4 Glacial Landform Assemblages of Mainland Nunavut, West of Hudson Bay and Their Palaeoglaciological Significance

Chris R. Stokes

Abstract

This case study focuses on the glacial landscapes of the low-relief tundra of the Canadian Shield of mainland Nunavut, west of Hudson Bay. Having been studied for well over a century, the glacial geomorphology of the region is spectacular and is widely recognised as being of international significance to palaeo-glaciology. At the global Last Glacial Maximum, the region lay close to the centre of the former Laurentide Ice Sheet and was submerged beneath a large (>3 km thick) ice dome, and with a major ice divide orientated roughly north-east to south-west (the Keewatin Ice Divide). This was an important influence on the distribution of glacial landforms and the aim of this chapter is to provide a concise overview of the key landform assemblages and their zonation around the Keewatin Ice Divide with a particular focus on hummocky terrain, ribbed moraines, drumlins, mega-scale glacial lineations, meltwater features, end moraines, and raised beaches and shorelines related to proglacial lakes and marine transgression. It is clear that whilst the distribution of landforms is a complex interplay between the overriding ice sheet properties (thickness, velocity, basal thermal regime) and the underlying geology, a detailed analysis of these landscapes has led to some important advances in our broader understanding of ice sheet dynamics and glacial geomorphology, particularly in the area of drift dispersal, the migration of ice divides, ice streaming and the evolution of subglacial meltwater drainage systems.

Keywords

Glacial geomorphology • Keewatin • Drumlins • Ribbed moraine • Mega-scale glacial lineations • Laurentide Ice Sheet

C. R. Stokes (✉)
Department of Geography, Durham University, Durham, DH1 3LE, UK
e-mail: c.r.stokes@durham.ac.uk

4.1 Introduction

This case study focusses on the glacial landscapes of mainland Nunavut, west of Hudson Bay, which is characterised by the low to moderate relief terrain of the north-western Canadian Shield (see Fig. 4.1). The shield itself is geologically complex, with rocks that were deposited, formed, and deformed over a period that encompasses over three quarters of the known geological history of the Earth (Aylsworth and Shilts 1989a). Most of the region is characterised by massive granitoid gneiss, but this is interrupted in places by plutonic bodies, sedimentary rock and volcanic strata (Aylsworth and Shilts 1989a; Dyke and Dredge 1989; Dyke et al. 1989). Of particular significance to the glacial geomorphological record are the Dubawnt Group of Upper Proterozoic unmetamorphosed sediments that outcrop as the Thelon sedimentary basin near the centre of the study area (Fig. 4.1). Unlike the physically hard rocks that characterise most of the Canadian Shield and that are generally resistant to glacial erosion and yielded very little debris, the Upper Proterozoic sedimentary and volcanogenic rocks are much less consolidated and more susceptible to glacial erosion. As such, these outcrops are known to have exerted a major influence on the nature and distribution of glacial landforms and sediments (Shilts et al. 1987; Aylsworth and Shilts 1989a).

The underlying geology is also an important control on the relief of the study region, which is typically flat to gently undulating topography with low broad hills scoured by glacial erosion of the hard crystalline bedrock, and with negligible relief over the areas of unmetamorphosed sedimentary bedrock, such as the Thelon sedimentary basin (Shilts et al. 1987). Glacial erosion of the low-relief landscape (and/or the melting of ice blocks in areas of thicker drift) has led to myriad lakes (Shilts et al. 1987) and, across most of the study area, >20% of the land surface is covered by water (Dyke et al. 1989). The drainage of the major river systems is generally east towards Hudson Bay (e.g. the

O. Slaymaker and N. Catto (eds.), *Landscapes and Landforms of Eastern Canada*, World Geomorphological Landscapes, https://doi.org/10.1007/978-3-030-35137-3_4

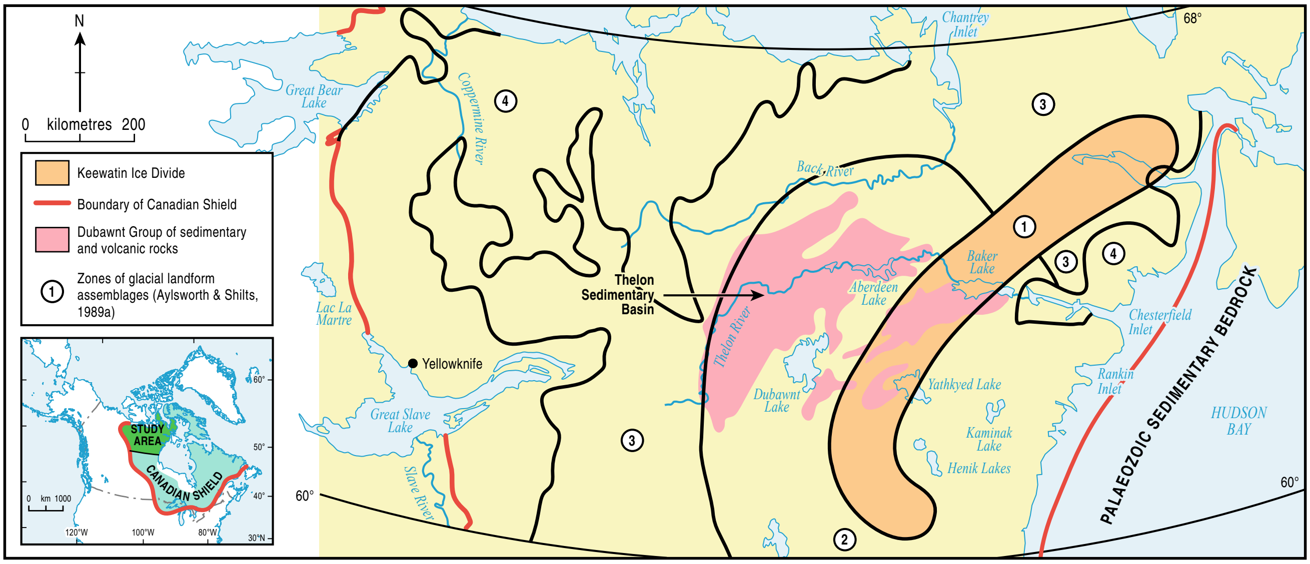

Fig. 4.1 Location map of the study area (inset) showing the four zones of glacial landform assemblages defined by Aylsworth and Shilts (1989a, b). Note the location of the last inferred position of the Keewatin Ice Divide (Zone 1) and the location of the Thelon Sedimentary Basin predominantly within Zone 2

Thelon River) or north towards the Arctic coast (e.g. the Back River) (Fig. 4.1). Most of the study region is classified as tundra with continuous permafrost, such that water-saturated organic deposits (bogs, peats) typically developing in the active layer of low-relief terrain. Patterned ground (e.g. stone circles) is also found throughout the region, although pingos are much rarer on the shield, e.g. compared to the fine-grained sediments to the north-west in the Mackenzie River corridor.

The surficial geology is dominated by glacial geomorphology and the glacial landscapes of the northwestern Canadian Shield are some of the most spectacular and important in the world, and have a long history of research (Tyrell 1897; Bird 1953; Dean 1953; Fyles 1955; Taylor 1956; Lee 1959; Blake 1963; Craig 1964, 1965; Falconer et al. 1965b). The region lay close to the centre of the former North American Laurentide Ice Sheet (LIS), which has traditionally been referred to as the 'Keewatin Sector' (e.g. Tyrell 1897; Dyke and Prest 1987; Dyke et al. 1989). Of particular note is that the distribution of glacial landforms is thought to be intimately associated with the Keewatin Ice Divide (KID) (Shilts et al. 1987; Aylsworth and Shilts 1989a; Tarasov and Peltier 2004), which represented the highest dispersal centre/dome of the ice sheet and which was recognisable as a distinct feature throughout much of deglaciation, even though it may have migrated by as much as 500 km (Lee et al. 1957; McMartin and Henderson 2004). During deglaciation, this region also was subjected to rapid ice flow in the form of ice streaming, which created some of the most spectacular landforms created by the LIS and has resulted in one of its best-studied palaeo-ice streams: the Dubawnt Lake palaeo-ice stream (Stokes and Clark 2003a, b). Superimposed on these landscapes are isolated moraine systems and numerous landforms and deposits related to subglacial meltwater activity (e.g. eskers) and the development of proglacial lakes (e.g. relict shorelines). In addition to its palaeo-glaciological significance, the Keewatin Sector has also been the focus of extensive and ongoing drift prospecting programmes, e.g. by the Geological Survey of Canada since 1970 (Cunningham and Shilts 1977; Shilts 1977; Shilts et al. 1979; Shilts 1980; McMartin and Henderson 2004).

The aim of this chapter is to provide a concise overview of the key glacial landform assemblages that occur on the northwestern Canadian Shield, with a particular focus on hummocky terrain, ribbed moraines, drumlins, mega-scale glacial lineations, meltwater features, end moraines, and glacial lake shorelines. Section 4.2 describes the characteristics of these features and their spatial distribution in relation to the KID, leaning heavily on the instrumental work of William (Bill) Shilts and co-workers who were the first to recognise an obvious zonation of glacial landforms in the region (Aylsworth and Shilts 1985, 1989a, b, c; Shilts et al. 1987). Section 4.3 then briefly highlights how the landform assemblages and patterns described in Sect. 4.2 are of international interest in terms of our understanding of the palaeo-glaciology of the LIS, and with particular reference to the KID (Sect. 4.3.1), meltwater drainage systems (Sect. 4.3.2) and fast-flowing ice streams (Sect. 4.3.3).

4.2 Glacial Landform Assemblages

In a series of publications in the 1980s, Shilts and co-workers noted that a striking feature of glacial landforms on the northwestern Canadian Shield was the apparent arrangement of landforms in broadly concentric zones around the location of the last inferred position of the Keewatin Ice Divide (KID) (e.g. Aylsworth and Shilts 1985, 1989a, b, c; Shilts et al. 1987) (Fig. 4.1). Ice divides are analogous to hydrological (drainage) divides in fluvial systems and they represent broad ridges or domes in the ice sheet surface topography (see e.g. Dyke and Prest 1987). The KID has long been recognised as one of the major dispersal centres in the LIS (Lee et al. 1957; Lee 1959), representing a northeast–southwest orientated feature several hundred kilometres long and lying approximately 200 km west of the coast of western Hudson Bay (Fig. 4.1; Dyke and Prest 1987). It has been inferred as one of the highest parts of the LIS (Prest 1969; Dyke and Prest 1987; Boulton and Clark 1990; Tarasov and Peltier 2004) and recent relative sea level data and glacial isostatic modelling indicate that it was likely around 3.4 km thick during the Last Glacial Maximum (Simon et al. 2014).

Aylsworth and Shilts (1989a) identified four main zones of glacial landform assemblages associated with the KID, with Zone 1 representing the innermost zone that includes the footprint of the ice divide itself (Fig. 4.1). It is generally characterised by featureless till plains (i.e. lacking prominent glacial landforms) or low hummocky terrain and with a general absence of eskers. Zone 2 encircles most of Zone 1 (Fig. 4.1) and is characterised by the presence of well-developed linear trains or ribbed moraine, that radiate out from the location of the former ice divide, but which are interspersed with fields of glacial lineations, again showing radial flow away from the ice divide. Beyond Zone 2, Zone 3 is characterised by a fairly continuous cover of till and numerous drumlin fields that contain some highly elongate glacial lineations, but where till thins to a discontinuous veneer in the outer portion of the zone. The outermost Zone 4 comprises extensive areas of massive crystalline bedrock that is typically devoid of any till cover.

Aylsworth and Shilts (1989a) noted that the arrangement of these zones around the location of the inferred ice divide suggests that they are likely to be broadly related to the dynamics of the ice sheet and the style of deglaciation, but they also noted the importance of the underlying geology in modulating the influence of the ice sheet dynamics, e.g. with the major sedimentary basins typically being associated with the best developed and most spectacular glacial landforms. The following subsections describe the glacial landforms assemblages within each of these zones (Fig. 4.1), followed by a discussion of major moraine ridges and glacial lake shorelines that typically span these zones.

4.2.1 Zone 1

The innermost Zone 1 includes the footprint of the KID (Fig. 4.1) and is characterised by areas of low-relief hummocky terrain and featureless plains of generally thin till cover. The hummocky terrain (see e.g. in Fig. 4.2) is most obvious towards the southern end of the inferred ice divide where it is comprised of small (<100 m across), low (<5 m high), rounded hummocks of till (Aylsworth and Shilts 1989a). Aylsworth and Shilts (1989a) noted that the hummocks appear to be underlain by till similar in composition to material formed in nearby drumlins (e.g. in Zone 2 and sporadically in Zone 1), but do not have the dense surface cover of bounders that typically cover ribbed moraines in Zone 2. Aylsworth and Shilts (1989a) also noted that although the hummocky terrain is best developed in Zone 1, it is by no means restricted to that area and that isolated patches occur elsewhere in the study area, e.g. west of Dubawnt Lake in Zone 2.

A characteristic feature of Zone 1 is the general absence of glacial landforms. Ribbed (sometimes known as 'Rogen') moraines are absent and streamlined features, such as drumlins, are rare. Where drumlins occur they indicate radial ice flow away from the last inferred position of the divide (Aylsworth and Shilts 1989a). Likewise, isolated crag-and-tails and roche mountonnées have been reported in this region and generally indicate radial ice flow away from the position of the ice divide (e.g. Bird 1953; Lee 1959). It has also been noted that Zone 1 is characterised by a marked absence of eskers, which are otherwise very common across the northwestern Canadian Shield (Aylsworth and Shilts 1989a; Storrar et al. 2013, 2014a, b). Storrar et al. (2014a), for example, analysed >20,000 mapped eskers across the bed of the LIS, but noted that they were absent from regions of inferred ice divides in both Keewatin and Labrador. They identified an esker-free area approximately 100–150 km wide and covering an area around 93,000 km^2 in Keewatin, noting its association with the last inferred position of the KID. Indeed, Zone 1 is almost completely devoid of any glacifluvial deposits, except for minor accumulations of outwash in some valleys (Shilts et al. 1987).

4.2.2 Zone 2

Aylsworth and Shilts (1989a) demarcated Zone 2 as a 200–250 km wide, U-shaped belt that wraps around the southern

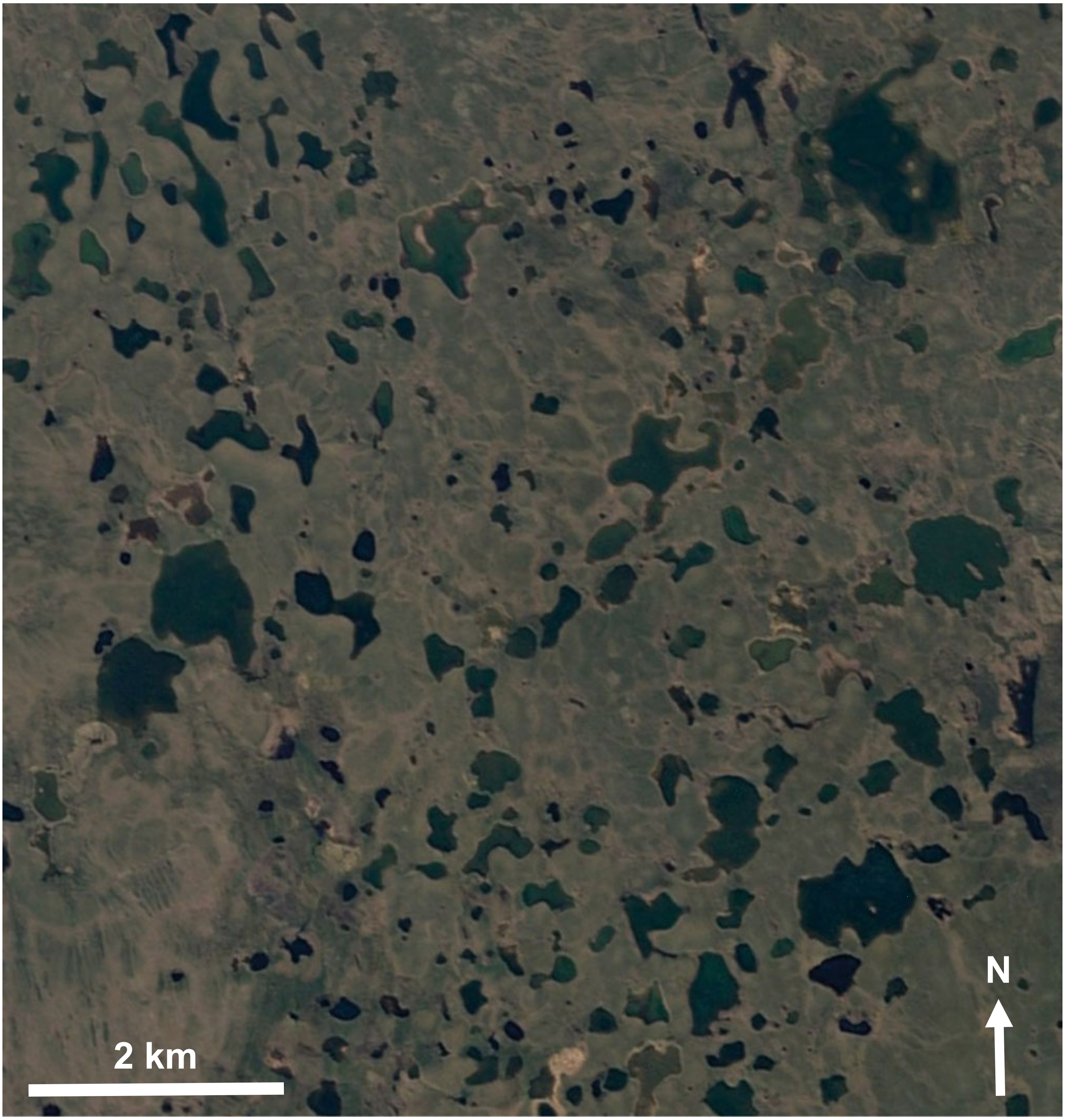

Fig. 4.2 Landsat image (data from Google Earth: Landsat/Copernicus) showing low-relief hummocky terrain in Zone 1 at the southern end of the of the last inferred location of the Keewatin Ice Divide, 20 km north-west of Yathkyed Lake (located on Fig. 4.1). Image is centred on 62°49′46″ N, 99°05′04″ W

end of the last inferred position of the KID (Fig. 4.1). The zone is characterised by extensive linear trains of ribbed moraine which radiate from the location of the inferred ice divide and which are interpreted to extend along former ice flow-lines. The areas of ribbed moraines (e.g. in Fig. 4.3) are typically comprised a nested series of short (<2 km long), low (<10 m high) sinuous ridges of irregular height that trend perpendicular to the inferred ice flow direction (Aylsworth and Shilts 1989a). Aylsworth and Shilts (1989a) noted that they appear to be asymmetric in cross section, with a steeper down-ice side, but this has not been systematically observed or quantified. They also noted that these 'classic' ribbed moraines can also be associated with less well-defined ridges and/or with irregularly shaped hummocks, albeit with a similar composition and relief.

The development of ribbed moraines in Zone 2 is abrupt, with virtually no gradational transition from the featureless terrain in Zone 1. Where the ribbed moraines are extensive, their distribution does not appear to be linked to topography, occurring on both lower and higher ground. That said, Aylsworth and Shilts (1989a) noted that isolated patches tend to occur in lower points in the landscape, which was also noted by Bouchard (1989) who suggested that they are formed by glacitectonic shearing and stacking of subglacial sediments under a compressive flow regime in topographic depressions.

In terms of their composition, Aylsworth and Shilts (1989a) reported that the ribbed moraines are commonly formed in till but are often associated with a dense concentration of boulders, which may mantle the till or be intermixed. In particular, they noted that some large fields of ribbed moraines, for example, west of Rankin Inlet (Fig. 4.1), extend down-ice from outcrops of post-glacially shattered granitic bedrock. They suggested that large quantities of felsenmeer (blockfields) are likely to have existed in these regions prior to the last glaciation and which have subsequently been incorporated into the existing trains of ribbed moraine. They also reported prominent trains of ribbed moraines that are composed of disaggregated debris derived from the friable sandstones and conglomerates of the Thelon sedimentary basin. Aylsworth and Shilts (1989a) also noted that there is little erratic material in these ridges and they are almost exclusively composed of orthoquartzite clasts, quartz sand and kaolin, with so few nutrients that vegetation is scarce. This lack of vegetation, especially on the well-drained ridge crests, enables the ribbed moraines and other streamlined features, such as drumlins, to be easily identified on satellite imagery and aerial photographs throughout the region.

Aylsworth and Shilts (1989a) drew attention to the fact that fields of ribbed moraines are closely associated with linear fields of drumlins and other highly elongate streamlined landforms in Zone 2 that also radiate from the inferred position of the ice divide. In most cases, there is a sharp lateral contact between ribbed moraines and alternating parallel trains of drumlins and other streamlined features (Fig. 4.4), which is common elsewhere on the Canadian Shield (Shilts et al. 1987; Dunlop and Clark 2006). It has also been observed that there are places where the ribbed moraines have been heavily streamlined with drumlins superimposed on ridge crests (e.g. Fig. 4.3), and where streamlined features have themselves been broken up into ribbed moraines (Fig. 4.5). It is important to note that whilst ribbed moraines are more common in Zone 2, the drumlins and streamlined features extend into Zone 3 and, although much less common, Zone 4.

More recently, Greenwood and Kleman (2010) reported a much larger scale of landform in Keewatin than had hitherto been recognised. They mapped a series of large till 'belts' or 'ridges', apparent only on satellite imagery and lying beneath the drumlins, flutes and ribbed moraine (Fig. 4.6). Planform and crestline mapping from remotely sensed imagery allowed them to map a population of >2500 individual landforms, whose dimensions are on average $\sim 10°$ km long and $\sim 1.5°$ km wide and which form extensive and coherent patterns throughout the Keewatin region, i.e. not just restricted to Zone 2. Some of these features were interpreted as over-printed mega-scale glacial lineations or moraines that predate the LGM (see also Kleman et al. 2002, 2010), but a group, comprising a significant number of the Keewatin population, did not fit any existing category of glacial landforms (Fig. 4.6). They referred to these as 'mega-scale transverse bedforms' and postulated that their close spatial integration with ribbed moraines suggests a similar mode of genesis.

Common throughout Zone 2 is an extensive network of hundreds of eskers that also radiate from the edge of the location of the ice divide (Bird 1953; Shilts 1984; Aylsworth and Shilts 1989a; Storrar et al. 2013, 2014a). Towards the east of the study area and below the marine limit occupied by the much larger ancestral Hudson Bay (known as the Tyrell Sea), eskers have been reworked and appear more subdued (e.g. Shilts et al. 1987; Aylsworth and Shilts 1989a). Elsewhere, most of the northwestern Canadian Shield is covered by an integrated pattern of eskers radiating outward from the area of the KID (Prest et al. 1968; Storrar et al. 2013, 2014a) (Fig. 4.7). Aylsworth and Shilts (1989a) noted that typical esker systems begin as a series of hummocks or short, flat-topped segments which pass down-ice into a relatively continuous and larger esker ridges or (occasionally) beaded eskers. They also noted that discontinuous eskers are common where material has either been removed or, in numerous cases, flooded by lakes (see example in Fig. 4.3a).

More recent work by Storrar et al. (2014a) found that gaps typically represent around 35% of the total length of esker systems that are obviously traceable despite the gaps. The length of the largest unbroken esker ridges can be traced for up to 75 km (Fig. 4.7), but the largest systems can be traced for several hundred kilometres (Storrar et al. 2014a). Indeed, Dyke and Dredge (1989) noted that the 'Thelon Esker', at >800 km long, including gaps, is probably the longest esker in the world (see Fig. 4.7). It is unlikely that some of the very longest eskers were formed rapidly (e.g. in tunnels of several tens of kilometres in length). Rather, it has been argued that it is more likely that eskers were deposited in short meltwater channels a few kilometres in length and

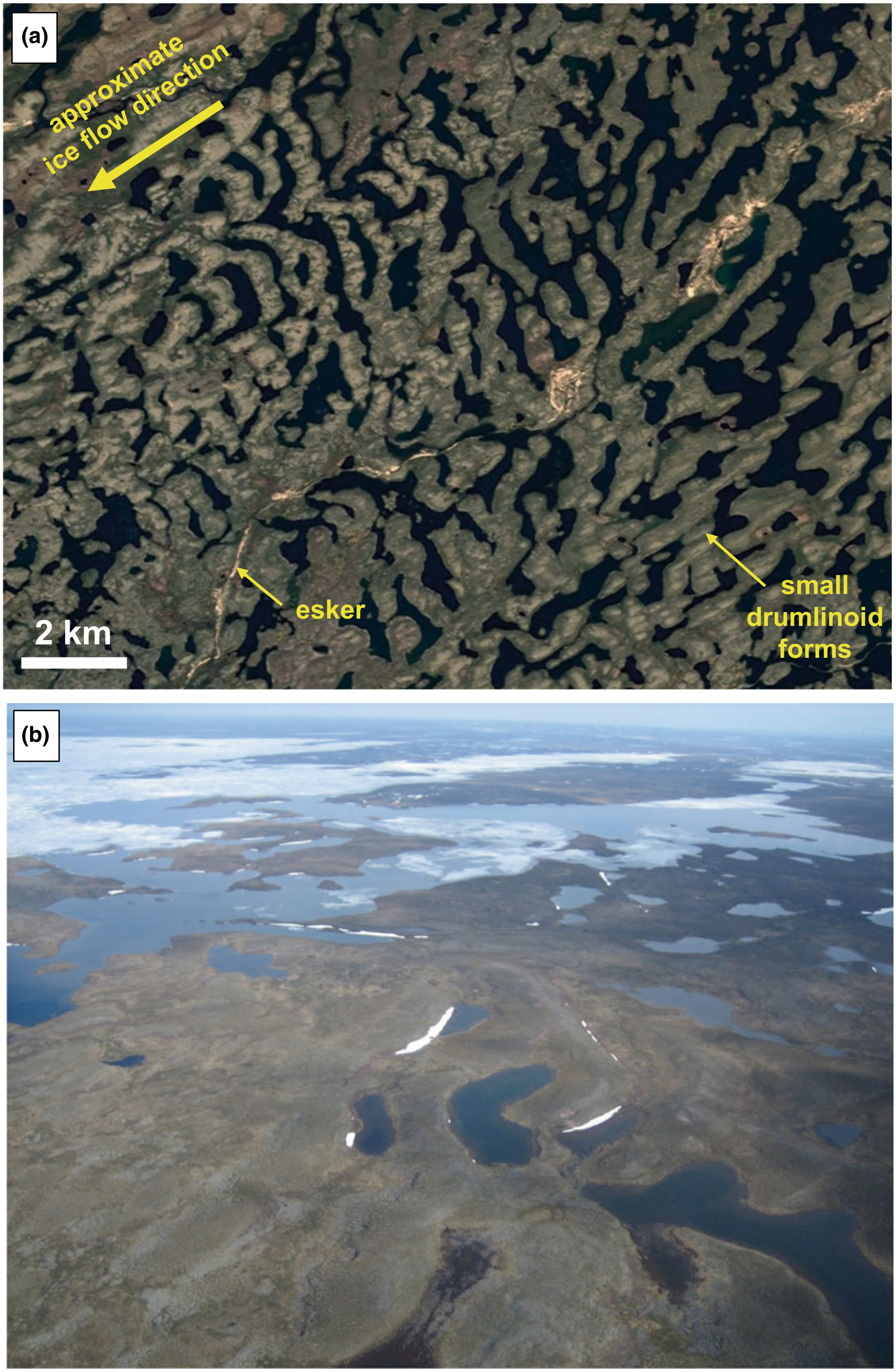

Fig. 4.3 **a** Landsat image (data from Google Earth: Landsat/Copernicus) showing classic ribbed moraines in Zone 2, around 170 km south-west of Dubawnt Lake (located on Fig. 4.1). Image is centred on 61°22′30″ N, 103°54′26″ W. Note the small drumlinoid forms superimposed on the ribbed moraines towards the south-east of the image. Also note the large esker system trending north-east to south-west broadly aligned with the approximate direction of ice flow. **b** An oblique aerial photograph of classic ribbed moraine from Zone 2. *Photograph* C. Stokes

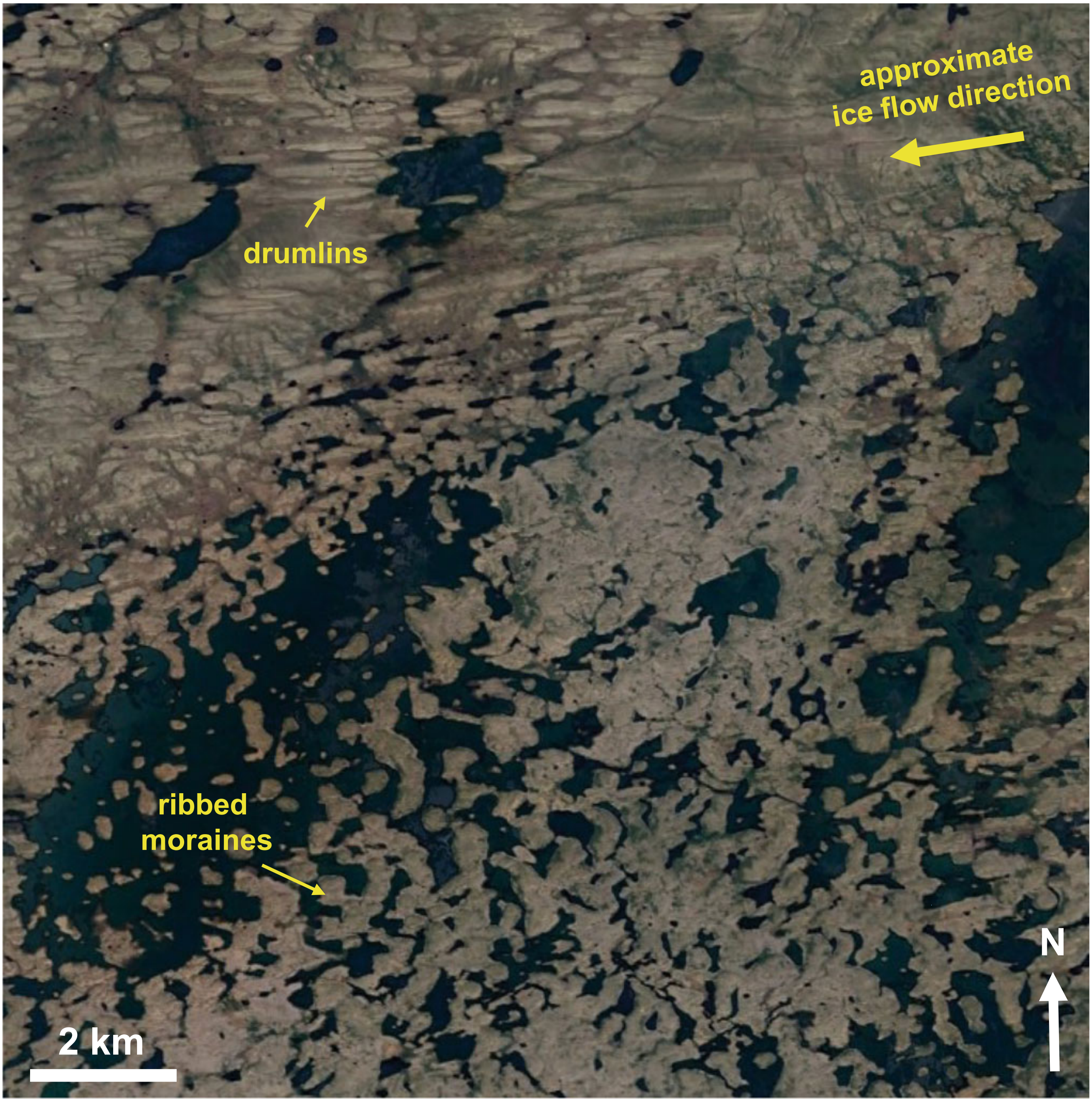

Fig. 4.4 Landsat image (data from Google Earth: Landsat/Copernicus) showing classic ribbed moraines with a sharp lateral contact to a drumlin field in Zone 2, around 70 km south-west of Dubawnt Lake (located on Fig. 4.1). Image is centred on 62°07′40″ N, 102°17′06″ W

largely fed by surface meltwater in the ablation zone of the retreating ice sheet (Aylsworth and Shilts 1989a; Storrar et al. 2014a; Livingstone et al. 2015). The fact that many eskers can be traced over much longer distances likely represents the headward propagation of surface meltwater streams (Livingstone et al. 2015).

Aylsworth and Shilts (1989a) reported that most eskers are sharp-ridged and up to 40 m high, with conical knobs projecting well above the average elevation of the crests in some locations. They also noted that eskers may be interrupted by bulges where a single ridge splits into multiple ridges before coalescing down-ice, which may be related to blockages in subglacial drainage and the formation of bypass tunnels in/under the ice. Aylsworth and Shilts (1989a) noted that above the marine limit, and beyond areas inundated by proglacial lakes (see Sect. 4.2.6), eskers are commonly flanked by glacifluvial outwash terraces that are scarred by channels and disrupted by kettle holes. These have also been

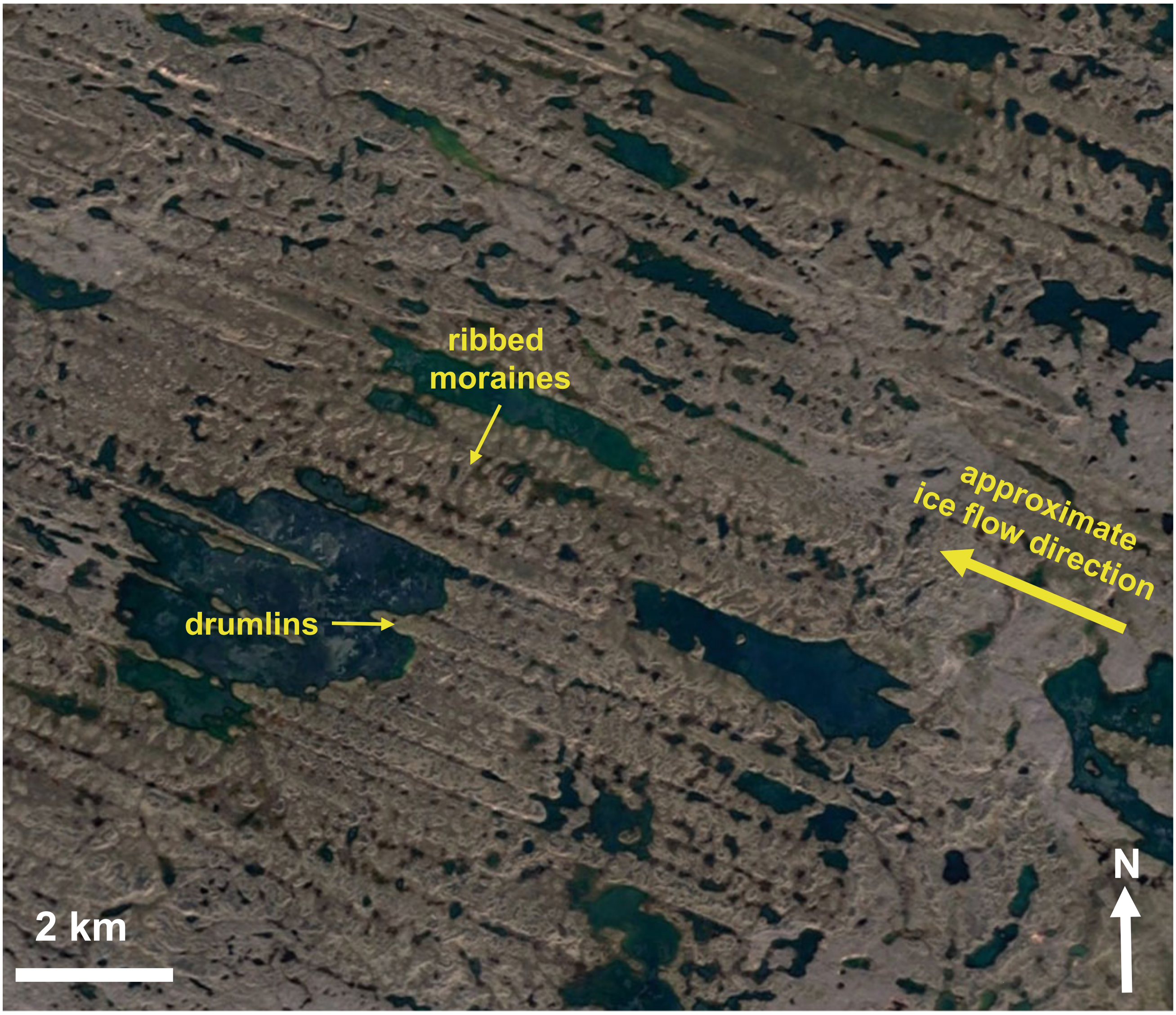

Fig. 4.5 Landsat image (data from Google Earth: Landsat/Copernicus) showing examples of where small ribbed moraines are superimposed on (and have partially 'broken up') glacial lineations in Zone 2, around 70 km north of Dubawnt Lake (located on Fig. 4.1). Image is centred on 63°57′58″ N, 101°33′46″ W. These elongate lineations form a part of the onset zone of the Dubawnt Lake palaeo-ice stream (see Sects. 4.2.3 and 4.3.3)

referred to as 'glacifluvial corridors' (St-Onge 1984) or 'glacifluvial corridor hummocks (GCHs)' (Utting et al. 2009), which are wide (several km) tracts of glacifluvial sand and gravel, often located in erosional corridors (Fig. 4.8) (Utting et al. 2009; Storrar and Livingstone 2017). They have been shown to occur extensively throughout the study area and, like eskers, can occur in quasi-dendritic systems that can extend for several tens of kilometres in length and regularly spaced 5–10 km apart (Utting et al. 2009). Detailed analysis of the hummocks associated with GCHs using both ground penetrating radar and pit excavations has revealed a single lithofacies consisting of coarsely stratified, matrix-supported gravely sand to a depth of several metres (Utting et al. 2009). Utting et al. (2009) suggested that this composition is similar to the 'sliding bed facies' commonly observed in esker sediments and other hyperconcentrated flow deposits, both of which are attributed to high meltwater discharges.

4.2.3 Zone 3

Zone 3 is characterised by a continuation of the dendritic networks of eskers, but here they are more commonly superimposed on extensive fields of drumlins and other streamlined forms, particularly in the inner part of the zone

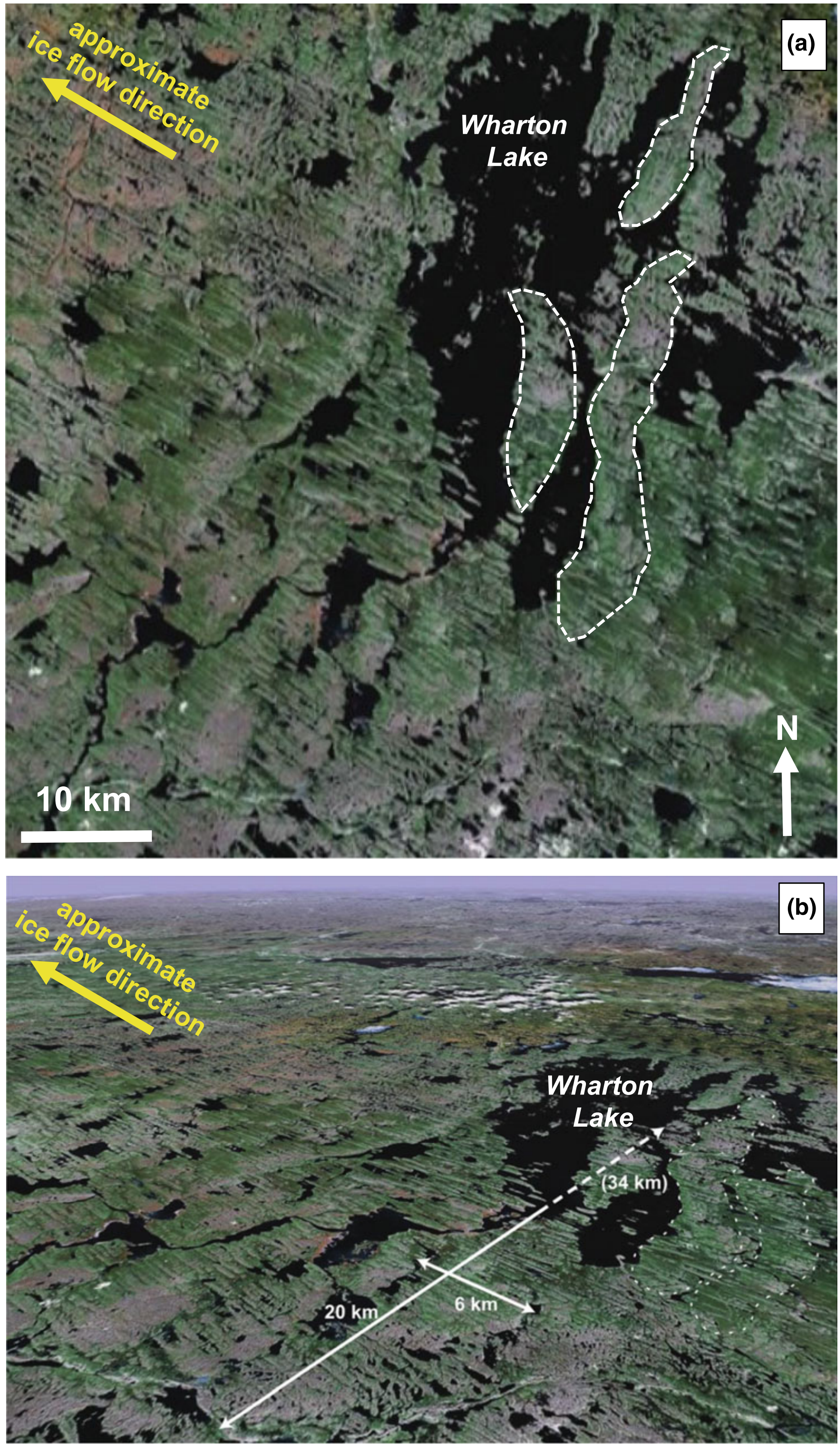

Fig. 4.6 Landsat images (data from Google Earth: Landsat/Copernicus) showing examples (white dashed polygons) of large 'till belts' or 'mega-scale transverse ridges' north-west of Dubawnt Lake. These features were first identified by Greenwood and Kleman (2010) who suggested that they form a coherent pattern in the landscape beneath the much smaller drumlins and ribbed moraines (examples shown in Figs. 4.3, 4.4 and 4.5). **a** Shows a Landsat image centred on 63°56′02″ N, 100°00′20″ W and **b** is an oblique view of the same area. Figures modified from Greenwood and Kleman (2010)

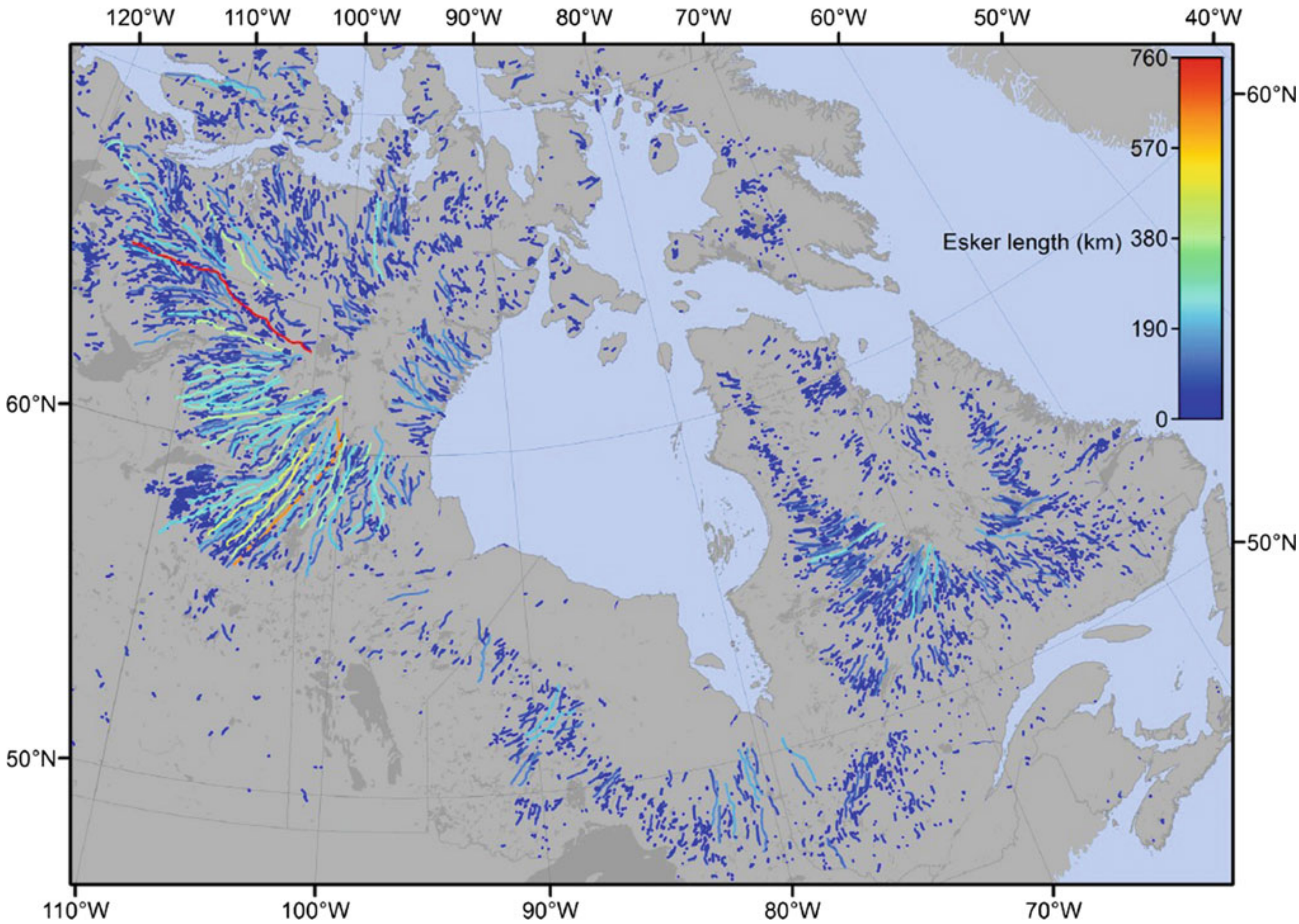

Fig. 4.7 Map of interpolated esker systems in Canada (i.e. tracing gaps conservatively between individual esker segments: from Storrar et al. 2014a). The northwestern Canadian Shield contains a high concentration of long esker systems and the large esker coloured red is probably the longest esker in the world (the 'Thelon esker': see Dyke and Dredge 1989) Also note the absence of eskers from under the inferred position of the Keewatin Ice Divide (Zone 1 on Fig. 4.1). Image taken from Storrar et al. (2014a)

where they are contiguous with similar features in Zone 2 (Shilts et al. 1987; Aylsworth and Shilts 1989a). Zone 3 is also characterised by a general decrease in till thickness from a fairly continuous mantle in the innermost portions to a discontinuous veneer towards the outer limits of the zone (Aylsworth and Shilts 1989a). Ribbed moraines also exist in Zone 3, but generally, occur as widely scattered fields of narrow linear trains in between the streamlined terrain (Shilts et al. 1987).

The streamlined glacial lineations in Zone 3 have attracted much attention in the literature, largely because of their extreme length (Tyrell 1897; Bird 1953; Dean 1953; Craig 1964; Ayslworth and Shilts 1989a; Stokes and Clark 2002; 2003a; Stokes et al. 2013). Although drumlins of classic form exist, it has been recognised for a long time that numerous features in Zone 3 are much more elongate (Bird 1953; Dean 1953; Craig 1964), with lengths approaching 20 km, but with widths <250 m wide (Aylsworth and Shilts 1989a). Some of the most spectacular features occur north of Dubawnt Lake (Fig. 4.9), where elongation (length:width) ratios reach a maximum of 149:1 (Stokes et al. 2013). Indeed, Stokes et al. (2013) found that ~25% of lineations in this region have elongation ratios >10:1. Bedforms with these extreme lengths and elongation ratios are typically referred to as 'mega-scale glacial lineations' (MSGLs: Clark 1993). In this location, MSGLs exist in an area of highly convergent flow, and Aylsworth and Shilts (1989a) noted that individual lineations can appear curved. It is also worth noting that much smaller (drumlin-like) features often occur interspersed with the much larger mega-scale glacial lineations (Stokes et al. 2013). Analysis of sediment exposures cut into the features by fluvial erosion (see Fig. 4.10) revealed a sandy till formed by a combination of non-pervasive subglacial sediment deformation and lodgement and which appears to have been reworked from pre-existing outwash sediments (Ó Cofaigh et al. 2013).

The most spectacular MSGLs on the Canadian Shield form a 'flow-set' that is developed on and down-ice from the

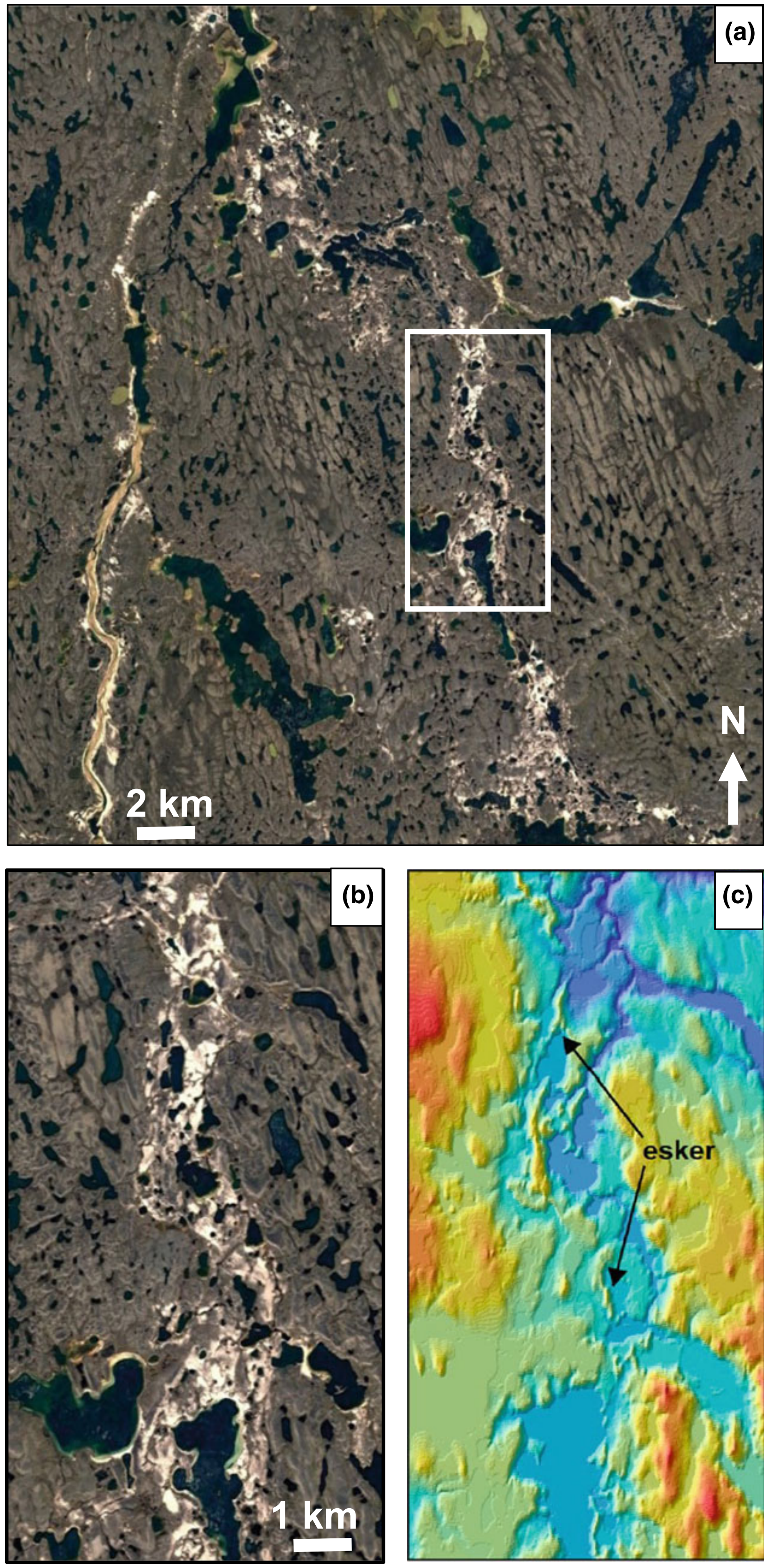

Fig. 4.8 Corridors of glacifluvial material ('glacifluvial corridor hummocks' (GCHs): Utting et al. (2009) characterised by a broad channels of sands and gravel and typically associated with esker ridges throughout Zones 2 and 3 (modified from Storrar and Livingstone 2017). Images in **a** and **b** from Google Earth (Landsat/Copernicus), with image (**a**) centred on 65°36′47″ N, 100° 30′37″ W, just south of the Garry Lake/Back River. Image (**c**) is a digital elevation model taken from Storrar and Livingstone (2017)

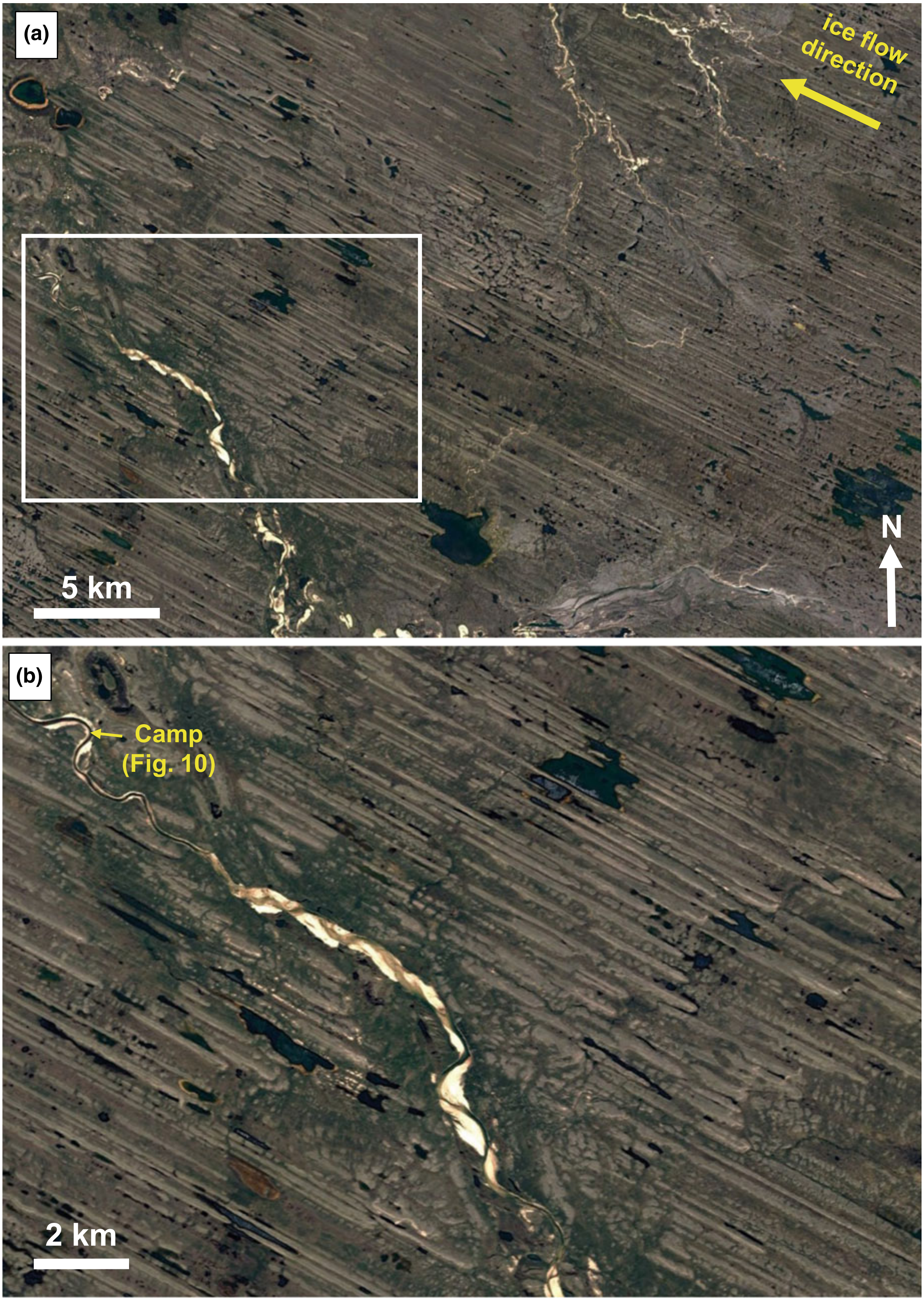

Fig. 4.9 Landsat images (data from Google Earth: Landsat/Copernicus) showing examples of mega-scale glacial lineations that exist in Zone 2 and Zone 3, north and northwest of Dubawnt Lake. **a** Illustrates the a large view of the landscape (image centred on 64°01′20″ N, 10202′51″ W) with **b** showing a close-up of features in the vicinity of the Finnie River (see also Fig. 4.10). Note that some of the features are >20 in length and are the largest glacial lineations identified above sea level anywhere on Earth

Fig. 4.10 **a** Photograph of a field camp on the Finnie River (see Ó Cofaigh et al. 2013) looking southwest across the field of mega-scale glacial lineations that form part of the Dubawnt Lake Ice Stream (location of photograph shown in Fig. 4.9. *Photograph* C. Stokes. **b** Shows an oblique aerial photograph of similar features nearby. *Photograph* C. Stokes

Thelon sedimentary basin (Figs. 4.9 and 4.10) where intense glacial erosion of poorly consolidated, unmetamorphosed clastic sedimentary rocks occurred. Aylsworth and Shilts (1989a) argued that the presence of highly elongate lineations may simply reflect an underlying geological control, in addition to any glacidynamic control related to increasing ice flow velocities away from the KID. In contrast, more recent work has recognised the importance of rapid ice flow, in the form of ice streaming, in helping explain the presence of MSGLS, with the flow-set being referred to as the Dubawnt 'surge fan' (Kleman and Borgström 1996) or the Dubawnt Lake Ice Stream (Stokes and Clark 2003a, b), see discussion in Sect. 4.3.3.

The eskers in Zone 3 are particularly well-developed (Fig. 4.7) and, as noted above, are the longest to be observed anywhere on the bed of the LIS (Storrar et al. 2014a). Shilts et al. (1987) noted that eskers are commonly associated (in space if not in time) with the scattered fields of ribbed moraine in Zone 3 (see also Stokes and Clark 2003a). They also noted that the spacing of eskers in Zone 3 was closer together (around 8 km) than in Zone 2 (13 km) and that although the number of eskers is greater in Zone 3 than Zone 2, there size (height and width) tends to decrease in Zone 3. Storrar et al. (2014a) also noted the remarkably consistent lateral spacing of eskers, which they measured to be around 12 km apart. Aylsworth and Shilts (1989a) observed that eskers in Zone 3 often cross-cut other landforms that were created by actively flowing ice, and that esker tributaries often join esker trunks at steep angles. They argued that this was evidence that the ice sheet had locally stagnated at the time of esker deposition.

4.2.4 Zone 4

The outermost Zone 4 is dominated by extensive areas of massive crystalline bedrock outcrop that is nearly everywhere devoid of till cover (Shilts et al. 1987; Aylsworth and Shilts 1989a) (Fig. 4.11). As such, streamlined features are rare and even eskers become gradually less well-developed and then absent towards the westward limit of Zone 4. Aylsworth and Shilts (1989a) attributed the lack of till cover and the general absence of glacial landforms to the fact that the underlying crystalline bedrock was generally resistant to glacial erosion and sediment production. As such, they argued that although the ice sheet dynamics may have been similar in this zone as in Zone 3, the depositional record is absent because of the scarcity of subglacial debris.

Aylsworth and Shilts (1989a, b, c) also noted that the lack of eskers in the westernmost part of the study area undoubtedly reflects a sediment poor part of the ice sheet related to the resistant bedrock and the lack of far-travelled material from Zone 3. Where eskers do occur (e.g. Fig. 4.11), it has been noted that they tend to be isolated, less-regularly spaced, and with few or no tributaries (Storrar et al. 2014a).

4.2.5 Major Moraine Ridges

Major end moraines, distinct from accumulations or aprons of from glacifluvial outwash, are not common on the northwestern Canadian Shield, although some occur in the northern and northwestern parts of the area (Aylsworth and Shilts 1989a). The largest end moraine system is the MacAlpine Moraine (Blake 1963, 1966; Falconer et al. 1965a). It comprises a complex of end moraine ridges associated with aprons of coalescing glacifluvial fans and deltas and clusters of small eskers (e.g. Fig. 4.12). Blake (1963) noted that in some places it is a sharp ridge that is only a few tens of metres wide, whereas at other points it may be several kilometres wide. It is, however, a conspicuous feature on the otherwise low-relief terrain, with Blake (1963) noting it reaches up to 70 m in height. A similar landform assemblage forms part of the Chantrey Moraine system to the north of the region and beyond onto Baffin Island (Falconer et al. 1965a; Dyke 1984). Together, these moraines are probably the world's longest system of end moraines (Falconer et al. 1965a, b; Dyke and Dredge 1989), which Falconer et al. (1965a, b) interpreted as the last climatically induced stillstand or readvance of the LIS prior to its demise. No other substantial end moraines lie up-ice of these systems in the study region (Dyke and Dredge 1989).

Aylsworth and Shilts (1989a) also drew attention to the Twin Jugs Moraine (Blake 1963), which is a prominent feature of the landscape in the western part of the study area. This is characterised by an extensive area of hummocky terrain with a few large eskers intersecting it. Further west, moraine systems are not obvious and, where present, may resemble eskers and areas of glacifluvial outwash, except for the fact that they are orientated parallel to the assumed ice margin positions (Ayslworth and Shilts 1989a). Aylsworth and Shilts (1989a) argued that the scarcity of major end moraines in the region is probably a function of both the passive nature of ice margin retreat and the paucity of debris available for moraine construction.

4.2.6 Features Related to Glacial Lakes and Marine Transgression

Wave-cut cliffs, beach ridges, spits, bars and other related features are common throughout the study area (e.g. Figs. 4.13 and 4.14), often in association with, but elevated above, present-day lake shorelines such as Dubawnt Lake (Tyrell 1897; Bird 1953; Lee 1959; Craig 1964; Craig and

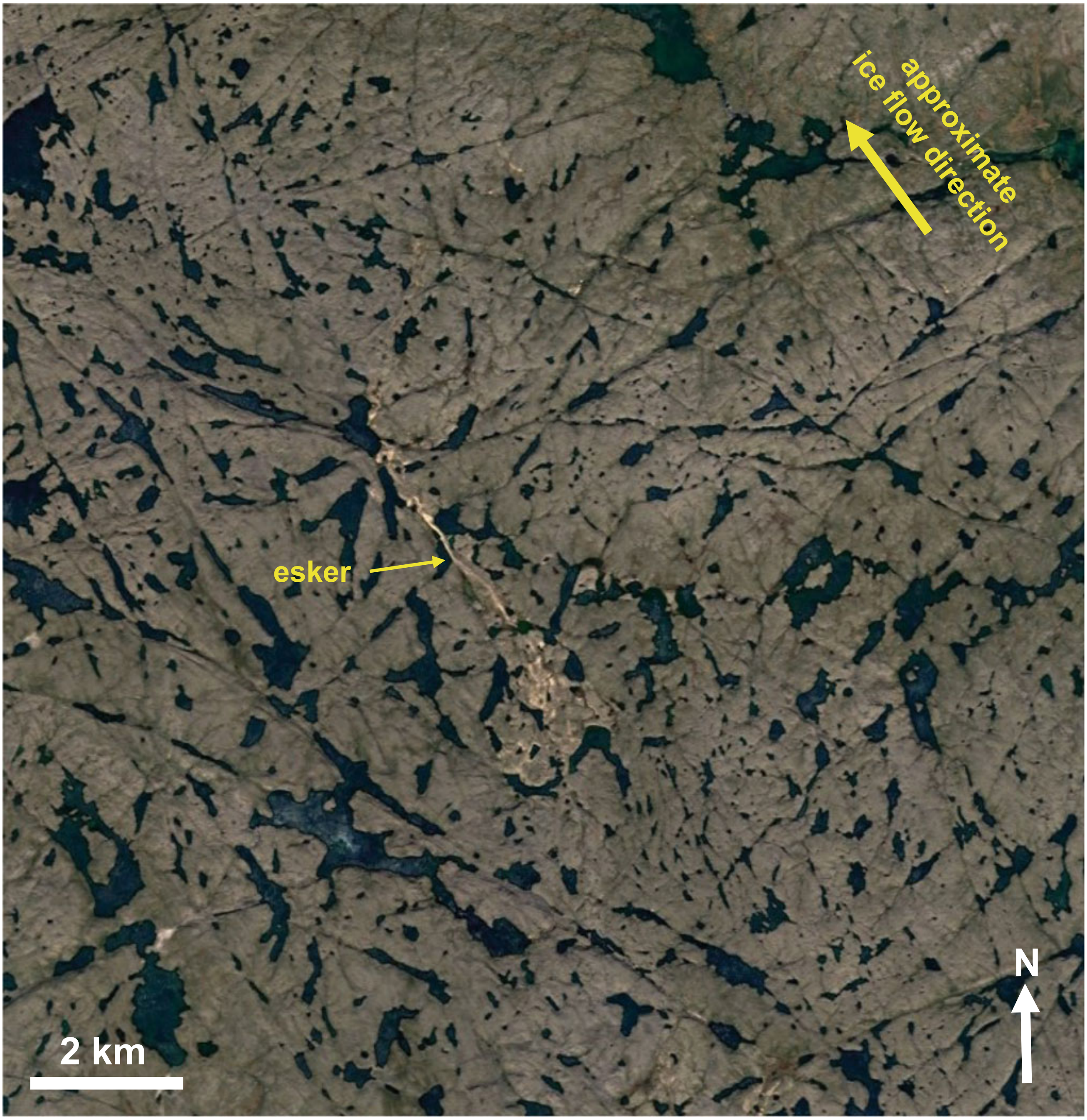

Fig. 4.11 Landsat image (data from Google Earth: Landsat/Copernicus) showing a typical area within Zone 4 of crystalline bedrock with very little till cover in the far north-west of the study area. Image is centred on 66°46′40″ N, 111°35′15″ W. Note the absence of any glacial lineations, but that a short esker likely reflects the ice flow direction towards the north-west during deglaciation

Fyles 1960; Stokes and Clark 2004). For a long time, it was recognised that proglacial lakes formed as the ice front retreated from the north-west to towards Hudson Bay and blocked the eastward drainage of meltwater (e.g. Tyrell 1897; Bird 1953; Craig and Fyles 1960). J. B. Tyrell was the first to record the elevations of some of these lakes, which he suggested were up to 225 m above sea level near Dubawnt Lake (Tyrell 1897). In the Thelon River Valley, Craig and Fyles (1960) noted that the ice sheet would have dammed the natural eastward drainage and that numerous beaches indicate a progressive eastward lowering of an ice-dammed lake as successively lower outlets were uncovered (see also

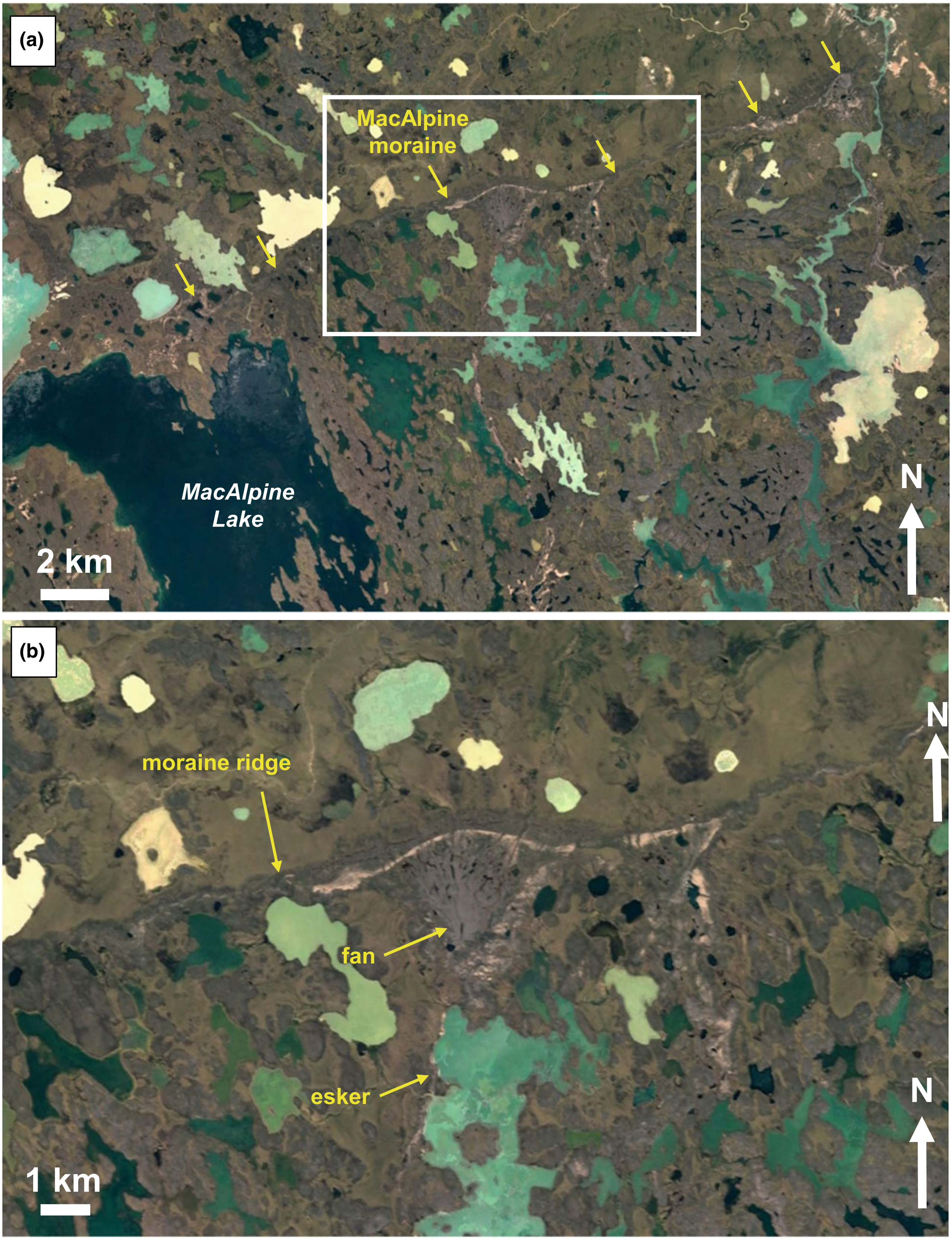

Fig. 4.12 Landsat images (data from Google Earth: Landsat/Copernicus) showing the MacAlpine moraine system immediately north-west of MacAlpine Lake in Zone 3. **a** Illustrates the moraine ridge running approximately south-west to north-east (image centred on 66°43′52″ N, 102°36′18″ W) and **b** is a close-up of an esker and glacifluvial fan that terminates at the moraine ridge

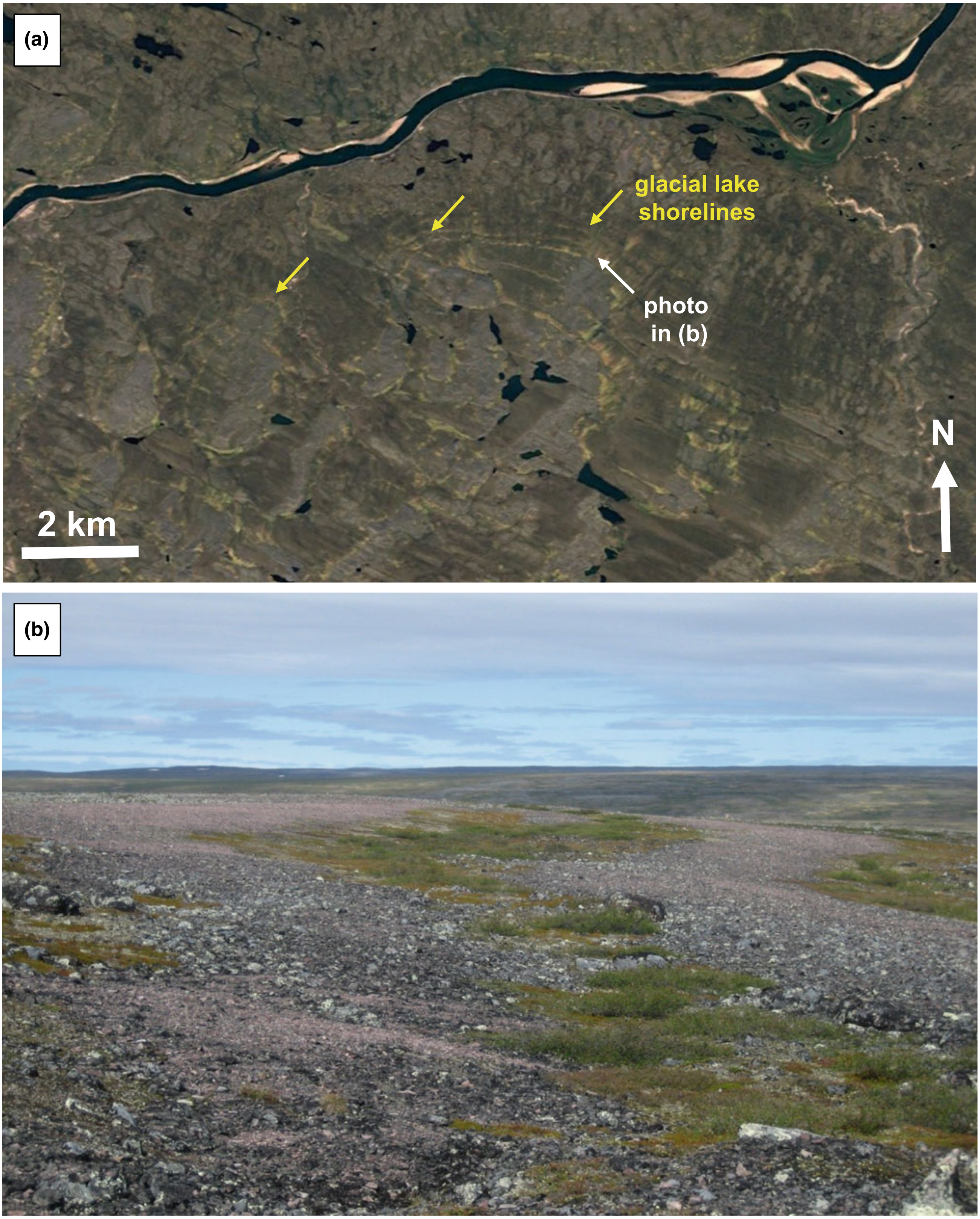

Fig. 4.13 **a** Numerous glacial lake shorelines in Zone 2, just south of the Thelon River (image is centred on 64°29′54″ N, 100°59′05″ W). Note glacial lineations in the bottom right indicating ice flow to the north-west. **b** Photograph of the beaches/shorelines looking north-west (approximate location shown on (a). *Photograph* C. Stokes)

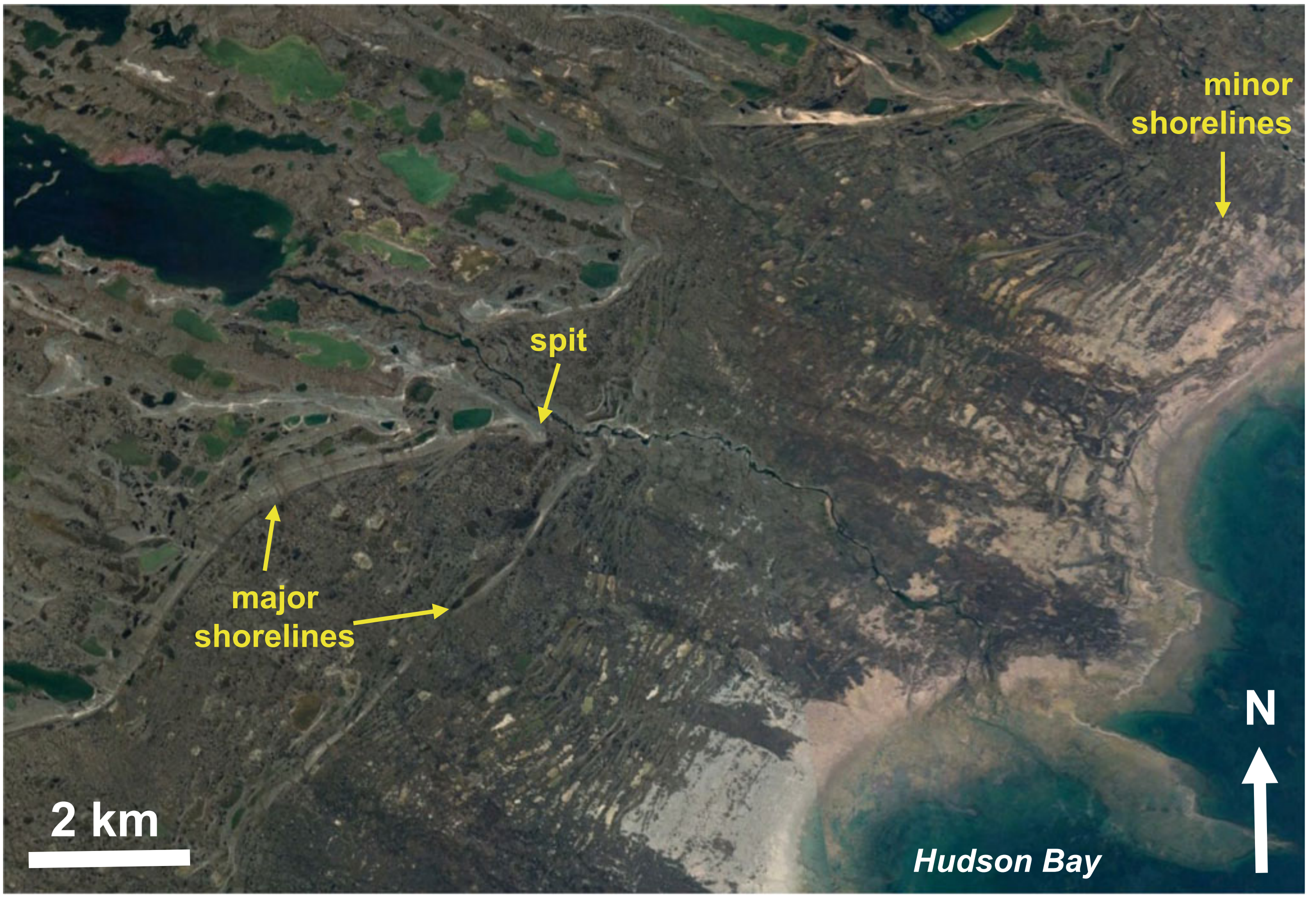

Fig. 4.14 Landsat image (data from Google Earth: Landsat/Copernicus) showing relict shorelines along the west coast of Hudson Bay, around 50 km north of Arviat (image is centred on 61°35′39″ N, 93°55′17″ W)

Bird 1953) (Fig. 4.13). Sediment stratigraphy also supports the existence of proglacial lakes, with several workers noting lacustrine sediments separate by organic layers and which may indicate ice advance through former proglacial lake basin (e.g. Taylor 1956).

The largest lake in the study area has been termed Glacial Lake Thelon (Craig 1964; Stokes and Clark 2004). Evidence from shorelines and spillways indicates that this lake was dammed by an ice lobe that flowed north-westward through the Thelon Sedimentary basin and which contains some of the most spectacular MSGLs described in Zone 3. Craig suggested that the lake may have initially drained to the south but that, in its later stages, abandoned channels indicate spillways to the north were occupied and drained into the Back River basin (Fig. 4.1). More recent work by Stokes and Clark (2004) used a reconstruction of ice margin positions (from Dyke and Prest 1987) and a digital elevation model to predict the evolution of lakes during ice margin retreat. This prediction of where lakes *should* exist supported previous work (e.g. Craig 1964) and the location of lakes matched well with the distribution of lacustrine sediments mapped on the Glacial Map of Canada (Prest et al. 1968). Their reconstruction also suggests that Glacial Lake Thelon was a large (7,500 km^2) and deep ($\sim$ 130 m) lake that would have exerted an important control on ice dynamics through its influence on calving and the drawdown of ice from the KID (see Sect. 4.3.3).

In addition to glacial lake shorelines, the isostatically depressed land surface was subjected to marine incursion during and immediately following deglaciation. The ancestral and much larger Hudson Bay was known as the Tyrell Sea and numerous strandlines and raised shorelines exist in the study area, with particularly remarkable flights of shingly beach ridges encircling present-day Hudson Bay (Dyke and Dredge 1989) (Fig. 4.14).

4.3 Implications for Laurentide Ice Sheet Dynamics

This section briefly highlights how the landform assemblages and patterns described in Sect. 4.2 have increased our understanding of the palaeo-glaciology of the LIS during the last glacial cycle, with particular reference to the Keewatin

Ice Divide (KID) (Sect. 4.3.1), meltwater drainage systems (Sect. 4.3.2), and fast-flowing ice streams (Sect. 4.3.3). It is important to note that many of these advances hold implications that extend well beyond the glacial history of the study area.

4.3.1 The Keewatin Ice Divide (KID)

The zonation of landforms described in Sect. 4.2 has been particularly important in terms of constraining the location and migration of the KID, which was the largest and highest feature in the ice surface topography of the LIS (Dyke and Prest 1987; Tarasov and Peltier 2004). The general absence of features in Zone 1, for example, is one of the main lines of evidence used to infer the presence of the ice divide, where ice thicknesses would have been large, but where ice velocities would have been negligible. Such conditions would have precluded subglacial erosion and transport and, where low-relief hummocky terrain exists in Zone 1 (Fig. 4.2), it has been suggested that it was formed by debris released from basal ice that was almost totally stagnant (Aylsworth and Shilts 1989a).

The general absence of eskers in Zone 1 is also conspicuous, but consistent with the location of an ice divide. For example, Storrar et al. (2014a) argued that any meltwater produced in that location would need to accumulate for a certain distance before it was able to form channels. They noted that the absence of esker ridges at the final location of the ice divide is rather abrupt, and the total width of the esker-free area is relatively constant at 100–150 km. They argued that this might relate to an underlying control on how far from ice divides eskers can form, perhaps related to the hydraulic potential and meltwater supply, or the availability of a sufficient quantity of sediment to build eskers. They also noted that the ice was likely cold-based under the ice divide until the final stages of deglaciation (Kleman and Glasser 2007), thus routing any surface-generated meltwater over the ice surface, rather than reaching the bed to form a channelized system, which may also explain their absence.

Aylsworth and Shilts (1989a) also noted that the presence of ribbed moraines in Zone 2, encircling the divide, may provide clues as to their mode of formation. They argued that the morphology of the ribbed moraines resembled inclined plates of sediment thrust one on top of the other and that this might reflect a compressive flow regime (see also Bouchard 1989) whereby englacially thrusted debris subsequently melted out during ice sheet stagnation. As such, they argued that the distribution of ribbed moraine in Zone 2 reflected the melting of a largely stagnant ice sheet. They also noted that esker tributaries in Zone 2 commonly join trunk eskers at right angles, implying that ice flow as sluggish or non-existent (see also Aylsworth and Shilts 1985). In contrast, Hättestrand and Kleman (1999) have argued that ribbed moraines are formed when a previously frozen till sheet is fractured apart during the transition to a warm-based, extensional, ice flow regime. Thus, Kleman and Hättestrand (1999) suggested that the widespread occurrence of ribbed moraines around the KID (and other core areas of the Laurentide and Fennoscandian ice sheets) implies that large areas of these ice sheets were cold-based at the LGM and, therefore, more likely to be high-domed and stable. In support of these arguments, it has also been recognised that there are numerous swarms of landforms on the Canadian Shield that must predate the LGM and that were protected by cold-based ice (Kleman et al. 2002, 2010). Kleman et al. (2010), for example, have used these relict flow-patterns to piece together the growth and evolution of ice dispersal centres prior to the LGM and they argued that they were dynamically independent for most of the Wisconsinan glaciation.

Although the zonation of landforms around the former location of the KID has long been recognised, more recent work has shown that the ice divide must have migrated to this final position (e.g. McMartin and Henderson 2004). The early work used ice flow indicators, particularly a dispersal train of red Dubawnt Supergroup rock types from the inferred position of the ice divide southeast towards Hudson Bay, to argue that it was a relatively static, long-lived feature (Shilts et al. 1979; Shilts 1980). Several authors have earlier recognised that the ice divide may have migrated towards its final position (e.g. Cunningham and Shilts 1977), but Aylsworth and Shilts (1989a) argued that 'it probably migrated no more than 100 km' (p. 3). More recently, however, McMartin and Henderson (2004) used the orientation of glacial landforms and small-scale glacial erosion features (e.g. striae) from over 800 sites to reconstruct the sequence of ice flows in the region. They argued that the profusion of multifaceted bedrock outcrops and intersecting striations, superimposed on streamlined landforms and stacked till units, was clear evidence of the migration of the main ice divide by as much as 500 km. Thus, whilst Aylsworth and Shilts' (1989a) Zone 1 (Fig. 4.1) is a reasonable approximation of the last position of the central axis of the KID, it is unlikely to reflect its position throughout much of glaciation and likely represents its final position having migrated from a position to the north-west (Cunningham and Shilts 1977; McMartin and Henderson 2004).

4.3.2 Subglacial Meltwater Drainage

As noted above, individual esker systems in the study area can be very long (several hundred kilometres when small gaps are taken into account: Fig. 4.7) and are developed into an integrated pattern with up to fourth-order tributaries that resembles a Hortonian drainage system (Shilts et al. 1987; Storrar et al. 2014a). It is unlikely that this whole system was formed subglacially across such a large area and, rather, several authors have inferred that it developed incrementally during deglaciation as the ice margin retreated (Shilts et al. 1987; Storrar et al. 2014a; Livingstone et al. 2015). They have argued that it was likely fed predominantly by surface meltwater reaching the bed in the ablation zone and that tunnels migrated headward (up-ice) and likely controlled by surface drainage patterns (Shilts et al. 1987; Livingstone et al. 2015).

The pattern and distribution of eskers throughout the study area have led to some important insights regarding the evolution of subglacial meltwater drainage systems during ice sheet deglaciation. Closer inspection of the esker patterns, supported by an extensive quantitative analysis (Storrar et al. 2014b), has also shown that there is a clear increase in esker density during deglaciation (e.g. from Zone 4, through to Zone 3 and into Zone 2: Fig. 4.1), but that the number of eskers with tributaries decreases. Storrar et al. (2014b) suggested that these patterns might be explained by increased surface meltwater production during deglaciation which led to a more efficient drainage system beneath the ice sheet as the ice margin retreated. An important implication, if correct, is that meltwater drainage systems of ice sheets appear to evolve in response to changes in meltwater input over hundreds of years, reminiscent of seasonal changes observed in much smaller valley glaciers (Hubbard and Nienow 1997).

4.3.3 Ice Streaming

The exceptionally long and spectacular glacial lineations north and north-west of Dubawnt Lake (see Sect. 4.2.3; Figs. 4.9 and 4.10) have attracted the attention of glacial geologists for over 100 years. Tyrell (1897) was probably the first to describe them and several studies in the 1950s and 1960s drew attention to their exceptional length and parallel conformity, particularly north and west of Dubawnt Lake (e.g. Dean 1953; Bird 1953; Craig 1964). Their highly elongate appearance was also noted by Aylsworth and Shilts (1989a), who were the first to speculate about the possible role of rapid ice flow (usually referred to as 'ice streaming' in an ice sheet setting) in explaining their great length. They also noted that, together, the highly elongate lineations form a distinctive 'bottle-neck' flow pattern that is clearly identifiable on the Glacial Map of Canada (Prest et al. 1968), i.e. with a broad zone of flow convergence narrowing into a main 'trunk' and then diverging to form a lobate terminus. Aylsworth and Shilts (1989a) also noted that parts of this flow pattern coincide with the relatively 'softer' rock of the Thelon sedimentary basin and they suggested that this geological influence might be more important than any change in the ice dynamics. That is, the highly elongate lineations might simply be a reflection of a softer and more erodible/deformable material beneath the ice, rather than a reflection of more rapid ice flow (ice streaming) in this region.

The distinctive bottle-neck flow was also mapped in Boulton and Clark's (1990) impressive synthesis and they attributed its formation to a late glacial event but did not speculate on the possible role of ice streaming. The interpretation that it represented a late-glacial event was supported by Kleman and Borgström (1996), who went one step further in suggesting that the flow-set was an exemplar of a 'surge fan'. Surge fans are thought to form during the decay stages of an ice sheet, often in relation to proglacial lake basins, and lineations are thought to form nearly synchronously over the whole fan area, rather than time-transgressively during steady ice margin retreat (Kleman and Borgström 1996). They referred to it as the 'Dubawnt fan', after the nearby Dubawnt Lake (see Fig. 4.1).

Building on this body of work, Stokes and Clark (2003a) used remote sensing techniques to undertake a detailed analysis of the landform assemblages comprising the Dubawnt fan/flow-set. They argued that the flow-set exhibited many of the criteria that would be expected to have formed beneath a spatially restricted zone of rapidly-flowing ice, i.e. a palaeo-ice stream (Stokes and Clark 1999, 2001). They named it the Dubawnt Lake Ice Stream and reconstructed its dimensions at ~ 450 km in length and ~140 km wide in the narrowest part of the bottle-neck. They also argued that it probably operated for only a few hundred years during deglaciation, and just prior to 8.2 ^{14}C ka BP, which is the age obtained for the MacAlpine moraines where the flow-set terminates (see Fig. 4.12 and Sect. 4.2.5).

Stokes and Clark (2003a, b) noted that although parts of the ice stream are underlain by the softer sedimentary rocks, including the extensive sandstones of the Thelon sedimentary basin, it is unlikely that the landform assemblage is simply a function of the underlying geology. In part, this is because there are gradual and predictable transitions in bedform elongation that match the expected velocity profile through the bottle-neck flow-set, rather than abrupt changes being observed at underlying geological boundaries (Stokes an Clark 2002). The shape of the flow-set itself is also clear evidence for spatially restricted rapid ice flow, which is the hallmark of ice streaming. Thus, they argued that there was clear evidence for ice streaming taking place during deglaciation and which likely played an important role in driving the south-eastward migration and eventual demise of the Keewatin Ice Divide (cf. McMartin and Henderson 2004).

At that time, the identification of a major palaeo-ice stream on the Canadian Shield challenged the paradigm that ice streams were largely restricted to soft-bedded areas of the ice sheet (see Clark 1992, 1994). Given the low relief, the location of the ice stream begged the question as to what may have triggered the fast flow. Stokes and Clark (2003a, b) had speculated that it might have been related to the development of proglacial lakes along this part of the ice sheet margin (see Sect. 4.2.6) and a more detailed analysis revealed that this was indeed the only part of the western margin of the Keewatin Sector where large and deep lakes formed. The ice stream is, therefore, thought to have been triggered by the development of Glacial Lake Thelon which induced high calving rates and the drawdown of ice from further up-ice (Stokes and Clark 2004).

As noted above, parts of the ice stream bed (e.g. in Zones 2 and 3) are characterised by ribbed moraines that are superimposed on the elongate bedforms produced by the ice stream (Fig. 4.5) (e.g. Aylsworth and Shilts 1989a; Stokes et al. 2006). This presents somewhat of a conundrum because ribbed moraines are not usually associated with rapid ice flow and yet they cover around 7% of the ice stream bed, with individual fields of ribbed moraine ranging in extent from <1 km^2 up to $\sim$2,500 km^2 (Stokes et al. 2006). Their patchy appearance on the ice stream bed is intriguing and their presence implies a marked change in ice dynamics following ice stream activity. Using observations of their morphometry, internal structure and sedimentological characteristics, Stokes et al. (2008) suggested that they formed in areas of the ice stream bed that stiffened as a result of dewatering of the till, e.g. through changes in subglacial meltwater drainage pathways or as a result of basal freeze-on (Christoffersen and Tulaczyk 2003). As such, they argued that patches of ribbed moraines represent sticky spots that developed during ice stream shut-down.

In summary, the identification of the Dubawnt Lake Ice Stream has proved important to a number of areas of research, with implications that extend beyond the glacial history of the Keewatin Sector of the LIS. Firstly, it is one of the largest and best-preserved palaeo-ice stream imprints that has so far been discovered on land and it has proved a useful template to identify other ice streams, including on the Canadian Shield (see review by Margold et al. 2015), but also further afield. Secondly, its large size and unusual location on the shield has enabled a broader understanding of the triggers and controls on ice streaming in ice sheets (cf. Winsborrow et al. 2010), demonstrating that ice streaming can occur over relatively hard bedrock terrain, likely triggered by proglacial lakes and facilitated via rapid basal sliding. Thirdly, the large size and reasonably well-constrained age bracket of the ice stream offers a useful analogue to understand the duration and propagation of ice sheet instabilities (Kleman and Applegate 2014) and the role of ice streams during ice sheet deglaciation (Stokes et al. 2016). It also offers a useful test for numerical ice sheet models that generate ice streams and can which can be tested against their 'known' location (e.g. Stokes and Tarasov 2010).

4.4 Conclusions

The glacial landform assemblages of mainland Nunavut, west of Hudson Bay, are some of the most spectacular and best-studied anywhere on Earth. At the Last Glacial Maximum, this region lay close to the centre of the former Laurentide Ice Sheet and was covered by a large ice dome (the Keewatin Dome) that was >3 km thick (Simon et al. 2014). This high point in the ice surface was associated with a major ice divide (the Keewatin Ice Divide) that was orientated north-east to south-west and which gradually migrated toward the south-east as deglaciation progressed (Lee et al. 1967; McMartin and Henderson 2004). It has, for a long time, been recognised that the Keewatin Ice Divide exerted an important control on the distribution of glacial landforms on the northwestern Canadian Shield (Ayslworth and Shilts 1985, 1989a, b, c; Shilts et al. 1987). These have been identified as four broadly concentric zones that include and encircle the location of the last inferred position of the ice divide (Fig. 4.1). Zone 1 is the innermost zone that includes the footprint of the ice divide itself. It is generally characterised by featureless till plains that lacking any obvious glacial landforms and/or low hummocky terrain and with a general absence of eskers. Zone 2 encircles most of Zone 1 and is characterised by the presence of well-developed linear trains or ribbed moraine that radiate out from the location of the former ice divide, but which are interspersed with fields of glacial lineations. Major esker systems dominate the landscape in Zone 2, where they reflect an extensive dendritic network in a pattern that also radiates away from the location of the ice divide. Individual esker ridges are some of the longest in the world with many tens of kilometres in length, and with some traceable for hundreds of kilometres with gaps. These esker systems continue into Zone 3, which is characterised by a fairly continuous cover of till and numerous drumlin fields that contain some highly elongate mega-scale glacial lineations. Towards the outer limits of Zone 3, till cover thins to a discontinuous veneer. Ribbed moraines also occur in Zone 3 but they are much less frequent than in Zone 2, and typically occur in isolated patches as linear trains interspersed with or superimposed on drumlins and mega-scale glacial lineations. The outermost Zone 4 comprises extensive areas of massive crystalline bedrock that are typically devoid of any till cover and with far fewer obvious glacial landforms and only sporadic eskers. Major moraine ridges are not common within the study area, but the largest and most obvious moraine systems (the MacAlpine and Chantrey Inlet moraines) can be traced for several hundred kilometres and likely represent a pause or readvance of the ice sheet during overall retreat (Falconer et al. 1965a, b). Throughout all four zones, relict shorelines, beaches, strandlines and spillways occur that demarcate the growth and drainage of numerous proglacial lakes, in addition to marine transgression of the ancestral Hudson Bay (Tyrell Sea).

Aylsworth and Shilts (1989a) noted that the distribution of glacial landforms on the northwestern Canadian Shield reflects "a complex and yet imperfectly understood interplay between dynamic conditions within the ice sheet and the geology of the bed across which it flows" (p. 19). It is clear that the location of the Keewatin Ice Divide was an important control on the broad-scale zonation of landforms, including both their formation and preservation (e.g. in relation to cold-based ice); but it is also clear that the underlying geology influenced the amount of sediment that could be eroded and remobilised at the ice sheet bed, with some of the most spectacular features forming over the relatively less resistant sedimentary outcrops (Aylsworth and Shilts 1989a). More recent work has reinterpreted some of these landscapes within a new paradigm that recognises the importance of ice streams in palaeo-ice sheet dynamics (Stokes and Clark 2001). Specifically, the Dubawnt Lake palaeo-ice stream (Stokes and Clark 2003a, b) is one of the largest and best-studied ice streams in the LIS and its rapid ice flow is now known to have created a distinctive bottle-neck flow-set through the centre of the study area. Its activity is able to explain the formation of the spectacular mega-scale glacial lineations north of Dubawnt Lake (Stokes et al. 2013) and the final southwestward migration of the ice divide (McMartin and Henderson 2004). The impressive network of eskers has also been subjected to detailed quantitative analysis (Storrar et al. 2014a, b), which has revealed that increased surface meltwater production during deglaciation led to a more efficient drainage system beneath the ice sheet.

Acknowledgements I would like to acknowledge several colleagues who I have had the pleasure to work with on the glacial history of the northwestern Canadian Shield (and some of whom have shared in some memorable fieldwork in the region), namely, Chris Clark, Olav Lian, Slawek Tulaczyk, Colm Ó Cofaigh, Matteo Spagnolo, Rob Storrar and Martin Margold. These collaborations were largely funded through grants from the NERC (NER/M/S/2003/00050 and NE/J00782X/1), and a Philip Leverhulme Prize awarded to CRS.

References

Aylsworth JM, Shilts WW (1985) Glacial features of the west central Canadian Shield. Curr Res Part B Geol Surv Can Paper 85-1B, 375–381

Aylsworth JM, Shilts WW (1989a) Glacial features around the Keewatin Ice Divide: Districts of Mackenzie and Keewatin. Geol Surv Can Paper 88–24, p 21

Aylsworth JM, Shilts WW (1989b) Glacial features around the Keewatin Ice Divide: Districts of Mackenzie and Keewatin. Geol Surv Can Map 24-1987, scale 1:1,000,000

Aylsworth JM, Shilts WW (1989c) Bedforms of the Keewatin Ice Sheet, Canada. Sediment Geol 62:407–428, 25:215–230

Bird JB (1953) The glaciation of central Keewatin, Northwest Territories, Canada. Am J Sci 251:215–230

Blake W Jr (1963) Notes on glacial geology, northeastern District of Mackenzie. Geol Surv Can, Paper 63-28

Blake W Jr (1966) End moraines and deglaciation chronology in northern Canada, with special reference to southern Baffin Island. Geol Surv Can Paper 66–26, p 31

Bouchard MA (1989) Subglacial landforms and deposits in central and northern Quebec, Canada, with emphasis on Rogen moraines. Sediment Geol 62:293–308

Boulton GS, Clark CD (1990) A highly mobile Laurentide ice sheet revealed by satellite images of glacial lineations. Nature 346:813–817

Christoffersen P, Tulaczyk S (2003) Signature of palaeo-ice stream stagnation: till consolidation induced by basal freeze-on. Boreas 32:114–129

Clark CD (1993) Mega-scale glacial lineations and cross-cutting ice-flow landforms. Earth Surf Process 18:1–29

Clark PU (1992) Surface form of the southern Laurentide Ice Sheet and its implications for ice sheet dynamics. Geol Soc Am Bull 104:595–605

Clark PU (1994) Unstable behavior of the Laurentide ice sheet over deforming sediment and its implications for climate change. Quat Res 41:19–25

Craig BG (1964) Surficial geology of east-central District of Mackenzie. Geol Surv Can Bull 99

Craig BG (1965) Glacial Lake McConnell, and the surficial geology of parts of Slave River and Redstone River map-areas, District of Mackenzie. Geol Surv Can Bull 122:33

Craig BG, Fyles JG (1960) Pleistocene geology of Arctic Canada. Geol Surv Can, no 60-10, p 21

Cunningham CM, Shilts WW (1977) Surficial geology of the Baker Lake area, District of Keewatin. Geol Surv Can, Paper 77-1B, pp 3111–3314

Dean WG (1953) The drumlinoid land forms of the Barren Grounds. Can Geogr 1:19–30

Dunlop P, Clark CD (2006) The morphological characteristics of ribbed moraine. Quat Sci Rev 25:1668–1691

Dyke AS (1984) Quaternary geology of Boothia Peninsula and northern District of Keewatin, Central Canadian Arctic. Geol Surv Can Memoir 407:26

Dyke AS, Prest K (1987) Late Wisconsinan and Holocene history of the LaurentideIce Sheet. Géog Phys Quat 41(2):237–263

Dyke AS, Dredge LA (1989) Quaternary geology of the northwestern Canadian Shield. In: Fulton RJ (ed) Quaternary Geology of Canada and Greenland. Geology of Canada Series, no. 1. Geological Survey of Canada, Ottawa, pp 189–214

Dyke AS, Vincent J-S, Andrews JT, Dredge LA, Cowan WR (1989) The Laurentide Ice Sheet and an introduction to the Quaternary geology of the Canadian Shield. In: Fulton RJ (ed) Quaternary Geology of Canada and Greenland. Geology of Canada Series, no 1. Geological Survey of Canada, Ottawa, pp 178–189

Falconer G, Ives JD, Loken OH, Andrews JT (1965a) Major end moraines in easternmn and central Arctic Canada. Geogr Bull 7:137–153

Falconer G, Andrews JT, Ives JD (1965b) Late-Wisconsin end moraines in Northern Canada. Science 147:608–610

Fyles JG (1955) Pleistocene features. In: Wright GM (ed) Geological notes on central District of Keewatin, Northwest Territories. Geological Survey of Canada Paper, 55-17
Greenwood SL, Kleman J (2010) Glacial landforms of extreme size in the Keewatin sector of the Laurentide Ice Sheet. Quat Sci Rev 29:1894–1910
Hättestrand C, Kleman J (1999) Ribbed moraine formation. Quat Sci Rev 18:43–61
Hubbard B, Nienow P (1997) Alpine subglacial hydrology. Quat Sci Rev 16:939–955
Kleman J, Borgström I (1996) Reconstruction of palaeo-ice sheets: the use of geomorphological data. Earth Surf Process 21:893–909
Kleman J, Glasser NF (2007) The subglacial thermal organisation (STO) of ice sheets. Quat Sci Rev 26:585–597
Kleman J, Hättestrand C (1999) Frozen-bed Fennoscandian and Laurentide ice sheets during the Last Glacial Maximum. Nature 402:63–66
Kleman J, Applegate PJ (2014) Durations and propagation patterns of ice sheet instability events. Quat Sci Rev 92:32–39
Kleman J, Fastook J, Stroeven AP (2002) Geologically and geomorphologically constrained numerical model of Laurentide Ice Sheet inception and build-up. Quat Int 95–96:87–98
Kleman J, Jansson K, De Angelis H, Stroeven AP, Hättestrand C, Alm G, Glasser N (2010) North American ice sheet build-up during the last glacial cycle, 115-21 kyr. Quat Sci Rev 29:2036–2051
Lee H (1959) Surficial geology of southern District if Keewatin and the Keewatin Ice Divide, northwest Territories. Geol Surv Can Bull 51
Lee HA, Craig BG, Fyles JG (1957) Keewatin Ice Divide (abstract). Geol Soc Am Bull 51, 42 pp
Livingstone SJ, Storrar RD, Hillier JK, Stokes CR, Clark CD, Tarasov L (2015) An ice sheet scale comparison of eskers with modelled subglacial drainage routes. Geomorphology 246:104–112
Margold M, Stokes CR, Clark CD (2015) Ice streams in the Laurentide Ice Sheet: identification, characteristics and comparison to modern ice sheets. Earth Sci Rev 143:117–146
McMartin I, Henderson PJ (2004) Evidence from Keewatin (central Nunavut) for paleo-ice divide migration. Géog Phys Quat 58 (2/3):163–186
Ó Cofaigh C, Stokes CR, Lian OB, Clark CD, Tulaczyk SM (2013) Formation of mega-scale glacial lineations on the Dubawnt Lake Ice Stream bed: 2. Sedimentology and stratigraphy. Quat Sci Rev 77:190–209
Prest VK (1969) Retreat of Wisconsin and Recent Ice in North America. Geological Survey of Canada Map, 1257A, scale 1:5,000,000
Prest VK, Grant DR, Rampton VN (1968) Glacial Map of Canada. Geological Survey of Canada, Map, p 1253A
Simon K, James TS, Forbes DL, Telka AM, Dyke AS, Henton JA (2014) A relative sea level history for Arviat, Nuanvut, and implications for Laurentide Ice Sheet thickness west of Hudson Bay. Quat Res 82:185–197
Shilts WW (1977) Geochemistry of till in perennially frozen terrain of the Canadian Shield—application to prospecting. Boreas 6:203–212
Shilts WW (1980) Flow patterns in the central North American Ice Sheet. Nature 286:213–218
Shilts WW (1984) Esker sedimentation models, Deep Rose Lake map-area, Distruict of Keewatin. Current Research in Canada, Part B. Geological Survey of Canada Paper 84-1B, pp 217–222
Shilts WW, Cunningham CM, Kaszycki CA (1979) Keewatin Ice Sheet —re-evaluation of the traditional concept of the Laurentide Ice Sheet. Geology 7(11):537–541
Shilts WW, Aylsworth JM, Kaszycki CA, Klassen RA (1987) Chapter 5: Canadian shield. In: Graf WL (ed) Geomorphic systems of North America. Boulder, Colorado, Geol Soc Am, Centennial Special Volume 2, pp 119–161
St-Onge DA (1984) Surficial deposits of the Redrock Lake area, District of Mackenzie. Curr Res Part A, Geological Survey of Canada, Paper 84-1A, 271–277
Stokes CR, Clark CD (1999) Geomorphological criteria for identifying Pleistocene ice streams. Ann Glaciol 28:67–74
Stokes CR, Clark CD (2001) Palaeo-ice streams. Quat Sci Rev 20:1437–1457
Stokes CR, Clark CD (2002) Are long subglacial bedforms indicative of fast ice flow? Boreas 31:239–249
Stokes CR, Clark CD (2003a) The Dubawnt Lake palaeo-ice stream: evidence for dynamic ice sheet behaviour on the Canadian Shield and insights regarding the controls on ice stream location and vigour. Boreas 32:263–279
Stokes CR, Clark CD (2003b) Laurentide ice streaming on the Canadian Shield: a conflict with the soft-bedded ice stream paradigm? Geology 31:347–350
Stokes CR, Clark CD (2004) Evolution of late glacial ice-marginal lakes on the northwestern Canadian Shield and their influence on the location of the Dubawnt Lake palaeo-ice stream. Palaeogeogr Palaeoecl 215:155–171
Stokes CR, Tarasov L (2010) Ice streaming in the Laurentide Ice Sheet: a first comparison between data-calibrated numerical model output and geological evidence. Geophys Res Lett. https://doi.org/10.1029/2009GL040990
Stokes CR, Clark CD, Lian O, Tulaczyk S (2006) Geomorphological map of ribbed moraine on the Dubawnt Lake Ice Stream bed: a signature of ice stream shut-down? J Maps 2006:1–9
Stokes CR, Lian OB, Tulaczyk S, Clark CD (2008) Superimposition of ribbed moraines on a palaeo-ice stream bed: implications for ice stream dynamics and shutdown. Earth Surf Process 33:593–609
Stokes CR, Spagnolo M, Clark CD, Ó Cofaigh C, Lian OB, Dunstone RB (2013) Formation of mega-scale glacial lineations on the Dubawnt Lake Ice Stream bed: 1. Size, shape and spacing from a large remote sensing dataset. Quat Sci Rev 77:190–209
Stokes CR, Margold M, Clark CD, Tarasov L (2016) Ice stream activity scaled to ice sheet volume during Laurentide ice sheet deglaciation. Nature 530:322–326
Storrar RD, Livingstone SJ (2017) Glacial geomorphology of the northern Kivalliq region, Nunavut, Canada, with an emphasis on meltwater drainage systems. J Maps 13(2):153–164
Storrar RD, Stokes CR, Evans DJA (2013) A map of Canadian eskers from Landsat satellite imagery. J Maps 9(3):456–473
Storrar RD, Stokes CR, Evans DJA (2014a) Morphometry and pattern of a large sample (>20,000) of Canadian eskers and implications for subglacial drainage beneath ice sheets. Quat Sci Rev 105:1–25
Storrar RD, Stokes CR, Evans DJA (2014b) Increased channelization of subglacial drainage during deglaciation of the Laurentide Ice Sheet. Geology 42:239–242
Tarasov L, Peltier WR (2004) A geophysically constrained large ensemble analysis of the deglacial history of the North American ice-sheet complex. Quat Sci Rev 23:359–388
Taylor RS (1956) Glacial geology or north-central Keewatin, Northwest Territories, Canada. Bull Geol Soc Am 67:943–956
Tyrell JB (1897) Report on the Dubawnt, Kazan and Ferguson Rivers. Canada Geological Survey Annual Report, 1986, Part E
Utting DJ, Ward BC, Little EC (2009) Genesis of hummocks in glaciofluvial corridors near the Keewatin Ice Divide, Canada. Boreas 38:471–481
Winsborrow MCM, Clark CD, Stokes CR (2010) What controls the location of ice streams? Earth-Sci Rev 103:45–59

Chris Stokes is a Professor in the Department of Geography at Durham University, UK. Following a Ph.D. (Geography) at the University of Sheffield (UK), he was appointed to a Lectureship at the University of Reading (UK). In 2007, he moved to take up a position as Reader in the Quaternary Environmental Change research group in the Department of Geography at Durham University, UK, where he was awarded a Personal Chair in 2014. His research focusses on glaciers and their interaction with the climate system, and ranges from monitoring small mountain glaciers over the last few decades to large-scale reconstructions of ice sheets over tens of thousands of years. His main research tool is satellite imagery, which affords a large-scale view of glaciers and glacial landscapes, but he has also visited the field in Arctic Canada, Norway, Svalbard, Russia and Patagonia. He has authored over 100 scientific papers and his work has been covered in both national and international media. He was awarded a prestigious Philip Leverhulme Prize (in 2009) in recognition of his research contributions, and has been awarded the British Society for Geomorphology's Gordon Warwick Medal (2013) and Wiley Award (2008). He currently serves on the Editorial Boards for two international journals and as an Editorial Advisor for a third. He is also a former President of the British Branch of the International Glaciological Society.