2 Genesis of Islands

The larger the island of knowledge, the longer the shoreline of wonder.

Ralph W. Sockman

Abstract

This chapter deals with the various primary processes of island genesis (volcanic, tecto-orogenetic, sedimentary or coralline) before addressing secondary processes such as isostatic subsidence, emersion, marine ingression, erosion, horizontal and vertical drift as well as doming. In most cases islands are formed as a result of several processes, although this varies with island size: the smaller the island, the more likely it is that a single dominant process is responsible for its genesis. Examples are provided for each of the different island types.

Keywords

Primary and secondary island formation processes • Plate tectonics • Volcanism • Zoogenous processes • Emergence • Built-up • Subsidence • Dislocation • Doming

Although it may seem obvious, islands are not simply there. Islands are indicators of underlying natural processes – outposts so to speak of the physical global realm. Physical-geological processes come into play just like biological natural processes. A rough classification of island-forming processes would include volcanism, tectonic uplift or folding and sedimentation on the physical side and the formation of coral as a biogenic process. Secondary processes such as lowering sea levels, isostatic uplift, dislocation, salt tectonics and erosion can give rise to islands that are continental outposts in the sea. Often, islands are formed by a combination of processes whose specificity depends on the island's size. The smaller the island, the more likely it is that only one dominant process is responsible for its genesis (e.g. volcanic, tectonic, biogenic). The larger the island, the more physical processes are likely to come into play, e.g. the drifted orogenic folding of Corsica or the folding, volcanism and sedimentation that gave rise to Cuba. The small island of Nauru is also of mixed type, where the primary genetic process is emersion, but the overall island

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B.M.W. Ratter, *Geography of Small Islands*, https://doi.org/10.1007/978-3-319-63869-0_2

is best described as a raised coral atoll. Processes of island formation interact with global changes such as changing sea levels across time and space. The result is a hotchpotch of large, medium and small islands that come and go, scattered in their great variety across the oceans of the world (see Depraetere and Dahl 2007; Nunn 2007, 2009).

First attempts to categorise islands were made by Charles Darwin in his famous book *On the Origin of Species* (1859). Alfred R. Wallace (1880) picked up on this, choosing to categorise islands according to their geological structure. This led him to differentiate between *continental islands* ('detached sections of continents') and *oceanic islands* ('never formed part of a continent or any large mass of land'). This differentiation between continental and oceanic islands is still frequently used today. But then, is this the real difference? Put very simply, an island is a piece of land surrounded by water. Looking at the global distribution of islands, beyond the islands and archipelagos of the Pacific, the majority of islands are located off the continental coasts, linking them to natural processes occurring on the continental edge.

From the 1960s onwards, following the discovery of the mid-ocean ridges and recognising the significance of Alfred Wegener's theory of continental drift proposed in 1912, scientists began to understand that many questions surrounding the formation and distribution of islands could be explained through modern plate tectonics (Rücklin 1963).

Several scientists have attempted to classify islands based on their genesis. Yet astonishingly few classification attempts draw on the fundamentals of island formation (e.g. Arnberger and Arnberger 1988; Grigor'yev 1971; Klug 1985; Depraetere and Dahl 2007). A particularly well-founded example goes back to the Russian geographer Grigor'yev who was critical of Darwin's differentiation: 'although such a division reflects the basic structural differences between continental and oceanic islands, it tends to be a rather arbitrary distinction' (Grigor'yev 1971). Disagreeing with the simplicity of Darwin's approach, Grigor'yev developed an alternative genetic classification based on the character of the underlying crust. Each type of crust is associated with particular types of islands, leading to continental, oceanic and intermediate islands. These three groups, in turn, are broken down further based on their structural geology and where they evolved on the respective crust types. The group of continental islands, for example, includes two structural types (shelf islands and fragments of continental platforms), the intermediate group three types (continental island arcs, oceanic island arcs and volcanic islands on the continental slope) and oceanic islands a single volcanic type (see Grigor'yev 1971: 586).

The German geographer Heinz Klug (1985) developed a more detailed geographical island categorisation based on dominant natural processes. He not only considered the original material of which the island is formed but also subsequent processes of isolation and change. Excluding freshwater islands, he distinguishes nine general island types and four types of genesis:

1. *Dislocation* of part of the mainland through marine ingression, erosion or tectonic dislocation
2. *Doming* of a small part of the seafloor as a result of salt tectonics or underground intrusion
3. Extensive *emersion* of the seafloor on account of isostatic or epeirogenic compensatory movement
4. *Build-up* from the seafloor through tectonic or volcanic events or zoogenous processes

These four general types were differentiated further according to the specific processes involved, ranging from marine ingression and regression to erosion, accumulation processes and shift-causing tectonics. In all, 16 island types were identified.

Coastal morphologist Ludwig Ellenberg has pointed to certain weaknesses of Klug's systematic typology. Despite the clear intention of differentiating islands according to their genesis, he argues that this typology cannot be consistently adhered to as islands are not the result of a single process. Further difficulties result from the distinction between salt tectonics and other tectonic lifting processes including orogenesis or upthrust, as well as the problem that land destruction and isolation cannot be separated as strictly as Klug's categories may suggest. As ever, the devil is in the detail. Islands – just as coasts – are particularly difficult to classify, resisting even broad typologies. Formed by the interplay of terrestrial and littoral processes, coasts and islands are comparable in their genetic complexity.

Despite these difficulties, I find the straightforward distinction between continental and oceanic islands so common in the anglophone literature too simple. For island researchers, a little more differentiation is called for. A typology of islands should at least distinguish four primary genetic processes (see Table 2.1), i.e. volcanic islands, tecto-orogenetic islands, sedimentary islands and coralline islands, as well as five secondary principles of isolation (see Table 2.2), namely, isostatic uplift and subsidence, plate tectonic processes, sea level fluctuations and resulting processes. Each of these processes will be discussed in the following in more detail.

2.1 Volcanic Islands

Many islands and groups of islands are of volcanic origin. Many volcanic islands, such as Réunion or Mauritius in the Indian Ocean, have a long history; these specific examples formed about 3.5 million and 8 million years ago, respectively. The Pacific Galapagos Islands are probably more than five million years old, but some of their westernmost islands, which are the most volcanically active, may only be hundreds of thousands of years old – in fact, they are still being formed today. Measured from the seafloor upwards, volcanic islands are among the highest mountains on Earth, with about 90% of their massive volume hidden below the sea. The volcanic island of Gunung Api (the name simply

Table 2.1 Typology of islands – part 1: primary processes

		Processes	
Island type	General genesis	*Special form*	Examples
Primary processes			
Volcanic islands	Volcanism	Subduction, convergence, divergence	Lesser Antilles, Mariana Islands, Réunion, Mauritius, Tonga, Solomon, Japan
		Island arc volcanism	
		Hotspot activity	Hawaii, Comoros, Ascension, St. Helena
Tecto-orogenetic islands	Orogenic folding	Tecto-orogenetic formation of strata with denudation	New Zealand, Puerto Rico, Balearic Islands
		Faulting tectonics	Hispaniola
Sedimentary islands	Accumulation by sedimentation	Marine sand aggradation, wind action, formation of dunes	Norderney, Padre Island, Mississippi-Alabama Barrier Islands
		Geest core sedimentation initialisation	Sylt, Amrum, Föhr, Agalega Islands
		Silt aggradation, silt sedimentation islands	Gröde, Langeneß, Oland, German Halligen
		Hallig	
Coralline islands	Zoogenic formation	Fringing reef or barrier reef formation	Barrier Reef Belize, Great Barrier Reef Australia
		Shelf reef islands	
		Ring-shaped reef or chain of islands; *evolution* on rim of a subsided deep-sea volcano	Bikini, Maldives, Chagos Archipelago, Diego Garcia, Henderson
		Atoll	

Table 2.2 Typology of islands – part 2: secondary processes

Island type	General genesis	Processes *Special form*	Examples
Secondary processes			
Subsidence islands	Isostatic land subsidence, submergence	Epirogenesis processes, separation from continental land mass through subsidence	Bissagos Islands, Sansibar, Pemba, Mafia
Ingression islands	Marine ingression	Eustatic sea level changes, sea level rise	Rottnest, Garden Island, Gotland, Oland, Hiddensee, Ilha do Governador, Djerba
Emerging islands	Isostatic emersion	Isostatic rebound and/or epigenetic uplift	Santiago, São Nicolau, Finnish skerries
Residual islands	Erosion	Erosion in combination with marine regression	Pellworm
		Morphologically resistant outliers of the mainland *Outlier island*	Gorée, Guernsey, Jersey
Dislocation islands	Horizontal drift	Tectonic separation and migration of peripheral plate parts *Drift island*	Madagascar, Seychelles, Corsica, Sardinia
	Vertical movement	Vertical lifting of blocks *Horst island*	Bornholm, Malta, Gozo, Comino, Lampedusa
Diapir islands	Tectonical doming	Doming caused by salt tectonics	Helgoland, Zarqua

means ‘volcano’ in Indonesian) dominates the entire Indonesian Banda group. This active volcano only rises 640 m above sea level but is located within two calderas that are part of a huge submarine volcanic cone (see Fig. 2.1) (Ritchie and Gates 2001: 21).

The Hawaiian archipelago is one of the best known groups of volcanic islands, with Mauna Kea not only representing the largest volcano in the world but also the highest mountain on earth, measuring 10,205 m from the seafloor. Even above sea level, this stately shield volcano still reaches 4205 m (see Wolfe et al. 1997). Iceland is the largest volcanic island in the world. Here, like on Hawaii, the volcanic activities of the earth’s lower layers are plain to see on the surface. The smallest volcanic islands are also some of the youngest. One island emerged in 2013 off Japan near the well-known Bonin Trench. Initially it was less than 2 km^2 in size, but it has grown further since and even swallowed a neighbouring island. The islands, christened Nishinoshima, were a source of joy to the Japanese because they extended their sovereign territory (Nishida and Ichihara 2016; Allen and Simmon 2014). Another island formed even more recently in January 2015 near Tonga. Several submarine eruptions occured at Hunga Tonga-Hunga Ha’apai, throwing up lava and ash that eventually reached above sea level to create a new island. How long it will remain there is uncertain. Most likely, it will simply be washed away within months because it seems mainly to be composed of loose material as opposed to more resistant solid lava flows.[1]

[1]For more information and constant updates, see *www.volcanodiscovery.com/de/hunga-tonga-hunga-haapai.html*. Last accessed 4.1.2017.

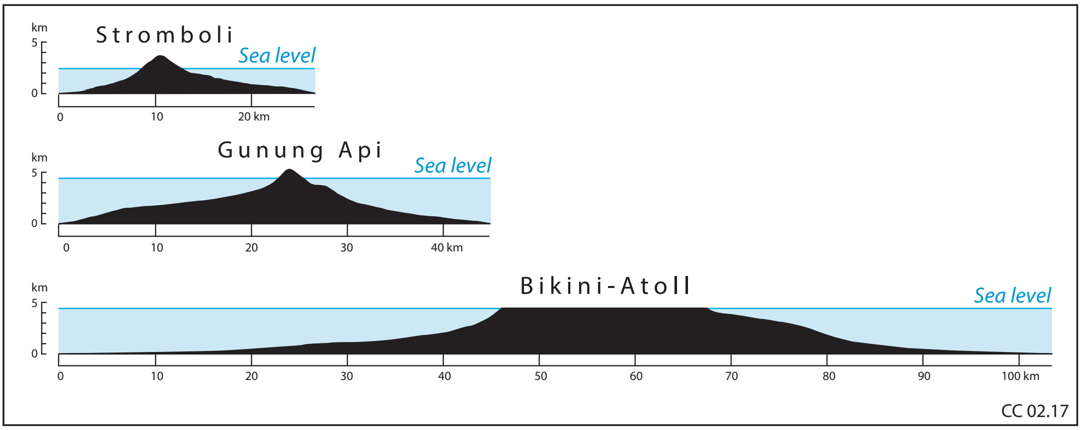


Fig. 2.1 Volcanic islands and giant atolls compared in size (After Menard 1964)

Put simply, volcanism means that the liquid interior of the earth is coming to the surface. This spilling of magma, lava or ash can only occur where the earth's crust is thin or broken, and fracture systems enable them to escape. The distribution of volcanic islands across the seas suggests a close relationship with geodynamic processes, especially plate tectonics and the organisation of the earth's surface. Plate tectonics explains the fundamental interrelationships that also play a key role in the formation of tecto-orogenetic and uplift/emergent and submergent/dislocation islands (see Sects. 2.2 and 2.5).

The Internal Structure of Earth

To understand volcanism, it is necessary to understand the structure of the earth. Earth is not a homogenous sphere – strictly speaking it isn't even a sphere – but consists of several layers of varying densities and aggregate states (see Textbox 2.1 Layered structure of the earth and Fig. 2.2). It is this which enables dynamic change and various natural processes on the surface of the planet. Geodynamics is the term describing fundamental geotectonic processes, comprising plate tectonics, continental drift and orogenesis – translated from the Greek (*oros* for 'mountain' and *genesis* for 'creation' or 'origin'), meaning the formation of mountains. Geodynamics refers to global dynamic processes which are subject to constant change. Orogenesis tends to be concentrated along certain long stretches of the earth's crust, unlike plate tectonic processes which affect the entire globe. Still, the deeper layers of the continental crust all consist of rock that has undergone orogenesis at some point of the earth's history (Frisch and Meschede 2013: 9).

Earth is the only planet in the solar system with an outer layer of large rocky plates that rub and press and move against each other. This outer crust is between 30 and 40 km thick; below lies the fluid-like upper mantle which is about 400 km thick. The origins of plate tectonics are vividly debated, with some experts assuming the process began shortly after the formation of the earth around 4.5 billion years ago but others suspecting a much later start about 800 million years ago. Scientists around Ming Tang of the University of Maryland have recently investigated the amount of magnesium, nickel, cobalt, chrome and zinc contained in the outer layers of rock from 18 different regions of the earth. Their analysis was able to determine that plate tectonics began three billion years ago, confirming an earlier study which analysed inclusions in diamonds and came up with a similar date (Ming et al. 2016; Willems 2016).

Plate tectonics led to a major paradigm shift in the 1960s as it was able to bring together all the dynamic phenomena of the earth in a unified theory. A precursor theory was the theory of continental drift that was proposed by German meteorologist and geophysicist Alfred Wegener (1880–1930). He not only noted that the continents on both sides of the Atlantic fit together like a jigsaw puzzle but also their geological relationships (Frisch and Meschede 2013: 10). Wegener of course was not the first to note the matching contours of the African west coast and the South American east coast[2], but using new evidence, he was able to question the geotectonic models of his day. Natural radioactivity, just discovered, showed that the earth had a heat balance, a finding which contradicted the idea of a planet that was gradually cooling and shrinking. His comparative analysis of American and European fossils and rocks yielded further evidence that the earth is anything but a solidified firm sphere.

Wegener presented his theory of continental drift in 1912. It posited that 200 million years ago, all continents were united in a supercontinent called Pangaea. This then broke apart, and individual elements began to drift apart horizontally on a plastic interior, leading to the formation of the Atlantic and Indian Ocean. In the 1920s and 1930s, Wegener's theory of continental drift caused much controversy as the scientific community doubted the ability of continents to move horizontally. What made matters worse was that Wegener was unable to come up with plausible forces that could cause the movement of plates. Based on the ideas of his colleague Robert Schwinner, he finally put forward thermal convection currents in what was to be the last edition of his book *The Origin of Continents and Oceans* (Wegener 1929). Due to his untimely death in November 1930, he did not live to see proof that convection currents are indeed the driving force behind the movement of plates.

Still during his lifetime, in 1924 and 1927, respectively, echo soundings carried out in the Atlantic on the research vessel Meteor gave first insights into seafloor topography. These soundings also confirmed the existence of thermal convection currents. Astonishingly, they additionally showed a mountain ridge in the middle of the Atlantic extending all the way across the globe from north to south (Press et al. 2011; Press and Siever 1982).

In subsequent years similar north-south mountain ridges were discovered on all ocean floors – the mid-oceanic ridges. In the central rift of these mid-oceanic ridges, divergent movements are taking place which are known as seafloor spreading. The central ridge continues to be opened up and re-forms on account of the oceanic plates moving sideways. Oceanic lava consisting of basalt and gabbro are brought to the seafloor. With a total length of about 70,000 km, the mid-oceanic ridges are a system of volcanic mountains that cross all oceans. Measurements on the mid-Atlantic ridge have shown that heat convection is particularly strong here; in contrast, measurements on the Acapulco Trench show that heat convection is significantly reduced (Wilson 1965; 1986: 20 f.).

These new insights, in particular the discovery of magnetic bands on both sides of the mid-oceanic ridges, laid the foundation of plate tectonics which finally became widely accepted as a theory in the 1960s (Hess 1962). At that time, geologists and oceanographers came to agree that mid-oceanic ridges form where convection currents move upwards and deep-sea trenches where they move downwards. Contrary to what Wegener thought, continents do not actively migrate as isolated slabs that 'plough through the sea floor like an icebreaker' (Wilson 1986: 21). Rather, they passively ride on the earth's crust which is moving horizontally (see Textbox 2.1 Layered structure of the earth). The seafloor drifts apart horizontally from the mid-oceanic ridges and sinks down into the deep-sea trenches. Continents are dragged along passively as they form part of larger plates that also comprise some oceanic crust and upper mantle (Frisch et al. 2011: 12; Frisch and Meschede 2013: 12; Press et al. 2011: 27).

When plate tectonics first became accepted as a theory, the total number of plates was assumed to be about 12. These were thought to consist of combinations of continents and ocean basins and were believed to have moved around on the Earth's surface through much of geological time

[2]Frank Bursley Taylor also noted that continents did not sink but slowly drift apart. Around the turn of the century, geologist Eduard Suess presented a map that brought together all southern continents in a single giant continent; see Suess 1892 and original map: http://blogs.scientificamerican.com/history-of-geology/files/2012/01/SUESS_1909_Antlitz_Erde.jpg. [last access 4.1.2017].

Textbox 2.1: Layered Structure of the Earth

In an idealised picture, Earth is made up of spherical shells, each consisting of material with very different density. This is the result of planetary differentiation, where denser materials sink to the centre to form the core. The shell with the lowest density is the outermost shell; the shell at the core of the earth – a sphere really – is the densest.

The Earth's outermost shell – the lithosphere – is divided into the crust and upper mantle. The outer crust can be further divided into a continental and oceanic crust. The continental crust is normally about 30 to 40 km thick, reaching up to 70 km in mountainous regions. This crust type mainly consists of granite and gneiss; feldspar, quartz and mica are the most important minerals. The oceanic crust is only between 5 and 8 km thick and mostly consists of basic rocks such as basalt and gabbros; here the dominant minerals are feldspar and pyroxenes.

The lithosphere lies on top of further layers of mantle, which are divided into the fluid-like upper mantle (asthenosphere) and lower mantle. The upper mantle has a high proportion of olivine and is very viscous. The lower mantle mainly consists of silicates and oxides. The core of the Earth mainly consists of iron and nickel. Due to the temperature difference between the earth's surface and outer core, there is convective material circulation in the mantle. This consists of the slow, creeping motion of the earth's silicate mantle across the surface, carrying heat from the interior to the surface. Hot material rises to the surface while cooler, heavier material sinks beneath.

Movement and convection are responsible for Earth's volcanic and seismic activity.

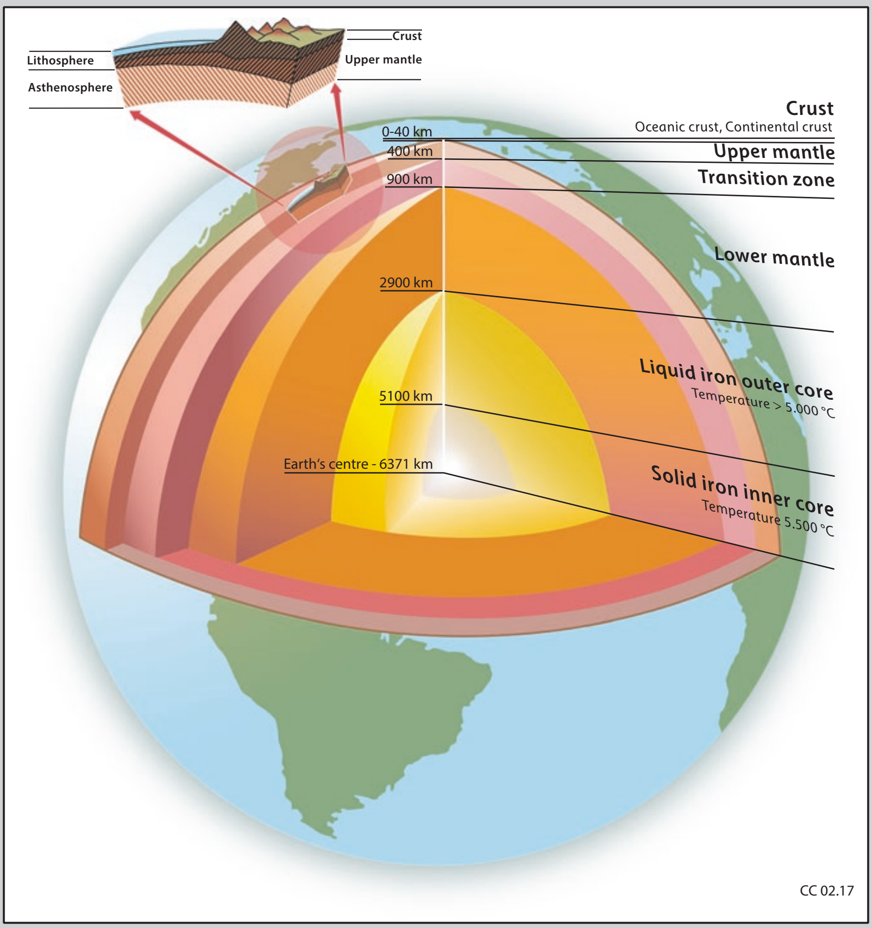


Fig. 2.2 Layered structure of the Earth (After Press and Siever 1982: 10)

(Anderson 1989). The upper parts of plates consist of either oceanic crust, continental crust or both sitting side by side. But since the start of the plate tectonic era, and especially relying on new techniques of more accurate earthquake epicentre location, modern ways of measuring ocean bathymetry using swath mapping and the use of space-based geodetic techniques, the number of plates thought to exist has grown considerably. A study by Bird (2003) proposed 52 plates. Because of the pattern of areas of these plates, he suggested that there should be more small plates than he could identify. Christopher Harrison in a recent study proposed a total of 107 new plates, giving 159 plates in all (Harrison 2016).

The greater the number of plates, the more boundaries there are between plates and with these potential activities that can lead to the formation of islands. Tectonic boundaries are the separation lines between plates, and along them tectonic movements almost always occur. Tectonic boundaries are the weak zones of the lithosphere where earthquakes, volcanism and orogenesis take place. There are three types of tectonic plate movement:

1. Divergent movements: Two plates are drifting apart at an average speed which can vary from 2.6 to 16.1 cm/year. This creates a rift zone, most commonly observed on the seafloor.
2. Convergent movements: Two plates are moving towards each other, with one plate sliding underneath the other. This is termed subduction.
3. Transform movements: Two plates are sliding past each other.

The *Pacific Ring of Fire*, or circum-Pacific belt, clearly shows the effects of the geodynamic processes occurring along tectonic boundaries. The Pacific Ring of Fire is not an enclosed ring but a horseshoe shape that runs along the western, northern and eastern edges of the Pacific Ocean along a total length of about 40,000 km (Fig. 2.3). The western and northern parts are characterised by a series of volcanic island arcs and continental volcanic belts, beginning with New Zealand's North Island (Taupo Volcanic Zone), continuing on to the New Hebrides, the Solomon Islands, New Guinea, the Philippines and the Mariana Islands and extending further to the Ryukyu Islands, the main Japanese islands, the Kuril Islands, the Kamchatka Peninsula and the Aleutian Islands. To the east, the belt of active volcanoes begins in Alaska and extends along the western side of North and Central America to the South American Andes, all the way to the southern tip of Patagonia.

The Pacific Ring of Fire encompasses the tectonic boundaries of the Pacific Plate and those of other smaller plates. A number of subductions are occurring concurrently. The Nazca Plate and the Cocos Plate are being subducted beneath the westward-moving South American Plate, resulting in the eastern section of the Ring. In Central America the Cocos Plate is being subducted beneath the Caribbean Plate and along with the small Juan de Fuca Plate; part of the Pacific Plate is being subducted beneath the North American Plate. The north-westward-moving Pacific Plate is being subducted along the northern section beneath the Aleutian Island Arc which is also part of the North American Plate. To the west, the Pacific Plate is being subducted southwards along the Kamchatka Peninsula past Japan. A number of smaller tectonic plates are in collision with the Pacific Plate at the southern part; this encompasses the Mariana Islands, the Philippines, Bougainville, Tonga and New Zealand. Australia lies at the centre of its tectonic plate. Indonesia lies between the Ring of Fire along the northeastern islands adjacent to and including New Guinea and the Alpide belt along the south and west from Sumatra, Java, Bali, Flores and Timor.

The effects of these geodynamic plate processes are twofold: volcanism and earthquakes. Earthquakes are generated constantly, multiple times a day but mostly too small to be felt. A severe earthquake on New Zealand on 20 January 2014 reached a magnitude of 6.2 on the Richter scale and was named Eketahuna earthquake due to its epicentre 15 km east of Eketahuna in the south-east of New Zealand's North Island. A total of 1112 aftershocks were recorded, ranging between magnitudes of 2.0 and 4.9 on the Richter scale. It was felt strongly down the country, from

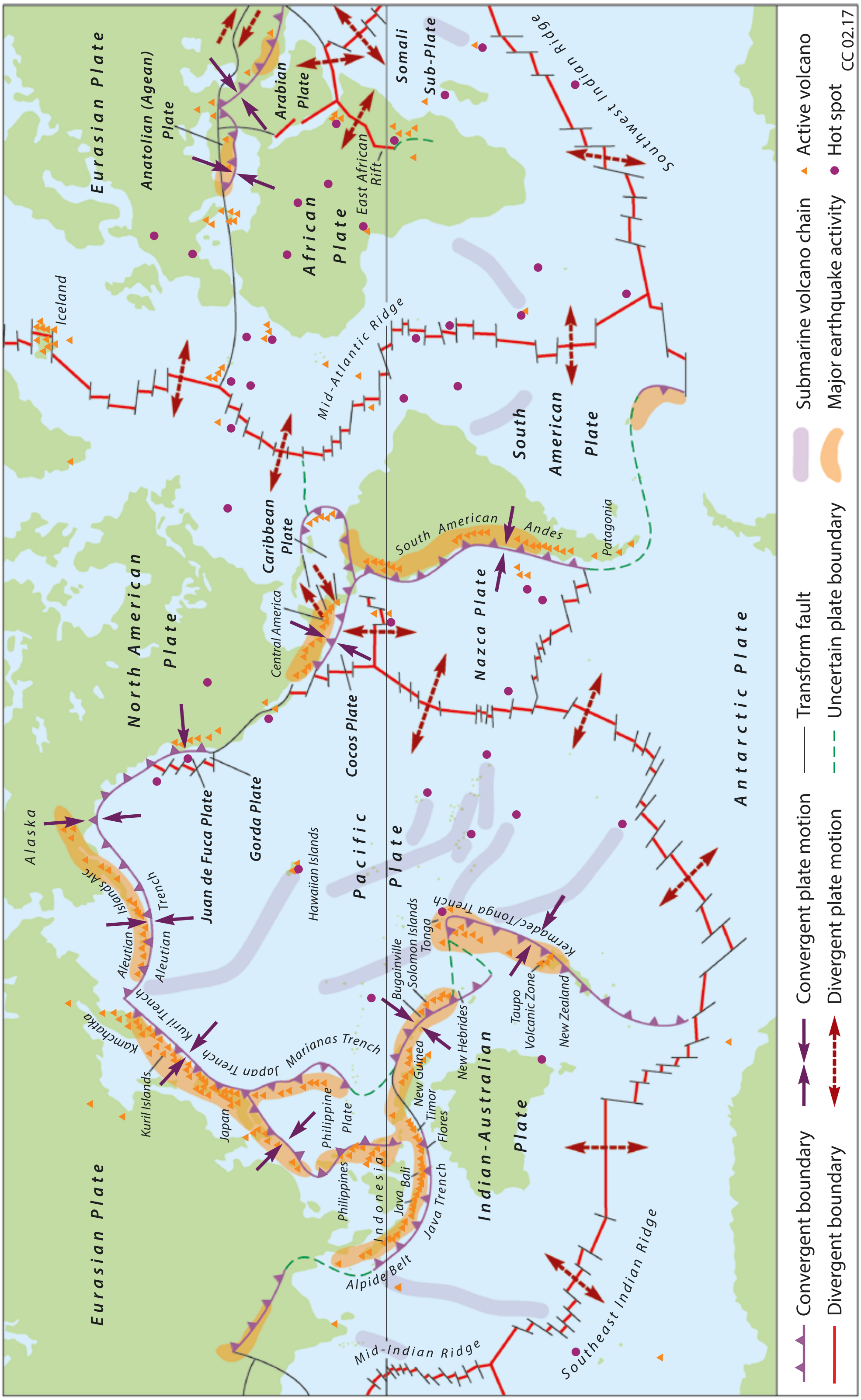


Fig. 2.3 Plate tectonics and Pacific Ring of Fire (After Press and Siever 1982; Frisch and Meschede 2013)

Auckland in the north to Dunedin in the south, and more than 9000 reports were submitted by the public to GeoNet[3], the geological hazards monitoring network. About 90% of the world's earthquakes and 81% of the world's largest earthquakes occur along the Ring of Fire (USGS 2016) – a considerable number of them happening on islands.

Volcanic islands form as a result of three different types of volcanism. As shown in Fig. 2.3, divergent and convergent plate movements lead to volcanic islands along the subduction margins of the continents or the mid-oceanic ridges. But some islands, especially in the Pacific (including Hawaii and Galapagos Islands), form on account of volcanism within lithospheric plates, on top of the so-called hotspots. And then there are special cases such as Iceland, where hotspot volcanism coincides with the volcanism of a mid-oceanic ridge.

Island arcs can result on three types of convergent plate boundaries. In intra-oceanic subduction zones, where an oceanic lithosphere is subducted beneath another, volcanic island arc systems are built on oceanic crust. Examples for intra-oceanic, ensimatic island arc systems (*sima*, an artificial word – silicon and magnesium – first used by A. Wegener to characterise the ocean floor and Earth's mantle, Frisch et al. 2011: 91) include the Mariana Islands, the Lesser Antilles in the Atlantic and the South Sandwich Islands (see Fig. 2.4). If oceanic lithosphere is subducted beneath continental lithosphere, an island arc may form on a continental base (a so-called ensialic island belt – *sial* for silicone and aluminium and the continental crust). These island arcs are generally separated from the continent by a marine basin underlain by oceanic crust. Examples for island arc systems underlain by continental crust include the Japanese islands and the eastern Sunda Arc (Frisch and Meschede 2013).

Subduction of an oceanic lithosphere beneath a continental lithosphere also occurs along active continental margins. The continental margin is connected directly to the rest of the continent, although there may be a shallow marine basin behind the volcanic arc (Frisch et al. 2011: 91). Examples for active continental margins include the Andes, Alaska and the western and central Sunda Arc (Sumatra, Java). This plate margin system is dominated by a volcanic zone on top of the subduction zone that often takes the shape of an island arc. The volcanic arc, which has an average width of 100 km, is the central part of the island arc or the active continental margin and is characterised by significant magmatic activity. Volcanism begins sharply on the forearc margin (the volcanic or magmatic front) and gradually peters out towards the backarc margin.

Why the arched shape? Frisch, Meschede and Blakey explain the principle as follows:

> A thumb pushed against a rubber ball causes the normal convex bulge of the ball to become a concave dent. The line of bending marks a circular line on the surface of the ball. Before a plate enters a subduction zone its curvature mirrors that of the earth. When it dives into the subduction zone this curvature inverts and convex becomes concave. Adjacent arcs commonly display a catenary-like pattern as observed in the Western Pacific where one island arc drapes next to the other (Frisch et al. 2011: 94).

Hotspot volcanism is a special form of volcanism not related to plate fringes. Such intraplate volcanic islands are fixed-point sources of magma that occur within continental or oceanic plates. They owe their existence to mantle diapirs or plumes (from Greek diapeirein, to penetrate, or from the French for feather) – in other words, hot fingers in the Earth's mantle that rise up from great depths. Once they arrive below the plates, they cause melt and, in the long term, volcanic eruptions. As the position of hotspots is stable, linear chains of volcanoes can result as plates slowly glide over the hotspot (Frisch et al. 2011: 11 f. & 80; Press et al. 2011: 49; Klug 1985: 204).

These mantle plumes (convection cells) below the lithosphere also occur along constructive plate boundaries, although they are more common within plates. They cause large-scale doming of the Earth's crust. Only about 5% of volcanoes are hotspot volcanoes, owing their existence to man-

[3]GeoNET is the official source of geological hazard information for New Zealand http://www.geonet.org.nz

Fig. 2.4 Island arc – example: Lesser Antilles

tle diapirs or plumes. Nevertheless, they play an important role in the mantle's convection system as they are responsible for about 5–10% of the energy the Earth gives off to the outside. Island chains of this type mostly occur within the Central Pacific.

When plates slide across a hotspot, long chains of volcanoes form. The hotspot is marked by the active end of the chain. Chains of islands that formed more than 40 million years ago tend to be oriented more north to south, while younger chains are arranged more east to west. This difference in orientation is caused by a change of direction in the movement of the Pacific Plate (Morgan 1971; Jackson et al. 1972). The classic and best-known example for this phenomenon is the Hawaii Islands that find their continuation in the Emperor seamount chain (Keating et al. 1987) (see Fig. 2.5). The Marshall Islands, Tuamotu, Samoa, Tubuai and the Caroline Islands have also formed in this way. In line with the plate's movement, islands grow ever older and show an increasing tendency to subside. The geomorphological form of the oldest volcanic islands has thus undergone some erosional change. A good example is the Comoros Islands group in the Indian Ocean with aging erosion processes increasing from Grande Comore to Moheli, Anjouan and Mayotte.

In this type of volcano, the molten rock is viscous and only flows slowly from the core of the Earth. This causes large flat shields to form - so-called shield volcanoes that are large in size but low in profile. Volcanic growth and erosion are often parallel processes on the same island. Older volcanic islands often have a fully formed barrier reef, while younger ones may even lack a fringing reef (Klug 1985: 204).

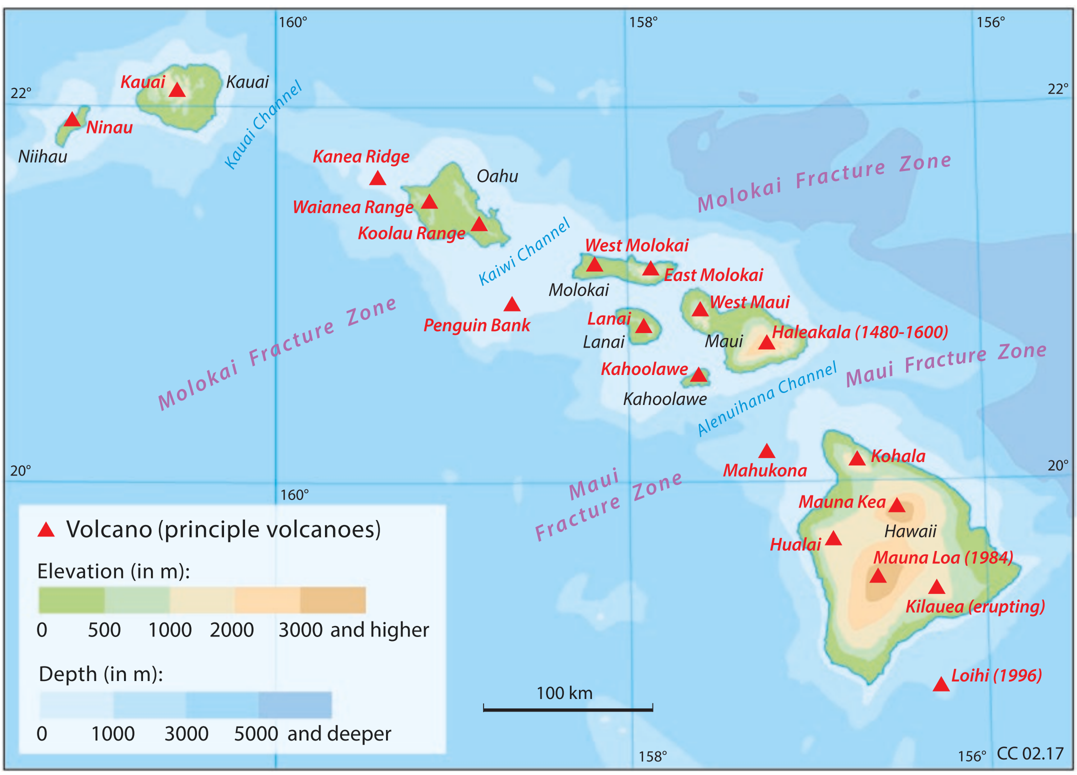


Fig. 2.5 Hotspot volcanism – example: Hawaii

2.2 Tecto-Orogenetic Islands

Apart from initialising volcanism and earthquakes, the collision of lithospheric plates can also lead to uplift and the subsequent building of mountains and islands. Puerto Rico or Hispaniola in the Caribbean or New Zealand in the Pacific are examples of islands created by tecto-orogenic events. Strictly speaking, these islands are the result of multiple geological processes. Orogenesis is preceded by sedimentation processes, and several events occur at once or sequentially when tectonic plates collide and folding takes place. The exact nature of events depends on the type of plates colliding and the manner of collision. In the *orogenic stage* of mountain building, accumulated sediments become deformed by the compressional forces resulting from the collision of tectonic plates.

This tectonic convergence can be of three types: arc-continent, ocean-continent or continent-continent. In an ocean-continent convergence, the collision of ocean and continental plates causes the accretion of marine sedimentary deposits on the edge of the continent. Arc-continent convergence occurs when an island arc collides with the edge of a continental plate. In that instance, the ocean plate area between the arc and the continent is subducted into the asthenosphere, and the volcanic rocks and sediments associated with the island arc become accreted to the margin of the continent over time. Continent-continent convergence occurs when an ocean basin closes and two continental plates collide to form mountain systems.

In all three types of tectonic convergence, layered rocks that were once located in the marine basin are squeezed into a smaller and smaller area. This compression causes the once-flat sedimentary beds to be folded and uplifted. When the compressional and tensile forces become greater than the rocks' ability to deform, faulting occurs. Compressional forces typically result in reverse and overthrust faulting. Another consequence of the orogenic stage is regional metamorphism and the incursion of magma plumes, plutons and volcanoes into the growing mountain range (Frisch et al. 2011: 93 & 95; Press et al. 2011: 75).

New Zealand is an instructive example of the complicated genesis of tecto-orogenic islands, of which there are many around the globe. New Zealand lies on the boundary of the Australian and the Pacific Plates, but the islands forming New Zealand developed as part of a broader continental shield made up of Antarctica and Australia. A clear delineation of the various processes involved is particularly difficult, as there is always a sequence of sedimentation, uplift, folding and orogenesis.

New Zealand's geological history can be divided into three main periods of sedimentation and three main periods of mountain building (orogeny):

1. *The early sedimentation depositional phase* in the Cambrian to Devonian period about 545 to 370 million years ago

The area now known as New Zealand began when the earliest major recorded rock formation was taking place. The oldest rock is found on the western coast of the South Island. It all happened just off the coast of Gondwana. After they became extinct, some volcanic islands were covered in sand and mud washed down from the land and built up. Sometimes the land uplifted only to be worn down again and pushed back into the sea.

2. *The Tuhua Orogeny* in the late Devonian to Carboniferous period about 370 to 330 million years ago

A long period of sedimentation ended with a period of pressure and uplift. Seafloor sediments were pushed up, folded and melted together to form mountains. Rocks were completely changed and regrouped into new minerals under great heat and pressure. Sandstones and mudstones became schist, known for the parallel layering of minerals. Plutonic intrusions formed granite and in some places diorite. All these activities happened long before today's New Zealand existed, and therefore the exact mountain building details are not exactly known.

3. *The New Zealand Geosyncline* in the Carboniferous to Jurassic period about 330 to 142 million years ago

This period is characterised by an enormous accumulation of sediment, extending northwest from New Zealand to New Caledonia and south far below the South Island. The rocks of this second cycle of deposition have formed much of the foundations of New Zealand. Two main groups of rocks can be identified from this period: the Torlesse Supergroup on the east mostly made up of greywacke with only very few fossils and the Murihiku Supergroup in the west, with a good series of fossils, with sediments rich in volcanic debris.

4. *The Rangitata Orogeny* in the early Cretaceous period about 142 to 99 million years ago

The previously deposited sediments were compressed and folded during this orogeny. The western rocks were deformed in open simple folds, and the eastern block was severely deformed in a stack of folds with complex faulting. Some seafloor was caught in the folding and later exposed when the orogeny had finished and erosional forces had levelled the mountains.

5. *The break-up* in the Cretaceous to Oligocene period 99 to 24 million years ago

Weathering and erosion of the mountains followed the preceding orogeny, so much so that some places were reduced to areas of low relief – the so-called peneplains. About 85 million years ago, a rift valley formed to separate the New Zealand region from the rest of Gondwana, resulting in the formation of a new ocean floor by means of *seafloor spreading*. Marine transgression and the following Oligocene period (about 35 million years ago) led to a sinking of the land and resulting in characteristic Cenozoic marine deposits: calcareous and fossiliferous, with common limestone.

6. *The Kaikoura Orogeny* in the Miocene to Quaternary period 24 million years ago to modern

A build-up of strain in the southwest Pacific crust in this period led to vertical and transcurrent fault movements. This resulted in uplift of central Westland and produced the majestic range of the Southern Alps, with their steep, straight western front of the Alpine Fault. Widespread tectonic activity continued from ten million years ago to the modern period; during this time the principal mountain ranges of both islands were uplifted, and New Zealand began taking its modern shape. The subduction of the Pacific Plate caused much volcanism in the North Island, starting initially in Northland in the early Miocene and moving south over time until it reached its present position along the Taupo Volcanic Zone (Nelson et al. 2003). Cenozoic intermingled with Cretaceous sediments dominates the North Island whereas Cretaceous and Palaeozoic metamorphic rocks are predominant on the South Island (see Fig. 2.6).

2.3 Sedimentary Islands

The genesis of sedimentary islands depends on the respective coastal morphology and related parameters such as tidal range, wave energy and basement control. A basic requirement is the availability of sediment – mostly sand – as well as a shallow shelf area which lends itself to accumulation processes. Sedimentary islands therefore only occur in shallow coastal seas. The group of sedimentary islands primarily includes sand spits and barrier islands, such as the Frisian Islands in Europe or the islands on the east coast of the USA from Florida to Rhode Island. According to Smith et al. (2010), 'chains of barrier islands can be found along approximately thirteen percent of the world's coastlines'. Excepting the tidal inlets that separate the islands, a barrier chain may extend uninterrupted for over a hundred kilometres, the longest and widest being Padre Island in Mexico (Garrison et al. 2010).

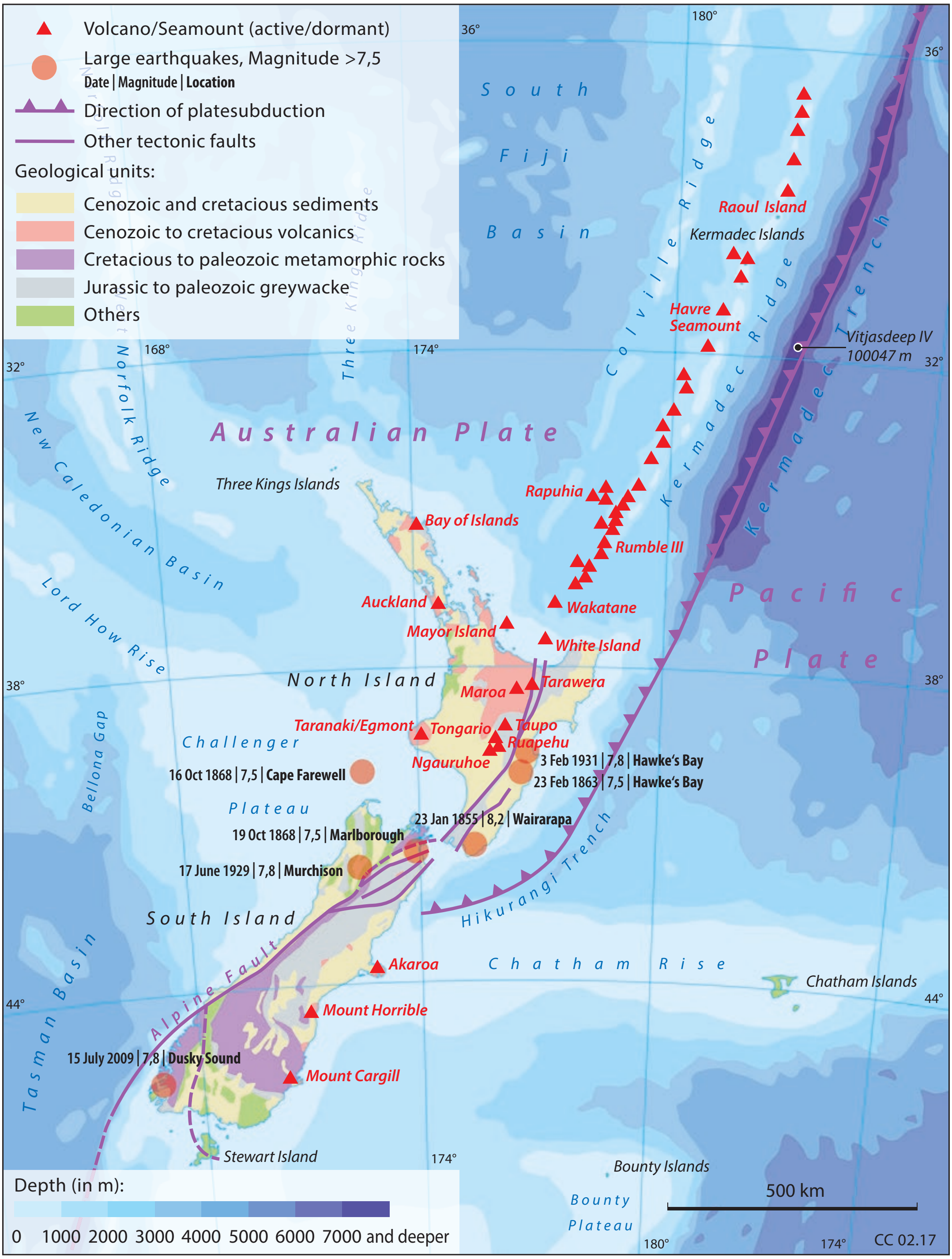


Fig. 2.6 Tecto-orogenetic island – example: New Zealand

Sedimentary islands result from the accumulation of loose sediment. In the coastal areas of shallow seas, the surf creates a plethora of different island forms. A key force is the vertical and horizontal deformation of deep ocean waves as they reach shallow areas. On approaching the beach, the rounded peak of the wave becomes more pointed (shoaling) before rising and then breaking towards the shore. Kelletat describes this as follows:

> Deformation begins when the wave reaches a critical water depth, which corresponds to about half the length of the wave and therefore varies. Upon reaching the critical water depth, the orbital trajectories of the lower water particles meet the seafloor, where they are slowed down or obstructed. In the upper part of the wave the orbital motion initially continues. This results in the characteristic breaking of the wave. (Kelletat 1999: 127 f.)

Depending on the wave's energy, water can spill up the beach quite a long way, depositing the transported sediment in that direction. In an interplay of swash and backwash, the energy of the water is shooting up the beach and then running back down which leads to the shifting, layering and sorting of sediment in processes that are collectively termed beach dynamics.

When a wave approaches the coast obliquely rather than head on, the nearshore part reaches shallow water first and slows down before the rest of the wave. This creates the so-called refraction effect, or a veering of the wave towards the coastline. While the wave and therefore the swash meet the beach at a certain angle, the backwash loses energy on account of friction and the deposition of sediment, causing it to run directly down the beach in the direction of gravity. This means particles are transported down the beach in a zigzag pattern. The displacement of sediment at an oblique angle is termed longshore drift (see Fig. 2.7).

Accretion bodies usually adjoin a coastal promontory. With frequently changing wave directions and subsequent changes in the direction of longshore drift, seasonal modifications of forms can occur. Sand spits can occur when sandy deposit is built up into a landform which projects into a body of water. Beach ridge accumulation results in islands gradually becoming attached to the mainland and the creation of ridge-like links between islands.

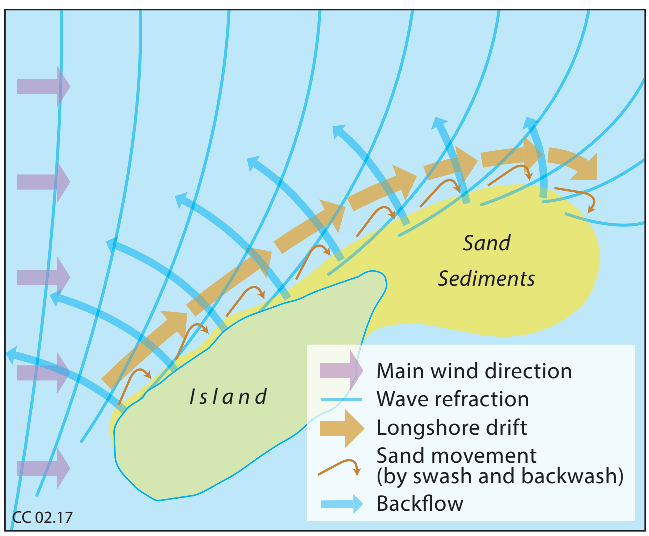


Fig. 2.7 Longshore drift and shifting island (After Klug 1985)

Such formations are termed tombolo. Kelletat shows an extensive representation for St. Christopher in the Caribbean (Kelletat 1999: 134). Another visible effect of longshore drift is the uneven accumulation of sediment on obstacles such as groynes. Coastal residents often make use of this effect when reclaiming new land on the coast, for example, on the Frisian coast in the Wadden Sea.

Since surf zones not only develop near beaches but also in shallows, quasi-stationary areas result with strong resuspension and accumulation of sediment. Mostly in tidal areas but also before tideless coasts, submerged bars of sand can thus grow above sea level. Due to the varying reach of the waves, accumulation coasts are normally characterised by extensive areas of dry sediment that are bare of any vegetation. This creates a point of attack for wind and is a prerequisite for the formation of dunes, as can be observed on the North Sea coast, for example. Once the upper areas of the beach dry out, freely migrating, barchan dunes initially form. On encountering obstacles, irregular and above all higher accumulation can occur. This mostly coincides with the first colonisation by salt-resistant plants (such as marram grass) with fast-growing and widely branching root systems, contributing to the further stabilisation and accumulation of sand. Over time, dune ridges develop, which in humid parts of the world are often covered in thick vegetation.

Along coasts with a larger tidal range, swash lines can shift horizontally, sometimes over considerable distances, especially along flat coasts. Ridge-like structures accumulate before the coast which initially grew above the low-water mark and eventually above the high-water mark, forming elongated islands. On mesotidal coasts, tidal inlets are kept open by the huge water masses streaming in and out with the tide which prevents islands from linking up. In these cases, free spits and barrier islands commonly develop, such as the chain of islands that lines the Dutch and German North Sea coast. Their distance from the coast is influenced by the critical, surf-forming water depth of the last transgression. In line with the predominantly westerly winds and waves that abound in this region, material is transported eastwards in parallel to the coast, causing a tendency of the East Frisian Islands to shift or migrate east-south-east.

Shifting Islands in the Wadden Sea

Entire islands shift from west to east and landwards. Only islands with an old Pleistocene core are safe from being shifted by the forces of nature. The predominant winds along the German North Sea coast are westerlies or northwesterlies, and the high tide always streams in from the west. The Wadden Sea extends between the mainland and the barrier islands, composed of finely sorted accumulation material. Twice a day, millions of cubic metres of water flow in and out between the islands through the small tidal inlets that connect the Wadden Sea to the open sea. Much sand is transported in this way. While it is washed onto the beach during high tide, it actually promotes steeper sloping of the beach when sucked out again by the reverse flow of water. In line with longshore drift, the coastal current leads to sand shifting eastwards in parallel to the beach. The western sides of the islands are eroded, and sandy hooks form on the eastern sides (see Fig. 2.8). Norderney, Spiekeroog and Wangerooge are classic examples of shifting Frisian barrier islands (Sindowski 1973).

Over the last 100 years, Spiekeroog has grown by about 4 km in an easterly direction. In 1894 the western side of the island was secured by human intervention (groynes and sea walls) to prevent storm surges from breaking through the western chain of dunes. The classic tripartite structure of dune core, salt marsh and eastern plate is readily visible on Spiekeroog. The eastern plate is occasionally flooded. Large amounts of sand are transported along the East Frisian Islands in an easterly direction and deposited, forming sandy hooks. The islands and the Wadden Sea are estimated to be at least 1500 years old (Pott 1995).

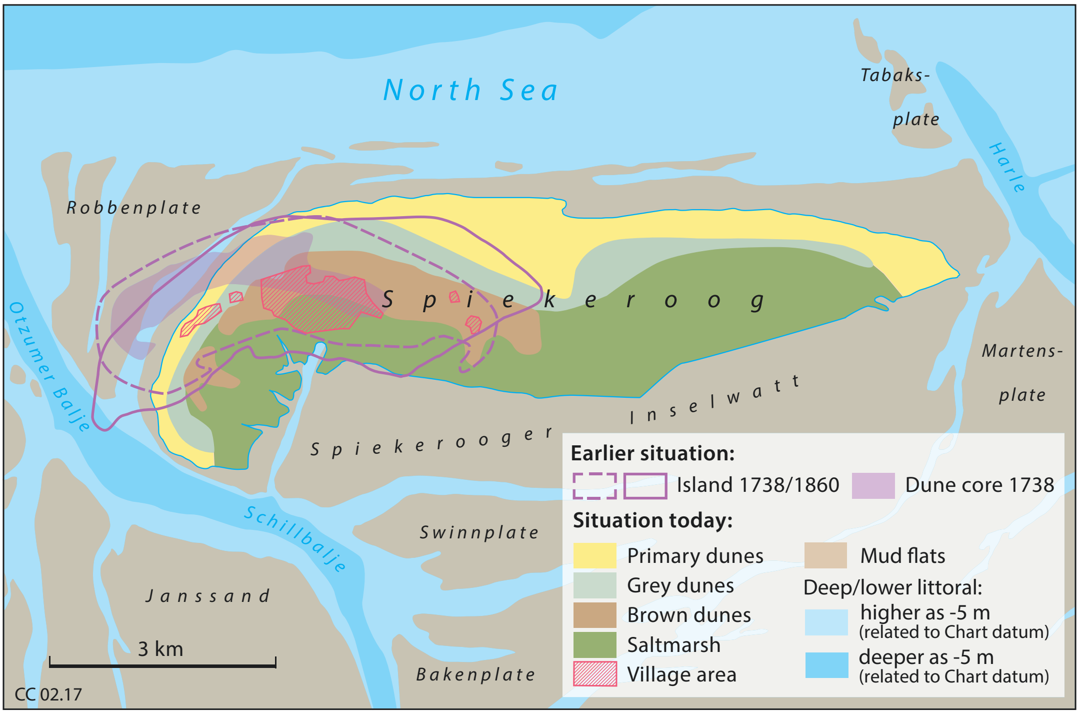


Fig. 2.8 Shifting Frisian barrier island – example: Spiekeroog (After Kelletat 1989)

Wangerooge today is a pure dune island. Since medieval times the island has lost its old core due to the eastward shift of the Harle inlet. The church tower, built in 1595 at the centre of the first village, is now in the sea west of the island. Wangerooge's West Tower was built around 1600 on the eastern shore; by 1793 it had moved to the centre of the island. Before the First World War, its western side already stood in water.

In contrast to the East Frisian Islands which are modern land formations, the North Frisian Islands are remains of mainland that was broken up by storm surges. Here, islands can be found with and without a Geestland core, the latter representing pure accumulations of sediment and the former depositions of sediment on an existing core of former mainland. Sylt, Amrum and Föhr are examples of islands with a Neogen[4] Geestland core around which coastal sediment was deposited over thousands of years (Streif 1990).

[4]The tertiary period is divided into the Neogene and Paleogene. The Saale glacial stage had pushed together the Geest cores.

Last not least, sedimentary islands also include islets that have gradually grown up through silt aggradation or sedimentation in areas of shallow water. According to the inhabitants of the region, these islets are not really islands but 'Halligen'. Contrary to what has long been assumed, these silt aggradation islands are not the remains of former mainland but gradually accumulated in layers after the destructive Rungholt storm surge in 1362. During this huge and destructive event, also known as the 'Große Mandränke' (literally 'big drowner of men'), much of the cultivated land on the North Frisian coast was lost. Formerly agricultural land became mudflat, or a shallow shelf area where sand and mud was deposited twice daily by the flow of the tide. With growing aggradation, the new deposits were eventually only flooded during spring tides. The second catastrophy in 1634, known as the 'Zweite Mandränke', again destroyed parts of the coastal mainland and islands. 'After the disastrous storm surge, inhabited mounds that had not been destroyed and still rose above the Wadden Sea by a few decimetres – the Hallig islands – continued to grow' (Mieth and Bork 2009: 116). New dwelling mounds were created on the fresh sediment that was flooded ever more rarely. With every submersion, the Hallig islands grew in elevation and changed their area. This is how these specific 'pieces of land surrounded by water' rather than islands came into being.

Due to the lack of coastal defence, or weaknesses in its execution, flooding events were more common in the past. Two large-scale changes to the coastline occurred in history, and many more Hallig islands used to exist that frequently changed shape. Some only lasted for a short period until a shifting tidal creek caused them to shrink and disappear again. Others grew due to sediment deposition and merged, such as Nordmarsch and Langeneß which became today's Langeneß. The ten remaining Hallig islands are grouped in a circle around the island of Pellworm, which is not a Hallig itself. Nordstrandischmoor, Gröde, Oland, Langeneß and Hooge as well as the smaller Halligs of Habel, Südfall, Süderoog, Norderoog and Hamburger Hallig are today's cultivated Hallig islands in North Frisia. Seven out of the ten are inhabited by a total of around 230 people.

Ecologically, barrier islands play an important role in mitigating ocean swells and other storm events. Not only do barrier islands create a unique environment of relatively low energy, brackish water, but they also serve as natural coastal defence. On the back-barrier side, multiple wetland systems form such as lagoons, estuaries and marshes depending on the specific context and conditions. These wetlands could not exist without barrier islands – they would be destroyed by the daily action of waves and storm events. Prominent examples of such wetlands can be found off the coast of Louisiana as well as off the Frisian coasts.

Islands Come and Go…

Coastal change and island creation is a continuous process. In the Lower Saxonian part of the Wadden Sea, a new island, the so-called Kachelotplate, was reported some 10 years ago when the sedimentation process created a dune island with primary dunes. Large parts of the Kachelotplate are now no longer flooded during high tide, meaning the sand bar, which measures about 3 km^2, is now officially an island. It is situated between the islands of Borkum and Juist and near the island of Memmert which is a bird sanctuary. The current process of island formation is caused by changing ocean currents and changes in the shift of sand. The sand deposited on the Kachelotplate originates from the West Frisian Islands off the Dutch coast. The next storm surge may once again destroy the Kachelotplate, but should it persist, it will almost certainly be dedicated to nature conservation as it is situated within Zone I of the Wadden Sea National Park (Wehrmann and Tilch 2008; CWSS 2013). Islands come and go; this holds true not only for the uninhabited islands of the Wadden Sea.

2.4 Coralline Islands

Apart from geological and physical processes, reefs and islands can also form as a result of zoogenic processes, specifically the growth of corals. A coralline reef takes thousands of years to grow, a process which is influenced by many geological and physical factors. Fluctuating sea levels, continental drift, subsidence and changes of temperature influence the shape and character of a reef more strongly than the different species of coral responsible for its construction. Zoogenic superstructures in the form of coralline reefs are particularly widespread in warmer waters.[5] Corals are primarily found in the tropics and subtropics between the 30th parallel north and

[5]Some stone corals build cold water reefs and live at depths between 50 and 1000 m. Despite the icy water temperatures of 4–12 °C, they grow by up to 2.5 cm a year. Much less is known about these so-called deep water or cold water corals than their tropical relatives. The Darwin Mounds, a huge belt of coral reefs extending from Norway to Portugal, was only discovered in 1998 (Roberts et al. 2006; Kiriakoulakis et al. 2004).

the 30th parallel south – where they are estimated to extend over a total area of 250,000 km^2 or approximately 0.1% of the world's oceans (Burke et al. 2011: 13; Spalding et al. 2001).

The development of coral reefs is dependent on hydrological factors, particularly water temperature, although relatively low water depth is also critical. Corals are related to jellyfish and sea anemones; together they form the phylum of Cnidaria. Corals are colony-forming, sessile Cnidarians which are subdivided into distantly related classes. Anthozoans include corals, sea anemones, sea pens and sea pansies; two important subclasses are the Octocorallia and Zoantharia. Octocorallians include gorgonian corals, sea pens, sea pansies, organ-pipe corals and soft corals (order Alcyonacea); zoantharians include the reef-building corals, which are also known as hard or stony corals. All corals have tube-like hollow bodies; the mouth is used for both taking in food and excreting wastes. Tentacles surrounding the mouth are used at night to catch food. More or less the entire body is given over to digestion.

Stone corals form skeletons by depositing calcium carbonate in their tissue. Coral banks or reefs are created by living organisms constantly overgrowing dead skeletal material. Stone corals are the master builders of the tropical seas. Their skeletons are largely composed of crystals of aragonite ($Ca[CO_3]$), a calcium carbonate excreted through the basal plate or epidermis in order to lend support to the colony. Individual skeletons tend to be branched like a tree. Colourful polyps sit at the tips of the branches where growth takes place, reinforcing the impression of submarine flowering plants. Coral growth largely depends on the species. Many soft corals grow faster than their hard relatives, with some growing as little as 1 cm and others over 100 cm per year. Growth essentially depends on sufficient levels of oxygen, nutrients and light, which is why corals are usually restricted to maximum water depths of around 40 m.

Stony corals are important ecosystem engineers, which regulate the functioning of the entire reef via the production of limestone structures and the release of organic substances such as carbohydrates and mucus. But soft corals have their advantages too. A recent investigation discovered a pecularity of soft corals of the Xeniidae family. Their feathery polyp tentacles open and shut like birds' wings. This pulsating movement generates considerable advantages for their metabolism and food supply – and ultimately probably affects growth positively. Another effect of pulsation could be of great importance; during photosynthesis, reactive oxygen radicals are also produced, which are very harmful to coral metabolism. When oceans warm, oxygen radicals induce corals to release their symbiotic algae, in turn causing the corals to bleach and often to die. Through pulsation, the radicals are probably transported away effectively. Therefore, it is likely that pulsating soft corals are particularly resistant to coral bleaching. Such robustness and the favourable energy balance create a significant competitive advantage for soft corals in the reef (Wild and Naumann 2013; Kremien et al. 2013).

Like most sessile marine organisms, corals are filter feeders that filter microplankton, nutrients and trace elements from the water. Corals that live in shallow water not only rely on filtering plankton but also obtain nutrients – in some cases predominantly –from symbiotic algae stored in their skeletons. These so-called zooxanthellae are responsible for the vivid colour of living coral tissue. The photosynthetic activity of these single-celled algae is an essential part of the coral's metabolism. A symbiotic relationship ensues: Since the algae need sunlight for their metabolism, corals grow towards the sun. When corals are shaded out by other corals, they branch and, just like the branches of a tree, reach out towards the light. As a result, genetically identical polyps can form entirely different shapes of coral. Depending on the available plankton, the size of coral polyps can also vary greatly, reflected in the distinction between large polyp corals (LPS – Large Polyp Sclerantinia) and small polyp corals (SPS – Small Polyp Sclerantinia). Polyp size ranges from fractions of a millimetre to several centimetres.

Corals have existed for over 400 million years, explaining their importance for palaeoclimate reconstruction. Coral reefs are so significant because they are home to thousands of plants and

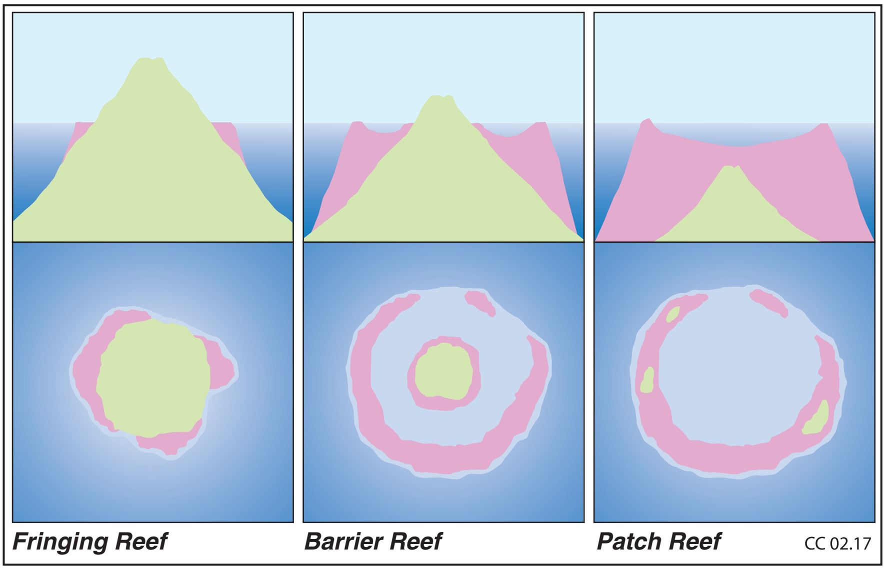


Fig. 2.9 Types of coral reefs: fringing reef, barrier reef and patch reefs (After Kelletat 1989)

animals. Crustaceans, starfish and other small animals live in their nooks and crannies; many fish use coral reefs as nurseries. In terms of biodiversity, the Indo-Pacific is roughly ten times more diverse than the Western Atlantic. For example, there are approximately 60 species of corals inhabiting the coral reefs of the Western Atlantic compared to an estimated 500–600 species in the Indo-Pacific. Coral reefs are rare or absent in the tropical Atlantic of South America and Africa mainly because of the great influx and circulation of freshwater and silt from the Amazon and Congo River systems.

2.4.1 Reef Types: Fringing Reefs, Barrier Reefs and Patch Reefs

Coral growth usually reflects the relief of the seafloor. Reef formations can be classed into four main types: fringing reefs, barrier reefs, platform reefs and atolls (see Fig. 2.9).

Fringing reefs form close to the coast, sometimes following the contours of the shore for kilometres. How far they extend out to sea depends on the decline of the seafloor and on water quality, as corals need clear, unclouded water and enough light to ensure the survival of the zooxanthella. Fringing reefs are the most common type of reef; they are predominantly found in the Red Sea, Southeast Asia, the Indian Ocean and the Caribbean.

Erosion may form a channel between the shore and the reef, giving rise to a fringing reef with an intervening shallow lagoon.

Contrary to a fringing reef whose lagoon is only a few metres deep, *barrier reefs* are separated from the coast by a lagoon with a water depth between 30 and 70 m. While fringing reefs always originate from the mainland, gradually moving into the sea, barrier reefs usually form in the open sea. Different environmental conditions are responsible for the different dimensions of the reef. If sea levels rise over the course of millennia, the reef responds by growth, ensuring the algae remain just below the water line to still receive enough sunlight. The same response occurs when the seafloor subsides. Barrier reefs are much rarer than fringing reefs due to the particular interplay of geological change and reef growth. The largest barrier reef is the Great Barrier Reef of Australia, followed by the barrier reef islands before the Belize Coast in the Caribbean (Gierloff-Emden 1980: 973 f.; Klug 1985: 205; Kench et al. 2005, 2009).

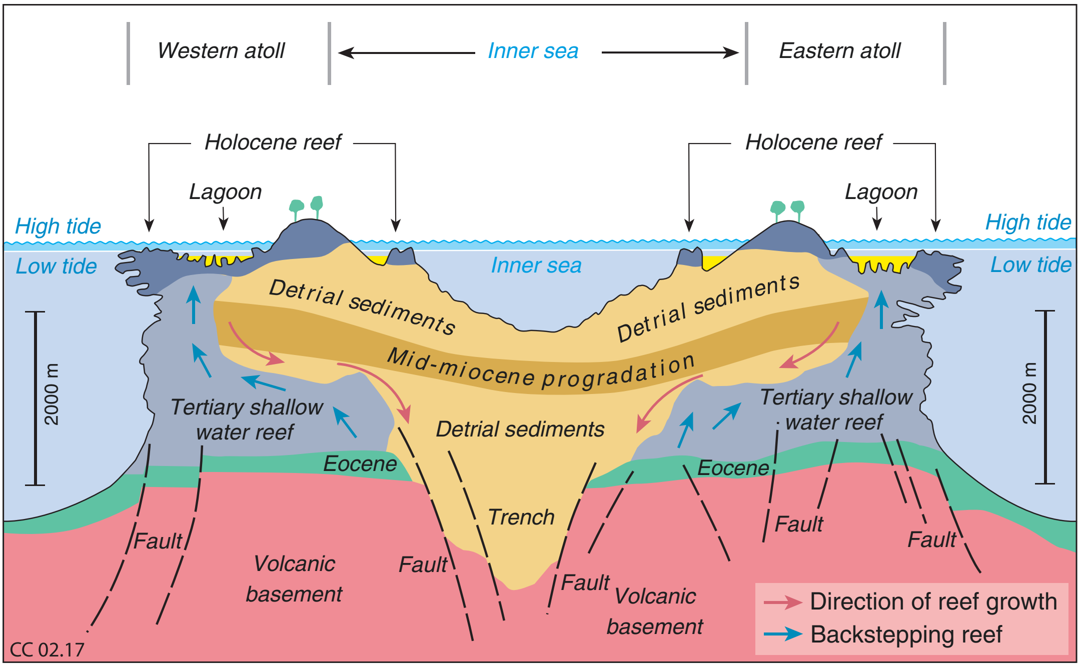


Fig. 2.10 Scheme of an atoll – example: Maldives (After Kench 2011)

Patch reefs form where the seafloor is close enough to the surface for corals to obtain enough sunlight. In terms of their form, patch reefs, like atolls, are ring-like. Unlike atolls, however, their development is not linked to a sunken island, which is why they lack the characteristic deep lagoon of atolls. Patch reefs can form anywhere, even hundreds of kilometres from the coast. Some patch reefs, for example, those that lie within the Great Barrier Reef, can reach a diameter of up to 15 km.

Shelf reef islands mostly trace the relief of the seafloor. Directly on the continental slope, they are often ribbon-shaped; behind the continental slope, they are mostly at a right angle to the shore. Islands spreading out from the shore, for example, may mirror a shelf relief characterised by the arms of a drowned river delta. Such hardened sediments, including also coquina formations and beach rocks, form suitable substrates for reef-forming corals. The growth of shelf reef islands is mainly driven by glacial eustatic variations of sea levels. Three reef types are commonly understood as precursors to islands: apron reefs, pseudo-atolls and patch reefs. Their inner sides often accumulate sand and bivalve debris which often solidifies and reaches above the water level at all times.

2.4.2 Atolls

Atolls are ring-shaped reefs that mostly sit on the slopes of (sometimes huge) submarine volcanoes. One of the first theories on the formation of atolls goes back to Charles Darwin (1809–1882) and his work *The Structure and Distribution of Coral Reefs* (1842). His theory, which became known as the subsidence theory, holds that atolls begin as fringing reefs surrounding a volcanic island that slowly turn into barrier reefs. Over time, the volcanic island gradually sinks into the sea as a result of erosion, seafloor subsidence or rising sea levels; at the same time, the reef continues to grow. It finally becomes an emerging reef and forms a new island. Atolls are typically ring-shaped with a central lagoon. Examples include the Line Islands in the Pacific and the Maldives in the Indian Ocean (see Fig. 2.10).

If the island remains unchanged, and if the surrounding water level allows, the reef continues to grow outwards into the sea. The same applies to rising islands, where reef terraces result from the inner fringe of the reef being lifted above sea level. If the island subsides, a barrier reef forms, separated from the main island by an increasingly deep lagoon. All of these processes assume that sea levels do not change by more than 10 mm per year, as this is the speed at which corals still grow well. Once the island has completely subsided beneath the water, a ringed reef remains, a so-called atoll.

'Atoll' is derived from the Dhivehi word *atolhu* (އަތޮޅު) (a Maldive language). The coral reef forms a fringe of mostly very narrow islands, which are often termed 'Motu' (the Polynesian word for island). Within the lagoon, remnants of the former peak of the volcano may also still appear as islands. Finally, only the reef emerges above sea level, where it forms a ring of small islands. Ringed reefs take up to 10,000 years to grow; if conditions are favourable, they can continue to expand for another 100,000 years. Atoll formation can take up to 30,000,000 years in total. Glacial eustatic changes in sea level undoubtedly also contributed to the formation of these islands (Klug 1985: 204 f.).

Another theory of atoll formation was put forward by oceanographer and zoologist Hans Hass (1919–2013). He suggested that atolls form from originally cone-shaped reefs whose innermost corals die back due to insufficient water supply; only the outer corals continue to grow, leading to a ring-like structure (Hass 1962a, b).

Darwin's theory was verified in 1951/1952 by analysing the results of deep core drilling at Bikini and Eniwetok Atolls in the Indo-Pacific Ocean. At a depth of around 1300 m, the drill went through the limestone of the reef and hit volcanic basalt. This confirmed that the island had gradually subsided while the reef had continued to grow towards the light. Eventually, the reef had overgrown the completely submerged island, forming a perfectly circular atoll around a semi-enclosed lagoon. This process took place over a period of 60 million years (Gierloff-Emden 1980: 976). In 2014 an international team of scientists took two sea containers of drilling equipment to the Southern Seas in order to retest the subsidence theory and to understand past climate fluctuations in order to better understand current climate change. The team was able to prove the subsidence of volcanoes and the concurrent growth of the reef, offering an explanation of how fringing reefs become barrier reefs and eventually atolls. They were also able to confirm subsidence on Bora Bora (Harper et al. 2015) which is at least 6 m and at most 16 m lower than 117,000 years ago. Below the current reef, which began to form 10,000 years ago, the researchers came upon the 117,000 year-old fossile reef which formed before the last ice age. One aspect Darwin was unable to take into account is the vast fluctuation of sea levels. The melting of the ice sheets after the most recent ice age around 20,000 years ago caused a huge rise in sea levels by about 120 m, implying that the volcano and the fossile reef of Bora Bora were simply flooded. It was also shown that fluctuations in sea levels overlie subsidence and that rising sea levels have greater influence on the formation and growth of reefs than subsidence.

The distribution of atolls is determined by their specific genetic origin around a volcanic island. Originally defined for Pacific reefs, these types of coralline islands mostly occur in the tropical waters of the Pacific and Indian Ocean. Darwin (1842) believed that there are no true atolls in the Atlantic. However, by defining an atoll as a geomorphic form (rather than by origin, as Darwin did), Bryan (1953) listed 27 Atlantic atolls, 26 of which are in the Caribbean (Milliman 1969).[6]

Most of the Caribbean atolls are found offshore in Central America, from Yucatan to Nicaragua. Caribbean atolls are somewhat misshapen, lacking the nearly perfect circular or oval symmetry seen in many of their Indo-Pacific counterparts. This is due to the simple fact that

[6] Prior to 1966, five Caribbean atolls had been extensively studied: Alacran Reef (Kornicker and Boyd CR03211962; Hoskin 1963); The Belize atolls, Lighthouse Reef, Glover's Reef and Turneffe Islands (Stoddart 1962); and Hogsty Reef (Milliman 1967a, b) (see Milliman 1969).

most (if not all) appear to have developed in ways other than through the subsidence of mid-oceanic volcanic islands (as is typical of Indo-Pacific atolls) (see Alevizon 2015).

Bermuda is termed a *pseudo-atoll* because its general form, while resembling an atoll, came about very differently – an isolated outpost of Caribbean reef development. Reef life survives there year-round only because of continual exposure to the warm waters of the Gulf Stream. Atoll-like reefs which are not of Darwin's type tend to form around isolated highs formed by local tectonics. Modern examples of exposed (i.e. non-lagoonal) patch reefs occur off the north coast of St. Croix in the Caribbean Sea (Gischler 1994).

The northernmost atoll is the Kure Atoll northwest of Hawaii at 28° 24′ N; the southernmost is Ducie of the Pitcairn Islands at 24° 41′S.[7] The largest continuous atoll is the Great Chagos Bank in the Chagos Archipelago which covers a total area of 12,642 km^2 but only has a dry land surface of 4.5 km^2. The atoll with the largest land area is Kiritimati in the Pacific island state of Kiribati, amounting to 389 km^2 of land surface (and 345 km^2 of lagoon); this is followed by Aldabra with 155 km^2. Many atolls have very small areas of dry land and no drinking water, rendering them uninhabitable. The Saya de Malha Bank, a ring-shaped coral reef, is entirely submerged and therefore not classed as an atoll; it has a total area of around 35,000 km^2 (not including the separate North Bank or Ritchie Bank).

The Maldives is a classic example of a nation state that only consists of atolls. The tiny country in the middle of the Indian Ocean is made up of about 1200 small coral islands grouped into 26 atolls. The Maldives owe their existence to a side ridge of altogether four mid-oceanic ridges, extending over a total length of about 2000 km. All this occurred about 200 million years ago when the former supercontinent of Gondwana in today's Indian Ocean broke up and, accompanied by powerful eruptions, drifted apart. The peaks of the volcanoes jutted up above the warm tropical sea, and corals built the first fringing reefs. Eventually, corals formed huge banks, similar to those found today in the area of the 'Maldive Ridge', which is properly known as the Chagos-Laccadive Ridge. Over the course of time, this ridge continued to sink, with its highest points today at 2200–2400 m below sea level. The Maldivian atoll of Thiladhunmathi-Miladhunmadulu (two names but one atoll geographically) has the largest total area of 3850 km^2, of which 51 km^2 are dry land (see Yamamoto and Esteban 2014).

Later stage events (such as volcanic elevation or lowering of sea levels) can cause atolls to be uplifted, causing the central lagoon to partly or fully dry out. Niue, Nauru and Henderson Island are examples for these processes.

Henderson Island is a classic example of a *makatea* formation – a large coral limestone mass that has been geologically uplifted high above sea level. Located about 170 km northeast of Pitcairn Island, the island takes the form of a raised coral plateau whose area of 37 km^2 makes it by far the largest island of the Pitcairn group. Originally formed as a volcanic structure rising from the seabed, the volcano has long since disappeared and is now capped by aeons of coral growth. It is these layers of fossilised reef that form the structure of the island down to great depth. Local seismic activity and global changes in sea levels have raised the island out of the water so that it is now largely surrounded by limestone cliffs. The island is thought to have been above sea level for the last 380,000 years. Its rocks have been used as a record of climate change, with some of its ancient reefs having being dated to around 600,000 years ago. The island was last inhabited by Polynesians in around 1600 AD, having been settled since 800 AD. The island is surrounded by steep but low cliffs of limestone rocks around 15 m in height – much of which has been undercut by wave action – forming overhangs and caves. Although largely cliff-lined, the island has three large sandy beaches that have formed around the northern coastline and which are nesting sites for green turtles (*Chelonia mydas*). Offshore there is a fringing reef of at least 200 m in width that surrounds the island on the north, northwest and northeast sides. Reefs off the

[7] Atoll-like structures further south are found on the submerged Elizabeth Reef in the Tasmanian Sea at 29° 58′ S.

north and northeast beaches are seaward sloping reef platforms without reef crests and are not typical fringing reefs. There are two narrow channels through the reef on the north and northwestern coasts. The interior plateau of the island has an elevation of up to 30 m and is largely composed of treacherous, dissected limestone formations and reef rubble. Much of the interior forms a central depression that is thought to be all that remains of its uplifted lagoon.

2.4.3 Threats to Coralline Islands

Coral growth is slow and highly sensitive, dependent on a symbiotic relationship with algae and good environmental conditions. Corals cannot grow without light and will die if their symbiotic algae cannot survive. As a result, coralline islands are susceptible to a wide range of man-made and natural threats. Waste water discharge into coastal waters contributes to coral bleaching by suspending faecal matter and coliform bacteria; litter turning into flotsam is a huge problem for humans and animals. Deep-sea fishing and unsustainable practices such as bottom trawling, bomb fishing or using bleach for lobster catching threaten corals. Excessive diving tourism in coral reefs and anchoring also destroy corals and negatively impact on the island's constituting ecosystem. In addition there are various threats arising from climate change. Some reefs have already died off; others are endangered. At present, one in five coral reefs subject to scientific monitoring in the Caribbean is found to be shrinking. The corals' symbiotic algae are very sensitive to temperature: In warmer water, they begin to produce toxins and are expelled by the corals; these then die off, leaving behind the white calcium carbonate skeleton – a process termed coral bleaching. Corals can recover from modest bleaching, but the global warming of the oceans reinforces the process to such a degree that they eventually die (Burke et al. 2011).

While there is much focus on the impacts of warmer ocean temperatures, there is another more direct effect of the burning of fossil fuels and deforestation. More than 30% of the carbon dioxide emitted by humans is dissolved into the oceans, gradually turning the water more acidic. Coral reef researcher Ove Hoegh-Guldberg explains the threat of ocean acidification: 'Evidence gathered by scientists around the world over the last few years suggests that ocean acidification could represent an equal – or perhaps even greater threat – to the biology of our planet than global warming' (Pelejero et al. 2010: 333). These authors label ocean acidification the 'evil twin' of global warming. As CO_2 dissolves in the oceans, it leads to a drop in pH and accelerates the process of acidification. This change in seawater chemistry affects marine organisms and ecosystems in several ways, especially organisms like corals and shellfish whose shells or skeletons are made from calcium carbonate. Today, the surface waters of the oceans have already acidified by an average of 0.1 pH units from pre-industrial levels; signs of its impact can be noted even in the deep oceans (see also Chap. 6 Island vulnerability and resilience).

2.5 Secondary Island Formation Processes

Proper 'outpost islands' are those that have not arisen through the physical or zoogenous processes described above but through processes that can be termed secondary. These secondary processes are mostly cut-off processes where parts of the mainland are separated off as islands, either through ingression or regression of the sea, the erosive force of the surf or tectonic activities. This group also includes islands that have arisen through diapirism (doming), lifting or dislocation (see Table 2.2). Geologically speaking such islands tell the story of the nearby mainland, as only secondary processes have surrounded them with water and turned them into islands. If it were not for the sea, they would be readily visible as outposts of the continent.

Some of these islands are classic outposts in that they have played an important role as physical outposts of the mainland or prison islands. Famous examples, and readily classified as continental outliers, include:

Île d'If/Chateau D'If – a small uninhabited island of about 2 ha, with a maximum elevation of 35 m. It is situated about 1.5 km off the coast of Marseille and part of the Frioul archipelago. Geomorphologically, Ile d'If is part of a submerged karst formation which was part of the mainland before the Holocene sea level rise (Collina-Girard 2004: 8).

Robben Island – situated about 12 km off the coast of Cape Town in Table Bay in the Atlantic. It has an area of 5.07 km^2 and a maximum elevation of 30 m (Minto Hill). In 2011 it was inhabited by 116 people. Robben Island can also be classed as a *continental outlier* as it is part of the Tygerberg formation and therefore part of the Malmesbury Group, the oldest sedimentary rock on the South African mainland (Rowe et al. 2010: 57; Department of Geological Science of the University of Cape Town 2016).

Alcatraz – consisting of sandstone and part of the Alcatraz terrace which also includes the hills around San Francisco. During the Pleistocene Alcatraz was linked to the Californian mainland. During the Holocene, sea levels rose, causing the island to become separated (Konigsmark 1998: 50). This marine ingression makes Alcatraz a continental outlier.

The following section takes a closer look at the various secondary processes of continental outlier islands or outposts.

2.5.1 Subsidence, Ingression and Emerging Islands

Subsidence, ingression and emergence are processes that cannot entirely be separated as secondary island formation processes often occur simultaneously. For example, an island can rise as a result of isostasis and be flooded by raising sea level at the same time.

Marine ingression can lead to the drowning of marginal mainland as a result of land subsidence (isostatic) or sea level rise (eustatic). Subsidence islands result from the gradual submergence of the seafloor, as in the case of Sansibar, Pemba and Mafia (on substrate that arches into the Somali Basin off the east coast of Africa (Valentin 1954: 70) or the Bissagos Islands that developed in a subsiding delta region on the coast of Guinea-Bissau (Buckle 1978: 218).

Ingression islands form when terrain formed by exogenous processes is flooded during a marine ingression, leaving behind higher areas of land as islands and groups of islands. As a consequence of global sea level rise after the last glacial period, all sorts of mainland terrain were caught out by marine ingression. Ingression islands of this type are therefore highly morphologically diverse. The Baltic islands of Gotland and Oland represent a drowned Precambrian escarpment, for example. Bodden islands such as Hiddensee in the Baltic are really the tips of moraines in a drowned peri-Baltic glacial landscape (Hurtig 1959: 47). Ilha do Governador in the Bay of Rio (Guanabara) is a tropical bell-shaped mountain surrounded by water, and Djerba Island is part of the southern Tunisian Djeffra Plain that was cut off by the last sea level rise (Klug 1973: 46; Klug 1985: 195).

Contrary to ingression islands, emerging islands form as a result of isostatic or epirogenetic movements that lift large areas of seafloor. Santiago and São Nicolau of the Cape Verde Islands exhibit a unique geological record rich in palaeo-markers of sea levels, thus allowing different synchronous uplift histories to be discriminated (Ramalho 2011). Examples include also the Scandinavian and Canadian skerries, which are rounded domes and small blocks of rock, partially also moraines or drumlines that jut out of the water off the mainland coast, giving away their origin as part of a former headland. Their origin goes back to the ice ages when inland ice originating from Scandinavia and North America flowed across and abraided the underlying rock. Skerries are often groups of many hundreds of individual islands. The idea of a 'yard' in front of a land mass has led to the name of *skärgård* (Swedish) or *skærgård* (Danish). The syllable 'gård' is related to the English word 'yard' but was understood in the German-speaking region as 'garden', leading to the skerries also being termed a 'skerry garden'.

Strictly speaking there are two types of skerries in Scandinavia where – depending on their position – marine ingression meets glacial isostatic emergence and meets subsidence. The Baltic skerries, such as those off Turku in Finland, are really rounded bosses that were flooded; as they are situated in glacial isostatic rebound areas, their current genesis is unaffected by the sea. As a result of their continued uplift, marine clay deposits are emerging that offer possibilities for agriculture on many skerries. The island fringe on the open tidal coast of western Norway from Stavanger up to the Vesteralen islands is part of a Strandflat, forming the foreshore of an inland glacier that was significant in the Pleistocene. These skerries are really denudation rumps that were formed below the shelf ice by constant tide-dependent freeze-thaw cycles (Gierloff-Emden 1980: 1214f; Tolvanen et al. 2004; Scheffers et al. 2012: 73ff). Whether formed by isostatic lift or marine ingression, the result is a garden of skerries that permits an impressive association: a sea of islands (see Fig. 2.11).

2.5.2 Residual or *Outlier Islands*

Outlier islands are the result of erosion processes that cut them off from the mainland. Residual islands are leftovers of mainland destroyed by transgression. Both types can be termed outlier islands, in analogy to the vocabulary describing the shapes of a scarp landscape. Quite often these islands are buttes or hard cores of rock that, unlike their surrounding land, were able to resist erosion.

The Channel Islands of Guernsey and Jersey are the remains of a former continental plateau; they owe their existence to the hardness and resistance of their constituting rock, which is mostly igneous and strongly metamorphic. The Oligocene brought with it erosion caused by transgression; however, the islands were only separated from the continent of Europe by rising sea levels at about 5000 BC during the new stone age (Power 1997: 276–277) .

The North Frisian Wadden Sea along the German Coast has several examples of residual islands. Föhr is the second largest and Pellworm the third largest island in the tidal marshes off this coast. The elevated Geest cores of the North Frisian Islands, situated within wide expanse of marshland, attracted people in the Neolithic period when sea levels in the North Sea rose. Up to the major storm surge in 1362, Föhr was not an island but part of the mainland, linked to the North Sea by deep channels of water. That same storm surge drowned the port town of Rungholt, later steeped in legend and formed a new island called Strand. Pellworm mostly consists of the western part of Strand which was destroyed in another storm surge in 1634. Pellworm, Nordstrand and some of the Halligen islands are all residuals of the former island of Strand (see Fig. 2.12).

2.5.3 Dislocation Islands (Horst and Drift Islands)

Islands can also form as a result of horizontal or vertical tectonic shifts, giving rise to drift and uplift islands, respectively. Vertical tectonic shifts can lead to the lifting of blocks, giving rise to a horst island. Horizontal tectonic movements are often linked to plate tectonics, leading to the separation and migration of peripheral parts of a lithospheric plate; in this case drift islands result. These are fragments of the mainland, forming part of the respective continental crust.

Bornholm is a typical example of a *horst island* that arose from tectonic lift. The entire rim of Fennoscandia was lifted as a predominantly crystalline block of rock (Blüthgen 1975: 193). In the Mediterranean, Malta, Gozo, Comino and Lampedusa are horst islands perched on the lifted shoulders of a trench on both sides of the Pantelleria reef system (Henning Illies 1980: 152) (see Fig. 2.13). Often, the erosive forces that cause islands to separate off from the mainland go hand in hand with vertical tectonic movements. Such mixed forms have shaped the island of Capri off the coast of Italy. The nearby Sorrentine peninsula is a rocky outlier in that it

Fig. 2.11 Subsiding and emerging islands – example: skerries east of Stockholm

forms a continuation of the mainland, but its shape indicates that it is also a tectonic horst.

Madagascar, the Seychelles, Corsica and Sardinia are examples of *drift islands*. Madagascar and the Seychelles split off from the African continental plate about 100 million years ago and gradually drifted into their current position as a result of a complex rift system (Smith and Hallam 1970). Sardinia and Corsica split off from the mainland during the late Tertiary. They reached

Fig. 2.12 Residual islands – examples: Föhr, Pellworm and the North Frisian Coast (After Umweltbundesamt 1998)

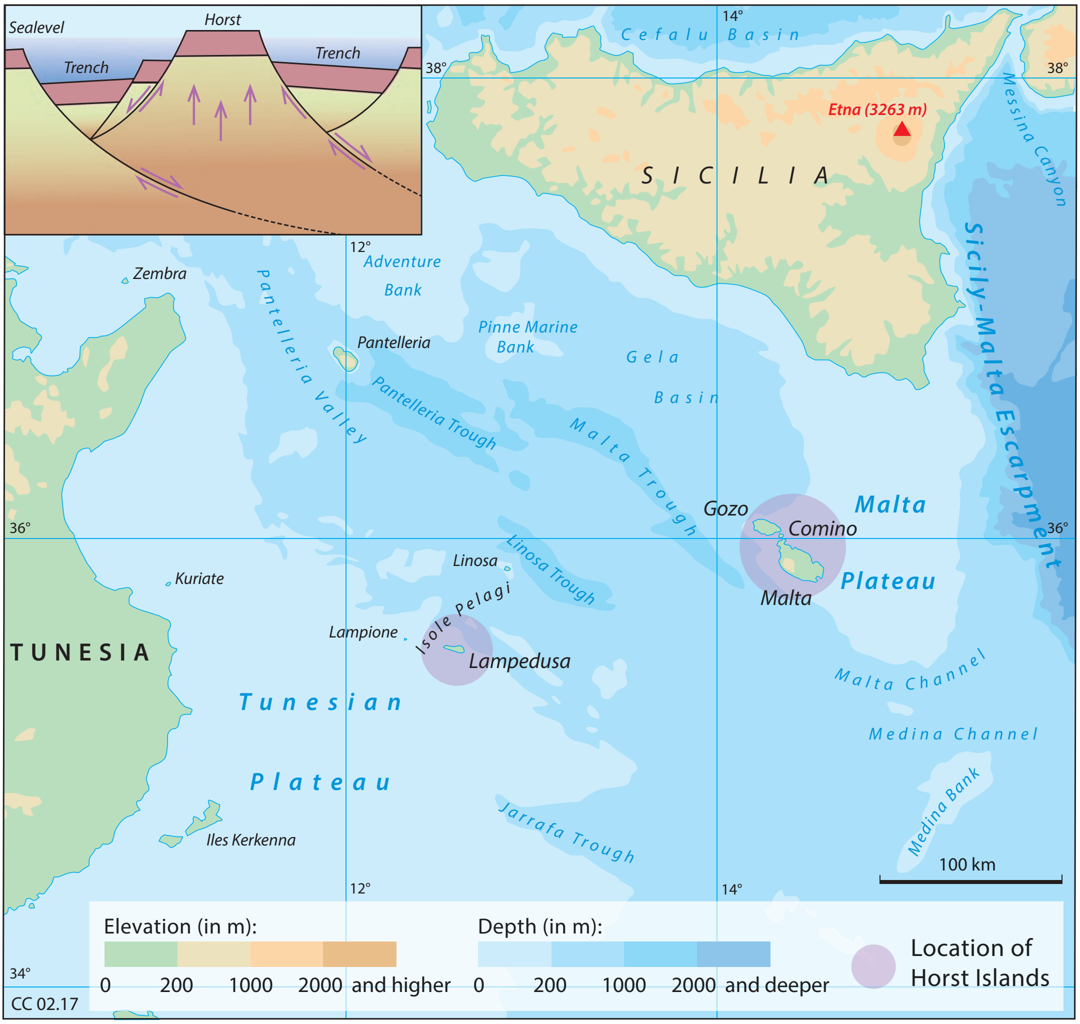


Fig. 2.13 Horst islands – examples: Malta, Gozo, Comino and Lampedusa

their current position following a rotation process which resulted from the development of Mediterranean microplates – a process that must have been completed six million years ago at the latest (Alvarez 1972) (see Fig. 2.14).

The South Shetland Islands represent a mixed type in that they were separated from the Antarctic Peninsula by a rift in today's Bransfield Strait and then tectonically lifted (Curl 1980: 11).

2.5.4 Diapir Islands

Some islands owe their existence to processes originating in the seafloor. Salt tectonics or intrusion can lead to the local emergence of a small area of seafloor through doming, leading to the formation of diapir islands. Helgoland is such an island, whose main island genetically represents a bunter (sandstone) horst lying on top of a Zechstein salt dome (Kremer 1983: 183) (see

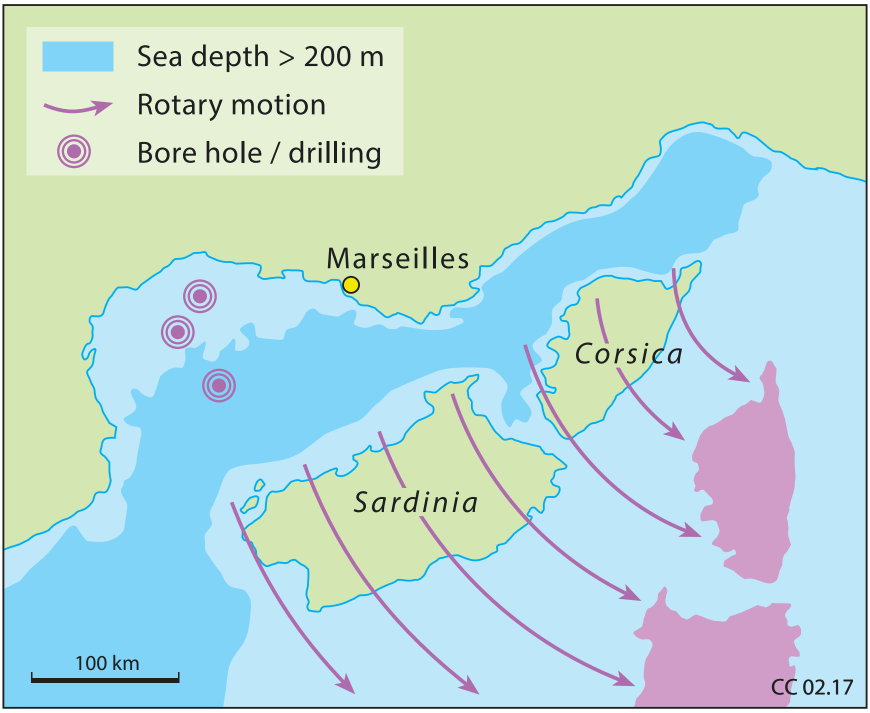


Fig. 2.14 Drift islands – examples: Sardinia and Corsica (After Klug 1985)

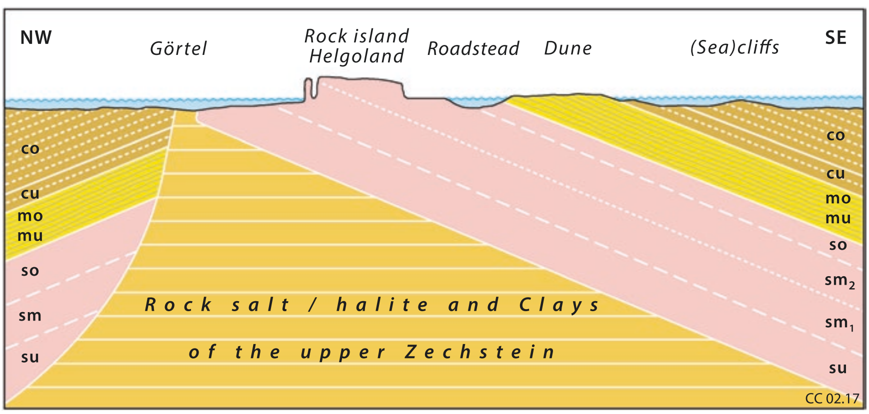
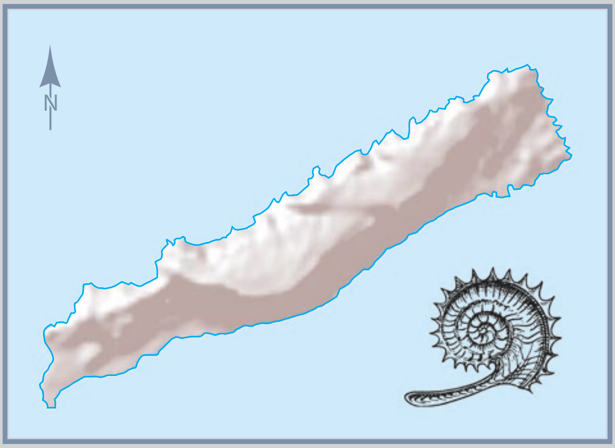


Fig. 2.15 Diapir islands – example: Helgoland (After Klug 1985)

Fig. 2.15). Many small islands in the Persian Gulf also form part of this category, such as Zarqua whose genesis, like Helgoland, is the result of a salt diapir below the seafloor.

Many islands across the world are of mixed genesis and could easily be assigned to several of the types described. This fact, or the progressing scientific understanding of complex physical processes, may be the reason why geographers, after their initial efforts in the 1970s, have been reluctant to attempt a general typification of islands. The larger the island, the more physical and biological processes interact in giving it its present face. These processes are never complete. Islands continue to emerge, erode and be destroyed. Islands come and go.

Island Brain Teaser 2



According to current geological knowledge, planet Earth is about 4.6 billion years old. But the visible surface of the Earth is comparatively young and continuously in motion, so that current topography is only ever a snapshot. Unsurprisingly, some islands are relatively old in Earth terms, while the birth of others, mostly in volcanic island regions such as Iceland or Tonga, can be witnessed live and in colour.

This chapter's mystery island is a silent witness of the formation of an entire tectonic plate. The plate initially formed about 150 million years ago through hotspot volcanism on the Pacific seafloor and subsequently moved east over thousands of kilometres – into an ocean which was only just emerging then. The plate reached its current position about 100 million years ago where it represents the foundation of an entire island region. On the island bizarre tuff and magnificent basalt formations along the rugged coast of the island are indicative of this history. Special forms of pillow lava point to the island's underwater origins at a depth of 3000–4000 m; marine microfossils are evidence of its long plate-tectonic journey along the Earth's surface.

Humans first settled on the island around 5000 years ago. In the middle of the last millennium, the island, which is today only 22 km^2 in size, was the desired anchoring point for European conquerors after their long and uncertain journey across the Atlantic. The indigenous population fought valiantly to prevent a European takeover, but the military forces of the colonisers were vastly superior. The last indigenous residents were eventually deported to a neighbouring island, and in consequence a plantation-based economy was introduced, typical at the time for the entire region.

Although the first European visitor sailed under the Spanish flag, today's 1500 residents mostly speak French and earn their living from agriculture, fishing and a little tourism. Above all, though, the island has remained a Mecca for interested geologists from all over the world. Which mystery island are we looking for?

For the solution please visit http://www.island-database.uni-hamburg.de/about.php

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