lake. Extensive lacustrine deposits in the Clyde and Irvine valleys of the western Midland Valley also indicate ponding of substantial lakes against the margin of the last ice sheet as it retreated (Finlayson et al. 2010).

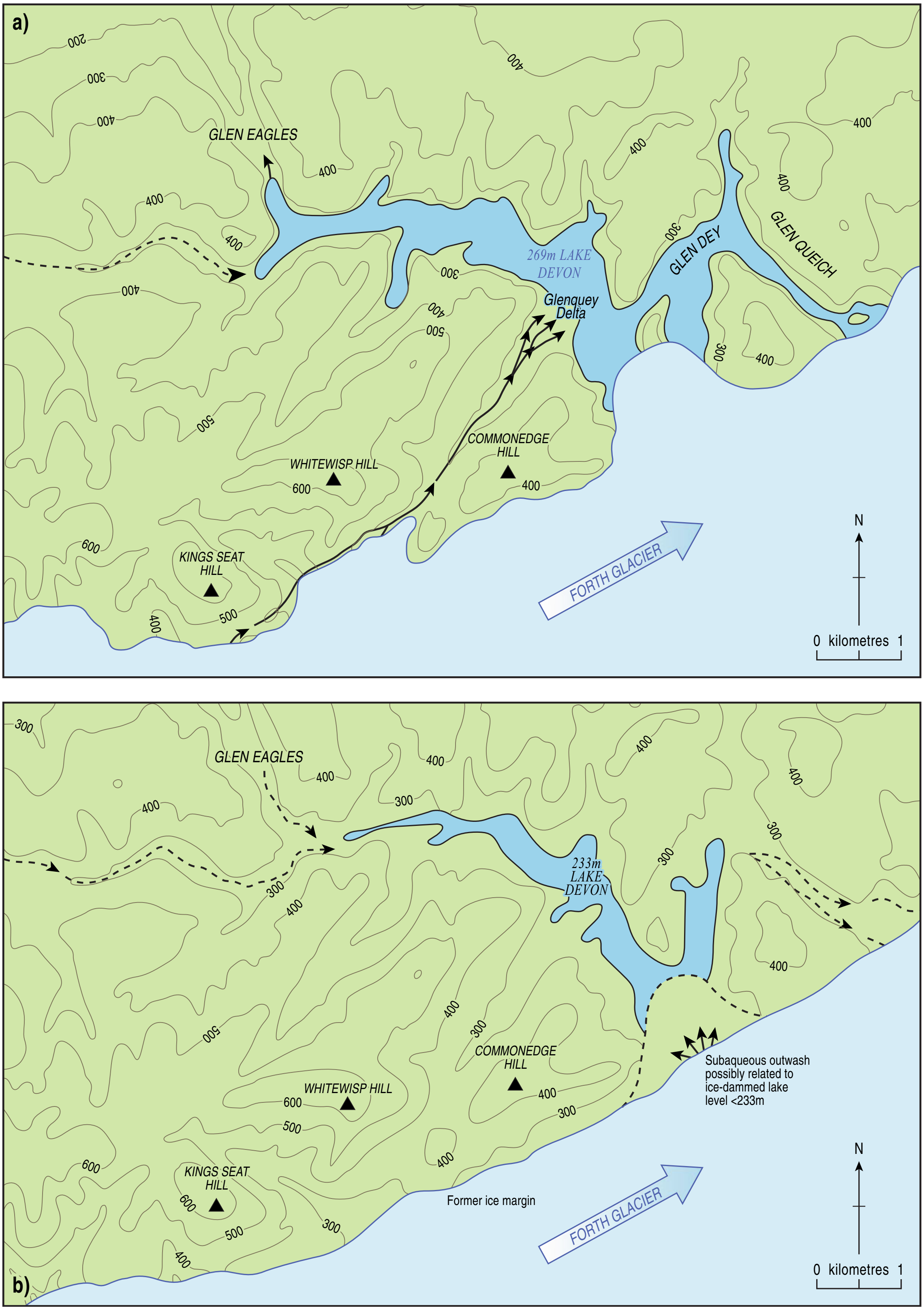

Fig. 25.8 Reconstructions of the stages of downwasting of the Forth Ice Stream that controlled the formation of Glacial Lake Devon and the Glen Eagles spillway. **a** The 269 m lake level, controlled by the Glen Eagles spillway and associated with the deposition of the Glenquey delta fed by ice-marginal meltwater. **b** The 233 m lake level, controlled by drainage across the col between Glen Dey and Glen Queich. (Modified after Russell 1995)

25.6 Conclusions

Four general styles of glacifluvial landsystem are identified in the Midland Valley. The first represents ice-sheet downwasting and marginal uncoupling in lowland terrain and is illustrated by the regionally extensive network of glacifluvial deposits that forms a WSW–ENE aligned corridor across the Midland Valley. It records the separation of the Southern Upland and Highland ice masses and is exemplified by the Carstairs Kames, a network of eskers that evolved into an ice-contact subaqueous depocentre on the floor of a proglacial lake and eventually into a supraglacial ice-walled channel and fan-delta complex. The second is representative of the development of subglacial meltwater erosion and esker sedimentation within a retreating and downwasting ice lobe and is illustrated by the Strathmore area, where the subglacial bedforms of the Strathmore glacier and its tributaries are overlain by the Alyth-Forfar esker belt, which is succeeded westwards by a corridor of multiple, parallel-aligned subglacial meltwater channels. This landform assemblage is interpreted as a relatively stable meltwater drainage pathway that was initiated after ice stream shutdown and which crossed local drainage divides; it likely represents the location of the former suture zone between ice moving eastwards along the Tay valley and ice flowing southeastwards from glens draining the SE Grampians. The third is representative of ice recession into upland topography and the sequential development of lateral, sub-marginal and subglacial meltwater channels and deposits due to thermal regime change. This assemblage is illustrated by the landforms of Strathallan, where the westerly recession of the lateral margins of the topographically confined Strathallan glacier lobe is demarcated by inset meltwater channels on the surrounding hillslopes. These features display a gradual change downslope from lateral channels to sub-marginal and eventually subglacial channels, recording a switch from supraglacial to increasingly englacial and subglacial drainage and the concomitant development of subglacially engorged eskers, kames, kame terraces and englacial or subglacial eskers in a landform assemblage characteristic of an expanding glacier karst. The incision of Kincardine Glen during the downwasting of the glacier lobe over the Strathearn–Strathallan watershed is a classic example of glacially focused drainage and the production of underfit streams, of which numerous examples occur throughout the Midland Valley. The fourth type is representative of ice-dammed lake formation and spillway incision and is illustrated by the Ochil Hills glacilacustrine landforms and the Glen Eagles watershed overspill channel, features relating to the temporary damming of upland drainage as the area became exposed by downwastage of ice in the adjacent Forth valley.

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The Midland Valley: Ice-Moulded Lowlands

26

David J. A. Evans

Abstract

The ice moulding of the Midland Valley is a product of ice streaming in the last and earlier ice sheets. The landform evidence of this glacier streamlining is a prominent aspect of urban topography, especially evident in Glasgow's drumlins and Edinburgh's crag-and-tail landforms, but is prevalent throughout much of the region in the form of a range of subglacial landforms that also includes drumlinoid drift tails, flutings, ribbed moraine, whalebacks and roches moutonnées. Sequential flowsets derived from the alignment of ribbed moraine and superimposed drumlins record dynamic glacier-flow switches during the last ice-sheet glaciation. Areas comprising both erosional and depositional streamlined forms, typified by the juxtaposition of erosional and depositional crag-and-tail features, are a product of the patchy emplacement of a subglacial deforming layer (till) and represent an important landsystem signature of a mixed (soft and hard) subglacial bed mosaic, which is characteristic of the boundary zone between upland and lowland glaciation.

Keywords

Subglacial bedforms • Drumlins • Ribbed moraine • Crag-and-tail features • Palaeo-ice stream • Flowsets

26.1 Introduction

The Midland Valley or Central Lowlands (Fig. 25.1) is an extensive area of glacier streamlining that reflects ice streaming in the British–Irish Ice Sheet, initiated by vigorous ice flow towards the Irish Sea Basin in the west and the Firth of Forth-northern North Sea Basin in the east (Sissons 1967; Burke 1969; Golledge and Stoker 2006; Clark et al. 2012; Livingstone et al. 2012; Hughes et al. 2010, 2014; Ballantyne and Small 2019). Evidence of streamlining is most prominent in the sediment-cored subglacial bedforms (drumlins) in the lowest terrain of the Clyde–Forth corridor and the bedrock erosional landforms (whalebacks and roches moutonnées) and crag-and-tail forms over large areas of upland topography such as the Ochil Hills, Sidlaw Hills and Pentland Hills and the numerous volcanic intrusions throughout the region (Haynes 1983; Figs. 26.1 and 26.2). Ice-flow directional changes within the last ice sheet are recorded by overprinted subglacial landforms (Hughes et al. 2010, 2014) and ribbed moraine (subglacial landforms with arcuate ridges aligned transverse to ice flow) created by the more extreme shifts in ice-flow vectors (Clark and Meehan 2001).

These subglacial landforms have been central to conceptual developments in glacial research, a prime example being the superimposed drumlins of the Clyde basin, which were the first to be recognised as unequivocal evidence of dynamic ice-flow switches during a single episode of ice-sheet glaciation (Rose and Letzer 1977). The predominance of till in the sediment-cored drumlins of the region is typical for such subglacial bedforms (Stokes et al. 2011), a characteristic that has been used to champion a subglacial deforming layer origin for these landforms (Boulton 1987; Dunlop et al. 2008; Stokes et al. 2013). The development of streamlined bedrock and crag-and-tail forms, and even non-till-cored drumlins, is explained in such deforming bed theories as the emergence of forms in settings where the deforming layer thins or becomes depleted (Clark 2010); this entails a more erosive set of processes, either those of abrasion and plucking (Sugden et al. 1992; Hallet 1996) and/or the operation of an erodent layer or excavational deformation (Eyles et al. 2016). Hence the ice-moulded lowlands of the Midland Valley, especially the urban

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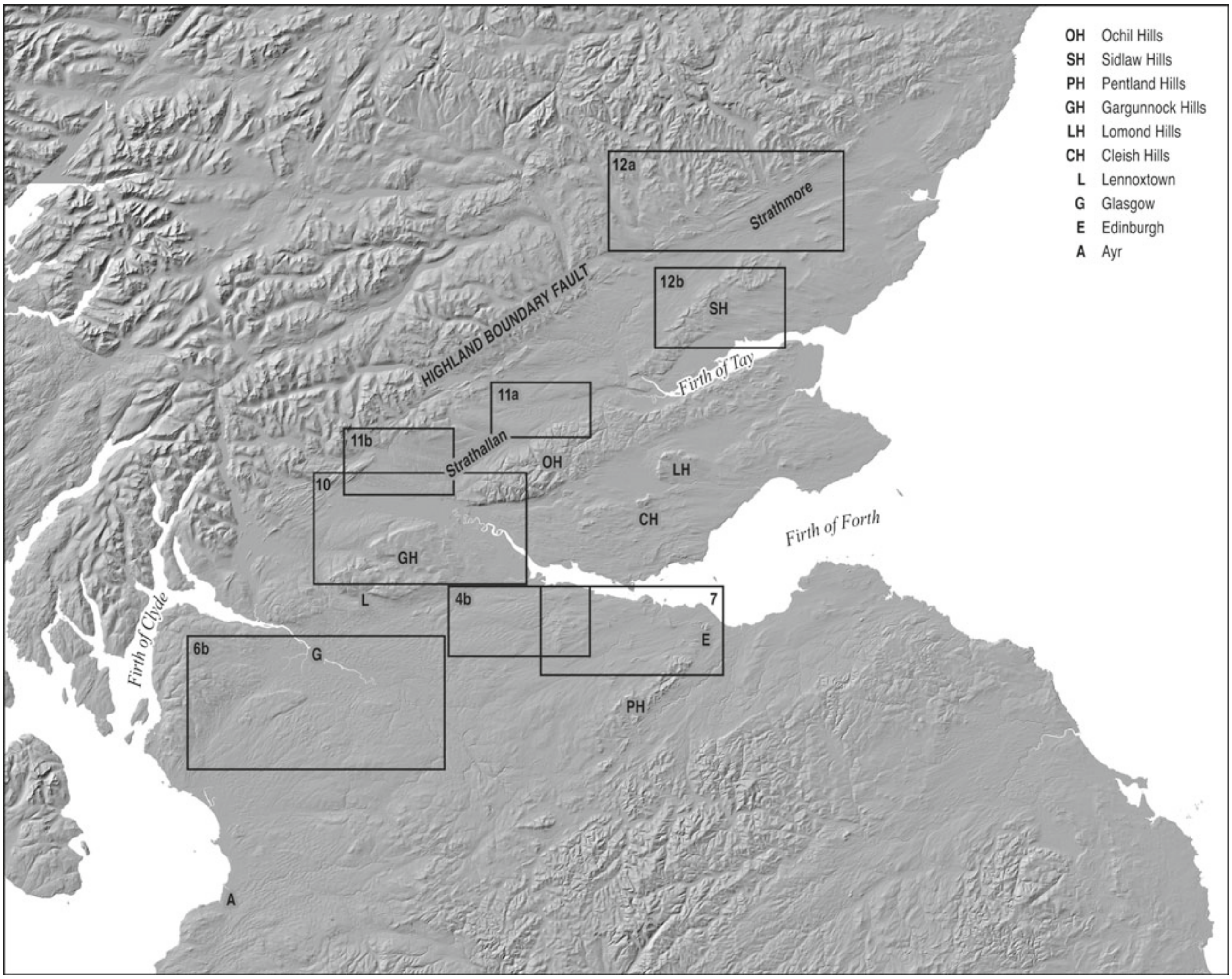

Fig. 26.1 Digital elevation model of the Midland Valley and adjacent areas of the Southern Uplands and Scottish Highlands. Smoothed terrain indicates areas of subglacial streamlining. Boxed areas outline the coverage of extracts used as case studies in this chapter. (NEXTMap Britain data from Intermap Technologies Inc, provided courtesy of NERC via the NERC Earth Observation Centre)

landscapes, display a range of subglacial bedforms representative of the former operation of glacial streamlining on a mosaic of soft and hard beds that reflects the availability and continuity of deformable substrate and till (Stokes et al. 2013; Eyles and Doughty 2016; Fig. 26.3).

The generalised flow directions within the last ice sheet over the Midland Valley and adjacent areas were initially reconstructed from the alignment of glacially streamlined hills or 'drums' and the distribution pathways of indicator erratics. Vigorous easterly flow through the Midland Valley is recorded by the prominent essexite erratic train, which has been dispersed from its source outcrop in Lennoxtown towards the Firth of Forth (Shakesby 1978). Complex changes in ice-flow directions in the last ice sheet over time are evident in this region in the form of overprinted subglacial bedform flowsets (Finlayson et al. 2010, 2014; Hughes et al. 2014; Fig. 26.4), of which there are three types: topographically unconstrained, topographically constrained and deglacial, the last relating to recessional lobate ice margins marked by end moraines. The most prominent unconstrained flowsets in the Midland Valley demarcate the Firth of Forth and Strathmore palaeo-ice streams (Fig. 26.4c).

26.2 Drumlins of the Glasgow Urban Area and the Clyde and Ayrshire Basins

Glasgow is renowned for its steep streets, a consequence of its development over drumlins up to 40 m high, with district names such as Hillhead, Jordanhill, Maryhill, Dowanhill and Firhill reflecting the prominence of steep-sided drumlins and thick, streamlined till sheets (Menzies 1996) interspersed with crag-and-tail forms such as Necropolis Hill near Glasgow Cathedral (Hall et al. 1998). The most accurate

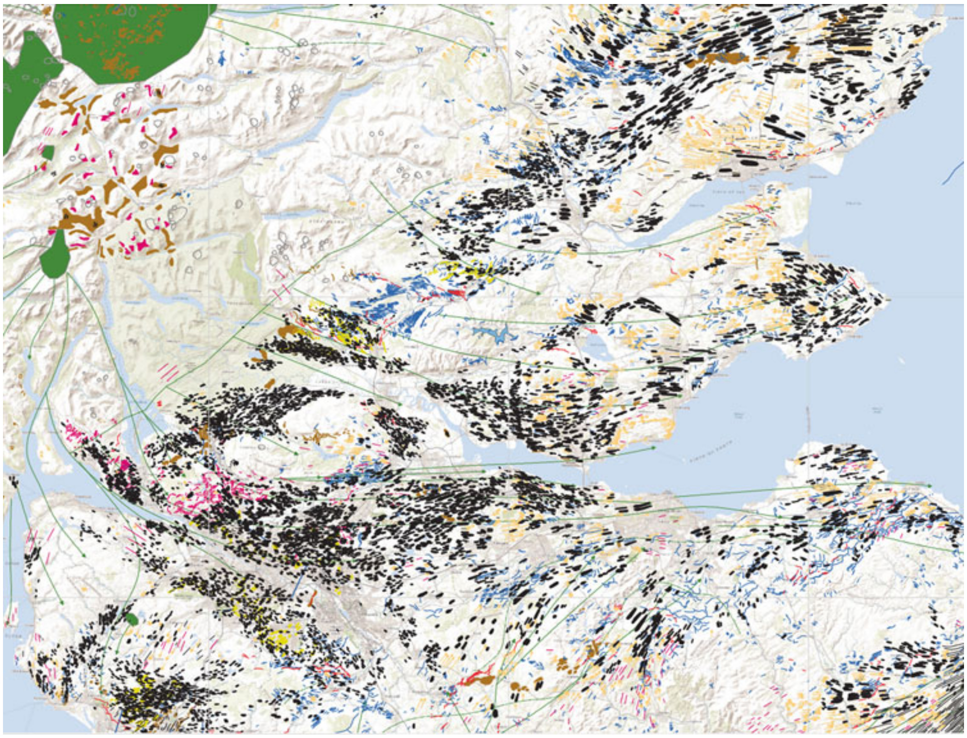

Fig. 26.2 Glacial landforms of the Midland Valley and adjacent areas as compiled in the BRITICE 2 database of Clark et al. (2018). Streamlined subglacial features are represented as follows: drumlins (black polygons); flutings and other lineations (black lines); crag-and-tail features (orange polygons); linear crag-and-tail features (orange lines); ribbed moraine (yellow polygons); linear ribs (yellow lines); glacially streamlined bedrock (dark pink polygons and lines). Moraines are demarcated as brown, meltwater channels as blue lines and erratic source areas and generalised ice-flow pathways as green.

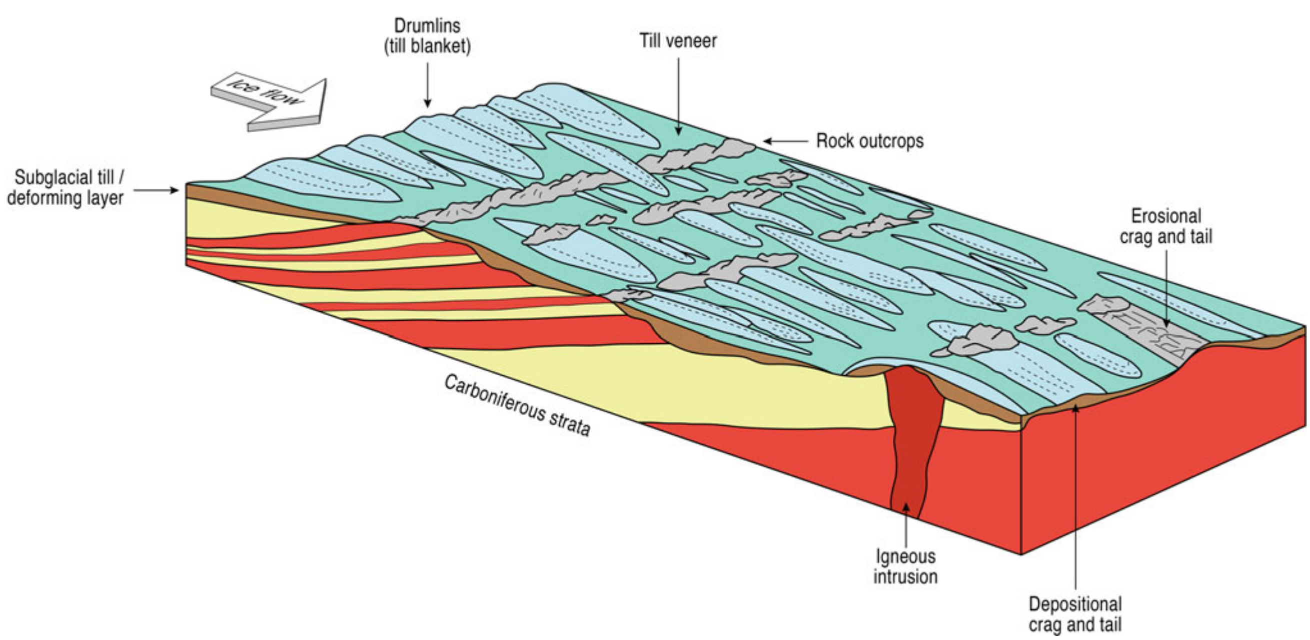

Fig. 26.3 Conceptual diagram showing the spatial distribution of hard- and soft-bedded elements within a mixed, streamlined subglacial bed. (After Eyles and Doughty 2016)

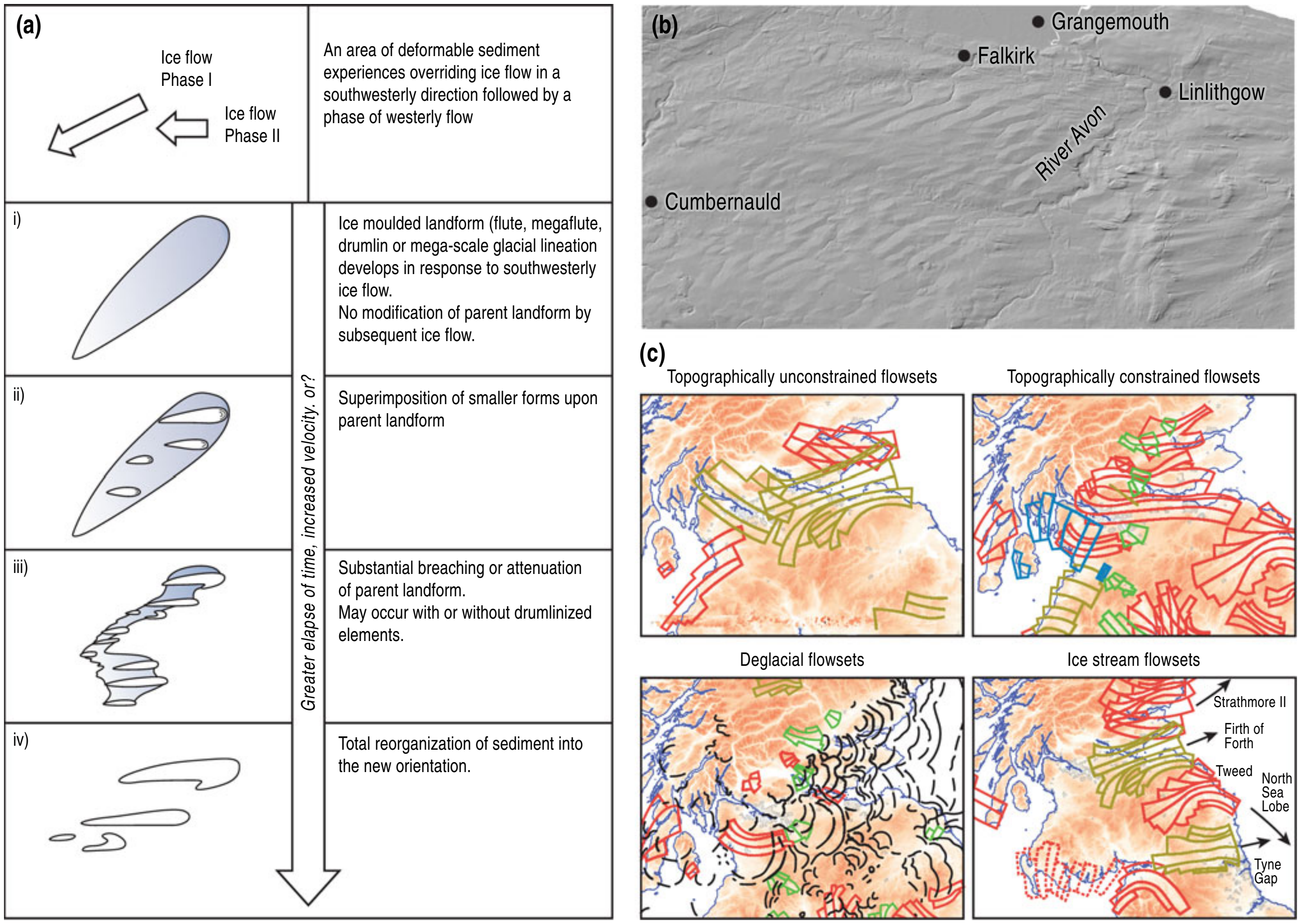

Fig. 26.4 Superimposition and moulding of subglacial bedforms. **a** Diagram showing the impacts of changing ice-flow directions on the development of streamlined overprinted bedforms (from Clark (1994); reprinted with permission from Elsevier). **b** Digital elevation model of the area south of the inner Firth of Forth, showing cross-cutting subglacial bedforms: NE-trending drumlins are overprinted by easterly-trending drumlins. (NEXTMap Britain data from Intermap Technologies Inc, provided courtesy of NERC via the NERC Earth Observation Centre). **c** Flowset types and their relationship to recessional ice margins and palaeo-ice streaming. (Modified from Hughes et al. 2014)

mapping of drumlins in the Glasgow area is that of Rose and Letzer (1977; Rose and Smith 2008; Fig. 26.5). They identified that the drumlins display complex morphologies: larger drumlin forms (megadrumlins) are superimposed by smaller drumlins with offset crest-line orientations, thus recording changing former glacier-flow directions. They also calculated mean areas of 0.147–0.337 km^2 for megadrumlins and 0.054–0.122 km^2 for superimposed drumlins; mean lengths and widths for megadrumlins are 0.671–1.193 km and 0.286–0.408 km, respectively, in contrast to 0.383–0.695 km and 0.156–0.217 km for superimposed drumlins. They found, however, that mean elongation ratios (and hence the degree of streamlining) are similar: 2.14–3.59 for megadrumlins and 2.50–3.27 for superimposed drumlins. Drumlin form quantification subsequently performed by Finlayson (2012) for the whole Clyde basin verified these findings for a wider area of the Midland Valley. Diminishing glacier energy and changing ice-flow directions were proposed by Rose and Letzer (1977) to explain the complex drumlin forms of the Glasgow area, although subsequent assessments of superimposed subglacial bedforms (e.g. Clark 1994, 1997) have promoted the concept that the duration and/or ice-flow velocity during individual flow directional events represent the prime control over bedform sizes.

Most local increases in till thickness in the Glasgow area are related to the occurrence of drumlins (Menzies 1981, 1996; Finlayson 2012; Kearsey et al. 2019) and hence are critical to the construction of subglacially streamlined bedforms, regardless of whether they are a product of

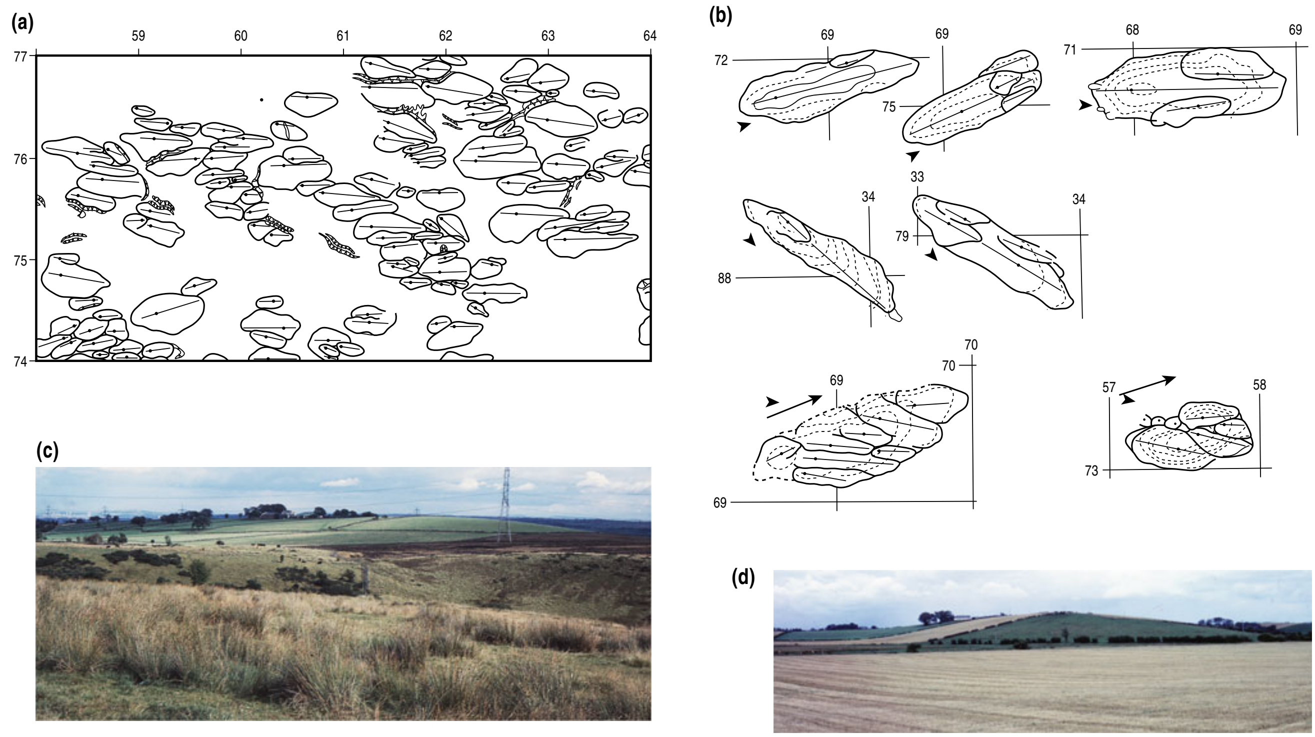

Fig. 26.5 Drumlins of the Glasgow area. **a** and **b** Extracts from the drumlin mapping of Rose and Letzer (1977), (reprinted with permission of the International Glaciological Society and Jim Rose), showing (**a**) the distribution and morphometries of one area of the drumlin field and (**b**) megadrumlins and superimposed drumlins. **c** and **d** Ground views of typical drumlins in the Glasgow area. (Images: Jim Rose)

constructional or excavational deformation. Finlayson (2012) calculated that 70% of the sediment in the Clyde basin is till, but its thickness varies greatly from 0.01 to >40 m. Thicker till tends to occur over free-draining sands and gravels, consistent with predictions that till thickening occurs wherever subglacial materials can dewater, hence creating sticky spots and potentially seeding drumlins (Boulton 1987).

Prominent within the overprinted subglacial bedforms of the Clyde and Ayrshire basins are areas of ribbed moraine (Finlayson et al. 2010; Fig. 26.6). Several hypotheses have been proposed for the generation of such features. Most of these relate to subglacial deformation and present contrasting views that areas of ribbed moraine represent either: (i) an early stage of transverse till-wave production due to instability at the ice-bed interface (Dunlop et al. 2008; Stokes et al. 2013); or (ii) partly re-moulded linear forms, which might be subglacial bedforms (drumlins or flutings) or quarried mega-rafts that are overprinted by less vigorous/shorter-lived ice-flow directional changes (e.g. Boulton 1987; Hattestrand and Kleman 1999), or overridden and streamlined ice-marginal moraines (e.g. Lindén and Möller 2005; Möller 2006). An overridden moraine origin has been proposed for the ribbed moraine of the Clyde and Ayrshire basins by Finlayson et al. (2010), on the grounds that they contain glacitectonised stratified sediments (McMillan and Browne 1983; Browne and McMillan 1989; Fig. 26.6c) and occur in areas where lobate ice margins (defined using flowset mapping) moved against adverse slopes and thereby dammed the drainage to form lakes that filled with stratified sediments. These sediments were then subject to glacitectonic disturbance (transverse ridge construction) due to compressive ice flow and then partial remoulding (ribbed moraine construction).

26.3 Subglacial Bedforms of the Edinburgh Urban Area and the Lower Forth Valley

The subglacial landforms of the Edinburgh urban area were central to the development of the glacial theory itself, for it was at Blackford Hill on 27 October 1840 that a group of Edinburgh geologists led the widely acknowledged father of the glacial theory, Louis Agassiz, to a striated rock surface, confirming to him that the region had been glaciated. Because of its significance, the site is now called Agassiz Rock (Gordon 1993).

The eastern part of the Midland Valley, especially the lowlands encircling the Firth of Forth, is characterised by strongly west–east aligned subglacial bedforms. Predominant are features that comprise a resistant bedrock outcrop, often a sill or volcanic plug and associated easterly tapering

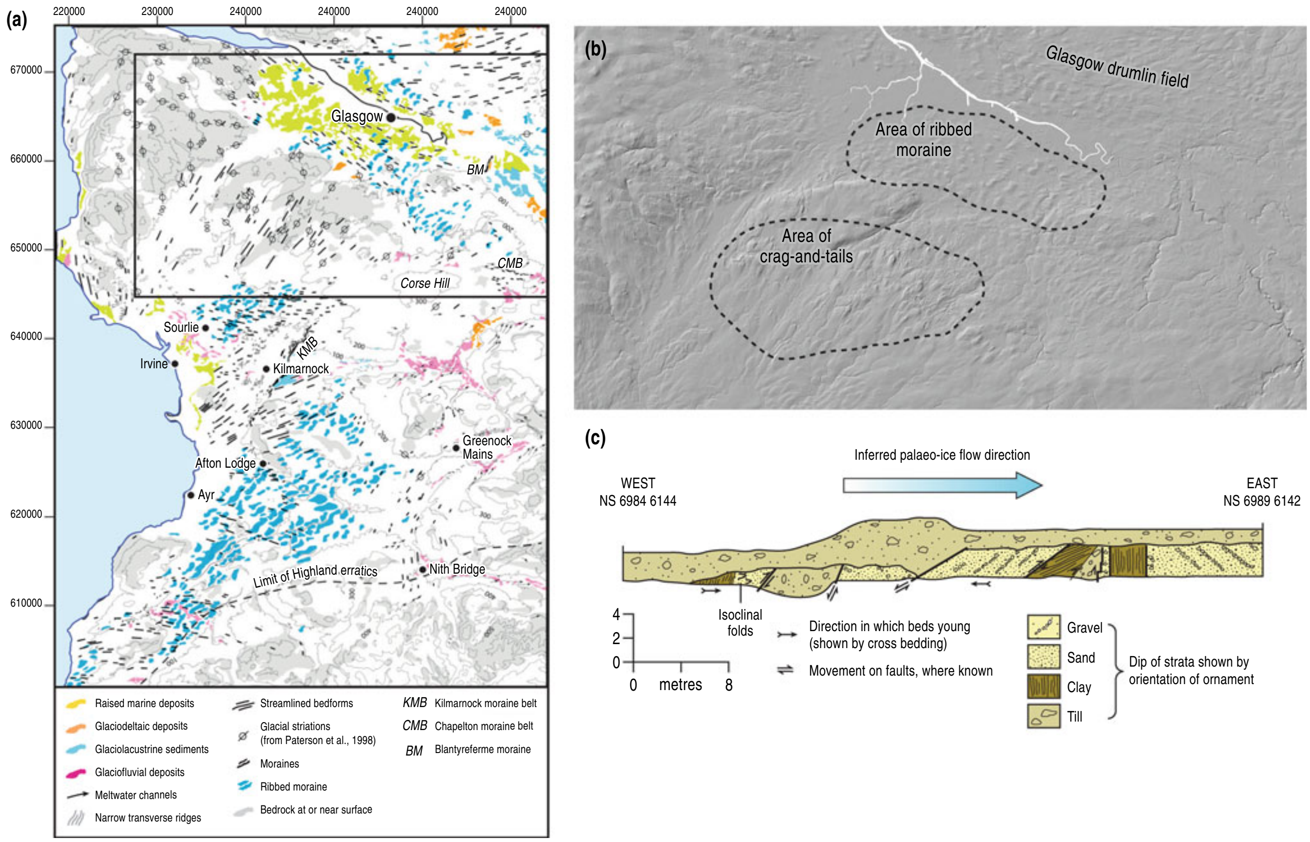

Fig. 26.6 Ribbed moraine of the Clyde and Ayrshire basins. **a** Map of glacial landforms. **b** Digital elevation model of the boxed area in (**a**) showing ribbed moraine south of the River Clyde and the moulding and incomplete streamlining of predominantly NE–SW aligned ridges by east-southeasterly ice flow. South of the ribbed moraine field are crag-and-tail features recording southwesterly ice flow. (NEXTMap Britain data from Intermap Technologies Inc, provided courtesy of NERC via the NERC Earth Observation Centre). **c** Exposure through ribbed moraine (from Finlayson et al. 2010, after McMillan and Browne 1983 and Browne and McMillan 1989), showing glacitectonically displaced stratified deposits capped by till. (**a** and **c** are reprinted from Finlayson et al. (2010); with permission from Elsevier)

tails or ridges (Fig. 26.7). In long section, these features display strong asymmetrical profiles with steep west-facing slopes and gentle eastern slopes. Like Agassiz Rock, the bedrock outcrops commonly display striations whose alignment parallels that of the tapering ridges. Such features around Corstorphine, to the southwest of Edinburgh city centre, were identified by Sir James Hall in the early nineteenth century (Hall 1815) and given by him the enduring term ‘craig-and-tail’ (crag-and-tail). In line with the traditional interpretations of that time, Hall proposed that these features were eroded by great ocean waves sweeping across the area from west to east.

We now understand that such features are the product of glacial erosion, especially under palaeo-ice streams (Eyles and Doughty 2016; Krabbendam et al. 2016). Crag-and-tail features form wherever resistant rock outcrops, such as the many volcanic intrusions of the eastern Midland Valley, have been worn down more gradually than the surrounding weaker materials, thereby forming a proximal stoss bump and distal or lee-side fluting. The potential for lee-side cavity fills and subglacial deforming till to accumulate on the protected lee side of bedrock bumps has prompted the recognition of both erosional and depositional types of crag-and-tail (Boulton 1987; Jansson and Kleman 1999; Fig. 26.8). Using the borehole log database of the Edinburgh area, Sissons (1971) demonstrated that the most celebrated of all crag-and-tail landforms, Castle Rock and the Royal Mile (Evans and Hansom 1996; Fig. 26.9), is covered by only a thin and patchy till and that the form outlined by the bedrock contours (Fig. 26.7c) is a product of differential glacial erosion of the resistant basalt plug of Castle Rock (crag) and less resistant sandstone of the Royal Mile (tail). The 1.5 km length of the tail represents a long subglacial pressure shadow, indicative of the high sliding velocities

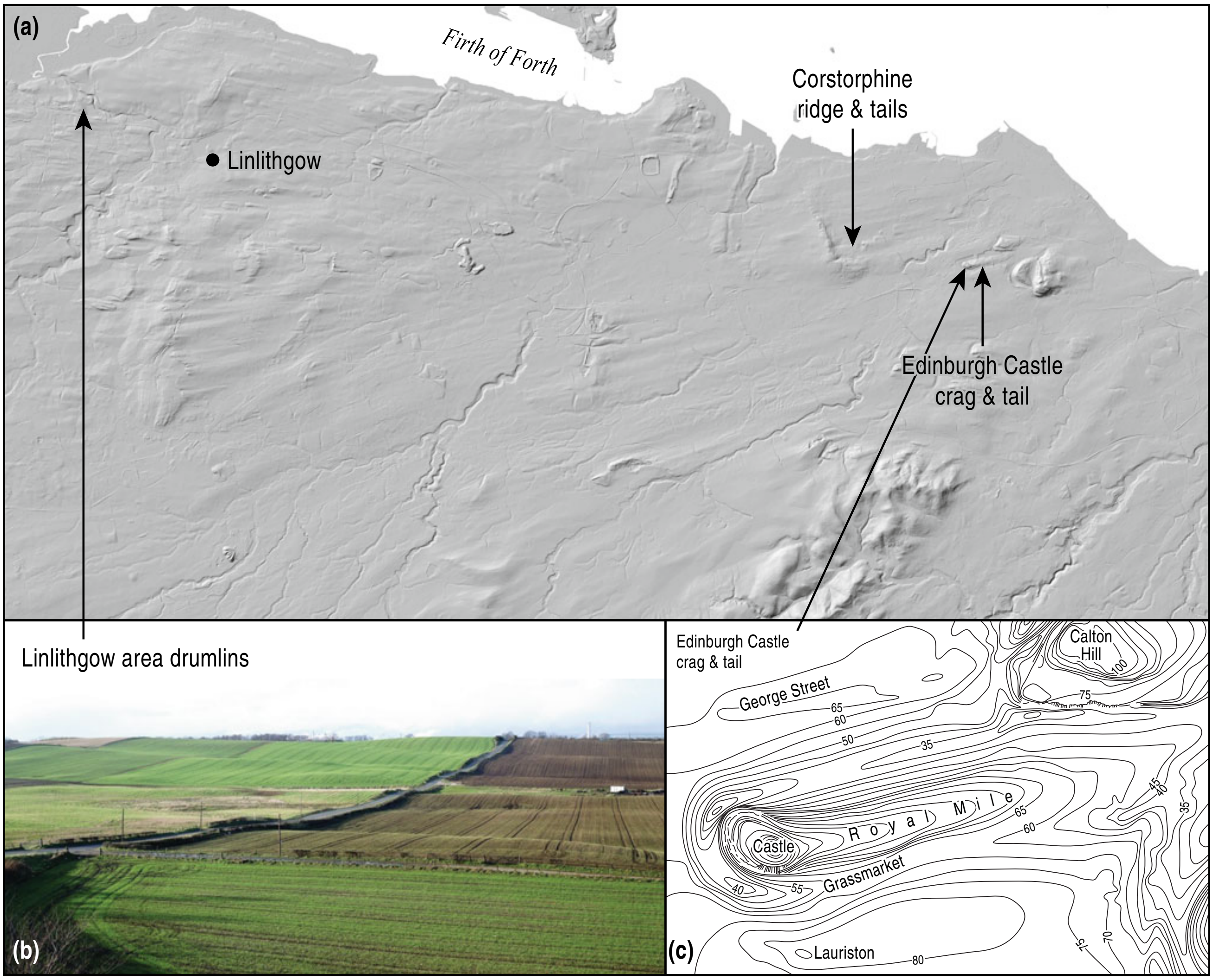

Fig. 26.7 Crag-and-tail features of the Edinburgh area. **a** Digital elevation model showing the signature of easterly ice flow represented by stoss crags and lee tails. (NEXTMap Britain data from Intermap Technologies Inc, provided courtesy of NERC via the NERC Earth Observation Centre). **b** Drumlins in the Linlithgow area. (BGS © UKRI, image P001215, reproduced with permission of the British Geological Survey and available at https://geoscenic.bgs.ac.uk). **c** Bedrock contour map derived from borehole data, depicting the form of the Edinburgh Castle crag-and-tail. (Modified from Sissons 1971)

beneath successive ice streams. Also, prominent in the bedrock contours of this and many other crag-and-tail forms is a curvilinear or horseshoe-shaped depression on the proximal side of the crag (Fig. 26.7c), attesting to concentrated erosion and overdeepening of the adjacent sedimentary bedrock. It is likely that such depressions were formed by the diversion and acceleration of debris-charged basal ice flow and its concentrated impact around the base of bedrock bumps.

In the central-southern part of the Midland Valley, subglacial bedform alignments indicate that ice flow converged on the Firth of Forth from the Southern Uplands, streamlining the interfluves of the Pentland Hills to create drumlinoid drift tails as well as valley-floor drumlins and flutings interspersed with crag-and-tail forms. In the north, Highland ice streamed southeastwards down the upper Forth valley between the Gargunnock and Ochil Hills (Fig. 26.1) and was coalescent with, and steered eastwards by, the vigorous flow of the Clyde–Forth corridor ice to form the Forth Ice Stream. North of the Firth of Forth, this ice overtopped and traversed the minor upland areas (Cleish Hills) south of Loch Leven and the numerous upstanding volcanic intrusions of southern Fife, creating widespread crag-and-tail features interspersed with elongate

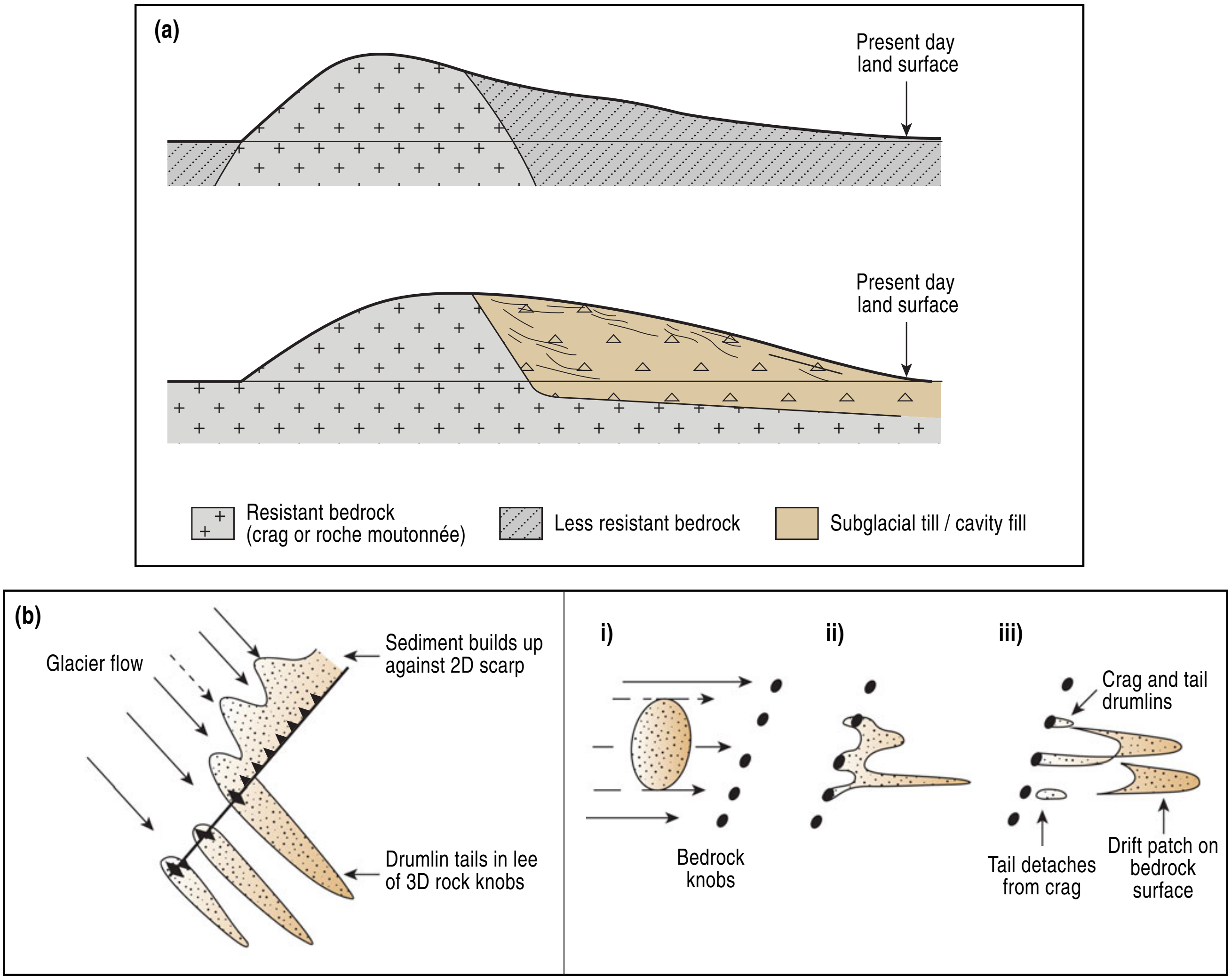

Fig. 26.8 Typical materials and forms of crag-and-tail landforms. **a** Erosional (upper) and depositional (lower) types of crag-and-tail. **b** Boulton's (1987) model for the production of crag-and-tail features by the movement of a subglacial deforming till patch over upstanding bedrock ridges and knolls. (From Boulton (1987) © 1987, reproduced with permission from Taylor & Francis Group)

low-amplitude drumlins, all indicative of generally thin and patchy till cover. Indeed, the eastern Midland Valley surrounding and immediately west of the Firth of Forth is characterised by extremely variable till thicknesses (Kearsey et al. 2019). Sissons (1967) employed numerous borehole datasets to show that tills in this area are mostly less than 6 m thick and form a veneer over a largely glacial erosional landscape. Localised thickening is remarkable, such as drumlins underlain by 160 m of glacigenic sediment, but clearly related to the glacial overdeepening of the Forth valley where a thick sediment infill was available for glacial reworking and moulding.

26.4 Subglacial Bedforms of the Highland Boundary

The Upper Forth, Strathallan and Strathmore lowlands contain subglacially streamlined landforms that were formed by Highland ice flowing southeastwards then eastwards towards the northern North Sea Basin (Golledge and Stoker 2006; Hughes et al. 2010; Roberts et al. 2019). Highland ice flowing from the upper Forth valley fed directly into the Forth Ice Stream, streamlining the horizontally bedded lavas of the Gargunnock Hills to form

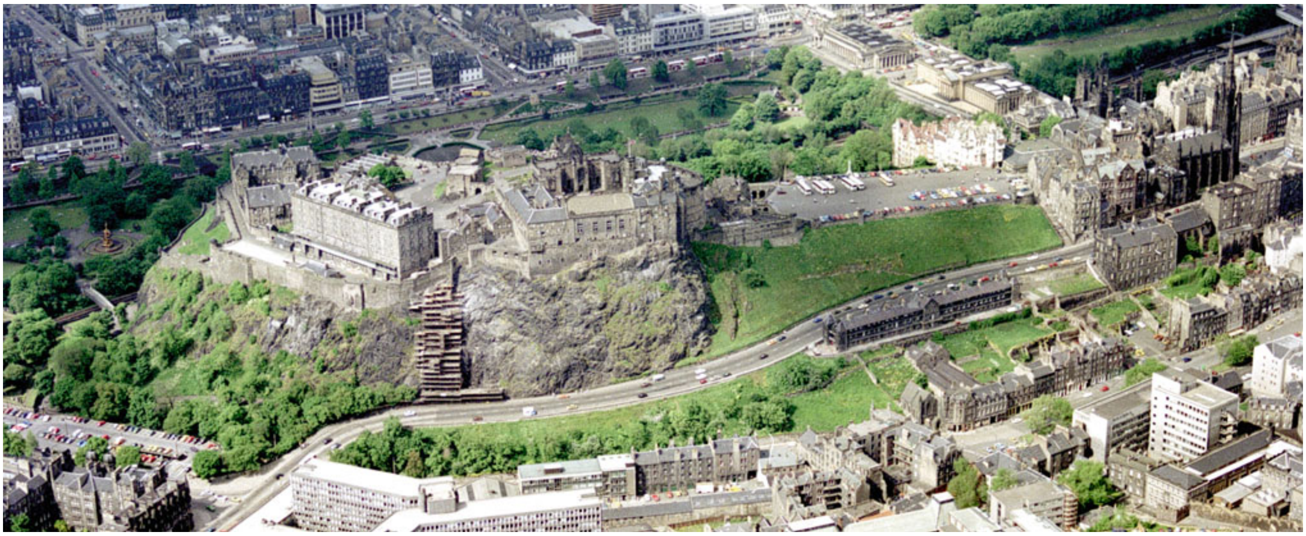

Fig. 26.9 Edinburgh Castle crag-and-tail. The 'crag' is the resistant basalt plug under the castle, and the 'tail' of weaker sedimentary rock underlies the car park and extends eastwards for over a kilometre. (Permit number CP20/039, Photograph P002937 digitised with grant-in-aid from the Scottish Cultural Resources Access Network, Reproduced with permission of BGS © UKRI. All rights reserved). Source: https://geoscenic.bgs.ac.uk/asset-bank/action/viewAsset?id=4228&index=84&total=1000&view=viewSearchItem

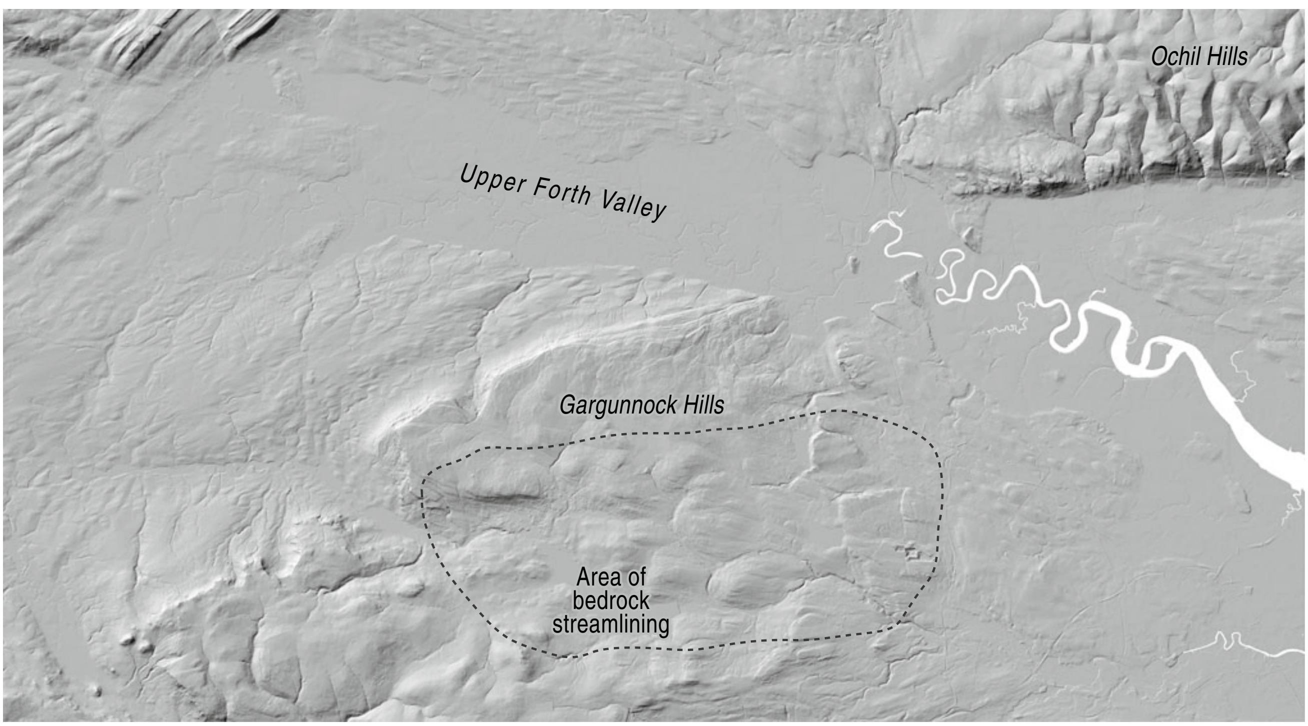

Fig. 26.10 Digital elevation model of the upper Forth valley and surrounding area, showing easterly orientated drumlins funnelling into the valley where ice flow was constrained between the Gargunnock Hills and Ochil Hills. The structural lineations of the Highland Boundary Fault are visible at top left. The overtopping of the Gargunnock Hills by the easterly flowing ice is recorded by the whalebacks, roches moutonnées and faint lineations on the horizontally bedded lavas at the bottom centre. (NEXTMap Britain data from Intermap Technologies Inc, provided courtesy of NERC via the NERC Earth Observation Centre)

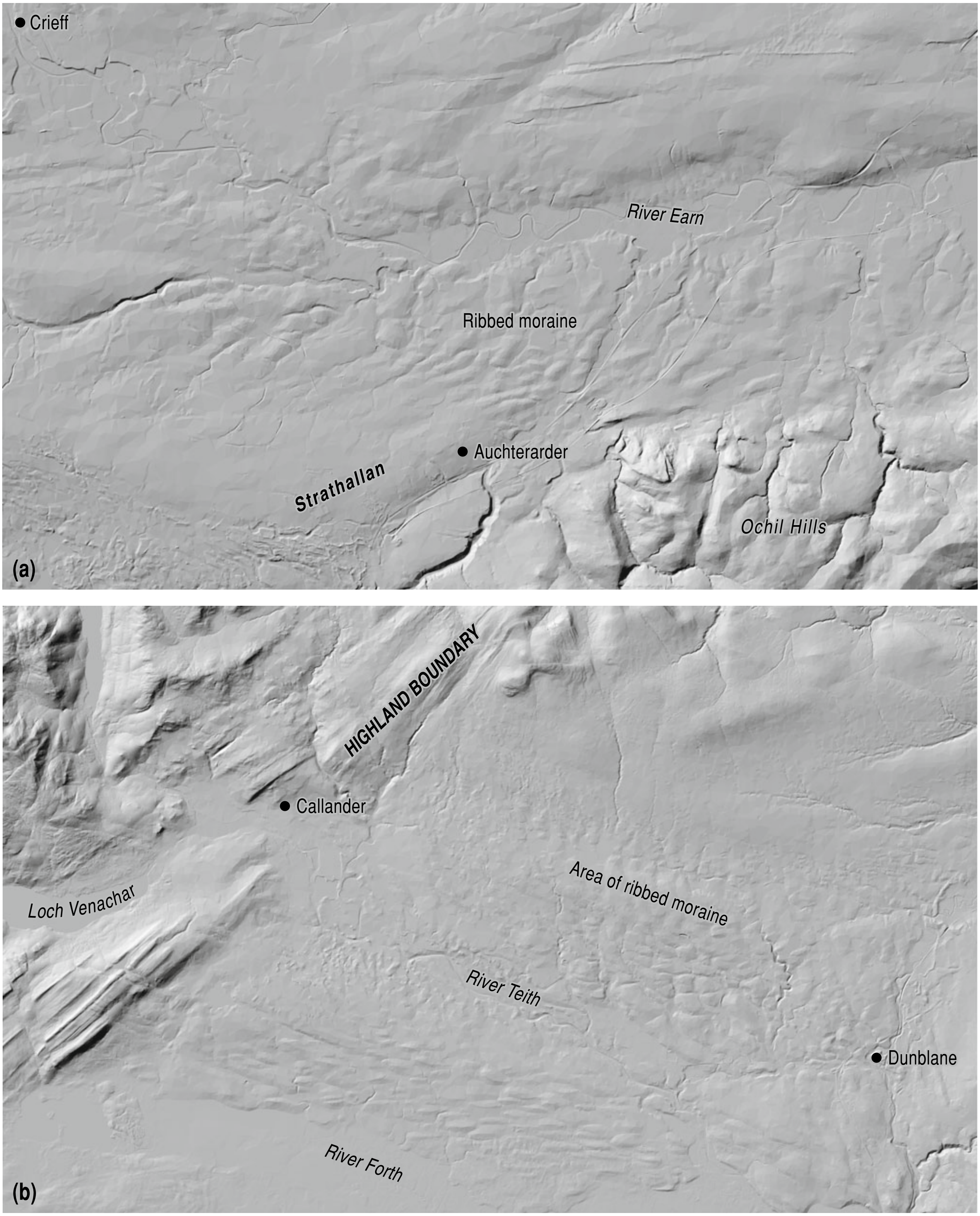

Fig. 26.11 Digital elevation models of ribbed moraine. **a** Strathallan area. **b** Lower Teith valley and upper Forth valley area. (NEXTMap Britain data from Intermap Technologies Inc, provided courtesy of NERC via the NERC Earth Observation Centre)

whalebacks or rock drumlins, roches moutonnées and faintly lineated bedrock, as well as partially rounded, up-ice-facing rock scarps (cf. Gordon 1981; Fig. 26.10). At its thickest, the upper Forth ice also flowed northeastwards into Strathallan, coalescing with ice from the Eastern Grampians to flow along Strathmore and southeastwards or eastwards across the eastern Ochil, Lomond and Sidlaw Hills (Fig. 26.1). The resulting bedforms form two flowsets that represent changing flow directions of the former Strathmore Ice Stream (Golledge and Stoker 2006; Hughes et al. 2010, 2014; Fig. 26.4c). Retreat of the ice margin was characterised by the separation of the Strathmore and Forth Ice Streams and the isolation of the Ochil Hills between the two ice masses (Roberts et al. 2019). A number of non-ice stream flowsets (deglacial flowsets) have been identified by Roberts et al. (2019), with a multi-lobate ice-marginal recession pattern being proposed by Clark et al. (2012) and Merritt et al. (2017).

Ribbed moraine is well developed in the upper Forth valley and Strathallan, where its distribution suggests that it likely originated as ice-marginal moraines, deposited by oscillating glacier lobes emerging from the Highlands, that were subsequently overridden by streaming ice. This is well illustrated near Auchterarder (Fig. 26.11a) by ribbed moraine that was likely deposited as latero-frontal moraines of the former Earn valley glacier then overrun by thicker, easterly flowing ice that produced the west–east aligned drumlins of the Strathmore Ice Stream. Similarly, the ribbed moraine of the lower Teith valley (Fig. 26.11b) could have originated as ice-marginal moraines demarcating an outlet glacier lobe that emerged from the upper Teith basin after being flow-constricted by high ground along the Highland Boundary Fault.

The patchy nature of the till within the subglacial bedforms of the former Strathmore Ice Stream is well illustrated where they are developed over bedrock highpoints (Fig. 26.12). Along the upstanding ridges of the Highland Boundary Fault, highly elongate drumlins are composed of glacigenic sediments (likely till), as evidenced by the fact that they locally mask the underlying bedrock structure (Roberts et al. 2019; Fig. 26.12a). Between these drumlins, bedrock structure appears in windows through the subglacial deposits, often giving rise to crag-and-tail forms; both erosional and depositional types are likely to coexist here and together with the drumlins constitute an excellent example of a mixed (soft and hard) subglacial bed imprint (Fig. 26.3) characteristic of the boundary zone between upland and lowland glaciation. A similar mixed-bed imprint on the Sidlaw Hills (Fig. 26.12b) is typical of bedrock masses that interrupt the down-ice thickening of subglacial deforming layer tills and constitutes an example of drumlin tails (sensu Boulton 1987; Fig. 26.8). An alternative interpretation of such lineations lying down-ice of rough bedrock ridges is that they are eroded into till by a grooving process (Clark et al. 2003); the grooving is effected by ice keels created by uneven carving of the ice stream sole as it passes over the bedrock bumps and then dragged through the soft substrate lying down-ice.

26.5 Conclusions

The Midland Valley contains extensive areas of glacier streamlining relating to ice streaming within the last and earlier ice sheets. Evidence of this includes a range of subglacial bedforms (drumlins, drumlinoid drift tails and flutings) that relate to subglacial deformation, streamlined bedrock erosional features, such as whalebacks and roches moutonnées, and crag-and-tail landforms of both erosional and depositional types. Also, prominent are ribbed moraines, which together with superimposed drumlins record dynamic glacier-flow switches and/or ice-margin oscillations during ice-sheet glaciation. Areas comprising both erosional and depositional streamlined forms reflect patchy till emplacement and are examples of a mixed (soft and hard) subglacial bed mosaic that is characteristic of the boundary zone between upland and lowland glaciation.

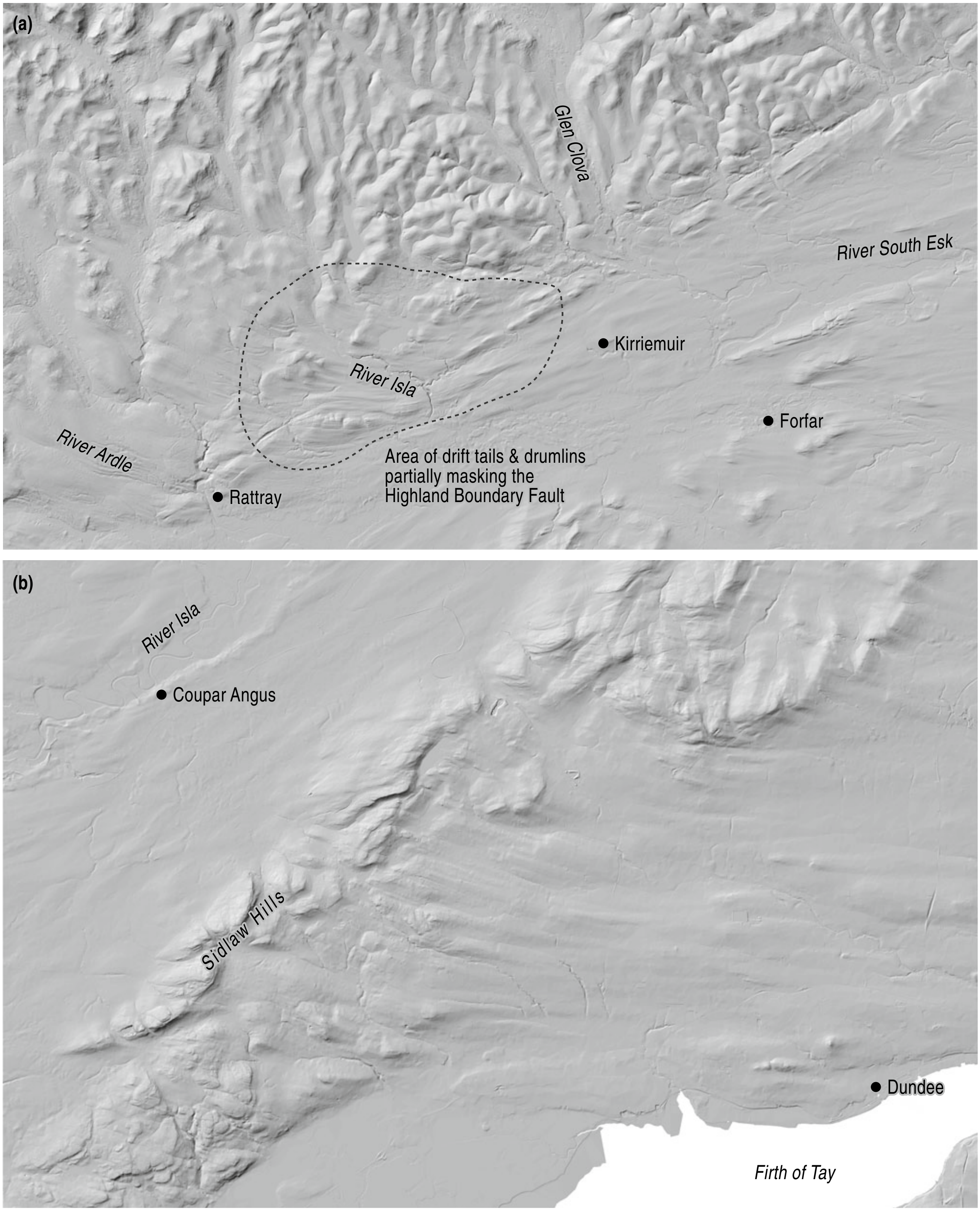

Fig. 26.12 Digital elevation models showing the patchy nature of streamlined deposits over bedrock (mixed-bed imprints) in the area affected by the Strathmore Ice Stream. **a** Drumlins locally masking the bedrock ridges of the Highland Boundary Fault. **b** Streamlined drift tails on the down-flow sides of the Sidlaw Hills. (NEXTMap Britain data from Intermap Technologies Inc, provided courtesy of NERC via the NERC Earth Observation Centre)

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Upland Landscapes and Landforms of the Southern Uplands

27

Colin K. Ballantyne

Abstract

The Southern Uplands form a tract of rolling, dissected tableland, underlain by steeply dipping sedimentary and metasedimentary rocks of Ordovician and Silurian age, into which several granitic plutons were intruded during the Devonian. The landscape is one of selective glacial erosion, but the effects of successive Pleistocene glaciations are muted: deep glacial troughs and cirques are largely restricted to the Galloway Hills in the southwest and the Tweedsmuir Hills in the centre of the region. Moraines are mainly limited to those deposited by Loch Lomond Stadial glaciers in the same areas, but meltwater channels are widespread and outwash terraces deposited during retreat of the last ice sheet flank the floodplains of both trunk and tributary rivers. Frost-weathered regolith covers summits and plateaux, suggesting that these were occupied by cold-based glacier ice during the last glacial cycle. Lateglacial solifluction of regolith and till has produced smooth valley-side slopes and contributed to the accumulation of periglacial valley fills; active solifluction is represented by lobes and ploughing boulders on some of the highest ground. Rock-slope failures occur mainly in clusters and exhibit a variety of forms conditioned by the underlying geological structure and lithology. There is evidence that settlement expansion and associated land-use impacts have caused gullying of hillslopes during the Late Holocene, resulting in the deposition of alluvial fans and debris cones, and transient floodplain aggradation due to increased sediment supply to rivers.

Keywords

Forearc accretionary complex • Selective linear glacial erosion • Loch Lomond Stade • Meltwater channels • Frost-weathered regolith • Solifluction • Periglacial valley fills • Holocene alluviation

27.1 Introduction

The Southern Uplands lie between the Southern Upland Fault and the political border between Scotland and England (Fig. 27.1). Much of this region consists of a rolling, dissected tableland, but it also includes extensive areas of low ground, namely the Galloway and Solway Lowlands in the southwest (Chap. 28) and the Tweed basin in the east. Both geologically and topographically, the upland landscapes of the region are unique in Scotland: geologically because the mountains are mainly composed of sedimentary and metasedimentary rocks of Ordovician and Silurian age, and topographically because the effects of successive Pleistocene glaciations are less marked across much of the region than is the case elsewhere in Scotland. Steep-sided glacial troughs, cirques, rock basins and ice-scoured landscapes are comparatively rare, though the region hosts numerous meltwater channels and outwash terraces flank most of the major rivers. The imprint of postglacial periglaciation is subtle, consisting of smoothing of hillslopes by solifluction and accumulation of periglacial valley fills. Holocene landscape evolution has been dominated by episodes of floodplain and fan aggradation and incision, some of which is attributable to human impacts that destabilized upland slopes and released sediment into fluvial systems.

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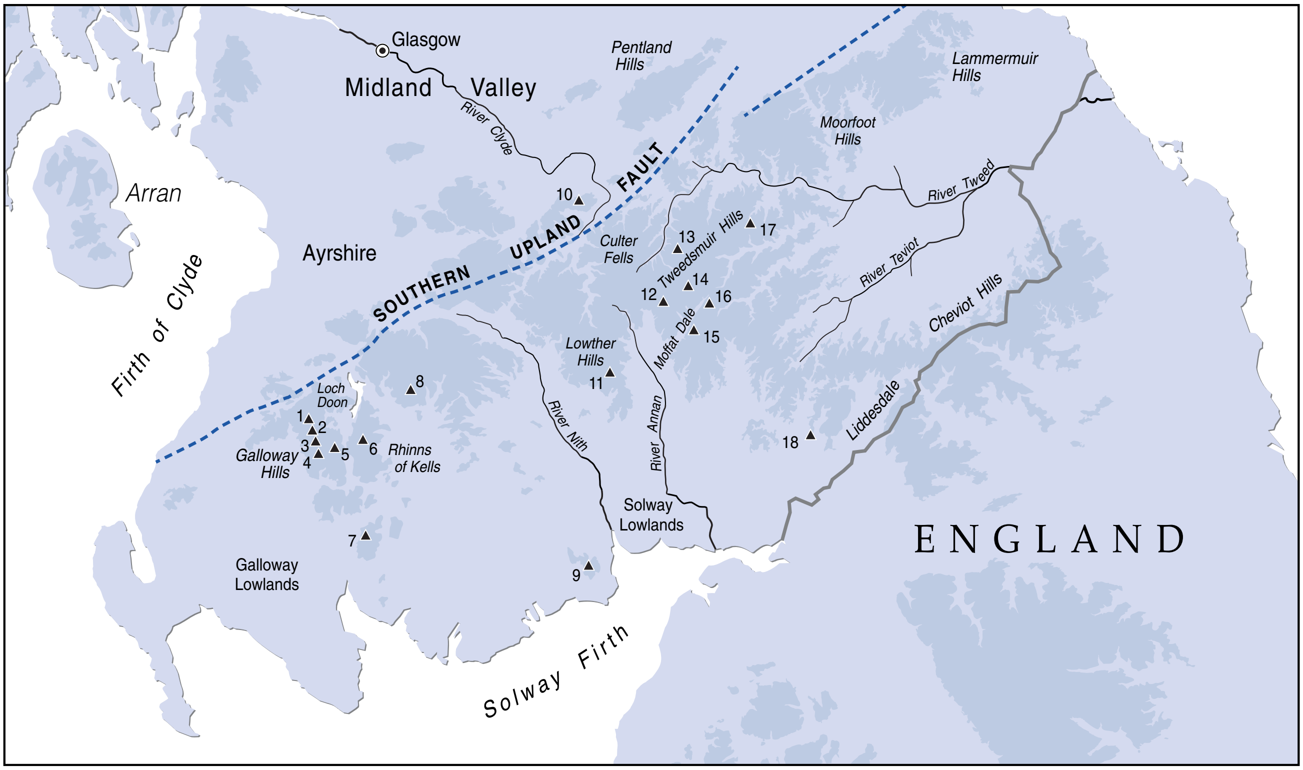

Mountains mentioned in the text

1. Shalloch on Minnoch (775 m)
2. Tarfessock (696 m)
3. Kirriereoch Hill (788 m)
4. Merrick (843 m)
5. Mulwharchar (692 m)
6. Corserine (814 m)
7. Cairnsmore of Fleet (711 m)
8. Cairnsmore of Carsphairn (797 m)
9. Criffel (569 m)
10. Tinto Hill (709 m)
11. Queensberry (697 m)
12. Hart Fell (808 m)
13. Broad Law (840 m)
14. White Coomb (822 m)
15. Capel Fell (678 m)
16. Bell Craig (624 m)
17. Deuchar Law (542 m)
18. Arkleton Hill (521 m)

4 ▲ Mountains
Scotland - England border
Land > 300 m
0 50 km

Fig. 27.1 The Southern Uplands, showing the location of key sites mentioned in the text

27.2 Geology

The region includes most of the Southern Uplands terrane, which extends from the Southern Upland Fault in the north to the Iapetus Suture Zone in the south. The latter is located a short distance south of the political boundary between Scotland and England (Fig. 27.1), and represents the zone where England was grafted onto Scotland as the Iapetus Ocean closed in the Late Silurian, bringing together the ancient continents of Laurentia and Avalonia (Chap. 2). Most of the rocks that underlie high ground in the Southern Uplands terrane are turbidite sandstones (greywackes), locally interbedded with siltstones and mudstones; some of these rocks have been altered to metasandstones and metamudstones by low-grade metamorphism. These form part of a forearc accretionary complex that initially accumulated at the margin of Laurentia during subduction of the Iapetus Ocean in Late Ordovician to mid-Silurian times and was subsequently thrust over the continental margin of Avalonia (Stone et al. 2012). As a result, thrust slices within the accretionary complex were rotated towards the vertical, forming a series of fault-defined tracts aligned southwest to northeast. The highest mountains composed of these rocks (such as those in the Tweedsmuir Hills and Lowther Hills) tend to be underlain by thick beds of steeply dipping, resistant, silica-rich sandstones and conglomerates.

During the final stages of the collision between Laurentia and Avalonia, melting of the base of the crust resulted in intrusion of granitic plutons in the west of the Southern Uplands. These also form high ground, notably Cairnsmore of Carsphairn (797 m), Cairnsmore of Fleet (711 m) and the isolated mass of Criffel (569 m) that rises boldly from the shores of the Solway Firth. The most interesting of these intrusions is the Loch Doon pluton, south of Loch Doon in the Galloway Hills; this outcrop is compositionally zoned from granite at the centre through an encircling mass of granodiorite to a marginal zone of tonalite and diorite. Here, the plutonic rocks form a basin of relatively low ground surrounded by some of the highest mountains in the region, such as Merrick (843 m) and Corserine (814 m). This apparent anomaly has resulted from thermal alteration of the sedimentary rocks surrounding the pluton, forming a metamorphic aureole of resistant hornfels (metamorphosed greywacke) that underlies the highest ground.

The high ground of the Southern Uplands is thought to have been uplifted during the period of intense Palaeogene magmatic activity that preceded opening of the North Atlantic Ocean (~62–55 Ma; Chaps. 2 and 3), augmented by a later Neogene (probably Miocene) uplift episode that may reflect the far-field effects of the Alpine Orogeny (Stone et al. 2012). These uplift episodes resulted in stripping of post-Silurian cover rocks from upland areas, and the formation and extensive fluvial incision of broad palaeosurfaces that now form a dissected tableland 400–600 m in elevation that declines gently northeastwards in average altitude from the high ground of the Tweedsmuir Hills to the Moorfoot and Lammermuir Hills (Fig. 27.2).

27.3 Glacial History

Although the large-scale glacial erosional features (cirques, troughs and rock basins) of the Southern Uplands developed over multiple Pleistocene glacial cycles, no unequivocal deposits predating the Devensian glacial stage have been found in the region. During the expansion, culmination and shrinkage of the last (Late Devensian) Scottish Ice Sheet (~35–14 ka), the Southern Uplands supported three confluent centres of ice dispersal, located over the Galloway Hills in the southwest, the Tweedsmuir Hills (Moffat Hills) in the centre and the Cheviot Hills in the east. The evolution of the ice divides during ice-sheet expansion and contraction has been reconstructed through identification of cross-cutting flowsets derived from the evidence provided by erratic transport and the alignment of drumlins, streamlined drift ridges, ribbed moraines and meltwater channels (Livingstone et al. 2008, 2012; Evans et al. 2009; Chap. 28).

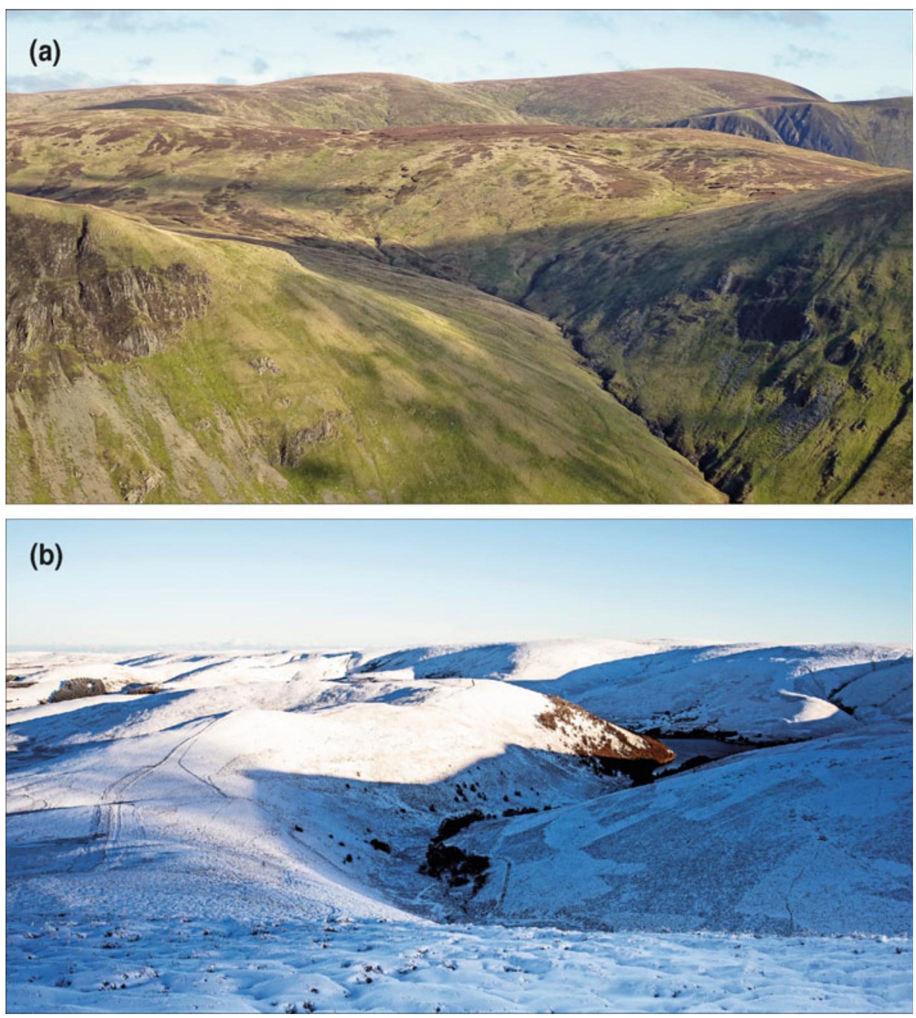

Fig. 27.2 Tablelands of the Southern Uplands. **a** The Tweedsmuir Hills: White Coomb (822 m) photographed from Hart Fell. **b** The Lammermuir Hills, looking east from the head of Hopes Water valley. (Images: **a** Colin Ballantyne; **b** John Gordon)

During the Last Glacial Maximum, a major ice divide became established over the western Southern Uplands between the Galloway Hills and the Tweedsmuir Hills, feeding radial ice movement to the northwest across the Firth of Clyde, to the northeast across the Midland Valley and to the south into the Irish Sea Basin. To the east and southeast, ice nourished over the Southern Uplands fed into the Tweed valley and into the Tyne valley in northern England. Both the Tyne and Tweed valleys were occupied by ice streams that contributed to a major ice lobe that extended southwards along the east coast of England (Everest et al. 2005; Livingstone et al. 2012, 2015). Radial ice movement persisted during the early stages of ice-sheet retreat, though the main ice divide in the east of the region appears to have migrated southwards to the Cheviot Hills. Oscillatory retreat of the southwestern sector of the Southern Uplands ice cap was interrupted by a readvance of the ice margin (the Scottish Readvance) across the Solway Lowlands and into the northern Irish Sea Basin at ~19.3–18.3 ka (Chiverrell et al. 2018). The Solway Lowlands were deglaciated sometime before 16.4–15.7 ka (Livingstone et al. 2015), and a basal radiocarbon age indicates deglaciation of the lower Annan valley (Fig. 27.1) prior to 15.9–15.2 cal ^{14}C ka (Bishop and Coope 1977). Recalibration of cosmogenic ^{10}Be exposure ages reported by Ballantyne et al. (2013) for two sites in the heart of the Galloway Hills indicates extensive deglaciation of these mountains by 15.0 ± 0.7 ka (Ballantyne and Small 2019). Collectively, the deglaciation ages cited above suggest that only limited ice cover remained over the Southern Uplands prior to rapid warming at ~14.7 ka, when mean July temperatures in SE Scotland rose by ~6 °C (Brooks and Birks 2000), ushering in the cool temperate conditions of the Lateglacial (Windermere) Interstade (~14.7–12.9 ka) and causing the disappearance of the last remnants of the Late Devensian ice sheet in southern Scotland.

At the onset of the ensuing Loch Lomond (≈Younger Dryas) Stade of ~12.9–11.7 ka, mean July temperatures on low ground fell to ~7.5 °C (Brooks and Birks 2000) and glaciers formed on the Galloway Hills and Tweedsmuir Hills. In the former area, Cornish (1981) employed a range of geomorphological evidence to reconstruct the former dimensions of 13 small cirque or valley glaciers with a total area of 9.9 km^2, though it is possible that some of these were fed by thin cold-based ice caps occupying adjacent plateaux. Mapping of moraines and meltwater channels in the Tweedsmuir Hills by Pearce (2014) indicates that during this time plateau ice caps formed on high ground both north and south of the Megget valley, feeding small outlet glaciers in the adjacent valleys; at their maximum extent, these ice caps probably occupied an area of over 60 km^2. One or two cirque glaciers also formed at this time on Cairnsmore of Carsphairn (797 m), northeast of the Galloway Hills.

Ice-wedge pseudomorphs occur in near-surface outwash gravels in lowland areas fringing the Southern Uplands (Stone et al. 2012) and provide evidence for extensive permafrost both during the retreat of the last ice sheet prior to ~14.7 ka and again during the coldest part of the Loch Lomond Stade (Ballantyne 2019). At these times, solifluction was widespread on upland slopes and periglacial valley-fill deposits accumulated on permafrozen terrain.

27.4 Glacial Landforms

27.4.1 Glacial Landforms of the Galloway Hills

The Galloway Hills (Fig. 27.1) are dominated by a ring of mountains composed of resistant metamorphosed greywacke that surrounds the comparatively low-lying plutonic rocks of the Loch Doon Basin. West of the basin lie the Merrick Hills, consisting of five summits over 600 m that culminate in Merrick (843 m); east of the basin are the Rhinns of Kells, a 15 km-long ridge that reaches its highest point in Corserine (814 m; Fig. 27.3a). The topography of the basin itself is complex, with a central north–south trending ridge of granite forming relatively high ground that culminates in Mulwharchar (692 m), flanked by valleys and shallow lochs on terrain mainly underlain by granodiorite.

The Galloway Hills formed an independent centre of ice dispersal throughout the lifetime of the last ice sheet (~35–14 ka). The evidence provided by striae and ice-moulded bedrock demonstrates former radial ice movement from the Loch Doon Basin (south of Loch Doon itself), and detailed mapping by Cornish (1982, 1983) has shown that allochthonous erratics are absent from the Galloway Hills, and that granite and granodiorite erratics radiate in all directions except east from their parent outcrop within the basin. Mapping of sequential flowsets derived from the alignment of subglacially streamlined bedforms on the adjacent lowlands has confirmed persistent radial ice flow to the west, southwest, south and southeast across the adjacent lowlands (Salt and Evans 2004). Large granite erratics are strewn across the highest summits west of the outcrop (Fig. 27.3b), implying that they have been elevated by up to 300 m over a distance of less than 4 km.

Over successive glacial cycles, ice moving radially from the Loch Doon Basin has excavated glacial breaches through the encircling high ground, notably the River Dee breach to the southeast, Glen Trool to the southwest and the Loch Doon breach to the north. Both the Glen Trool and Loch Doon breaches comprise narrow rock basins where they cut through the rocks of the metamorphic aureole: Loch Trool has a maximum depth of 17 m, and prior to damming in the 1930s Loch Doon had a maximum depth of 30 m. In each case, the deepest part of the rock basin coincides with the

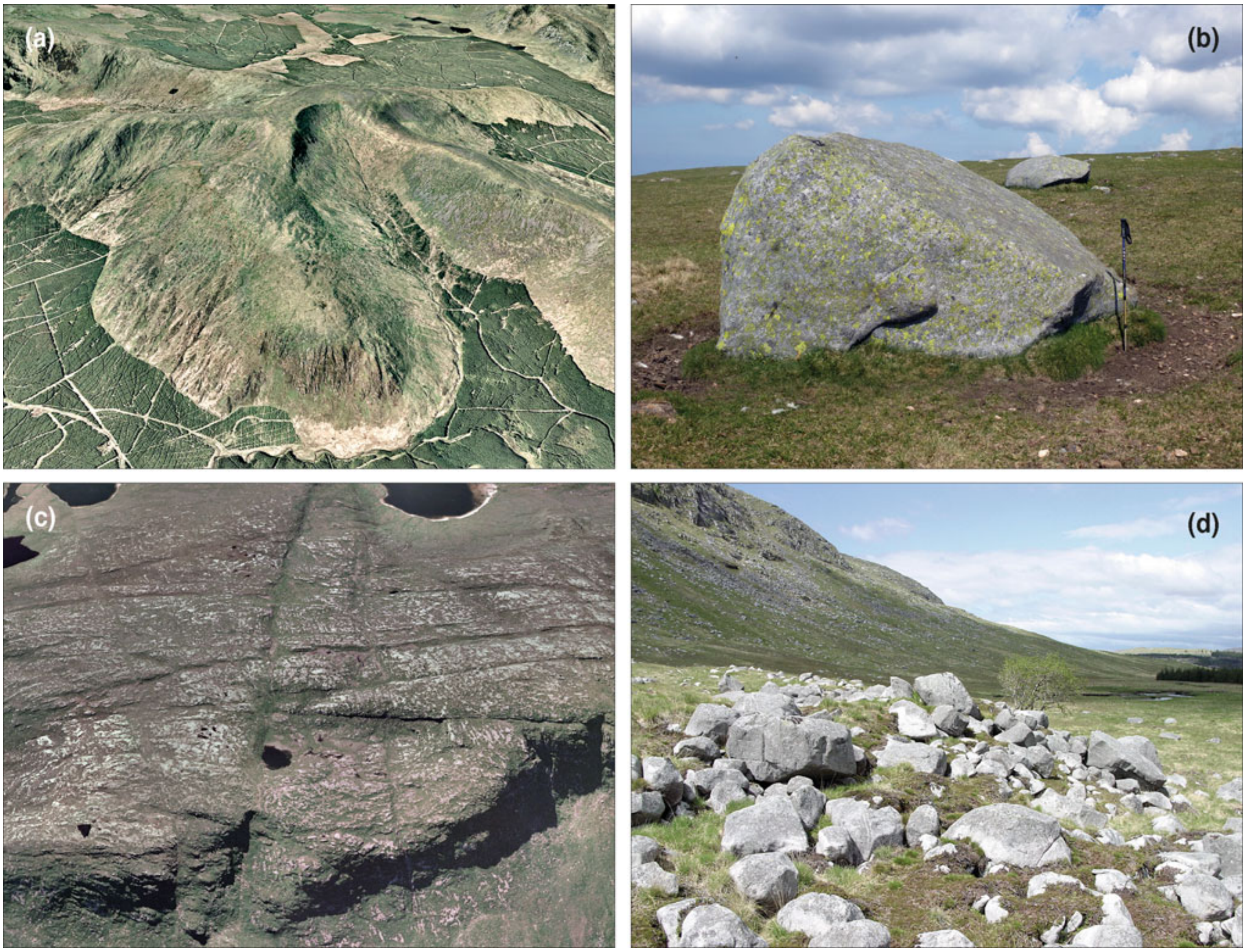

Fig. 27.3 Glacial landforms in the Galloway Hills. **a** Shallow, immature cirques on Corserine (814 m), the highest summit on the Rhinns of Kells ridge. **b** Granite erratics resting on frost-weathered regolith of metamorphosed greywacke near the summit of Shalloch on Minnoch (775 m) in the Merrick Hills. **c** Glacially scoured granite bedrock in the centre of the Loch Doon Basin. **d** The distal slope of the outermost Tauchers moraine. (Images: **a**, **c** Google Earth™; **b**, **d** Colin Ballantyne)

constriction imposed by the belt of metamorphosed greywacke. On terrain underlain by granite and granodiorite within the Loch Doon Basin is one of the few areas of ice-scoured bedrock topography in the Southern Uplands (Fig. 27.3c). As glacial erosion is generally considered to be ineffective at ice divides, it is likely that this reflects the operation of subglacial abrasion and quarrying during the early stages of ice-sheet growth (and possibly the final stages of ice-sheet retreat) when glaciers moved across the basin from their sources in the surrounding hills (Cornish 1982, 1983).

Cirques within the Galloway Hills are immature: most lack cliffed headwalls, and all have outward-sloping floors (Fig. 27.3a). Glacial deposits and depositional landforms within the area are largely limited to those associated with the 13 small glaciers that formed during the Loch Lomond Stade. The extent of these glaciers was reconstructed by Cornish (1981) from the evidence provided by drift limits, spreads of glacially deposited boulders and end, lateral and hummocky moraines. The most outstanding depositional landforms are a suite of superb converging fluted moraines deposited by a Loch Lomond Stadial glacier in a valley northwest of Merrick, and the Tauchers moraine, an end-moraine complex that occupies low ground (260–280 m) below steep granite cliffs (The Tauchers) that form the northeast flank of Mullwharchar (692 m) in the heart of the Loch Doon Basin. The outer complex of the Tauchers moraine is up to 200 m wide, rises up to 12 m above the adjacent terrain (Fig. 27.3d), extends over a distance of 1.3 km and is continued upslope towards the former glacier source area by well-defined bouldery lateral moraines. The complex consists of two continuous ridges and multiple discontinuous secondary ridges, suggesting that it was deposited during slow glacier-front recession interrupted by limited readvances of the ice margin. Cosmogenic ^{10}Be dating of samples chiselled from boulders on the moraine has yielded a mean age of 11.9 ± 0.8 ka, confirming that it was deposited during the Loch Lomond Stade (Ballantyne et al. 2013).

Cornish (1981) calculated that the glacier that deposited the Tauchers moraine had an area of $\sim$0.9 km^2 and an equilibrium line altitude (ELA) of $\sim$330 m. This is much lower than that of the ELAs he calculated for the six former glaciers that occupied the Merrick Hills to the west (497–670 m), a discrepancy he explained by nourishment of the former Tauchers glacier by snow blown from the high ground to the west. An alternative possibility is that the glacier was fed by a thin, cold-based ice cap on the adjacent plateau. The large dimensions of the moraine complex and of its constituent boulders (which reach lengths of 5 m; Fig. 27.3d) were attributed by Ballantyne et al. (2013) to either collapse of the Tauchers rockwall prior to the Loch Lomond Stade and reworking of the runout debris by the advancing glacier, or rockwall collapse directly onto the glacier surface. If the latter occurred, it is possible that loading by rock avalanche debris contributed to extension of the glacier front onto low ground.

27.4.2 Glacial Landforms of the Tweedsmuir Hills

As in the Cairngorm Mountains (Chap. 18) and the Eastern Grampian Highlands (Chap. 20), the upland landscape of much of the Southern Uplands is dominated by selective linear erosion: successive glaciers and ice sheets have excavated troughs along pre-existing trunk valleys, but extensive areas of pre-glacial plateau have been preserved. This type of landscape is exemplified in the Tweedsmuir Hills (Fig. 27.2a), an area of dissected tableland that culminates in Broad Law (840 m) north of the Megget valley, and White Coomb (822 m) and Hart Fell (808 m) to the south. Glacial troughs radiate from the high ground or follow SW–NE aligned faults, but in comparison with their Highland counterparts most are rather subdued features with comparatively gentle valley sides rising less than 300 m above trough floors (Fig. 27.4a). The most impressive is the 17 km-long Moffat Dale trough, which follows a fault that separates two geological formations and breaches the local drainage divide (Fig. 27.4b). Near its northern end a spectacular waterfall, the Grey Mare's Tail, marks the outlet of a hanging valley perched above the trough floor and indicates that formation of the trough involved $\sim$200 m of valley deepening over successive glacial cycles (Fig. 27.4c). Rock basins are rare in the area, with the exception of 47 m-deep St Mary's Loch near the head of Moffat Dale. Other troughs, such as the Talla and Fruid valleys, have been dammed to form reservoirs, though a shallow rock basin was present in the Talla valley prior to damming. Most tributary valleys lack headscarps, and well-developed cirques are rare in the area. The best examples are those at the heads or flanks of tributary valleys feeding southeastwards into Moffat Dale, such as that enclosing Loch Skene and the twin cirques that are scalloped into the southern spur of Hart Fell (Fig. 27.4d).

Moraines are largely restricted to valleys and cirques at the margins of the Tweedsmuir Hills and most represent readvance of local glaciers during the Loch Lomond Stade. The majority are recessional moraines, nested chevrons of ridges that define the former margins of outlet glaciers that underwent active, oscillatory retreat (Fig. 27.4e), and their location in valleys radiating from high ground implies that they were fed by ice caps on the adjacent plateaux (Pearce 2014; Pearce et al. 2014). These plateaux are covered by a mantle of frost-weathered regolith (Sect. 27.5), suggesting that they were occupied by thin, cold-based glacier ice that has left no trace on the plateau surfaces. The most interesting site is at Loch Skene, a shallow lochan (small lake) in a cirque on the eastern flank of the Moffat Hills. Here, a succession of hummocky recessional moraines defines the extent of a glacier that developed in the cirque. A 650 m-long, 2.5 m-high lateral moraine ridge known as The Causey extends northwards from the area of recessional moraines (Fig. 27.4f). As recessional moraines extend beyond the ridge, it probably represents the limit of a minor readvance that interrupted retreat of the Loch Skene glacier. Pearce (2014) calculated that the area-weighted palaeo-ELA of the south Tweedsmuir Hills ice cap and associated outlet glaciers was up to 100 m higher than that for coeval glaciers on the Galloway Hills, a contrast that suggests an eastward decline in snowfall during the Loch Lomond Stade.

27.4.3 Glacifluvial Landforms

Amongst the most widespread landforms in the upland valleys are suites of hundreds of meltwater channels that are cut in drift or bedrock along spurs or valley-side slopes (Fig. 27.5a). Many are ice-marginal channels cut by streams flowing along the lateral margins of the last ice sheet as it downwasted to form a transection complex of retreating valley glaciers (Fig. 27.5b). Others are of submarginal or subglacial origin, and some are col channels that represent superimposition of englacial streams across upland cols; the latter often have up-and-down long profiles that demonstrate flow across cols by subglacial streams under hydrostatic pressure (Fig. 27.5c). Research on these channels in the northern Lammermuir Hills and upper Tweed valley produced a series of classic papers that demonstrated that the channels were of ice-marginal, submarginal or subglacial origin, and not (as previously supposed) channels cut by the

Fig. 27.4 Glacial landforms of the Tweedsmuir Hills. **a** Dryhope valley, a tributary trough. The valley floor is partly occupied by dissected periglacial valley-fill deposits. **b** The Moffat Dale trough, excavated along a major fault. The rumpled slopes on the left are part of the Bell Craig rock-slope deformation. **c** The Grey Mare's Tail waterfall drops ~200 m from a hanging valley into the Moffat Dale trough. **d** The Upper Coomb Craig cirque, one of two on the SE spur of Hart Fell above the Blackhope valley. **e** Recessional moraines in the Talla valley. **f** Hummocky recessional moraines at Loch Skene. The prominent ridge in the background is The Causey, which apparently formed during net retreat of the Skene glacier. (Images: **a–c**, **e**, **f** John Gordon; **d** Colin Ballantyne)

Fig. 27.5 **a** Meltwater channels cut in glacial drift, Lowther Hills. **b** Marginal and submarginal meltwater channels, Culter Fells. **c** 'Up-and-down' col channels, Culter Fells. **d** Large col channel, Culter Fells. (Images: **a** John Gordon; **b**–**d** Google Earth™)

overflow from former ice-dammed lakes (e.g. Sissons 1958, 1961; Price 1960; Evans 2004). The largest bedrock channels are over 30 m deep (Fig. 27.5d) and probably reflect reoccupation by meltwater streams during successive glaciations. More recently, the pattern of ice-marginal meltwater channels has contributed to reconstructions of the overall pattern of regional ice-sheet retreat (Clark et al. 2012).

In some locations, meltwater channels feed into eskers and kame terraces on low ground (Sissons 1958, 1961). By far the most extensive glacifluvial deposits in the Southern Uplands, however, are represented by outwash terraces located on the broad straths of major rivers such as the upper reaches of the Nith, Annan, Clyde, Yarrow and Tweed and their major tributaries. South of Abington in the upper Clyde valley, for example, a suite of broad outwash terraces rises up to 20 m above the present floodplain. In most upland valleys, however, subsequent fluvial incision and channel migration have left only a fragmentary record of the extent of former outwash deposits. The surviving outwash terraces lack kettle holes, suggesting that they were deposited proglacially at the termini of actively retreating glaciers.

27.5 Periglacial Features

27.5.1 Relict Periglacial Features

The plateaux and smooth upper slopes of the Southern Uplands are extensively mantled by a cover of frost-weathered regolith typically 1–3 m thick. On hills above 450 m underlain by greywackes, this has a threefold structure: an uppermost rubble layer, 0.5–0.7 m thick, of angular clasts and interstitial sand; an underlying layer, ~0.2–0.7 m thick, of small clasts in a tough, compact, silt-rich matrix; and a lowermost layer of larger clasts derived from the underlying bedrock. The uppermost rubble layer was interpreted by Ragg and Bibby (1966) as the product of vertical frost sorting and the underlying compacted silt-rich layer as a fragipan or indurated soil horizon. Fragipans develop through illuviation of silt and fine sand and subsequent compression of soil peds by ice-lens growth. They form under severe periglacial conditions in the uppermost ice-rich zone of permafrost or at the base of the active layer immediately above the permafrost table (Ballantyne 2018). The regolith mantle thus represents mechanical weathering of bedrock by frost wedging and ice segregation operating on jointed bedrock, upward frost sorting of coarse debris in the former active layer and compaction of the underlying illuviated fines by ice-lens formation within or above former permafrost. The age of the regolith cover in the Southern Uplands is unknown, but studies of bouldery mountain-top regolith covers (blockfields) in the Scottish Highlands have shown that these developed prior to the build-up of the last ice sheet and were preserved under cold-based ice that was frozen to the underlying substrate and hence accomplished little or no erosion of the regolith cover (Fabel et al. 2012; Hopkinson and Ballantyne 2014). In the Southern Uplands, this interpretation is supported by the occurrence of large granite erratics resting on the hornfels-derived regolith that mantles the higher summits of some of the Galloway Hills (Fig. 27.3b).

On slopes, the regolith has moved downslope by solifluction, and clasts within the upper layers of the regolith mantle exhibit strong preferred downslope long-axis alignment (Ragg and Bibby 1966). On many slopes, Lateglacial solifluction has little or no surface expression: a featureless solifluction sheet covers the slope, merging imperceptibly with soliflucted till deposits near the slope foot. Elsewhere, however, solifluction has resulted in the formation of solifluction lobes and terraces. Some of these are active under present conditions, but others are relict Lateglacial landforms. The northern spur of Corserine in the Galloway Hills, for example, is festooned with hundreds of large, bouldery, stone-banked solifluction lobes.

In many parts of the Southern Uplands and Cheviot Hills, Lateglacial mass movement under cold conditions has contributed to the accumulation of periglacial valley-fill deposits. In broad upland valleys, these form terraces 20–300 m wide, with steep frontal bluffs formed by river incision (Fig. 27.4a). Some consist entirely of reworked till or regolith, and others comprise in situ till overlain by reworked till and regolith derived from adjacent slopes (Harrison 2002). Some are massive, others are crudely stratified, and some are intercalated with slopewash gravels (Mitchell 2008). Optically stimulated luminescence dating of such deposits suggests that they mainly accumulated during the Loch Lomond Stade (Harrison et al. 2010). Although the downslope accumulation of sediment on valley floors was initially attributed to solifluction, the thickness of some periglacial valley fills appears incompatible with transport by Lateglacial solifluction alone, prompting Harrison (2002) to suggest that they represent reworking of valley-side sediments by paraglacial debris flows or possibly by rapidly moving active layer failures over permafrost.

27.5.2 Active Periglacial Landforms

Evidence for present-day periglacial activity is limited to peat-free ground above ~500 m. The most widespread active periglacial landforms are vegetation-covered solifluction lobes and terraces with steep risers up to a metre high. These are best developed on Merrick (843 m) and Kirriereoch Hill (788 m) in the Galloway Hills, where they are accompanied by ploughing boulders; the latter also occur on some granite mountains, such as Cairnsmore of Carsphairn (Wilson 1993). Turf-banked terraces formed through retardation of creeping frost debris by bands of vegetation are rare in the Southern Uplands, but well developed above 700 m near the summit of Corserine (Ballantyne and Harris 1994) .

Extensive vegetation and peat cover limit the formation of active frost-sorted patterned ground. Just 5 km north of the Southern Upland Fault, however, the finest active sorted stripes in Scotland occur above 500 m on Tinto Hill (709 m), where stripping of the vegetation cover has exposed felsite regolith consisting of small clasts embedded in frost-susceptible, silt-rich soil. Individual stripes are typically 15–35 cm wide (Fig. 27.6) and have been shown to re-form over disturbed ground within 1–3 winters. Lateral sorting has been accomplished by differential needle-ice growth and near-surface frost heave, which cause the fine stripes to dome upwards during periods of severe freezing. The striped area forms a carpet of mobile debris that is moving downslope by a combination of frost creep, needle-ice creep and possibly surface wash over frozen ground; downslope clast displacement of 24–62 cm has been recorded over a single winter (Ballantyne 2001).

27.6 Landslides

Although only about 60 rock-slope failures (RSFs) have been identified on slopes between the Southern Upland Fault and the English border, these illustrate a variety of types that is partly conditioned by the nature of the underlying geology (Fig. 27.7). Most are rockslides or rock-slope deformations and occur in five clusters. Two of these clusters are in the Cheviot Hills north of the English border: in the north, there are arrested rockslides on andesite lavas, and in the western Cheviots there is a group of small rockslides caused by failure along a scarp of Carboniferous sedimentary rocks. In

Fig. 27.6 Active frost-sorted stone stripes, Tinto Hill. (Image: Colin Ballantyne)

the low hills west of Liddesdale, several distinctive rockslides represent scarp failure of interbedded sandstone and argillaceous rocks overlain by sandstone or basalt caprocks. The largest cluster of about 14 RSFs is in the Tweedsmuir Hills, on various sedimentary and metasedimentary rocks. Finally, a group of seven RSFs occurs in the Lowther Hills, mostly on greywacke. Rock-slope failures are absent on most of the Galloway Hills, though Cornish (1981) documented a large (~ 0.6 km^2) rockslide on the metamorphosed greywacke of Tarfessock (696 m), west of the Loch Doon Basin.

Figure 27.7 illustrates some of the variety of RSFs in the Southern Uplands. The Deuchar Law arrested rockslide in greywacke (Fig. 27.7a) is unusual in that it is located at the margin of a meltwater channel incised through a col in the Tweedsmuir Hills. By contrast, the Black Cove RSF on Arkleton Hill near Langholm (Fig. 27.7b) has resulted from failure of sandstones and argillaceous rocks underlying a basalt scarp and exhibits extended run-out in the form of a tongue of debris. The most extensive area of rock-slope failure in the Southern Uplands consists of four large rock-slope deformations affecting an area of ~ 2.0 km^2 along the steep eastern flank of the Moffat Dale trough. These are represented by tension cracks, rock benches and antiscarps (uphill-facing scarps) that extend along the slope (Figs. 27.4b and 27.7c). These impressive landforms result from gradual deep-seated movement of steep valley-side slopes underlain by metasedimentary rocks and represent over 25% of the total area of rock-slope failure in the Southern Uplands. The RSF at Capel Fell near Moffat Dale (Fig. 27.7d) is a partly collapsed rock-slope deformation, where movement was probably focused along the weak mudstone beds of the Moffat Shale Group.

Most postglacial RSFs in the Southern Uplands are paraglacial landforms that represent reduction of slopes to a state of critical stability as a result of internal rock damage caused by successive episodes of glacial loading, erosion and unloading, together with the effects hydromechanical and thermomechanical fatigue induced during deglaciation. Failure may have occurred during or immediately after final deglaciation, or centuries or millennia later as a result of progressive rock-mass weakening due to fracture of internal rock bridges, reduction of friction angles to residual values through strain deformation along joints, and possibly seismic fatigue (Chap. 14).

27.7 Holocene Fluvial Landforms

Numerous small alluvial fans and debris cones occur at the mouths of steep tributary valleys throughout the Southern Uplands. These represent gully erosion of sediment on valley-side slopes and its subsequent deposition along the flanks of major valleys (Fig. 27.8a). Radiocarbon dates obtained from organic soils or peat layers under or within the fan or cone deposits at several sites in the Tweedsmuir Hills have yielded Late Holocene ages (<4.2 ka), leading Foster et al. (2008) to link sediment accumulation to phases of enhanced gully erosion following cycles of settlement expansion and consequent land-use change in upland areas. Woodland clearance, burning and grazing pressure probably represent the most important factors preconditioning

Fig. 27.7 Rock-slope failures in the Southern Uplands. **a** Arrested slide at the margin of a col channel, Deuchar Law, Tweedsmuir Hills. **b** Collapse of a lava scarp and extended debris runout at Arkleton Hill. **c** The Bell Craig rock-slope deformation above Moffat Dale. **d** Partly collapsed rock-slope deformation on Capel Fell near Moffat Dale. (Images: **a**, **b**, **d** Google Earth™; **c** Colin Ballantyne)

sediment-mantled slopes to erosion, and they envisaged that such changes had lowered the stability of such slopes, rendering them vulnerable to gullying during extreme rainstorm events. For the Southern Uplands, they suggested that sporadic gullying events were initiated during the Late Bronze Age to Iron Age (~4.0–2.0 ka), but that the most intensive periods of gullying occurred at ~1.3–1.1 ka, 0.9–0.7 ka and after 0.55–0.45 ka. This interpretation is supported by dating evidence for similar fans and debris cones in both the uplands of NW England (Chiverrell et al. 2007) and the Scottish Highlands (Ballantyne 2019). Some fans, however, may simply represent the effects and after-effects of exceptional rainstorms: Tipping and Halliday (1994), for example, found no evidence for vegetation changes prior to the deposition of an alluvial fan in the upper Tweed valley at ~0.9 ka (Fig. 27.8b).

Although the floodplains of the Southern Uplands have experienced net postglacial incision and terrace formation, this was periodically (if locally) interrupted during the Holocene by periods of overbank sedimentation or floodplain aggradation. Cores through floodplain sediments in upper Annandale, for example, have revealed intercalated sequences of peat and overbank deposits. An upstream transect records overbank sedimentation at ~4.8–3.9 ka, but evidence for contemporaneous sedimentation is absent from cores retrieved farther downvalley. Renewed sedimentation, coeval with Iron Age settlement expansion, occurred between ~3.0 and ~2.1 ka, but is asynchronous across the floodplain; a further major aggradational event occurred at ~1.9–1.5 ka, and a final period of sediment accumulation at ~0.6 ka. Tipping et al. (1999) suggested that at least some of the sedimentation events recorded in upper Annandale may be linked to settlement expansion and associated land-use changes in the adjacent uplands causing episodes of sediment release into the river system. Studies at other sites in southern Scotland and northern England,

Fig. 27.8 **a** Late Holocene debris cone, Tweedsmuir Hills, showing erosion of slopes in the parent valley. **b** Hopecarton alluvial fan, upper Tweed valley. The large fan terrace was deposited at ~0.9 ka and subsequently incised. (Images: **a** Colin Ballantyne; **b** Google Earth™)

however, have produced different chronologies of Holocene floodplain aggradation. In the Bowmont catchment on the northern slopes of the Cheviots, for example, radiocarbon dating of roots at the base of terrace gravels yielded an age of 4.6–4.3 ka, which Tipping (1992) considered broadly synchronous with palynological evidence for Neolithic woodland clearance (~4.7–4.2 ka), but the younger terraces in this catchment appear to reflect floodplain aggradation and subsequent fluvial incision over the past 250 years (Tipping 1994). Thus although there is tantalizing evidence that episodes of Holocene floodplain aggradation may be linked to increased sediment release triggered by anthropogenic activity (deforestation, burning and livestock grazing), the available evidence suggests that this may be catchment specific.

27.8 Conclusion

The upland landscape of much of the Southern Uplands is dominated by rolling, fluvially dissected plateaux. The imprint of successive Pleistocene glaciations is represented by selective linear erosion of glacial troughs along trunk valleys; rock basins and glacially scoured bedrock are rare, cirques tend to be immature, and moraines are largely limited to those deposited by small cirque and outlet glaciers during the Loch Lomond Stade. Classic glacial upland landscapes are largely limited to the Galloway Hills and Tweedsmuir Hills. Frost-weathered regolith mantles plateaux and summits, suggesting that during the last glacial cycle high ground was occupied by cold-based ice that accomplished little or no erosion of the pre-existing regolith cover. Glacifluvial landforms are nonetheless abundant and include deep meltwater channels that were excavated in bedrock over multiple glacial cycles, hundreds of smaller meltwater channels cut in drift along valley sides during retreat of the last ice sheet and flights of outwash terraces along the margins of present floodplains. Lateglacial solifluction has smoothed valley-side slopes, locally forming lobes and terraces on high ground, and has contributed to the accumulation of periglacial valley fills in tributary valleys. Present-day periglacial activity is limited to small solifluction landforms on some mountains. Although rock-slope failures are mainly restricted to a few small clusters, these illustrate a range of contrasting types that is partly controlled by geological configuration. Holocene fluvial activity in some valleys appears to have been dominated by the effects of settlement expansion after ~4.0 ka, which resulted in anthropogenic land-use changes (woodland clearance, burning and grazing pressures). Such changes are thought to have destabilized upland slopes, causing gully erosion, deposition of small alluvial and colluvial fans and short-lived floodplain aggradation and overbank deposition of sediment.

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The Solway Lowlands and Coast

28

James D. Hansom and David J. A. Evans

Abstract

The Solway Lowlands and coast are characterised by a wide variety of glacial and glacifluvial landforms deposited by the British–Irish Ice Sheet during the last glaciation. These landforms display evidence of multiple ice-flow events and readvances of Scottish ice across the Solway Firth Lowlands. They include drumlins, crag-and-tail landforms, ice-contact outwash deposits, eskers and kame terraces. Reworking of terrestrial and submarine glacigenic deposits by waves and tidal currents has provided substantial volumes of sediment, enabling the formation of emerged beaches and estuarine flats (carse) and their modern counterparts of saltmarsh (merse), beaches and dunes. The saltmarshes and tidal sandflats of the inner Solway Firth are amongst the most extensive and well developed in Scotland and owe their development to eastward transport of sediment into the estuary by the dominance of flood tidal currents and a unidirectional wave climate.

Keywords

Subglacial bedforms • Drumlins • Glacial readvances • Glacigenic sediment • Sediment supply • Sea-level change • Emerged marine surfaces • Sandflats • Saltmarsh

28.1 Introduction

The Solway Firth extends ~130 km eastwards from its mouth between the Mull of Galloway on the southern coast of Scotland and St Bees Head in England to the estuaries of the River Esk and River Eden near the border between the two countries (Fig. 28.1a). A complex assemblage of subglacial and glacifluvial landforms makes this area critical to reconstructions of the evolution of the British–Irish Ice Sheet during the last (Late Devensian) glaciation; in particular, the eastern Solway Lowlands contain type examples of superimposed drumlins and ribbed moraine that are diagnostic of overprinted ice-sheet flowsets relating to multiple ice-flow events. Readvances of Scottish ice across the Solway Firth deposited significant landforms whose sediments became the source material for subsequent beach building over the Lateglacial and Holocene periods as relative sea level fluctuated. Bay-head beaches formed along the rocky, inlet-fretted north coast, with fringing beaches developing along the south coast. Dominant southwesterly waves have resulted in sediment transport into the Solway and accumulation along its inner coast, which is particularly notable for extensive tidal sandflat and saltmarsh development. The estuarine saltmarshes of the Solway (locally called *merse*) cover 36.16 km^2 (Pye and French 1993), accounting for almost 8% of the area of saltmarsh in Britain.

28.2 Geological and Physiographic Context

The Solway Lowlands form a low-lying (<100 m above sea level) basin open to the Irish Sea in the west and rising northwards into the lower valleys of the Scottish Rivers Nith, Annan and Esk and southwards towards the uplands of Cumbria. The head of the Solway Firth itself lies in England and is dominated by the estuary of the River Eden, which drains the eastern Lake District (Cumbria) and the western Pennines via the Vale of Eden (Fig. 28.1b). To the east, the

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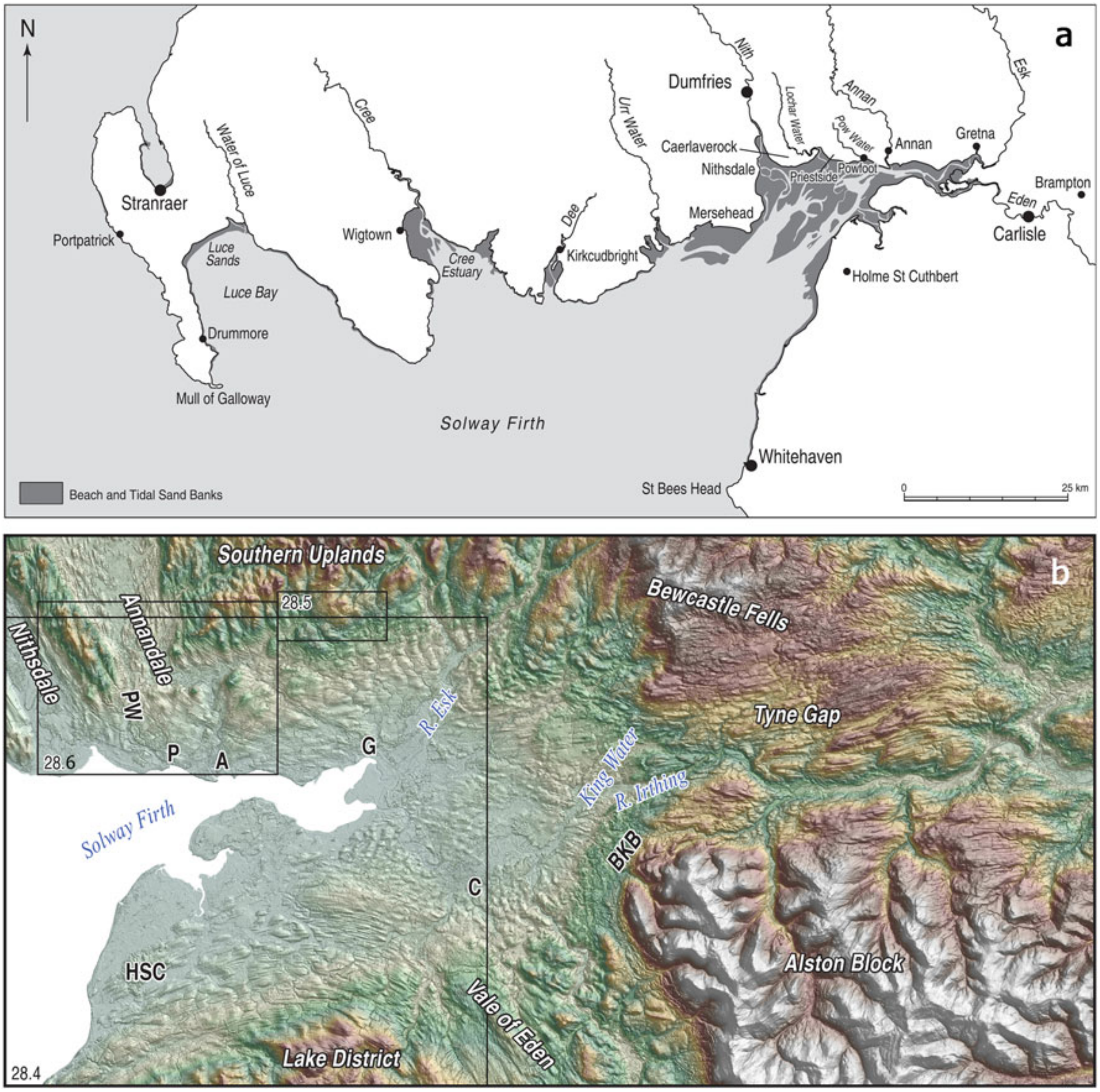

Fig. 28.1 **a** The Solway Lowlands and coast: coastal landforms and key locations mentioned in the text. **b** Digital elevation model showing physiography and glacial streamlining (drumlins) of the inner Solway Lowlands and surrounding region. Boxes outline areas of later figures. A—Annan; BKB—Brampton kame belt; G—Gretna; HSC—Holme St. Cuthbert; C—Carlisle; P—Powfoot. (NEXTMap Britain data from Intermap technologies Inc., provided courtesy of NERC via the NERC Earth Observation Data Centre)

Tyne Gap forms a corridor that separates the northern Pennines from the Cheviot Hills.

The inner Solway Lowlands form the Northumberland Trough, which is underlain by mudstones, siltstones, limestones and sandstones of Triassic and Jurassic age. To the north and east, these give way to a range of Carboniferous and Silurian sedimentary rocks. The Northumberland Trough served as a major ice-stream drainage pathway for the last (Late Devensian) British–Irish Ice Sheet ($\sim$35–14 ka) and probably during successive earlier periods of ice-sheet glaciation. The northern indented shore of the Solway Firth is underlain by Silurian greywackes and Permian sandstones; west of the River Nith these are intruded by the Criffel granite intrusion, which forms the only high (569 m) ground in the Solway region (McMillan et al. 2011). Along the north shore, bedrock structure and lithology and their influence on glacial modification of the landscape have resulted in a rocky and craggy coast; the indented bays of the western north shore and the low and more sheltered shore within the inner firth have provided ideal conditions for sandflat and saltmarsh development.

28.3 The Impact of Glaciation

The last (Late Devensian) British–Irish Ice Sheet was highly dynamic, exhibiting a temporal and spatial complexity that is outstandingly exemplified over the Solway Lowlands, where

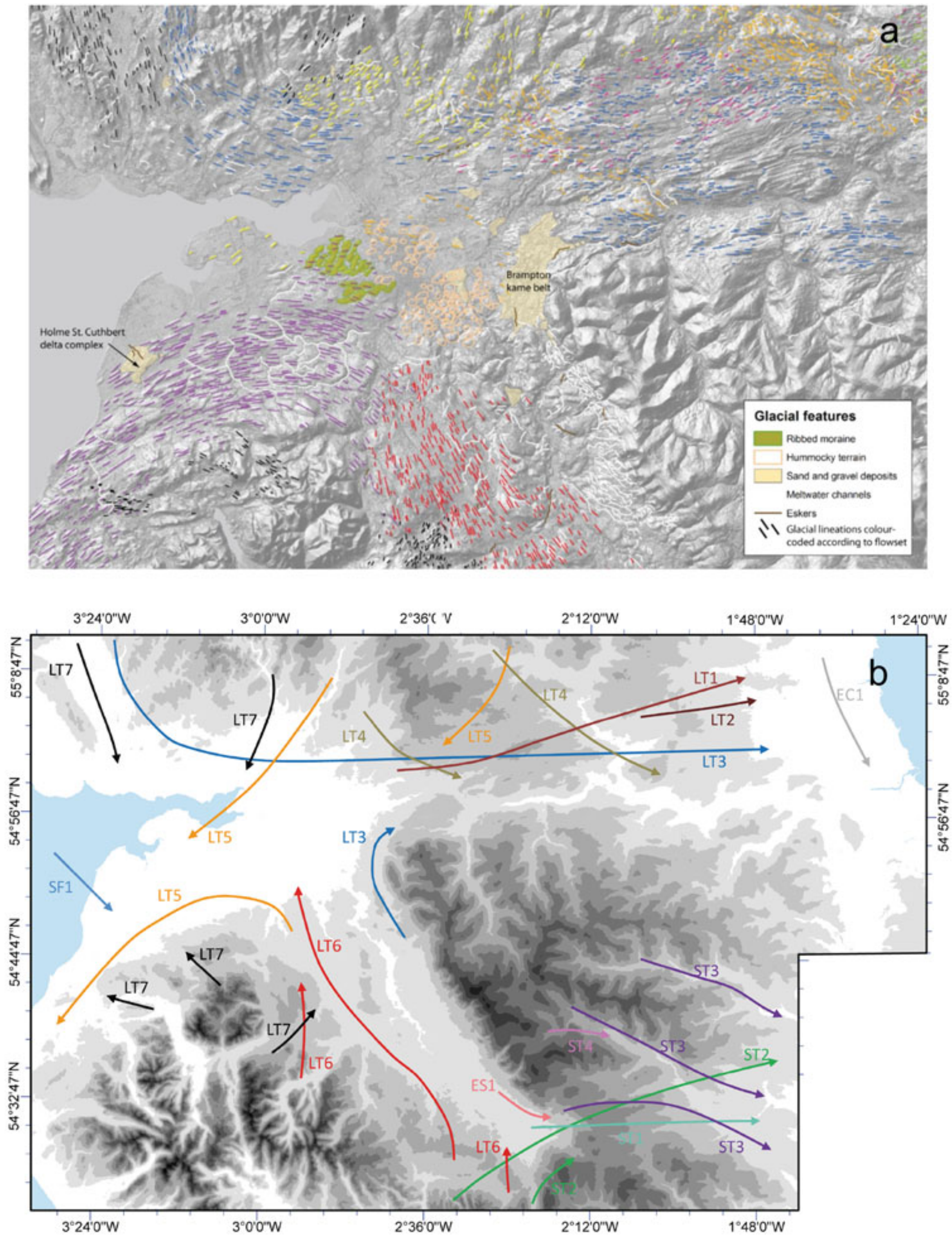

Fig. 28.2 **a** Solway Lowlands glacial geomorphology mapped onto NEXTMap imagery (From Livingstone et al. 2010a). **b** generalised ice-flow phases during the last (Late Devensian) ice-sheet glaciation. Flowsets are numbered sequentially according to cross-cutting relationships (From Livingstone et al. 2008). (NEXTMap Britain data from Intermap technologies Inc., provided courtesy of NERC via the NERC Earth Observation Data Centre)

convergence of major ice streams from regional upland dispersal centres occurred (Evans et al. 2009; Fig. 28.2). These ice streams shifted and even reversed their flow directions in response to migrating ice divides over a single glacial cycle, and it is likely that similar changes in ice-flow directions occurred during previous periods of ice-sheet glaciation (Chap. 4). This complexity is manifested in the regional erratic dispersal pathways and landform-sediment record of the Solway Lowlands from Galloway in the west to the Eden valley in the east. This area contains type examples of classic landforms diagnostic of multiple ice-flow events and overprinted ice-sheet flowsets, such as cross-cutting or superimposed subglacial landforms (Salt and Evans 2004; Livingstone et al. 2008; Fig. 28.2).

Major ice-marginal oscillations occurred during ice-sheet recession in the region, the most prominent of which were readvances of Scottish-sourced ice across the Solway Firth into northern England and the Irish Sea Basin. The earliest (>19.3 ka) was the Gosforth Oscillation, recorded on the Solway Lowlands by glacilacustrine deposits sandwiched between two tills (Livingstone et al. 2010c, 2012, 2015). A later readvance, known as the Scottish Readvance, involved extension of a glacier lobe into the northern Irish Sea Basin at ~19.3–18.3 ka (Livingstone et al. 2010a, 2012; Chiverrell et al. 2018). The Solway Lowlands were finally deglaciated before 16.4–15.7 ka (Livingstone et al. 2015), and by ~15.0 ka, remnants of the last ice sheet had retreated to upland source areas (Ballantyne and Small 2019; Chaps. 4 and 27).

28.4 The Legacy of Sea-Level Change

Sea level controls the position and configuration of the shoreline and the offshore supply of beach-building sediment. Relative sea level (RSL) in the Solway Firth was influenced by the interaction between local glacio-isostatic adjustment (GIA) and global (eustatic) sea-level change. During and immediately after deglaciation, high RSLs resulted in the deposition of beach and estuarine sediments along the Solway coast. As GIA subsequently outpaced eustatic sea-level rise, these Lateglacial emerged marine deposits now lie at or slightly above ~10 m ordnance datum (OD) along the west Solway coast (Wells 1999; McMillan 2000, 2011); for example, gravelly Lateglacial emerged beaches lie between 11 and 13 m OD near Southerness, west of the Nith estuary (McMillan et al. 2011). However, over the Lateglacial period GIA resulted in a fall in relative sea level; for example, near Caerlaverock on the Nith estuary, RSL fell from ~11 m OD to less than ~1 m OD during this time (Haggart 1999).

A second set of emerged marine deposits formed along the Solway north shore following Early Holocene sea-level rise. At the mouth of the Cree valley in the outer Solway Firth, relative sea level rose rapidly from about −1 m OD at ~9.5 ka to the Holocene marine limit at 7.7–10.3 m OD by ~7 ka (Smith et al. 2019). Elsewhere along the north shore of the firth, Holocene emerged beach gravels, tidal flats and saltmarsh deposits locally reach ~9 m OD and are intermittently backed by abandoned cliffs (Tipping 1999); those on the south shore are ~2 m lower (Lloyd et al. 1999), due to a southward decline in the rate of GIA (Smith et al. 2019). During the Middle to Late Holocene, shoreface beach deposits of sand and gravel and extensive estuarine (carse) deposits were formed. In the west at Luce Bay, for example, the estuarine shoreline had extended several kilometres inland by ~6.4 ka (Figs. 28.1a and 28.3), its emerged seaward edge being locally buried and enclosed by gravel beach-ridge barriers (Smith et al. 2020). Summarising modelling and empirical data for the Solway, Shennan et al. (2018) depict a slow fall in RSL over the Middle and Late Holocene towards present levels. Smith et al. (2020) suggest that this may have occurred in stages, favouring the development of dunes that now cover some emerged beaches and carse deposits, for example, at Luce Bay (Fig. 28.3) and Southerness (McMillan et al. 2011). Recent erosion of coastal landforms in the Solway, however, suggests that sea-level fall has now ended (Pye and French 1993; Firth et al. 2000). Tide-gauge records for the Mull of Galloway (Portpatrick) indicate a sea level rise of ~1.9 mm a^{-1} between 1968 and 2010 and ~4.0 mm a^{-1} between 1992 and 2010. Whether this short-term increase is sustained remains to be seen, but the UKCP18 high emission scenario (RCP 8.5: the most likely scenario based on current trends; Horsburgh et al. 2020), estimates current (2020) RSL rise at Mull of Galloway (Drummore) at ~3 mm a^{-1} above the 1980–2000 average (Palmer et al. 2018). By 2100, the UKCP18 (RCP 8.5) anticipates an average sea-level rise of ~0.6 m in the Solway area, at a projected rate of ~9 ± 5 mm a^{-1}.

28.5 Glacial Landforms

The Solway Lowlands are remarkable for subglacial bedforms that display localised patterns of overprinting, including elongated drumlins, crag-and-tail landforms, hummocky terrain and ribbed moraine (Livingstone et al. 2008; Figs. 28.1b and 28.2). Although these features dominate the glacial geomorphology of the area, elongate spreads of glacifluvial sediment-landform associations and meltwater channels are also well developed within the lower catchments of the Rivers Nith and Annan.

The cores of subglacial bedforms include bedrock, partial bedrock (crag-and-tail features), glacitectonised sediment (glacitectonite) and till. Elongation ratios mainly fall within the range 2:1–3:1, indicative of formerly low ice-flow velocities. The highest elongation ratios (up to 12:1) occur south of the Solway Firth and relate to late stage, fast ice flow into the Irish Sea Basin (Fig. 28.4).

The evidence provided by overprinted drumlins has been used to compile the flowset sequencing depicted in Fig. 28.2. The most prominent drumlin alignments were created during two different flow phases: LT3 on the north side of the Lowlands and LT5 on the south side. The earlier ice-flow phase (LT3) is represented in the inner Solway area by WNW-ESE orientated drumlins and upland crag-and-tail features indicative of a curving pattern of ice dispersal from a northwesterly source in the Southern Uplands (Fig. 28.2b). The trajectory of the LT3 flow phase through the Tyne Gap suggests that ice flow was diverted eastwards by ice nourished in the English Lake District and northern Pennines. Livingstone et al. (2012) concluded that the LT3 phase occurred during the maximum expansion of the last ice sheet, at which time an ice divide straddled the outer Solway Firth, with ice from the Galloway Hills feeding the Irish Sea Ice Stream but ice from the western Southern Uplands being diverted through the Tyne Gap (Chaps. 4 and 27).

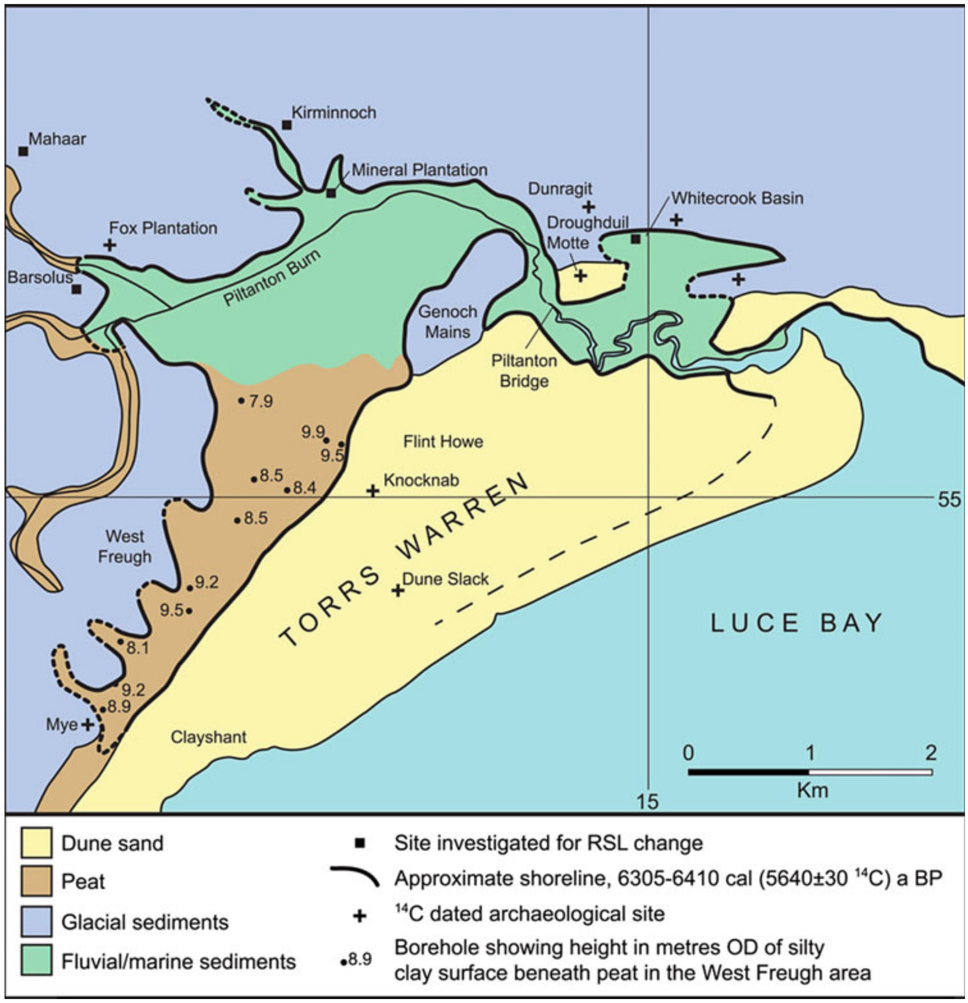

Fig. 28.3 Luce Bay, showing the approximate shoreline position at 6305–6410 cal BP and borehole altitudes for the height of emerged carse surfaces. (© Smith et al. 2020, Creative Commons Licence CC-BY 4.0)

Overprinting of these drumlins by the NE–SW aligned drumlins of flow phase LT5 records a subsequent ice flow reversal, when the Southern Uplands ice was drawn down into the Irish Sea Basin (Fig. 28.2).

An important subglacial landform assemblage on the Solway Lowlands is an area of ribbed and hummocky terrain in the lower Eden valley (Fig. 28.4). The ribbed terrain (moraine), centred on Burgh by Sands, forms an area of ridges aligned NNE–SSW, transverse to the former direction of ice flow, that grade eastward into increasingly less elongate mounds or ovoid forms, and eventually into hummocky terrain where elongation ratios approach unity. Together the hummocky and ribbed terrain constitute a continuum where pre-existing subglacial bedforms have been glacially overrun and streamlined by ice following flowset LT5 (Livingstone et al. 2008). No similar assemblage, however, occurs in the Scottish sector of the Solway Lowlands.

On the south side of the Solway Firth, the limit of the Scottish Readvance is demarcated by an ice-contact delta at Holme St. Cuthbert (Huddart 1991; Livingstone 2010a; Figs. 28.1b and 28.4). North of Gretna, extensive areas of hummocky terrain contain glacifluvial sands and gravels, including a pitted, flat-topped mound at Barnglieshead, interpreted as an outwash delta-esker complex due to its linkage to the Solwaybank Esker (McMillan et al. 2011; Fig. 28.5).

Glacifluvial landforms are also well developed on the north shore of the inner Solway Firth around Powfoot (Fig. 28.1b), where they have been variously interpreted as a large kamiform moraine belt (Charlesworth 1926a, b), recessional push moraines developed in ice-contact outwash (Huddart 1999) or eskers and kame terraces (McMillan et al. 2011), but none of these interpretations is mutually exclusive. Eskers in the area are generally fragmented, but a prominent ridge named the Hurkledale Esker forms the raised bed of the Annan to Dumfries railway line (Fig. 28.6). Corridors of kames, kettle holes and eskers also occupy lower Nithsdale, lower Annandale and the valley of the lower Kirtle Water around Gretna (Goodlet 1970; Cameron 1977). These are typical of valley-floor glacifluvial complexes developed within and over downwasting ice after it had become confined by topography (Huddart 1999), classically illustrated by the Brampton kame belt in Cumbria

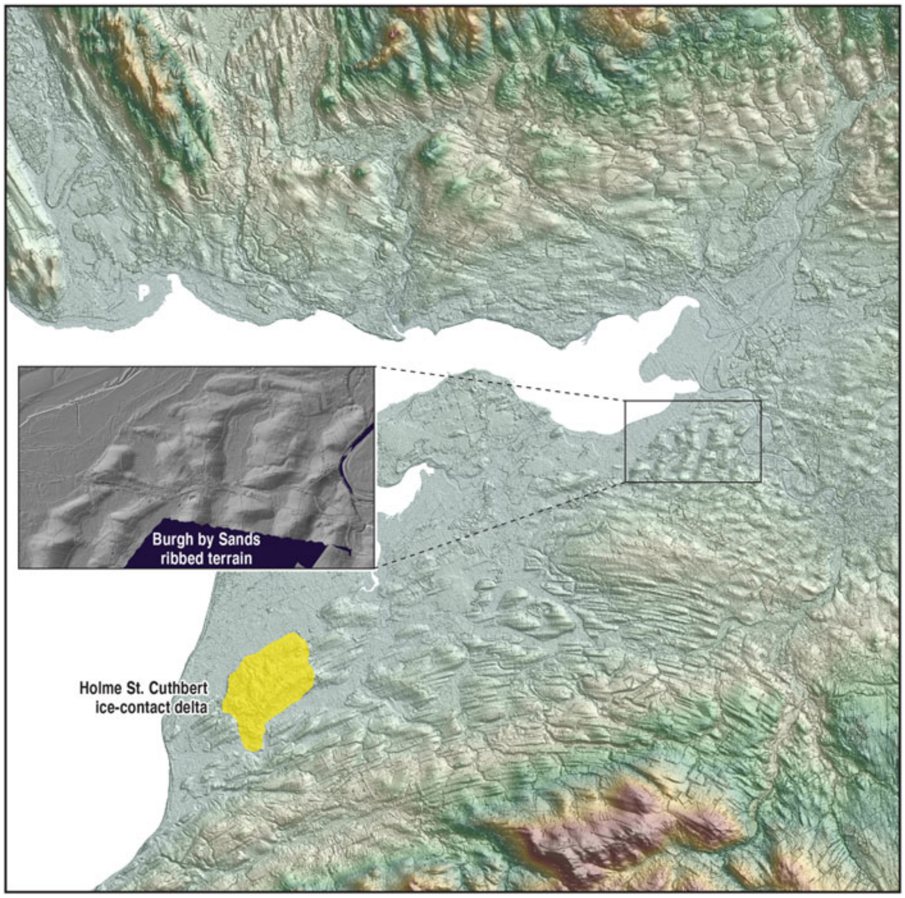

Fig. 28.4 NEXTMap imagery of the distribution of subglacial landforms around the Solway Lowlands. Inset is a LiDAR image of the ribbed moraine around the village of Burgh by Sands. (NEXTMap Britain data from Intermap technologies Inc., provided courtesy of NERC via the NERC Earth Observation Data Centre)

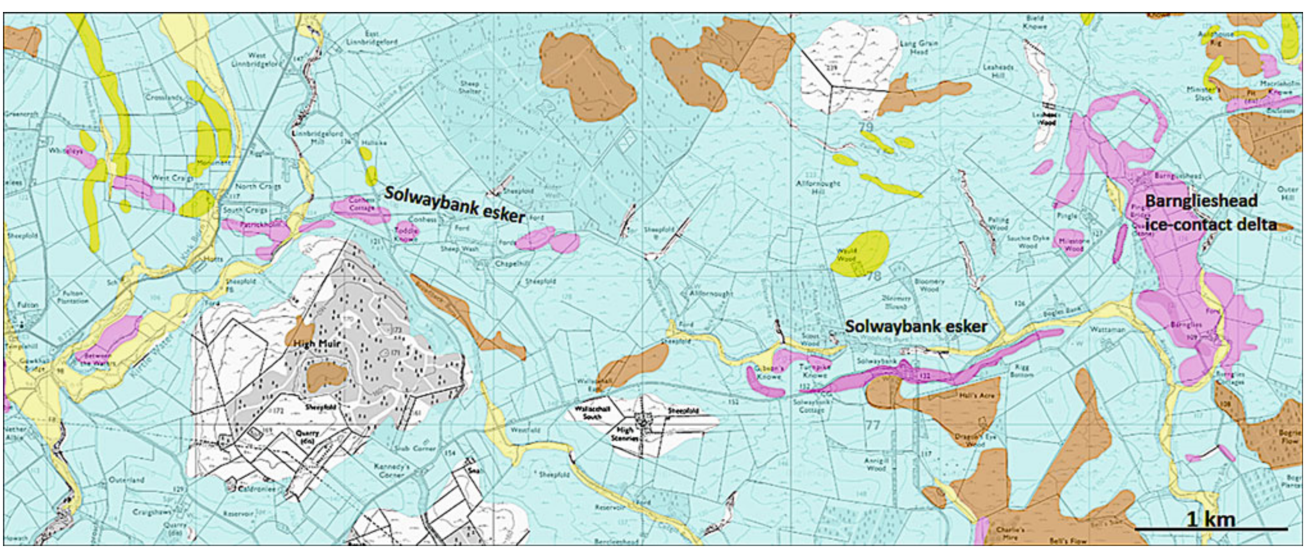

Fig. 28.5 Quaternary geology map (EDINA Digimap extract) of the area north of Gretna, showing the elongate assemblage of glacifluvial deposits (pink) forming the Solwaybank Esker and the Barnglieshead delta. Other surface deposits include glacifluvial sediments (green), till (blue), alluvium (yellow) and peat (brown). (Permit number CP20/090 BGS Geological Map data © UKRI. All rights reserved, sourced via BGS Digital Data under the Edina Licence)

(Livingstone et al. 2010b). McMillan et al. (2011) also identified an ice-contact face on the western side of the Powfoot landform assemblage, indicating that an ice margin occupied that area during glacifluvial deposition. Meltwater channels on the surrounding valley slopes and interfluves (Fig. 28.6) record successive positions of the lateral margins of a downwasting ice lobe as it retreated westwards towards lower Nithsdale.

28.6 Coastal Landforms

Almost all the modern coastal depositional systems in the Solway Firth lie seaward of abandoned Holocene cliffs or emerged marine deposits. Although the detail may differ slightly, the morphogenetic environment today provides a useful insight into coastal landform development over the Holocene. The Solway is macrotidal, and its slower ebb currents transport less sediment than the stronger flood tides from the west, thereby accentuating net transport of sediment into the firth. Dominant southwest waves and refracted Atlantic swell also render the Solway Firth a sediment trap, particularly within the inner firth (Ramsay and Brampton 2000), with little seaward sediment loss (Perkins and Williams 1966). As a result, the Solway saltmarshes are dominated by influx of marine sand with an unusually low (<4%) average clay content; average clay contents of British saltmarshes are commonly >30% and often >65% (Marshall 1962). Within the sand-dominated Solway, the elevation at which pioneer saltmarsh vegetation can establish lies higher in the tidal range than on muddy coasts (Gray 1992), possibly because sandy substrates tend to occur in more exposed areas (Pye and French 1993). Sand is more mobile than mud, and mechanical removal of seedlings by waves or tides may occur before adequate rooting has developed (Balke et al. 2014). Compared with saltmarshes farther south, areas of pioneer vegetation are not extensive; the Solway saltmarshes tend to be dominated by Atlantic marsh communities commonly found at higher elevations elsewhere (Haynes 2016).

Several locations are critical to understanding coastal landform evolution in the Solway: of these, Luce Sands, the Cree estuary, Mersehead, Caerlaverock and Priestside are located on the north (Scottish) shore and are discussed below. Extensive ($\sim$5.6 km^2) areas of saltmarsh also occur at the head of the Solway Firth in England, between and adjacent to the mouths of the Rivers Esk and Eden (Burd 1989). These experienced net expansion at an average rate of $\sim$15 ha a^{-1} between 1946 and 1975 (Rowe 1978) but by 1993 were contracting in area (Pye and French 1993), likely due to rising RSL and submergence replacing long-term emergence.

28.6.1 Luce Sands

Luce Sands, at the head of Luce Bay in the outer Solway Firth (Fig. 28.1), represents the largest beach and dune system in southwest Scotland. The head of the bay has a wide, sandy beach backed by sand dunes, emerged gravel ridges and estuarine surfaces and, farther inland, glacial deposits. A prominent abandoned Holocene cliff at $\sim$10 m OD, cut in till and glacifluvial sediments, frames the extremities of the bay, between which lie glacial sediments incised by meltwater channels and emerged estuarine sediments capped by peat. Coast-parallel emerged gravel ridges occur at 10–12 m OD in the west, with patches of gravel

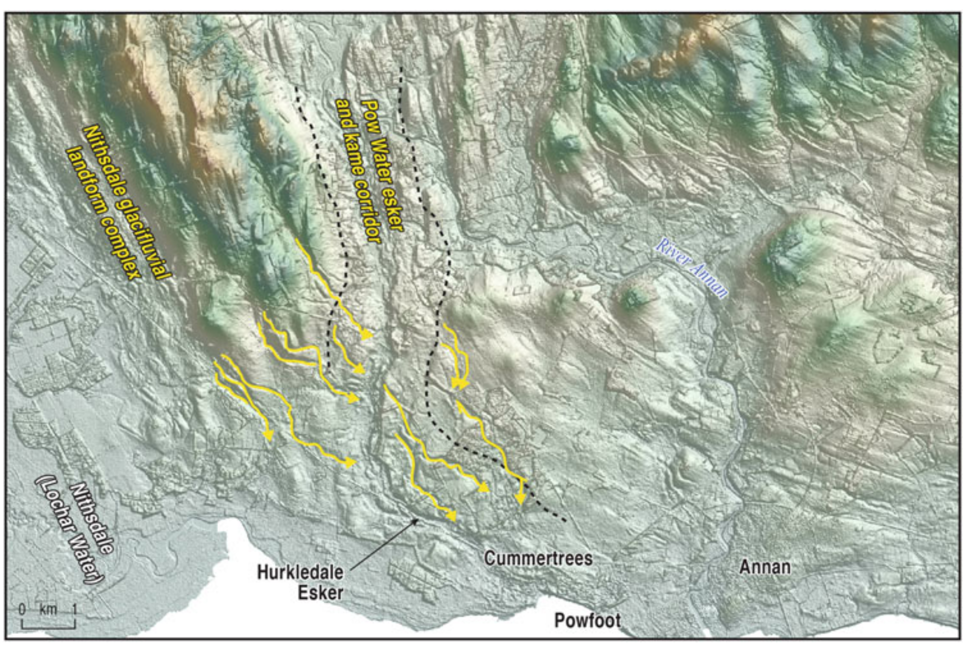

Fig. 28.6 NEXTMap imagery showing glacifluvial landforms on the north side of the Solway Lowlands. The Pow Water corridor of eskers and kames is demarcated by black-dashed lines, with major meltwater channels highlighted by yellow arrows. The subglacial streamlined terrain shows curving southeasterly ice flow from the Southern Uplands into the inner Solway Lowlands. (NEXTMap Britain data from Intermap technologies Inc., provided courtesy of NERC via the NERC Earth Observation Data Centre)

ridges within the sand dunes (Single and Hansom 1994). Smith et al. (2020) identified a large former marine basin that extends inland in the east along the lowlands of the Piltanton Burn (Fig. 28.3), where radiocarbon dating of peat interdigitated with estuarine sediment indicates a Middle Holocene (~6.3 ka) RSL highstand at ~10 m OD, with RSL falling intermittently thereafter. Present processes provide an insight into Holocene depositional conditions (Single and Hansom 1994). The wave climate of Luce Bay is unidirectional, driving sediment northwards across a shallow bay floored by unconsolidated glacigenic sand and gravel, and is augmented by northwards flood-tide sediment transport; the head of Luce Bay is therefore an effective sediment trap (Hansom 2003). Such onshore sediment transport by waves and tidal currents was probably important over most of the Holocene, with inland glacigenic sources providing additional sediment to the coast. The history of long-term progradation of Luce Sands is atypical of beaches in Scotland, many of which have experienced reduction of sediment supply as offshore sediment sources became progressively depleted.

28.6.2 The Cree Estuary

Landward of extensive tidal sandflats and saltmarsh, extensive emerged estuarine surfaces characterise much of the Cree estuary lowland (Fig. 28.7). Estuarine sedimentation at this location occurred over much of the Holocene, and most modern saltmarsh is backed by emerged estuarine carse up to 3.5 km wide that reaches an elevation of 7–10 m OD at its inland margin. Cores retrieved from the carse deposits contain at least 9 m of fine-grained sediments capped, at –2 to 0 m OD, by a thin buried peat layer, itself overlain by up to 9 m of estuarine sediments deposited as RSL rose to its Holocene highstand; these now form the modern carse surface (Firth et al. 2000). Subsequent RSL fall was not uniform, as indicated by younger peats within the carse sediments. A lower emerged shoreline formed during a Late Holocene stillstand or transgression, after which RSL in the Cree estuary and along the rest of the northern Solway coastline fell to present levels. The RSL fall has been accompanied by the formation of fringing saltmarsh, favoured by the shelter of flanking headlands and a unidirectional southerly wave regime. Some of the saltmarsh is relatively young; for example, the marsh on the west side of the estuary (Fig. 28.8a) mainly developed in the nineteenth century and near Wigtown at least 5 km^2 was artificially created for grazing around AD 1875.

28.6.3 Mersehead, Caerlaverock and Priestside

Within the inner Solway, Holocene emerged marine deposits and extensive modern saltmarsh are well developed along both north and south shores (Cressey et al. 1998; Tipping 1999), notably at Mersehead, Caerlaverock and Priestside, close to the Nith estuary on the north shore (Fig. 28.1a). West of the Nith estuary at Mersehead, tidal sandflat, beach and dunes extend for ~10 km along the coast. Former landward migration of the beach and dune system is evidenced by freshwater peat exposed on the shoreface (Fig. 28.8b) and by exposed networks of land drains extending onto the tidal flats beyond. Landward of the dunes, emerged carse terraces decline westwards into areas of enclosed freshwater marsh and saltmarsh that have developed behind the shelter of a prominent spit, largely composed of shells (Fig. 28.8c). On the saltmarsh areas, recent sedimentation rates locally average 27 mm a^{-1} (Harvey and Allan 1998; Harvey et al. 2007).

East of the River Nith, Caerlaverock Merse and Priestside Merse jointly comprise the largest saltmarsh area in Scotland at ~7.2 km^2, both dominated by a *Puccinellia–Festuca–Glaux* sward with small stands of *Phragmites* reeds (Haynes 2016), although areas of pioneer vegetation (*Salicornia*) are also present (Fig. 28.8d). Caerlaverock Merse extends for 8 km along the coast and is up to 1 km wide at its eastern end, its landward boundary being marked by a terraced bluff cut into Middle Holocene emerged carse deposits at 9 m OD. Seaward of these, and sitting on top of a lower carse surface at 5 m OD, four well-defined emerged gravel beach ridges formed in the twelfth and thirteenth centuries (Tipping and Adams 2007). The presence of granodiorite gravel within the ridges, sourced from west of the River Nith, suggests that the ridges were emplaced by extreme storms during the Medieval period and presumably before the formation of the extensive tidal sandflats that now reduce wave action and limit gravel movement.

The seaward edges of Caerlaverock Merse and Priestside Merse locally form a ~1.5 m high, often stepped, terrace edge cut into the fine saltmarsh sediment. After onshore storm wave events, the terrace surface shows extensive areas of stripped vegetation that extend up to 10 m inland, the junction between the sandflat and saltmarsh edge being adorned by blocks of eroded saltmarsh sediment (Fig. 28.8e). Sediment released by erosion during such events is transported landward, contributing to rapid accretion of the saltmarsh surface. At Priestside in March 2008, for example, a storm produced up to 60 mm of deposition spread across tens of metres inland from the eroding saltmarsh edge (Fig. 28.8f). Progressive accretion of the tidal sandflat involves colonisation by lower and middle marsh communities that locally appear to be replacing the pioneer communities, such as *Salicornia*, that were formerly extensive here (Haynes 2016). Surveys of recent changes of the saltmarsh extent at Caerlaverock have shown that marked erosion of its western (Nith) extremity is often counterbalanced by net accretion along parts of its seaward edge, leading to

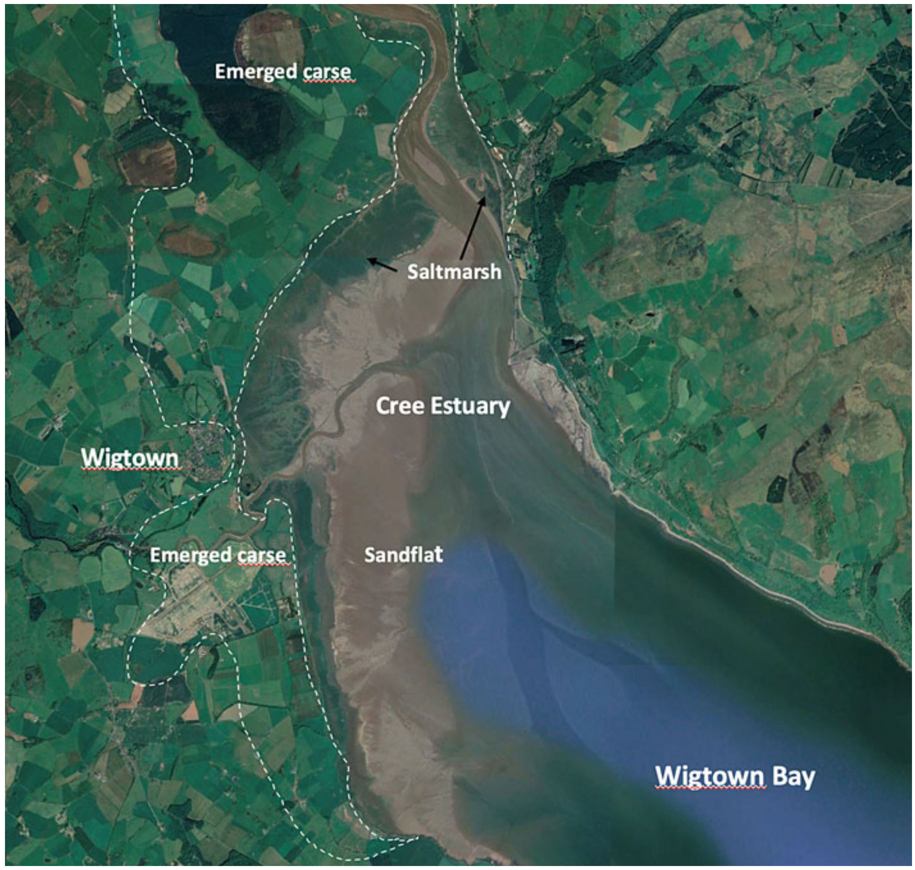

Fig. 28.7 Extensive emerged carse surfaces flank the Cree estuary, with fringing saltmarsh along the river channel and landward of tidal flats at the estuary mouth. (Image: Google Earth™)

saltmarsh extension eastward (Harvey 2000). This trend partly mirrors the migration of high water mark at Caerlaverock, which advanced seaward between the 1890s and 1970s and subsequently retreated inland in the west, but still lies seaward of its 1890 position in the east (Hansom et al. 2017). It seems likely that cyclic changes in the positions of the main outflow channels of the River Nith and the nearby Lochar Water determine changes in the accretion or erosion patterns of the adjacent tidal saltmarsh as well as the position of high water mark along this part of the coast.

28.7 Conclusion

The effect of the British–Irish Ice Sheet during the last glaciation imparted the Solway Lowlands with a wide variety of glacial and glacifluvial landforms that demonstrate multiple ice-flow events and readvances of Scottish ice across the area. These events are recorded by the subglacial bedforms of the area, which include elongated drumlins with localised patterns of overprinting, together with crag-and-tail landforms, hummocky terrain, and, more locally, ribbed moraines. Glacifluvial landforms, including ice-contact outwash landforms, eskers and kame terraces, are well developed in the inner Solway Firth area. Following deglaciation, extensive areas of terrestrial and submarine glacigenic sediment were reworked by waves and currents during periods of postglacial sea-level rise, permitting the development of beaches and estuarine flats (carse) that formed emergent features as sea level subsequently fell. The widespread development of emerged coastal landforms and their extensive modern equivalents of tidal sandflats and saltmarshes results from sediment being trapped within the Solway due to tidal currents and a unidirectional wave climate, whose overall character appears to have changed little during the Holocene.

Fig. 28.8 **a** View northeast over saltmarsh creeks on the west side of the Cree estuary; much of the saltmarsh formed in the nineteenth century. **b** View east along the Mersehead foreshore, showing freshwater peat beds exhumed on the foreshore due to landward migration of the beach and dunes. **c** Saltmarsh at Mersehead has formed in the shelter provided by a well-developed shelly spit; spit migration occasionally isolates areas of saltmarsh to seaward. **d** Areas of pioneer vegetation, such as occurs here at Caerlaverock, are relatively few and sparsely distributed. **e** Storm stripping of the saltmarsh edge at Priestside; the lower edge is strewn with eroded blocks of saltmarsh sediment. **f** Fine sediment released during a 2008 storm event at Priestside resulted in widespread sediment accumulation across the upper vegetated saltmarsh surface, raising it by up to 60 mm. (Images: **a** Shaun Bithell; **b** © Christine Johnstone (CC BY-SA 2.0); **c–e**, Jim Hansom; **f** Bruce Robertson)

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Part III

Geoheritage

29 Scotland's Geomorphological Heritage and Its Conservation

John E. Gordon, Vanessa Brazier, James D. Hansom, and Alan Werritty

Abstract

Geomorphological features and processes contribute significantly to the geodiversity and geoheritage of Scotland. Key sites identified through the Geological Conservation Review are mostly protected as Sites of Special Scientific Interest. These sites represent the variety of geological, glacial, periglacial, fluvial, coastal, mass-movement and karst features that distinguish the Scottish landscape. Scotland's geodiversity and geoheritage have additional value for educational, aesthetic, cultural and ecological reasons and in the delivery of ecosystem services. Greater recognition of these values is fundamental to more holistic approaches to nature conservation, the development of nature-based solutions to societal challenges and connecting people and nature.

Keywords

Geodiversity • Geoheritage • Geoconservation • Geological Conservation Review • Sites of Special Scientific Interest • Geoparks • Geotourism

29.1 Introduction

As outlined in the preceding chapters of this book, Scotland has a remarkable diversity of landforms and landscapes arising from global tectonics, a long and varied geological history, changing climates and the effects of multiple glacial-interglacial episodes and relative sea-level changes during the Quaternary (2.59 Ma to the present). Scotland's geomorphological diversity includes mountain glacial and periglacial landforms and mass-movement features; lowland glacial erosional landforms and deposits; upland and lowland fluvial landforms; coastal landforms comprising emerged shorelines and active beaches, dunes, machair and rock-coast features; and local areas of karst and caves (Figs. 29.1 and 29.2). Offshore, a range of glacial landforms, submarine landslides and active seabed forms characterise the continental shelf. Geomorphology therefore forms a significant component of Scotland's geoheritage—those elements of geodiversity considered to have significant heritage value and therefore meriting conservation. Historically, in Scotland as elsewhere, emphasis has focused principally on the conservation of geoheritage through the identification and designation of key localities that are protected for their scientific value (Gordon et al. 2019). However, there is now increasing interest, both in Scotland and internationally, in additional intrinsic, cultural, aesthetic and

The original version of this chapter was revised: The error in "Neogene/Early Pleistocene fluvial gravels" has been corrected as "Palaeogene fluvial gravels" in Page "487". The correction to this chapter is availableat https://doi.org/10.1007/978-3-030-71246-4_30

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Fig. 29.1 Key locations mentioned in the text. (Map base by Eric Gaba, NordNordWest, Uwe Dedering from Wikimedia Commons CC BY-SA 3.0)

ecological values, recognising that geodiversity is part of natural diversity and contributes significantly to the benefits and ecosystem/geosystem services that the natural world provides (Gray 2013; Brilha et al. 2018). In turn, this is leading to advocacy for more integrated approaches to nature conservation that link natural and cultural landscapes, thereby emphasising the interconnections between 'people and nature', rather than 'nature for people' (Knudson et al. 2018; Gordon et al. 2019; Slaymaker et al. 2020). Consideration of geomorphology is fundamental to such approaches, insofar as it forms the physical template for many cultural and scenic landscapes in Scotland and elsewhere.

29.2 Geoconservation in Scotland

29.2.1 Sites of Special Scientific Interest and National Nature Reserves

Statutory geoconservation in Scotland began in the mid-twentieth century following the enactment in 1949 of the National Parks and Access to the Countryside Act, which made provision for the designation and statutory protection of nature reserves and Sites of Special Scientific Interest (SSSIs) in Great Britain, including those of geomorphological importance. Subsequently, the UK Wildlife and

Fig. 29.2 Examples of protected sites for geomorphology in Scotland representing a range of outstanding glacial, fluvial, coastal and mass-movement features. **a** Loch Lomond Readvance moraine and drift limit, An Teallach, Wester Ross. **b** Active river and coastal landforms of the lower River Spey and Spey Bay. **c** The granite cliff coast near the Bullers of Buchan, Aberdeenshire. **d** Part of the Quiraing landslide, Isle of Skye. (Images: **a**, **c**, **d** John Gordon; **b** © Peter Gordon Smith)

Countryside Act (1981) and the Nature Conservation (Scotland) Act 2004 strengthened the conservation management of SSSIs. SSSIs are selected and designated for the particular scientific importance of their natural features—flora, fauna, geology and geomorphology—which are protected under this legislation. National Nature Reserves (NNRs) have a broader purpose and are areas of land set aside for nature conservation and enjoyment.

The systematic assessment of key geoheritage sites in the 1950s and 1960s culminated in the Geological Conservation Review (GCR), which was initiated in 1977 to identify all the nationally important sites representing the geology and geomorphology of Great Britain (Ellis 2011). GCR sites are selected solely for their scientific interest, and most but not all are designated as SSSIs. Currently, 227 sites in Scotland are listed in the GCR database for their geomorphological and Quaternary (including stratigraphic) importance and represent 25% of the 901 Scottish GCR sites (Gordon et al. 2019; Table 29.1; Fig. 29.3). Sites are selected: (i) for their international importance; (ii) because they contain rare or exceptional features; and/or (iii) because they are representative of a particular feature, process or event considered fundamental to understanding the geological history and geomorphology of Great Britain. Several of Scotland's NNRs also showcase significant geomorphological features, including Glen Roy (Chap. 16), the Cairngorms (Chap. 18) and Forvie (Chap. 23).

29.2.2 Other Statutory and Non-statutory Protected Areas

As in England and Wales (Goudie and Migoń 2020), several other, sometimes overlapping, site designations and other non-statutory forms of recognition incorporate geomorphological features. A site assessment exercise similar to the GCR identified 35 key areas for geomorphology and Quaternary interests in the seas around Scotland (Brooks et al. 2013; Table 29.1), some of which are now incorporated within Nature Conservation Marine Protected Areas (Gordon et al. 2016). These areas are intended to protect nationally important marine wildlife, habitats, geology and

Table 29.1 Quaternary and geomorphology GCR sites and marine key geodiversity areas in Scotland

Thematic block	No. of sites/key areas[a]	Principal landforms and deposits	Sources
Terrestrial			
Quaternary geomorphology	142	Glacial, periglacial and glacifluvial landforms and deposits; emerged and buried shorelines; non-glacial weathering and slope forms; deposits representing Quaternary stratigraphy	Gordon and Sutherland (1993)
Fluvial geomorphology	29	Bedrock and alluvial landforms and process systems	Werritty and McEwen (1997)
Coastal geomorphology	41	Rock coast landforms and beach, dune, machair and saltmarsh systems	May and Hansom (2003)
Mass movements	13	Rock-slope failures, including rock-slope deformations, arrested rockslides and catastrophic rock avalanches	Cooper (2007)
Karst and caves	2	Caves and limestone pavement	Waltham et al. (1997)
Marine			
Marine Quaternary	11	Landforms and deposits associated with the last and earlier Scottish Ice Sheet(s)	Brooks et al. (2013)
Submarine mass movement	4	Submarine slides	Brooks et al. (2013)
Marine geomorphology of the Scottish deep-ocean seabed	5	Contourites, sediment drifts and erosional features associated with deep-ocean currents	Brooks et al. (2013)
Seabed fluid and gas seep	2	Pockmarks and sand volcanoes	Brooks et al. (2013)
Cenozoic structures of the Atlantic margin	2	Large structural blocks (seamounts) and mud diapirs	Brooks et al. (2013)
Marine geomorphology of the Scottish shelf seabed	6	Non-tropical shelf carbonate systems and bedforms	Brooks et al. (2013)
Coastal geomorphology of Scotland (offshore)	2	Submerged coastal landforms (cliffs, caves and rock platforms)	Brooks et al. (2013)
Biogenic structures of the Scottish seabed	3	Biogenic sediment mounds and cold-water coral reefs	Brooks et al. (2013)

[a]A list of the sites is included in Gordon et al. (2019 Supplementary material)

seabed landforms. Geomorphological features are also represented within World Heritage Sites (WHS) and National Parks. Three of Scotland's six World Heritage Sites have significant geomorphological interests: New Lanark WHS includes the Falls of Clyde gorge and waterfalls (Fig. 29.4a); the Old and New Towns of Edinburgh WHS includes Edinburgh Castle crag and tail (Chap. 26); and the St Kilda WHS includes spectacular rock coasts (Chap. 9). However, New Lanark and Edinburgh are solely cultural properties and the presence of geomorphological features was not a reason for their designation; St Kilda is a mixed property, and while the statement of Outstanding Universal Value under selection criterion (vii) specifically mentions the geological and geomorphological underpinning of its superlative scenery, it was not deemed to qualify under criterion (viii) for its geological and geomorphological features alone.

Scotland's two national parks, the Cairngorms National Park and the Loch Lomond and The Trossachs National Park, contain a number of geomorphological SSSIs and in particular are noted for their landforms and landscape-scale geomorphology (Goodenough et al. 2008; Barron et al. 2011; Chaps. 18 and 24). Scotland's national parks are required to conserve and enhance their natural and cultural heritage, promote sustainable use of their natural resources, enhance understanding and enjoyment of their special qualities by the public and foster sustainable economic and social development of their communities. In addition, most of Scotland's 40 designated National Scenic Areas (NSAs) incorporate many

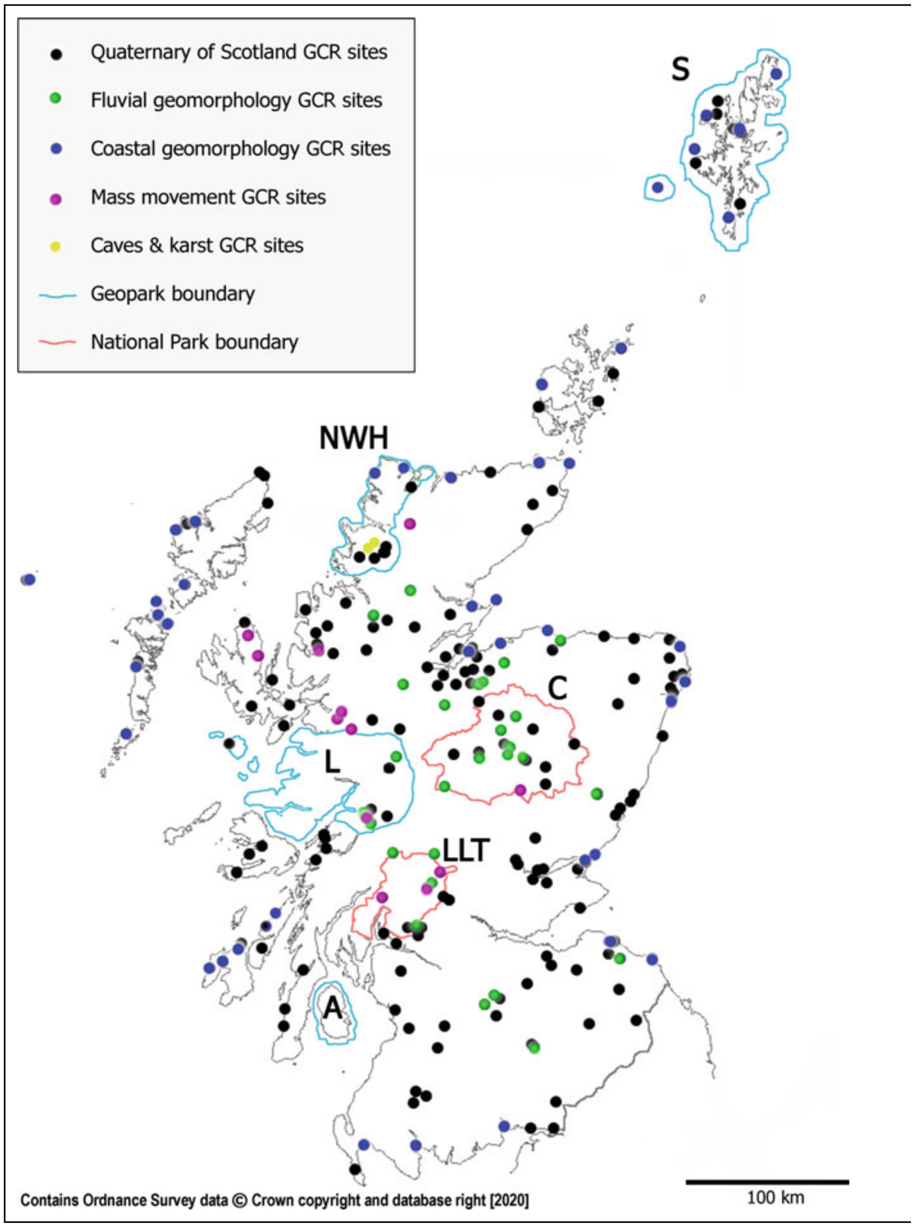

Fig. 29.3 Locations of GCR sites, geoparks and national parks in Scotland. A: Arran Geopark; L: Lochaber Geopark; NWH: North West Highlands UNESCO Global Geopark; S: Shetland UNESCO Global Geopark; C: Cairngorms National Park; LLT: Loch Lomond and The Trossachs National Park

coastal and mountain landscapes of outstanding geomorphological value (e.g. Wester Ross, Chap. 13; Ben Nevis and Glencoe, Chap. 17; and Upper Tweeddale, Chap. 27; https://www.nature.scot/national-scenic-areas-scotland-map-special-qualities). NSAs are safeguarded under Scottish Planning Policy, and although the designation is not a primary geoconservation mechanism, it nevertheless offers a potential means of addressing landform conservation at the landscape scale since geomorphology is an important, if under-recognised, factor underpinning the special qualities of these areas (Scottish Natural Heritage 2010).

At a regional level, Local Geoconservation Groups, local authorities and the British Geological Survey have undertaken audits to identify Local Geodiversity Sites (e.g. Arkley et al. 2011; Whitbread et al. 2015). These regionally important sites are selected using a broader range of criteria than for GCR sites or SSSIs; this may include their educational value, scientific interest, aesthetic value and historical significance. They are given discretionary protection by local authorities and the National Parks through the development planning system. These sites mostly cover south, central and eastern parts of Scotland. In the Lothians and Borders, for example, there are 36 sites that include landforms or Quaternary deposits (Edinburgh Geological Society 2020).

Scotland has four geoparks: in Shetland, the North West Highlands, Lochaber and Arran (Fig. 29.3). The first two are accredited UNESCO Global Geoparks. All include spectacular geomorphology (Chaps. 7, 10, 12, 13, 14, 16 and 17;

Fig. 29.4 Examples of the wider values of geodiversity and geoheritage. **a** The Falls of Clyde, a spectacular example of glacial drainage diversion and a significant visitor attraction since the late eighteenth century. The Falls lie within the buffer zone of the New Lanark World Heritage site. **b** The unique landscape of Assynt in the North West Highlands UNESCO Global Geopark has outstanding scenic value, supporting tourism and the rural economy. Grass and moss-heath communities are typical of the Torridonian sandstone mountains, whereas a complex of wet heath, blanket mire and open-water bodies occurs on the ice-scoured Lewisian gneiss basement. **c** Beach, dune and machair landforms of the Outer Hebrides support a rich variety of habitats for plants, animals and birds and benefit from traditional agricultural land use. At Tràigh Luskentyre and Corran Seilebost GCR site, part of the South Lewis, Harris and North Uist National Scenic Area, the habitats include shell-sand beach, sand dunes, calcareous machair grassland and saltmarsh, with moorland on adjacent rocky hillsides. **d** The diverse geomorphology of the Cairngorms supports habitats ranging from those of the alpine plateau environment to mosaics of wet and dry areas on lowland glacifluvial landforms and kettle holes. **e** Cover of Scotland's Geodiversity Charter. (Images: **a**, **d** © Lorne Gill/NatureScot; **b**, **c**, **e** John Gordon)

Fig. 29.4b). Although not a statutory designation, geoparks incorporate numerous areas protected under other designations. They promote best practice in geoconservation but must also be underpinned by sustainable economic and social development linked to geotourism and demonstrate partnership working with local communities. Scotland's geoparks have an important role in raising awareness of geoheritage and its links with natural and cultural landscapes through appropriate interpretation, education, training and public engagement activities (Gordon 2016).

29.3 Scotland's Geomorphological Diversity and the Scientific Basis for Geoconservation

The great diversity of Scotland's landscapes and landforms has outstanding geoheritage value for science and education, particularly those features relating to long-term landscape evolution, Pleistocene ice-sheet and mountain glaciation and Pleistocene and Holocene periglacial, fluvial, coastal and mass-movement and karst processes (Gordon et al. 2019; Table 29.1). Some sites or areas are internationally important because of their significance in the history of scientific discoveries (Chap. 1); for example, the association of Glen Roy with the development of the Glacial Theory (Chap. 16), the Forth valley with the theory of glacio-isostasy first proposed by Thomas Jamieson (1865) and the Cuillin Hills where James Forbes (1846) published one of the first detailed studies of glacial landforms in Britain and the role of glacial erosion in shaping the landscape (Chap. 10). Other sites display internationally rare landforms such as the machair (calcareous, floristically rich, coastal shell-sand plains) of the Western Isles (Chap. 9; Fig. 29.4c). Some are classic localities for particular landforms, such as the shore platforms of Islay or the emerged gravel beach ridges of Jura (Chap. 11), while others are representative of particular subsets of geomorphology such as the diversity of beach, dune and saltmarsh systems or river environments ranging from high-energy mountain torrents to lowland, low-energy floodplains. The site networks for the different GCR geomorphology subject blocks (Table 29.1) reflect the main research themes, but the site coverage in several of these blocks now requires updating to incorporate recent research findings (Gordon et al. 2019). Generally, sites vary in size from small rock or sediment exposures (e.g. Agassiz Rock in Edinburgh; Chap. 26) to extensive, mainly upland areas with a range of landforms (e.g. Glen Roy, the Cuillin Hills and the Cairngorms), extending up to 292 km^2 in the case of the Cairngorms SSSI. Most, however, are of intermediate size.

29.3.1 Pre-Quaternary Landforms

Elements of the present geomorphology of Scotland can be traced back ~400 Ma, in the form of large-scale landscape components inherited from the Caledonian Orogeny and subsequent events. Recognisable large-scale relict features with clear topographic expression include the four terrane-bounding faults (the Moine Thrust Zone, the Great Glen Fault, the Highland Boundary Fault and the Southern Upland Fault); Cenozoic palaeosurfaces and basins; inselbergs; scarp slopes along the margins of volcanic hill groups in the Midland Valley; and geological controls on drainage patterns (Hall 1991; Chap. 3). These features and occurrences of Palaeogene and Neogene-Early Pleistocene saprolites and Palaeogene fluvial gravels in Buchan in NE Scotland (Chap. 21), as represented at the Hill of Longhaven and Windy Hills GCR sites, provide insights into long-term weathering, erosion and landscape evolution and demonstrate that the present landscape is a palimpsest of inherited landforms of different origins and ages. From a geoconservation viewpoint, the landscape-scale features are generally robust and difficult to fit within the SSSI concept, which is more appropriate for smaller, discrete areas.

29.3.2 Quaternary Glacial and Non-glacial Landforms and Deposits

The Scottish landscape, both onshore and offshore, preserves a record of landscape evolution during successive Quaternary glacial and interglacial periods (Chaps. 4 and 6), and the associated landforms and deposits feature prominently in the 142 Quaternary of Scotland GCR sites (Gordon and Sutherland 1993). Large-scale landforms of glacial erosion were formed during the course of multiple mountain and ice-sheet glacial phases. The record of glacial and glacifluvial landforms and deposits, however, relates mainly to the last Scottish Ice Sheet (~35–14 ka) and the subsequent more limited glaciation during the Loch Lomond Stade (~12.9–11.7 ka). Spectacular glacial troughs, watershed breaches and cirques principally occur in the Highlands and are represented in GCR sites such as the Cuillin Hills (Chap. 10) and the Cairngorms (Chap. 18). Spatially variable basal-ice thermal conditions have produced upland landscapes of different character, ranging from those that display intensive glacial dissection (e.g. Chaps. 13 and 17) to those exhibiting selective or minimal glacial erosion (e.g. Chaps. 18, 20 and 27). Lowland landscapes in the Midland Valley are dominated by ice-streamlined topography, both erosional and depositional, with superimposed bedforms indicating changing flowline patterns (Chap. 26). Ice-scoured topography dominates the lowland landscapes of the Archaean foreland of the Outer Hebrides and NW Scotland (Chaps. 9 and 12), but in contrast, the Buchan area of NE Scotland has been minimally modified by glacial erosion (Chap. 21). These landforms and patterns of glacial erosion are important for elucidating the basal thermal regime and dynamics of Scottish palaeo-ice sheets and palaeo-ice streams, which have been an important focus of research (Hall et al. 2019). With a few upland exceptions, such as the Cuillin Hills, Cairngorms and Glen Roy, the landscape-scale features are generally poorly represented in the GCR and SSSI site coverage.

Research advances have also demonstrated that the geomorphological legacy of the last Scottish Ice Sheet is fundamental for understanding the history and dynamics of marine-based ice sheets and their responses to changes in climate and sea level. A large part of the record of the last and earlier ice sheets lies offshore (Bradwell et al. 2008, 2019; Chap. 6). Landforms include trough-mouth fans, end moraine complexes near the outer edge of the continental shelf, recessional moraines that mark the active oscillatory retreat of the ice margin and tunnel valleys cut by subglacial meltwater. Most of these features are represented in the 11 Quaternary marine key geodiversity areas (Gordon et al. 2016). Ice streams produced streamlined bedforms and fed huge shelf-edge fans, and both iceberg scours and plough marks are widely represented on the sea floor. The depositional record on land is characterised by a variety of landforms. Pulsed recession of the last ice sheet is indicated by readvance moraines in Wester Ross (Chap. 13) and along the northern flanks of the Cairngorms (Chap. 18). Meltwater landforms are common and include single channels and networks of channels, eskers and kame terraces along upland valley margins and floors (e.g. on the northern margins of the Cairngorms and in the Muir of Dinnet NNR), and in the Midland Valley and Southern Uplands (Sissons 1976; Chaps. 15, 20, 25 and 27). The final stages of ice-sheet wastage are represented in upland and lowland areas alike by areas of kame and kettle topography and terraced outwash deposits.

A variety of moraines and other landforms were produced by cirque, outlet and valley glaciers during the Loch Lomond Readvance, including end moraines (Fig. 29.2a) and nested assemblages of recessional hummocky moraines, together with valley sandar, and deltas and shorelines associated with ice-dammed lakes; fine examples of these features are included in GCR sites in Wester Ross, the Cuillin Hills, Lochaber, the Cairngorms, Eastern Grampians, the Loch Lomond area, the western Forth valley and the Tweedsmuir Hills in the Southern Uplands (Chaps. 13, 16, 17, 20, 24 and 27). These landforms have enabled reconstructions of palaeoclimatic conditions, the timing of glacier advance and retreat and the response of glaciers to climate changes (Chap. 4).

Periglacial landforms and processes add to the geomorphological diversity on many mountain summits and higher slopes. They include tors, blockfields and other regolith covers that developed prior to the last glaciation, together with Lateglacial solifluction landforms and patterned ground. Holocene periglacial features include active solifluction lobes and ploughing boulders, miniature frost-sorted patterned ground, deflation surfaces and high-level (niveo)aeolian sand deposits (Chap. 4). Periglacial features are particularly well represented in the Western Hills of Rùm, An Teallach and Cairngorms GCR sites (Chaps. 10, 13 and 18).

29.3.3 Fluvial Landforms and Processes

The diversity of river landforms and processes in Scotland reflects the climate, variations in relief and topography and the legacy of long-term landscape evolution and glaciation (Werritty and McEwen 1997). The asymmetric location of the main watershed has given rise to short, steep rivers in deeply incised valleys in the west and longer, less-steep rivers with extensive lowland reaches in the east. The glacial legacy includes alternating active alluvial reaches separated by stable bedrock reaches or gorges, the latter representing former glacial meltwater channels, and typically coarse bed material from lateral reworking of glacial and glacifluvial deposits. The full range of fluvial landforms and processes encompasses both mountain torrents and active wandering gravel-bed rivers in the uplands and low-energy meandering rivers in the lowlands.

The 29 GCR sites representing the fluvial geomorphology of Scotland are based on a channel and planform typology comprising bedrock rivers, alluvial rivers and alluvial channels with steep gradients (Werritty and McEwen 1997). Bedrock river GCR sites include those representative of slot gorges, wide bedrock reaches and waterfalls (e.g. Corrieshalloch Gorge; the Grey Mare's Tail in the Tweedsmuir Hills, Chap. 27; and the Falls of Clyde, Fig. 29.4a). Alluvial river GCR sites include a variety of meandering, wandering and braided types, notably the braided reaches of the River Feshie and River Spey (Chap. 19; Fig. 29.2b) and meanders on the lower River Clyde. Alluvial rivers with steep gradients where glacial sediments have been reworked by flash floods include mountain torrents (e.g. the Allt Mòr in the Cairngorms, Chap. 19), fluvially modified debris cones and alluvial cones (e.g. in Glen Coe and Glen Etive in Lochaber, Chap. 17) and alluvial fans (e.g. those of the River Quoich and the River Feshie, both in the Cairngorms National Park, Chaps. 18 and 19).

These GCR sites reflect the main foci of scientific studies, principally on alluvial systems. They include, inter alia, studies of braiding processes, downstream fining of sediment, the role of extreme events in triggering channel changes, the assessment of channel changes to detect secular trends and the reconstruction of past flow regimes from sedimentary and documentary archives (e.g. Ferguson et al. 2002; Werritty and Hoey 2004; Werritty et al. 2006; Wheaton et al. 2013; Chaps. 4 and 19). These and other sites also have a wider applied role in studies to assess the resilience of Scotland's rivers to future climate change. For example, whilst active gravel-bed rivers have been relatively resilient to climate trends over the last ~250 years (Chap. 19), natural approaches to the management of lowland rivers, such as the Eddleston Water, can reduce increasing flood risk and enhance aquatic and riparian biodiversity (Werritty et al. 2010).

29.3.4 Coastal Landforms and Processes

The 41 coastal geomorphology GCR sites in Scotland represent the main coastal landforms and processes and their various controlling factors: lithology and structure; complex glacio-isostatic and glacio-eustatic sea-level changes during the Quaternary; patterns of glacial erosion; and variable supplies of glacigenic sediment at or near the coast and available for reworking into beaches (May and Hansom 2003; Ritchie and Dawson 2010). The coasts of the north and west are generally hard and rocky and highly indented by fjords and sea lochs that occupy drowned glacial troughs (Chaps. 9 and 13). Exposure to Atlantic and North Sea storm waves has produced spectacular cliffs, stacks, arches, blow-holes and shore platforms, particularly in the Northern and Western Isles, Caithness and parts of Aberdeenshire; these are represented in the GCR by sites including St Kilda, the West Coast of Orkney, Papa Stour, Duncansby and the Bullers of Buchan (Fig. 29.2c). In the north and west, sandy beaches and dunes are locally confined to isolated bays, except on the west coast of the Western Isles where extensive suites of shell-sand beaches, dunes and machair are developed, notably in the GCR sites on the west coasts of North Uist and South Uist (Chap. 9; Fig. 29.4c).

In contrast, although interspersed by cliffs, the coasts of south and east Scotland display long stretches of beach and dune systems, large coastal forelands of emerged sand and gravel beach ridges (such as at the Spey Bay, Morrich More and Tentsmuir GCR sites; Chaps. 22 and 23; Fig. 29.2b) and saltmarshes in more sheltered locations, particularly along the Solway coast (as in the Upper Cree Estuary GCR site; Chap. 28). Present beach and dune systems are often formed in front of, or superimposed on, inherited coastal landforms and spreads of Lateglacial and Holocene gravel, some of which have been partly or fully remodelled by present processes (May and Hansom 2003; Smith et al. 2019).

Scotland's coast has been profoundly affected by glaciation and changes in relative sea level. Landforms include emerged and submerged shore platforms, some of which may in part pre-date the last glaciation, including the 'strandflat' of North Uist, Benbecula, South Uist, Coll and Tiree, and the high rock platforms of the Inner Hebrides (e.g. the Northern Islay and West Coast of Jura GCR sites; Dawson et al. 2013; Chaps. 9, 10 and 11). Lateglacial and Holocene emerged shorelines in central Scotland record the changing interplay between local glacio-isostasy and global sea level: where sediment supply was abundant, staircases of emerged gravel ridges or prograded forelands were produced (as at the Spey Bay, Morrich More and Tentsmuir GCR sites; Chaps. 22 and 23). In more peripheral areas where eustatic sea-level rise outpaced glacio-isostatic rebound, such as in the Northern and Western Isles, the trend has been one of coastal submergence (Chaps. 7 and 9). This has resulted in flooding of terrestrial peats that now lie in the inter- and sub-tidal zones, and the creation of numerous islands. Even so, where sediment is locally available, island-tying can occur in areas subject to continuing submergence (e.g. Sanday GCR site, Orkney; Chap. 8).

Much of the research attention has been on coastal evolution, particularly relative sea-level change and the formation and development of machair and active beach and dune forms. However, applied studies are increasingly focusing on how enhanced coastal resilience under climate change and projected sea-level rise can be achieved by identifying trends in coastal erosion and working with nature, for example using natural flood and erosion management, managed realignment and artificial beach feeding (Hansom et al. 2017). Another recent research focus has been on the formation of cliffs and cliff-top storm deposits during extreme events on rock coasts, particularly in Shetland, Orkney and Caithness (Hansom and Hall 2009; Chaps. 7 and 8).

29.3.5 Mass Movements

Scotland is notable for the diversity of rock-slope failures (RSFs) developed on the Neoproterozoic metamorphic rocks of the Highlands, the Palaeogene lavas of Skye and Mull and the Palaeozoic lavas of the Midland Valley (Cooper 2007; Jarman 2006, 2007; Cave and Ballantyne 2016; Chaps. 10 and 14). These reflect paraglacial responses to glacial loading and unloading of glacially steepened slopes and the relationships with structure, lithology and neotectonics. Recent recognition of the significance of debris-free failure scars, where runout debris has been removed by glacier ice, has highlighted the role of rock-slope failure during interglacials in widening glacial troughs, enlarging cirques, forming arêtes and generally shaping the morphology of many mountain ridges in the Highlands (Chap. 14).

The 13 Mass Movement GCR sites include representative and outstanding examples of postglacial rock-slope deformations, such as that on Beinn Fhada, and arrested rockslides, such as that on Ben Hee (Chap. 14). A small number of well-documented catastrophic RSFs include the Beinn Alligin rock avalanche in Wester Ross, exceptional for its excess runout (Chap. 14), and a rock avalanche in Glen Coe in Lochaber, which has blocked the valley of Coire Gabhail (Chap. 17). The most extensive and spectacular landslides occur on Skye, where stacks of Palaeogene lava flows have foundered over weaker Jurassic sedimentary rocks (Chap. 10; Fig. 29.2d).

In addition, postglacial debris flows and debris cones are widespread in the Highlands. Many are still intermittently active today, triggered by high-intensity, prolonged rainfall events (Winter 2020). Representative examples are included in the Quaternary of Scotland GCR, for example in Glen Coe (Chap. 17) and the Cairngorms (Chap. 18).

29.3.6 Karst and Caves

Karst geomorphology is a relatively minor component of Scotland's geodiversity associated with outcrops of dolomitic limestone of the Cambrian Durness Group in the NW Highlands and Skye, and Neoproterozoic limestones in the Dalradian Supergroup that crop out intermittently from Islay to east of Spey Bay on the coast of Aberdeenshire. Small areas of limestone pavement occur at Durness, Inchnadamph (Assynt), Rassal (near Kishorn, Wester Ross) and Strath Suardal (Skye), and in parts of Perthshire and on the Garvellach Islands. The largest and longest cave systems are developed in the Assynt karst which includes two GCR sites, the Traligill Valley and Allt nan Uamh Caves (Waltham et al. 1997; Chap. 12). Here, faunal remains and speleothems have also provided significant palaeoenvironmental records. Numerous smaller and shorter caves occur elsewhere (e.g. in Applecross, Wester Ross and the Appin area of Lochaber), while Smoo Cave near Durness is a spectacular example of a coastal cave in the Durness Group limestone. Tufas and marls are present in spring lines of carbonaceous sandstones and metamorphic rocks across Scotland (Faulkner and Brazier 2016). At Inchrory GCR site in upper Glen Avon in the Cairngorms, the tufa has preserved Early Holocene plant and mollusc remains, while the Starley Burn near Burntisland in Fife is the only known actively forming stromatolite deposit in Scotland.

29.3.7 Marine Geomorphology

In addition to the glacial landforms noted above, the marine geodiversity of Scotland's territorial and offshore waters includes features that represent the geological and geomorphological processes that have influenced the evolution and morphology of the Scottish deep-seabed and shelf and past and present marine and coastal processes. Specific interests of the marine geodiversity key areas range from large-scale landforms (e.g. submarine landslides, seamounts and trenches) to fine-scale active features (e.g. sand waves) (Gordon et al. 2016; Chap. 6; Table 29.1). Particular highlights include: large seabed fluid and gas-seep features (the Scanner-Scotia and Challenger pockmark complexes in the North Sea); the only known examples of mud diapirs in UK waters (the Pilot Whale Diapirs at the north end of the Faroe-Shetland Channel, rising >70 m above the adjacent seabed); and the contourites (extensive accumulations of muds and silts deposited and moulded by deep-sea currents into large sediment waves) that form a complex of bedforms north and west of Shetland.

Other notable features of the Scottish seabed are the non-tropical, shelf carbonate systems that form the sources of sediment for the dune and machair systems of the Western and Northern Isles. Mingulay Reef, at the entrance to the Sea of the Hebrides, is an exceptional example of an extensive, cold-water coral reef in UK waters. Tidal sand banks are represented by the spectacular Sandy Riddle Bank east of the Pentland Firth (Chap. 6), while excellent examples of now-submerged landforms formed by coastal processes during episodes of lower sea level include under-sea cliffs, caves and rock platforms, notably around St Kilda and Sula Sgeir (Chap. 9).

29.4 Threats to Geoheritage

Landforms and geomorphological process systems have been affected by, and continue to be vulnerable to, a range of threats and human pressures (Gordon et al. 2019). These principally include urbanisation, commercial, industrial and infrastructure developments, mineral extraction, land-use changes, afforestation, overgrazing, coastal defences, river defences and pollution (Chap. 5). Potential impacts include partial or complete destruction of landforms, fragmentation of landform assemblages and loss of site integrity, loss of site access and disruption or deactivation of geomorphological systems, especially where natural processes are severely constrained. Relict landforms are particularly vulnerable as the processes that formed them are no longer active; Brazier et al. (2017), for example, have provided a case study of the changing pressures and threats to the geomorphological heritage of the Parallel Roads of Glen Roy (Chap. 16). Submarine landforms may also become vulnerable as offshore windfarms are developed, and more microplastics enter seabed sedimentary systems.

Coastal and fluvial features are likely to be most vulnerable to climate change and sea-level change, both directly through loss of landforms and indirectly through erosion and flood protection schemes involving installation of hard coast defences and river engineering that will disrupt water flows and sediment transfer (Wignall et al. 2018). Positive nature-based solutions to climate change, such as regeneration of native woodlands and allowing sufficient space for both the coast and rivers to move position will restore more naturally functioning systems (Chaps. 5 and 19).

Despite the various pressures, the success of site protection in achieving practical on-site conservation, delivered both through the statutory planning system and through working in partnership with land owners and land managers, has been demonstrated in the programme of site condition monitoring undertaken for SSSIs by NatureScot (the operating name of Scottish Natural Heritage; Wignall 2019). This shows that the vast majority of GCR geomorphology sites continue to be in favourable condition, although some have required specific conservation management interventions, such as clearance of scrub or access improvements; only six sites are recorded as having some features partly destroyed. Some vulnerable sites on the coast and in active river systems have the capacity to self-heal once disturbance has been stopped, as in the case of the River Feshie (Chap. 19). However, irreversible damage may occur where all or part of a landform is removed, for example by quarrying or sand and gravel extraction as in the case of the Carstairs Kames esker system (Chap. 25). There is no comparable monitoring system for non-statutory sites, however. Since geomorphological heritage does not stop at the boundaries of statutory protected areas, the recognition of geoheritage values and the adoption of sound management principles and techniques is a pressing requirement across all categories of conservation area, not only those primarily designated for geoconservation, and in the wider landscape (Crofts et al. 2020).

29.5 Wider Values of Scotland's Geoheritage

As part of 'whole of nature' approaches, there is growing advocacy from the geoconservation community to recognise the role of geodiversity and its cultural, aesthetic and ecological values in helping to deliver natural solutions to socio-environmental challenges such as climate change adaptation; connecting people, place and nature; contributing to human well-being; ecosystem stewardship; and achievement of the UN Sustainable Development Goals (Brilha et al. 2018; Gordon et al. 2018; Schrodt et al. 2019). Many of these values are manifested through the contributions of geodiversity to cultural, supporting and regulating ecosystem services (Gordon and Barron 2013; Gray et al. 2013).

First, spectacular or distinctive landforms and landscapes contribute significantly to landscape character and the scenic qualities that are a prime attraction for visitors to Scotland (Fig. 29.4b). Often they have cultural and aesthetic values celebrated in literature, art or music and have wide public appeal as 'natural wonders'. Some have been visitor destinations since the late eighteenth century and were prominent in the early development of tourism in Scotland based on the aesthetics of sublime and picturesque landscapes and their associations with notable literary figures and artists such as Sir Walter Scott, the Wordsworths and J. M. W. Turner (Gordon 2012; Gordon and Baker 2016; Fig. 29.4a). The appeal of such spectacular and iconic sites today is enhanced by modern cultural associations through backdrops to films and promotion on social media (e.g. Skye and Glen Coe; Chylińska 2019; Le Coeur 2019). For example, the annual number of visitors to The Storr landslide complex on Trotternish in Skye now exceeds 200,000. Geotourism is also a fundamental driver for Scotland's four geoparks as a form of sustainable development with benefits not only for geoconservation, but also for people's health and well-being and the delivery of economic benefits for local communities.

Second, geomorphology provides the natural 'stage' for many habitats and species across a range of scales (Hjort et al. 2015). Large coastal and upland landform assemblages support valued habitats and species of national and international importance (Fig. 29.4b–d). Understanding the connections between geomorphology and biodiversity is also essential to informing effective adaptation to climate change in active process environments (Brazier et al. 2012). Such connections underpin the advocacy of 'conserving nature's stage' as a coarse-filter approach to support biodiversity conservation and the development of robust protected area networks in the face of climate change (Beier et al. 2015).

Third, geomorphological systems fulfil an important regulating function in mitigating natural hazards. For example, the value of the human assets protected by natural beaches and sand dunes on Scotland's 'soft' coasts has been estimated at £13bn, compared to only £5bn of assets currently protected by artificial defences (Hansom et al. 2017). Similarly, where rivers have been channelised to flow in narrow corridors by flood banks or other engineering works, reconnecting rivers and their wider floodplains is an important part of natural flood management, with additional benefits of enhancing biodiversity (Werritty et al. 2010).

Recognition of these additional values is clearly expressed in Scotland's Geodiversity Charter, a national framework and vision to promote geoheritage and deliver geoconservation objectives through engagement with stakeholders (Scottish Geodiversity Forum 2017; Fig. 29.4e). The Charter currently (2020) has been signed by 95 organisations spanning government agencies, local authorities and the voluntary sector.

29.6 Conclusion

Geomorphological features form a significant part of Scotland's internationally important geodiversity and geoheritage. The focus of geoconservation effort has principally been on protected areas for their scientific value, based on the GCR site evaluation process and SSSI designation as the principal conservation mechanism. As a site assessment process, the GCR has stood the test of time and provided a robust network of the most representative and scientifically most important sites, while SSSI designation has generally been an effective means for their statutory protection. However, some GCR sites still remain to be designated as SSSIs and some of the GCR site networks now require updating, particularly that for the Quaternary of Scotland, and new sites added to reflect recent advances in scientific research.

Other designations and non-statutory forms of protected area also include geomorphological features. Although this is not always explicitly acknowledged in the supporting documentation, these areas provide opportunities for extending geoconservation, as well as recognising the additional educational, aesthetic, cultural and ecological values of geoheritage. Better appreciation of these values, as reflected through ecosystem service benefits, should help to strengthen public support for geoconservation and for the integration of geoconservation within the developing nature conservation agenda, including implementation of nature-based solutions, the 'conserving nature's stage' approach and improving links with cultural heritage.

A notable gap in the conservation of geomorphological features is the lack of a natural heritage protection mechanism at the landscape scale and the unsuitability of SSSIs for other than small- to medium-scale site protection. National Scenic Areas could potentially play a part in filling this gap, but the association between scenery and underlying geoheritage values needs to be strengthened in the relevant remit for each area. Also, such designation does not fall within the International Union for Conservation of Nature (IUCN) protected area management categories. Geoparks play a different role and offer no statutory protection, and extending the network of geoparks and maintaining their activities are challenging in economically difficult times. Future directions in geoconservation to address the landscape scale and the connectivity between different geomorphological elements within catchments and coastal systems should therefore form part of the wider discourse about nature conservation at the landscape scale (Pepper et al. 2014). More generally, better understanding across the full range of protected area categories of the interdependencies of geomorphology, biodiversity and cultural heritage would enable wider stakeholder engagement and the development of new forms of conservation management that incorporate the integrity of nature and people and the links between natural heritage and cultural heritage at all spatial scales.

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Correction to: Scotland's Geomorphological Heritage and Its Conservation

John E. Gordon, Vanessa Brazier, James D. Hansom, and Alan Werritty

Correction to:
Chapter 29 in: C. K. Ballantyne and J. E. Gordon (eds.), Landscapes and Landforms of Scotland, World Geomorphological Landscapes, https://doi.org/10.1007/978-3-030-71246-4_29

The original version of the book was inadvertently published with error in "Neogene/Early Pleistocene fluvial gravels", and this has been corrected as "Palaeogene fluvial gravels" in Page "487" of Chapter "29".

The correction chapter and the book have been updated with change.

The updated version of this chapter can be found at https://doi.org/10.1007/978-3-030-71246-4_29

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