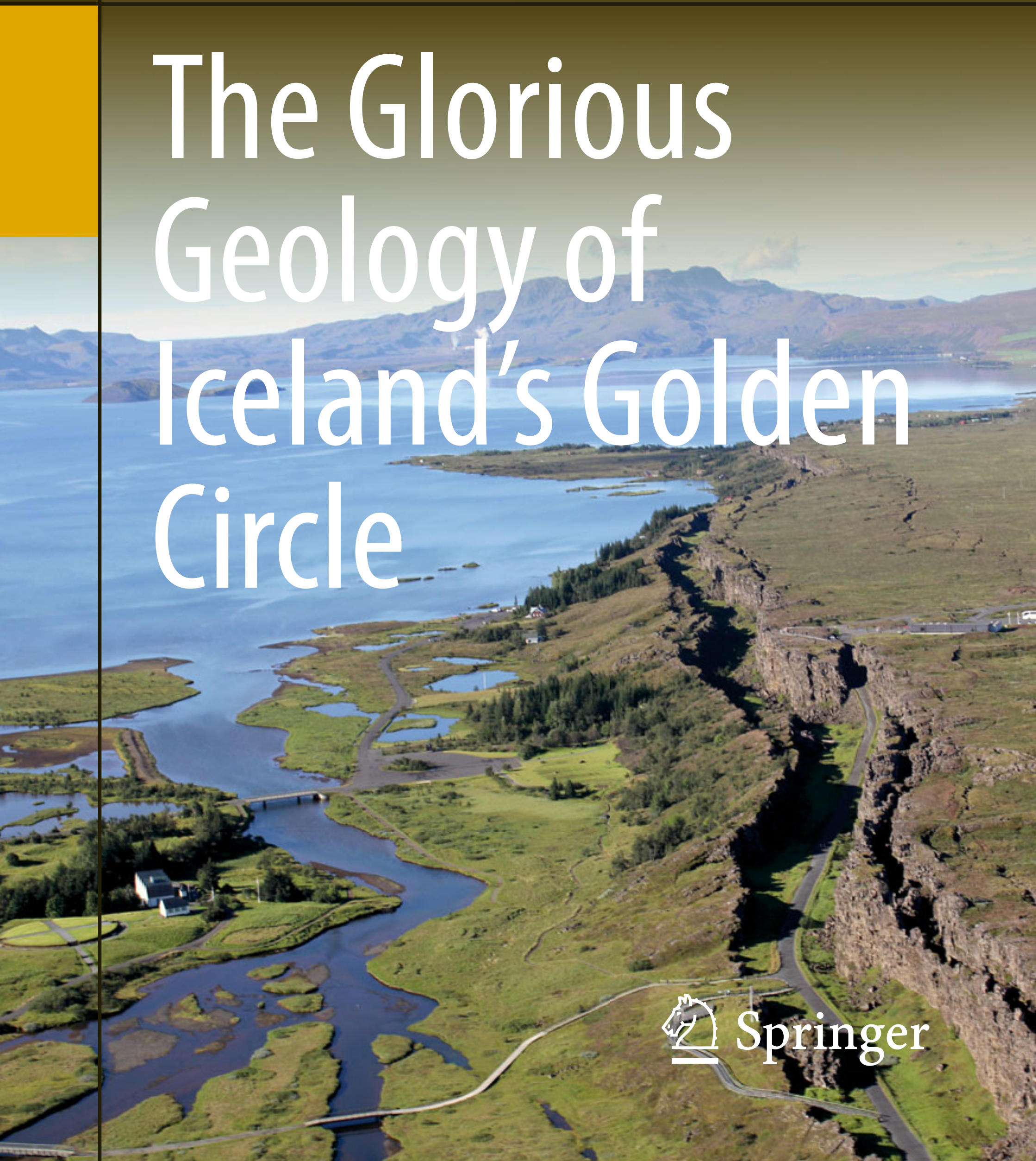
GeoGuide

Agust Gudmundsson

The Glorious Geology of Iceland's Golden Circle



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Preface

Iceland's Golden Circle provides a unique opportunity to observe and understand many of Earth's natural forces in action. These include those that move the Earth's tectonic plates, rupture the crust, and generate earthquakes and volcanic eruptions. Earthquakes and volcanoes, in turn, provide the paths and the heat sources for hot springs and geysers. These are internal forces which, in combination with external forces, are also responsible for forming the landscape. The external forces relate to the actions of glaciers and rivers that erode the Earth's surface so as to generate waterfalls, river canyons, and, eventually, mountains.

The Golden Circle does not only illustrate natural forces but also natural resources. Among these is plenty of exceptionally clean and readily available groundwater. As humankind moves towards renewable energy, potential or actual renewable energy resources become of gradually greater importance. Those you can see—some from a distance—while travelling the Golden Circle include waterfalls (hydropower), geothermal fields (geothermal power), waves (wave energy) and—perhaps occasionally less welcome—the wind (wind power).

The Golden Circle is normally travelled (driven) in one day. There are many other exciting places that can be visited during one-day excursions. Here I describe four additional excursions that are less well known than the Golden Circle and all of which offer a deeper understanding of the processes that shape our planet as well as places of great landscapes and beauty. Since most visitors to Iceland stay in the Greater Reykjavik area (the Capital Region), all the excursions described here begin and end in Reykjavik. However, these excursions can be made wherever you stay in the southwestern part of the country.

More specifically, the book describes five one-day excursions in the southwestern part of Iceland. The first excursion is the classic Golden Circle, which includes the well-known sites of rifting at Thingvellir (Þingvellir), the geothermal area of Geysir, the waterfall Gullfoss, and the volcanic crater Kerid (Kerið). The second excursion is to the beautiful fjord Halfjördur (Hvalfjörður), north of Reykjavik, where you can observe the deep interiors of volcanoes and volcanic zones as well as a variety of impressive landscape features. The third excursion focuses on the unique landscape, volcanic activity, and geothermal energy of the Hengill Volcano, south of Lake Thingvallavatn. The fourth excursion is to the Reykjanes Peninsula, south of Reykjavik, which contains the Blue Lagoon and the 'Bridge between two continents' and focuses on lakes, explosion craters, geothermal fields, volcanic fissures, and lava fields. The fifth excursion is to South Iceland and includes the main earthquake zone in this part of Iceland but focuses on the famous volcanoes Hekla (erupted in 2000), Eyjafjallajökull (erupted in 2010), and Katla (erupted in 1918), as well as the waterfalls, sandur plains, and the beautiful rock columns at the beach of Reynisfjara.

The book is written for the general visitor to Iceland. In particular, the book is for people who not only wish to enjoy Iceland's unique beauty, but also to appreciate and understand the processes that create that beauty. No geological knowledge is assumed. Technical terms are avoided as much as possible, and those that must be used are explained using non-technical language in main text and in a detailed glossary at the end of the book. I have been a guide on numerous geological excursions in Iceland, including all those described in the book. Some of the excursions have been primarily for people educated in geosciences, whereas others have been for people with no such background. While the book is aimed primarily at people with no background in geosciences, many geoscientists and students may benefit from the well-illustrated field examples and explanations of processes presented in the book. Many readers may neither have the time nor inclination to read the entire book. I have therefore written the chapters so as to make them comparatively independent of other chapters. It is thus possible to go for a single excursion and read the relevant chapter without having to read all the other chapters. For this reason there is considerable repetition of various terms, principles, and processes. To clarify points of interest or under discussion there is, however, much cross-referencing between chapters and, in particular, figures.

In many science books for the general public most of the figures are line drawings, cartoons. This book is unusual in that, while line drawings are used to illustrate certain geological processes, the focus is on explaining the geological structures and processes through annotated photographs. The advantage is that the processes are then explained in terms of structures as you really see them in nature. As a consequence the book has more than 240 illustrations, the great majority of which are photographs. Most of the photographs show the geological features exactly as you will see them during the excursions. There are, in addition, many photographs taken from aircrafts, providing aerial views of the same features. These are meant to explain processes and structures at a different scale from that seen on the ground, as well as to underline and emphasise the unique beauty of the geology of Iceland.

Egham, UK Agust Gudmundsson

Acknowledgements

In working on the book I have received much, and greatly appreciated, help from my wife Nahid Mohajeri. In particular, she has redrawn earlier illustrations and made all the original illustrations in the book. Furthermore, she has read the entire text and made many helpful comments. I would also like to thank Ines Galindo and Sonja Philipp for their meticulous reading of the entire text and for many helpful suggestions.

All the digital elevation models presented in the book are based on data from Landmælingar Íslands (Iceland Geodetic Survey). More specifically, these are Figs. 1.1, 1.5, 2.1, 2.3, 4.1, 10.1, 11.1, 12.1, 13.1, 14.1, 14.12b, and 14.18. Also, the aerial photographs in Figs. 8.2 and 13.7 are from Landmælingar Íslands. These data are made available for free by Landmælingar Íslands through an 'open government licence'. I thank Landmælingar Íslands very much for the free use of these data and aerial photographs.

I gratefully acknowledge being allowed to use photographs taken by the following persons. By Ines Galindo: Figures 13.20, 13.21, 13.23, 13.34, 13.35, 13.36, and 13.39; by Sonja Philipp: Figures 7.1, 9.4, 12.4, 12.5, 12.18, 12.19, 14.25 and 14.28; by Ævar Johannsson: Figure 13.27. All other photographs in the book were taken by me.

Some of the illustrations were modified from my earlier papers and books. The following figures were modified from figures in the listed papers. Figures 2.2, 2.4, 4.10, and 11.14 from the paper *The mechanics of large volcanic eruptions.* Earth-Science Reviews, 163, 72–93, 2016; Fig. 5.9 from the paper *Effects of mechanical layering on the development of normal faults and dykes in Iceland.* Geodinamica Acta, 18, 11–30, 2005; Fig. 11.3 from the paper *How local stresses control magma-chamber ruptures, dyke injections, and eruptions in composite volcanoes*. Earth-Science Reviews, 79, 1–31, 2006; and Fig. 14.20 from the paper

Strengths and strain energies of volcanic edifices: implications for eruptions, collapse calderas, and landslides. Natural Hazards and Earth System Sciences, 12, 1–18, 2012. Similarly, Fig. 13.26 is modified from a figure in the book *Rock Fractures in Geological Processes*, Cambridge University Press, 2011. All the other illustrations were particularly made for the book.

Agust Gudmundsson

Contents

1 Introduction

If you want to understand the natural forces that shape our planet, then the place to go to is Iceland. There is no equally small area on this planet that offers such a variety of easily observed geological processes as Iceland. You can observe the **tectonic plates** that constitute the Earth's surface moving apart, also known as **spreading**, and thereby giving rise to volcanic eruptions, earthquakes, geysers, and geothermal energy. Then there are the large rivers with beautiful waterfalls, as well as the ice caps and the outlet glaciers that have cut or eroded deep valleys and fjords into the land, the surface of the crust, forming tall mountains in-between. All these are easily observed and understood in Iceland.

Because the geological processes and structures are so clear and easily observed in Iceland they can be understood without any geological background. Just looking at the landscapes and the rocks, you should, with the help of this book, be able to recognise the main landforms, by which processes they are generated, and how the processes operate. The aim of the book is to illustrate and explain what main landforms and geological processes can be seen during five **one-day excursions** (described below) in a way that is understandable to those without a formal education in geology. With this in view, I keep technical concepts and jargon to a minimum and use everyday examples and analogies to explain the landforms and processes that you can see in Iceland. Every concept is defined when it is first introduced, and in addition there is a detailed **glossary** at the end of the book, summarising in simple terms the meaning of some common geological and other scientific terms. I use **boldface** type for emphasis and for words that are explained in the glossary. More specifically, boldface is used for important items, technical or semi-technical terms, particularly where first used, and for the stops during the excursions.

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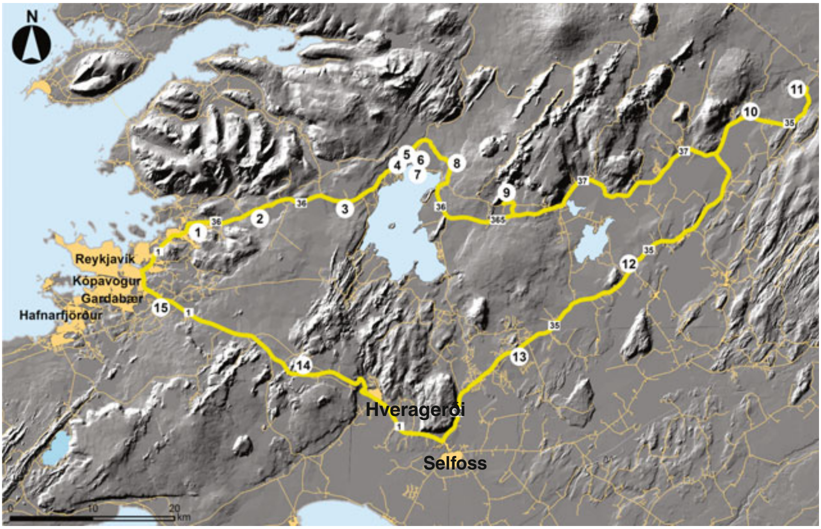

Fig. 1.1 The location of the Golden Circle, shown in yellow. The large encircled numbers, here and on other similar maps, refer to the main stops during the excursions. The small numbers in squares indicate the road numbers. The orientation of north is shown by an arrow in the upper left corner, whereas the scale is indicated in the lower left corner. The city of Reykjavik and the main towns are indicated in orange-brownish colour. The location of the Golden Circle within Iceland as a whole is provided in Fig. 1.5 and a larger version, with more details, in Fig. 4.1. Although the real geometry of the Golden Circle is that of a triangle rather than a circle, the name is traditional and use here. Fig. 4.1 is a larger version of this figure

All the processes mentioned above and their products can be seen in the southwestern part of Iceland, and many along the so-called 'Golden Circle'. While there are somewhat different definitions of the Golden Circle—which is geometrically not really a circle but rather closer to a triangle (Fig. 1.1)—to most people the Circle is composed of the following trips:

- From Reykjavik to the Thingvellir Graben.
- From Thingvellir to the Geysir geothermal field.
- From Geysir to the waterfall of Gullfoss.
- From Gullfoss to the crater of Kerid.
- From Kerid to Reykjavik.

Some versions of the Golden Circles include additional stops. For example, some prefer to stop at the Skalholt Cathedral in central South Iceland, an important historical site but geologically of little interest—and thus omitted here. Similarly, some include the town of Hveragerdi and the power plant of Hellisheidarvirkjun. Both are located in geologically interesting places, not only in relation to geothermal fields and geothermal power but also because they are located close to (Hveragerdi) or inside (Hellisheidarvirkjun) active volcanoes (Hengill) and zones of earthquakes. Hveragerdi and Hellisheidarvirkjun are both discussed briefly in the present version of the Circle, and are optional stops.

The Golden Circle described here is the one in Fig. 1.1. It includes 15 main stops or sites and thus many more than listed in the bullet points above. These additional stops are chosen because of their geological interest. I show photographs from some of the main sites in Fig. 1.2. But the descriptions are not confined to the main sites—Esja, Thingvellir, Geysir, Gullfoss, and Kerid. There are many interesting geological features that can be observed between the main sites. Most of these can be seen from the roads that constitute the standard Circle; for others you need to drive a short distance off the main road. For example, on Road 365 from Thingvellir to Geysir, a short drive to north off that road to Laugarvatnshellar (the Caves of Laugarvatn, the ninth stop) offers many geological features of interest. Not only are the caves themselves interesting (they were inhabited in the early part of the 20th century), but nearby is a section through an inactive volcano that allows you to understand how volcanoes and volcanic islands form during eruptions in water, such as the meltwater of ice caps. In particular, the ninth stop is the location of beautiful **pillow lava** (Fig. 1.3), the type of lava commonly formed at great water depths at mid-ocean ridges. The pillows are found here because the mountain formed in deep water, more specifically in the melt water within the ice sheet of the last ice age. Eruptions in deep water, in the sea or under ice caps, are still happening in Iceland. For example, one such eruption occurred in the Vatnajökull ice cap (located in Fig. 2.2) in 1996 (Fig. 1.4), and the 2010 Eyjafjallajökull eruption (Chap. 14) was partly within an ice cap.

The chapters related to the Golden Circle are broadly of two types. One type describes remarkable geological features and processes seen on the way between the main sites (for example, between Thingvellir and Geysir). The other type describes the main sites themselves, the interesting features that can be seen, and by which geological processes they form and develop.

There are many exciting and beautiful geological structures and activities that can be seen in the vicinity of **Reykjavik** in addition to those of the Golden Circle. I presume many who come to Reykjavik would be interested in seeing more of Iceland's fascinating geology than just the Golden Circle. I have therefore added

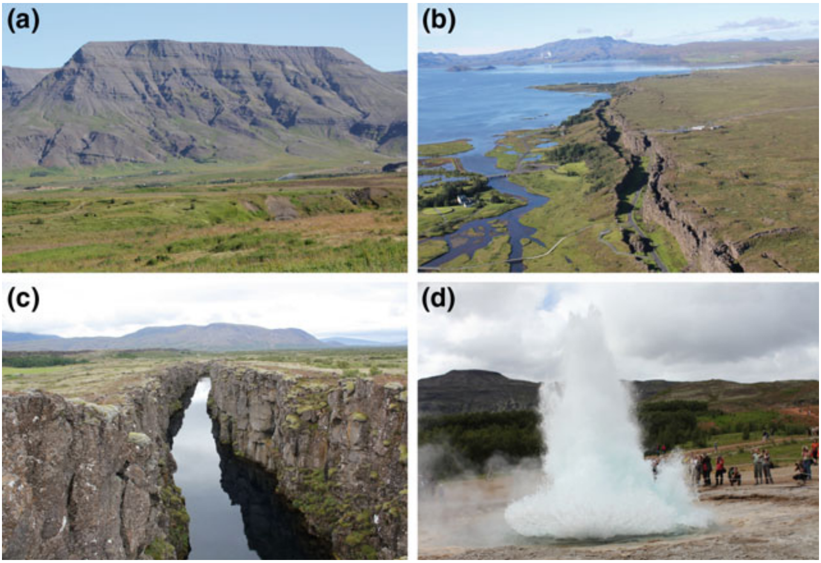

Fig. 1.2 Examples of geological and landscape features seen while travelling along the Golden Circle. All these features are shown again, and discussed in great detail, in Chaps. 4–9. **a** The mountain Esja, which can be seen from Reykjavik (first stop in Fig. 1.1). The rock layers that constitute this mountain originated tens of kilometres to the east, in the volcanic rift zone at Thingvellir, and have been carried through slow spreading (about 1 centimetre per year) to where they are seen now. **b** Part of the active volcanic rift zone at Thingvellir. The photograph is taken from an aircraft, view southwest across Lake Thingvallavatn to the central volcano (stratovolcano, composite volcano) of Hengill—the white 'smoke' is from geothermal fields. The large fracture is an earthquake fault, as wide (open) as 60 m, formed by spreading or plate-tectonic forces (the fourth, fifth, and sixth stop in Fig. 1.1). The land left (to the east) of the fracture has subsided by 40 m. **c** A tension (open) fracture at Thingvellir (the seventh stop in Fig. 1.1). The maximum opening (aperture) of the fracture is about 15 m and the maximum visible depth 25 m, but it may reach to a depth of several hundred metres. The fracture is filled with very clean groundwater. **d** The erupting hot spring, geyser, Strokkur (the tenth stop in Fig. 1.1). The fractures supplying the boiling water for the eruptions maintain their openings or apertures through earthquake activity. **e** The Gullfoss waterfall (the eleventh stop in Fig. 1.1). The total drop is about 32 m and occurs in two steps which follow the directions of the main earthquake fractures in South Iceland. Through erosion, the waterfall is gradually migrating further inland, by about 30 centimetres per year. **f** The crater (volcano) Kerid (the thirteenth stop in Fig. 1.1). It is a collapsed small lava pond, a pit crater, now partly filled with groundwater. Its maximum diameter is about 300 m and depth about 50 m

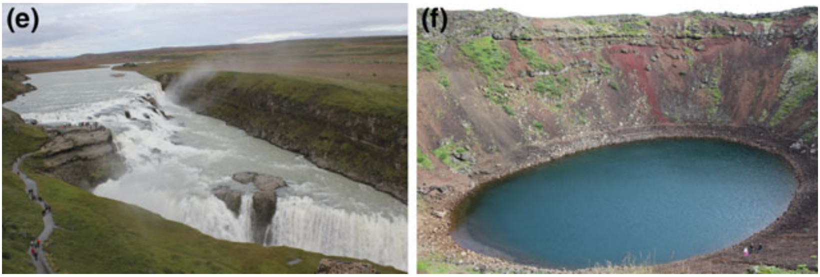

Fig. 1.2 (continued)

several one-day excursions to sites in the vicinity of Reykjavik, namely the following (Fig. 1.5):

- Reykjavik-Hvalfjördur.
- Reykjavik-Hengill.
- Reykjavik-Kleifarvatn-Reykjanes.
- Reykjavik-Eyjafjallajökull-Reynisfjara.

Some of the highlights from these additional excursions are shown in Fig. 1.6. In Hvalfjördur, for example, you can observe the magma-filled fractures (now frozen as solid rock, that is, **dikes**) that supply magma to volcanic eruptions. In Hengill, an active volcano, there is a very unusual and beautiful landscape of valleys and ridges, formed by eruptions and earthquake fractures, that is, **faults**. In the excursions to Reykjanes, you can explore a major geothermal field, explosion craters, the Blue Lagoon, and the fracture referred to as the 'Bridge Between Two Continents'. And in the excursion to Reynisfjara you can observe the famous volcanoes Hekla (from a distance), Eyjafjallajökull, and Katla, in addition to beautiful waterfalls and sets of fractures, referred to as columnar joints, formed when hot molten rock, magma, freezes or solidifies slowly.

Most people who visit Iceland arrive at the Keflavik Airport (Keflavikurflugvöllur). From there most of them drive to the capital Reykjavik or one of its surrounding towns. The Keflavik Airport is located on the Reykjanes Peninsula, which has many interesting geological features in addition to the well-known Blue Lagoon. Some of these features can be seen from the road while driving to Reykjavik. Since this drive provides the introduction of most people to the landscapes and geology of Iceland—although few would describe this part of the country as beautiful—it is logical to start the present book with a brief chapter on the geology seen from the road from Keflavik to Reykjavik (Figs. 1.5 and 2.1).

Before we begin that journey, however, a few words about numbers and spelling of names. As for numbers, particularly the ages of rocks, I commonly give only the first digits and then add the appropriate word, such as thousands or millions. For example, I write 13 thousand years old and 2 million years ago. When the number is presumed very accurate, then I write out the entire number, such as for an eruption that occurred in the year 1000 or the settlement of Iceland which is supposed to have happened in the year 874. I do not use commas in 4-digit numbers, including years. Thus, I write 1200 °C rather than 1,200 °C.

Fig. 1.3 Lava flows that resemble a stack of pillows, that is, ellipsoidal bodies of (mostly basaltic) lava are named pillow lavas (ninth stop in Fig. 1.1). Molten rock, magma, forms pillow lavas under water, the water being meltwater when the eruption occurs beneath a glacier (subglacial eruptions) like here. Individual pillows are commonly about 1 m or less in diameter. Pillow lavas form the lowermost layers in many volcanoes seen in Southwest Iceland, most of which are formed in subglacial eruptions and referred to as hyaloclastite (in Icelandic moberg) mountains. Pillow lavas are also very common at mid-ocean ridges, and some in Iceland are formed in eruptions in the sea (submarine eruptions)

Fig. 1.4 The 1996 Gjalp eruption in the ice sheet Vatnajökull (located in Fig. 2.2) in Iceland. This eruption melted through the glacier, forming hyaloclastite and presumably pillow lava (Fig. 1.3), and caused an enormous flood on the sandur plaines in southern Iceland. Vatnajökull and its volcanoes are outside the scope of the present excursions, but this photograph is a reminder that the processes forming the pillow lavas and most of the volcanoes that you see in the excursions in Southwest Iceland are still operating. The dark, fractured surface is the surface of the ice sheet and the depression and fractures are because of melting from the hot magma beneath the ice sheet. This photograph is from the beginning of the eruption

For 5-digit numbers and higher, however, I use comma, such as in describing a map scale as 1:17,000. As for names, Icelandic has ten letters that do not exist in English. These are the letters á, ð, é, í, ó, ú, ý, þ, æ, and ö. Notice that the letters ö and æ are specific letters and sounds, as are the letters with acute accents, such as á and é. In the book, I transliterate the letters ð as d and þ as th, as is normally done, and æ as ae. Also, I keep ö but omit all acute accents: that is, I write a for á and u for ú, and so forth (as in my own name, which is Ágúst in Icelandic but Agust in English). In the following chapters where the name occurs for the first time (and for some more often) in the main text I give the Icelandic spelling in parentheses following the English spelling. In some chapter headings, I also give the Icelandic spelling when it differs much from the English spelling, such as in the word

Thingvellir, which in Icelandic is Þingvellir. I have translated some of the Icelandic geographical names, particularly those that are geologically interesting, but most are untranslated. In the road maps showing the excursions I use the Icelandic spellings of names because that is the way the spelling is on the topographic and geological maps that you are likely to buy and use.

Which brings me to the topic of **maps**. While I show all the main roads associated with the excursions, and associated digital elevation (shaded topographic) maps, I do not provide detailed topographic maps. Nor do I provide geological maps, but I give full reference to the most important ones for this book in the list of references at the end of the book. Topographic and geological maps are readily available in bookstores and elsewhere in Reykjavik. In particular, there

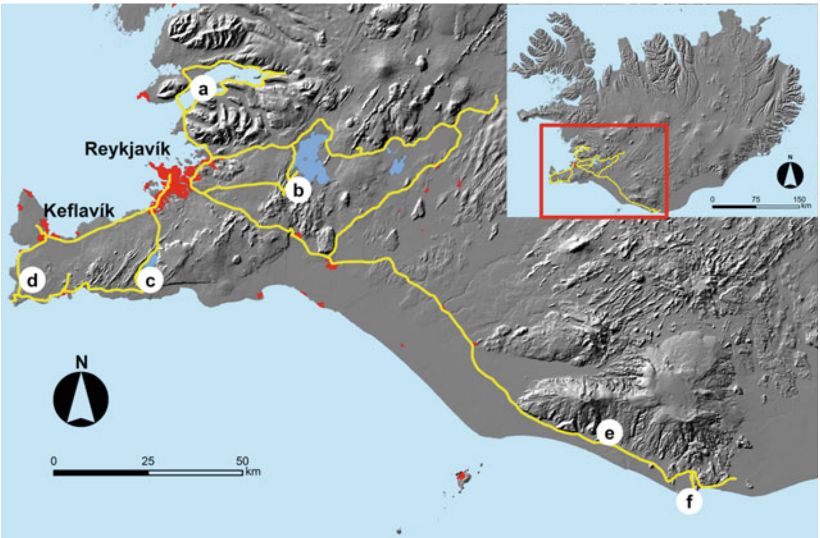

Fig. 1.5 The standard 'Golden Circle' is indicated in Fig. 1.1. To make the book more useful, it covers four additional excursions, shown here in addition to the Golden Circle, in relation to Iceland (inset map). These are to **a** the fjord north of Reykjavik (Hvalfjördur), **b** the Hengill Volcano (seen in Fig. 1.2b), **c**, **d** the Reykjanes Peninsula, and **e**, **f** Eyjafjallajökull (erupted in 2010) and the coast of Reynisfjara (cf. Fig. 10.1). These extra excursions allow you to see deep into the structure of (inactive or extinct) volcanoes, observe geothermal fields and explosion craters, see the famous volcanoes of Hekla and Eyjafjallajökull (located in Figs. 14.1 and 14.12 b) as well as exploring some of the most beautiful waterfalls and beaches of Iceland. Some of the photographic highlights of these excursions are located by the letters **a** to **f**, the corresponding photographs being shown in Fig. 1.6

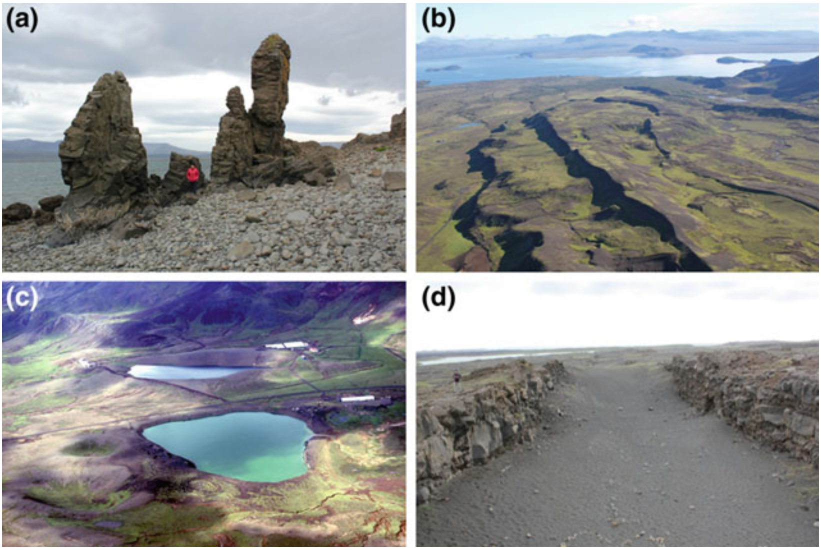

Fig. 1.6 Examples of geological and landscape features seen while making the extra excursions shown in Fig. 1.5. All these features are shown again, and discussed in great detail, in Chaps. 11–14. **a** Molten rock or magma is normally transported in the Earth's outermost solid layer, the crust, along fractures. When the magma subsequently freezes in the fractures, it forms structures named dikes (or dykes). Here is a dike in the fjord Hvalfjördur (located at **a** in Fig. 1.5) which forms a sea stack because it is harder (more resistant to erosion) than the surrounding host rock. The horizontal columns form during cooling of the magma (discussed in **f** below). **b** Aerial view of earthquake fractures (normal faults) in the Hengill area (located at **b** in Fig. 1.5). View north to Lake Thingvallavatn, the faults are large in the old rocks but become smaller in the young lava flow close to the lake. **c** Aerial view of explosion craters, also known as maars, near Lake Kleifarvatn (located at **e** in Fig. 1.5). The one with the green water has a maximum diameter of about 360 m and a depth of 45 m. **d** The tension fracture across which is the 'Bridge Between the Two Continents' (located at **d** in Fig. 1.5). The maximum opening or aperture of the fracture is about 30 m (15 m where the bridge crosses it). **e** The Skogarfoss waterfall (located at **c** in Fig. 1.5) falls or drops vertically about 60 m off an old sea cliff. The cliff was formed some 13 thousand years ago when the sea level was much higher than today. **f** When a magma body cools to form a solid rock (here a basaltic intrusion), the body shrinks and may generate beautiful columns. Here some of the rock columns in Reynisfjara (located at **f** in Fig. 1.5) are seen. The columns are vertical, indicating that they formed in a horizontal sheet-like magma-filled fracture known as a sill

Fig. 1.6 (continued)

exists a very detailed geological map (scale: 1:100,000, meaning that one centimetre on the map corresponds to one hundred thousand centimetres, or one kilometre, in nature) of Southwest Iceland. There are also available geological maps of Southwest and South Iceland (scale: 1:250,000) that cover all the excursions. For those who like great details, perhaps the best are the photomaps, that is, aerial photographs with all the place names and elevation contours shown. Such maps (scale: 1:17,000) are available of the Reykjanes Peninsula, Gullfoss and Geysir, and some other areas. The topographic map Reykjanes-Þingvellir (scale: 1:100,000) is very useful and includes a special and more detailed map of Thingvellir (Þingvellir) itself. In addition, there are many other geological and topographic maps, of parts of Iceland as well as the entire country.

2 Keflavik to Reykjavik

The road we drive from the **Keflavik (Keflavík)** Airport to **Reykjavik (Reykjavík)** is shown in Fig. 2.1. The airport is located on a basaltic lava flow. More specifically, the airport is situated on a smooth lava flow of the type referred to as **pahoehoe** lava. Basalt means that the molten rock, the **magma**, which generated the lava was very hot, about 1300 °C, and of comparatively low viscosity, that is, it flowed easily. When flowing on the surface, the lava itself had probably a temperature of around 1200 °C. The lava hosting the airport, however, lies outside the main active volcanic areas of Iceland (Fig. 2.1).

To clarify this: the parts of Iceland where the tectonic plates move apart and eruptions are possible are defined as volcanic zones. The main zones are mostly 20–80 km wide, covered with rocks formed in the past 800 thousand years, and indicated with a bright yellow colour in Fig. 2.2. Within these zones, however, volcanic eruptions and fracture formation (and earthquakes) are mostly confined to certain areas, named **volcanic systems** (Figs. 2.2, 2.3, and 2.4). The volcanic systems are those parts of the volcanic zones where eruptions have occurred in the past 10–11 thousand years. The volcanic systems thus mark the **main active parts** of the volcanic zones. Volcanic systems are supplied with magma from great depths; in Iceland from depths of 10–20 km or more (Fig. 2.4). In many volcanic systems, but not all, the most frequent eruptions occur in a certain part of the system. This part has a **shallow magma chamber**, commonly with a roof at 1–5 km below the surface, and develops a major volcano, which we refer to as a **central volcano**. The central volcanoes in the active volcanic systems in Iceland are indicated in Fig. 2.2. A typical central volcano, with an associated shallow magma chamber, is shown in the central volcanic system in Fig. 2.4 (cf. Figs. 4.10 and 11.3). Three of the volcanic systems on the Reykjanes Peninsula do not have central volcanoes; only the easternmost system has a central volcano, namely the volcano Hengill (Chap. 12). Most of the famous volcanoes in Iceland and elsewhere in the world are central volcanoes.

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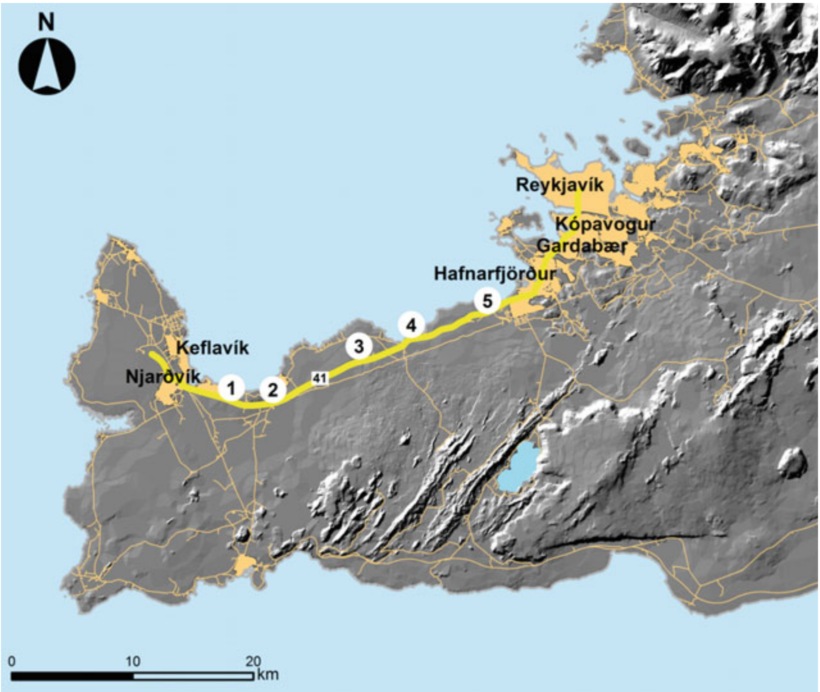

Fig. 2.1 Some geological features of interest can be seen from Road 41 on the way from the Keflavik Airport to Reykjavik. The numbers 1–5 indicate the approximate location along the road where these features are best seen

It is common to define a volcano or a volcanic system as **active**, that is, one that has a reasonable chance of erupting again sometime in the future, if it has erupted at least once during the past 10 thousand years. If no eruption has occurred in a volcano or a volcanic system for more than 10 thousand years, the probability that it will ever erupt again is low. We follow this definition and regard the volcanic systems as the main active parts of the volcanic zones. The other parts are regarded as essentially inactive. The areas outside the volcanic zones (pale yellow colour in Fig. 2.2) are regarded as inactive or extinct.

There are different ways of marking the outlines of the volcanic systems shown in Fig. 2.2. On the Reykjanes Peninsula there are four main volcanic systems. They are shown (more accurately drawn than in Fig. 2.2) in Fig. 2.3. The lava flow on which the Keflavik Airport is built is outside any of the active volcanic systems

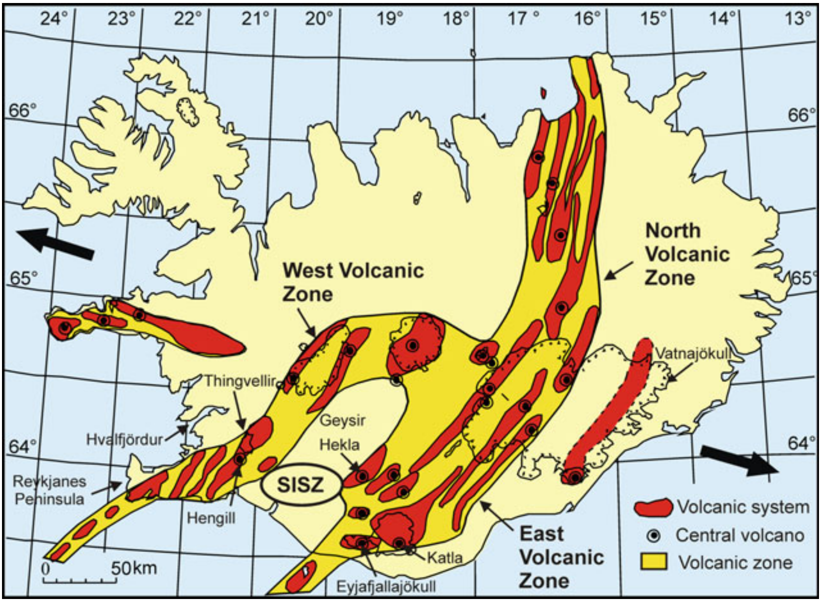

Fig. 2.2 Volcanic zones and systems of Iceland. The zones of possible volcanic eruptions are indicated in bright yellow colour. These are the West, East, and North Volcanic Zones, as well as the Snaefellsnes Volcanic Zone (Snæfellsnes Volcanic Zone), forming the central peninsula of West Iceland. The main active parts of the volcanic zones, however, are the volcanic systems, shown in red. These are mostly zones or swarms of volcanoes formed in single eruptions, such as volcanic fissures (crater rows), lava shields, and various types of single craters, as well as tectonic fractures (tension fractures and faults—discussed in Chap. 5). There are 28 volcanic systems shown here; their number and geometries vary somewhat depending on the criteria used to define them. All, however, have been active in the past 10–11 thousand years. Most volcanic systems, but not all (Fig. 2.4), develop central volcanoes (stratovolcanoes, composite volcanoes, collapse calderas) supplied with magma from shallow magma chambers. The central volcanoes are indicated by encircled dots. Some of the main central volcanoes discussed in the book are indicated (Hengill, Hekla, Eyjafjallajökull, Katla), as well as areas of interest such as the Reykjanes Peninsula, Hvalfjördur (Hvalfjörður), Thingvellir (Þingvellir), and Geysir. The main earthquake zone discussed in the book, the South Iceland Seismic Zone (SISZ) is also indicated. The thick, black arrows indicate how the west and east parts of Iceland are being pulled apart, that is, the regional direction plate-tectonic spreading

(Figs. 2.2 and 2.3). The lava flow is, in fact, well over 100 thousand years old, and may be much older. Volcanic eruptions are thus not expected at the location of the airport.

Driving along Road 41 towards Reykjavik, some of the geological features that can be seen from the car/bus are indicated by number in Fig. 2.1. In contrast to later excursions, we do not go to the structures seen on this drive. I presume you will either see them while driving or, if you stop, look at them from the main road (Road 41), that is, from a distance. However, even when you just look out of the window of a bus, it is worth mentioning some of the geological structures and features that you see (depending on visibility) on the way to Reykjavik.

The first remarkable feature is the comparatively young basaltic lava flow **Arnarseturshraun (1)**. Being mostly an **aa lava flow** (Fig. 2.5; see also Fig. 13.20a), it has rough surface with broken irregular blocks, and the same applies to the adjacent lava flow to the south, **Illahraun**. Both lava flows can be seen from Road 41, and Road 43 which leads to the town Grindavik (Chap. 13), passes through these lava flows. Illahraun is today very well known to millions of

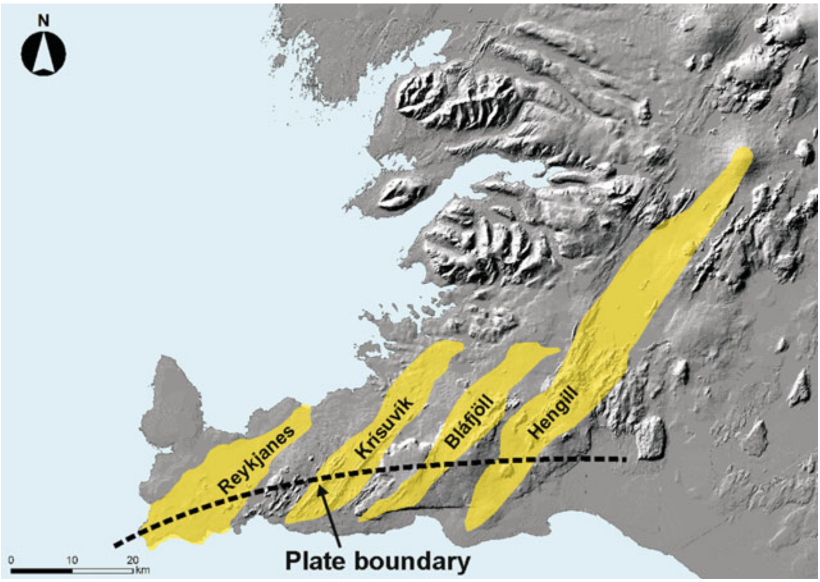

Fig. 2.3 Main volcanic systems on the Reykjanes Peninsula are Reykjanes, Krisuvik (Krísuvík), Blafjöll (Bláfjöll), and Hengill. The Vogar Fissure Swarm constitutes the northernmost part of the Reykjanes System, whereas the Thingvellir (Þingvellir) Graben (Chap. 5) constitutes the northern part of the Hengill system. The broken line shows the highly oblique plate boundary—where most of the earthquakes on the Peninsula occur

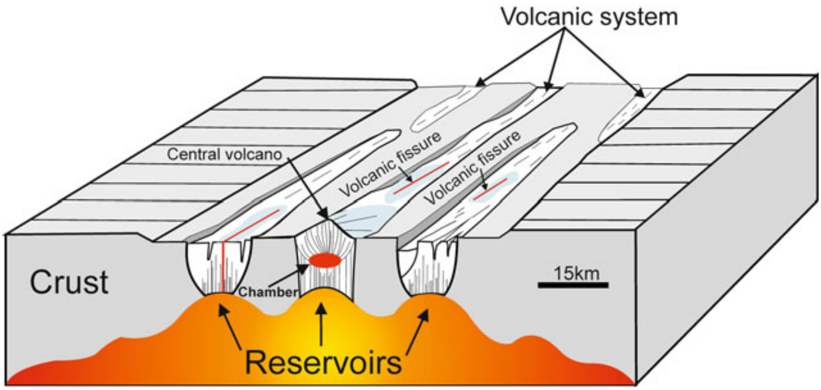

Fig. 2.4 Internal structure of volcanic systems. Only one of the systems, the one in the centre, has a shallow magma chamber in addition to a deep-seated reservoir. Because the central system has a shallow chamber which channels magma to a limited area on the surface, it also has a central volcano, shown here as a central volcano (stratovolcano, composite volcano) of a typical (cone) shape. In the other two systems the magma comes straight from the deep-seated reservoirs. At the surface, the volcanic systems are composed primarily of normal faults (see Figs. 2.5, 2.6, 4.5, 4.12, and 5.8 for clarification and examples), tension fractures (see e.g., Figs. 2.5, 2.6, 5.8, 5.11, 5.12, and 5.13 for clarification and examples) and volcanic fissures (see Figs. 12.21, 13.20e, 13.25, 13.27, and 13.32a for examples)

tourists who have visited the **Blue Lagoon (Bláa Lónið**; cf. Fig. 13.24; Chap. 13), whose geothermal waters derive from deep wells that pass through Illahraun. Both Arnarseturshraun (Fig. 2.5) and Illahraun are of similar age, formed in the period 1210–1240 CE, that is, somewhere around 800 years ago (CE means **Current Era**, which is also denoted by AD, Anno Domini. I normally drop CE/AD when giving the age of lava flows in this book. Thus, **CE is implied** unless stated otherwise.) These are among the youngest lava flows on the Reykjanes Peninsula. Surprisingly, given the size of the Reykjanes Peninsula and that it has at least four main volcanic systems (Figs. 2.2 and 2.3), there have been no known volcanic eruptions on the entire Peninsula since about 1340 CE, that is, for close to 680 years. Taking into account the general activity on the Peninsula, this is a long hiatus and eruptions are expected in the future.

The geological activity is partly reflected in numerous large earthquake **fractures (2)** in the lava flows on the Peninsula (Figs. 2.5 and 2.6). I discuss how these and other similar fractures form in detail in Chap. 5 about Thingvellir. Basically, all the fractures you see here are formed because Iceland is being pulled apart across the volcanic zones, and particularly across the volcanic systems (Figs. 2.2, 2.3 and 2.4). The pull-apart (or spreading) rate is mostly between one and two centimetres per year. This continuous movement stretches or strains the rock until it breaks and forms a fracture. When such a fracture forms, during the rupture of the rock, there is commonly an earthquake. Most earthquakes associated with the types of fractures you see here, however, are small. The fractures in Figs. 2.5 and 2.6 occur mostly in a pahoehoe lava flow forming a gently sloping, shield-shaped volcano named **Thrainsskjöldur (Þráinsskjöldur)**, which formed more than 10 thousand years ago and possibly as long as 14 thousand years ago. Volcanoes of this type are referred to as **lava shields** and are common in Iceland. They differ from the famous **shield volcanoes** on Hawaii, Galapagos, and many other islands

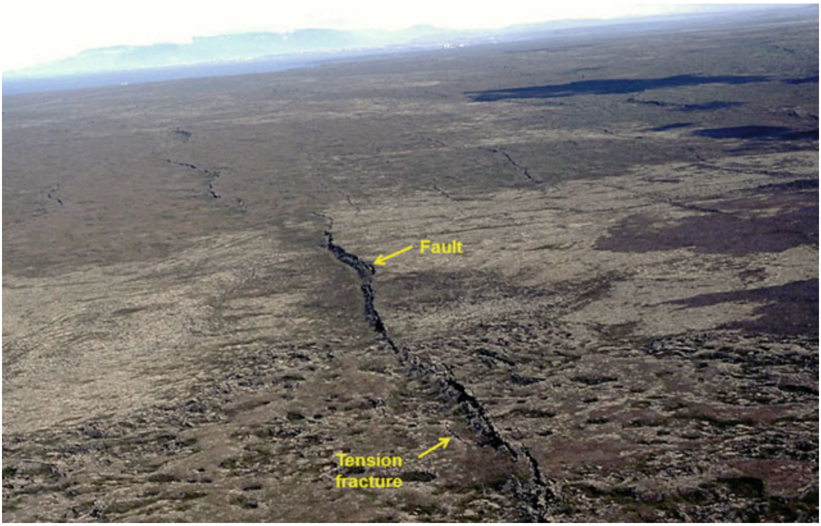

Fig. 2.5 Aerial view of fractures formed by plate-tectonic forces in pahoehoe (smooth-surface) basaltic lava flows of the Reykjanes Peninsula. View (facing or looking) northeast towards the capital region and the mountain Esja. The long fractures are normal faults, the shorter ones tension fractures (their formation is discussed in Chap. 5). The longest fractures reach lengths of many kilometres. These fractures, and those seen in Fig. 2.6, belong to the northwestern part of the Reykjanes Volcanic System (Fig. 2.3); more specifically to the so-called Vogar Fissure Swarm of that system—Vogar is a village close to the second stop in Fig. 2.1. Most of the lava seen here belongs to Thrainsskjöldur, but the closest lava flow, where the main fault in the centre is ending, is the margin of the Arnarseturshraun lava flow

Fig. 2.6 Aerial view of fractures in pahoehoe basaltic lava flows on the Reykjanes Peninsula. View southeast, the main fracture is a normal fault; the lava surface to the left of the fault (closer to you) has subsided by several metres relative to the lava surface to the right of the fault. Several other normal faults are seen here, but also tension fractures. These fractures are part of the Vogar Fissure Swarm of the Reykjanes Volcanic System (Fig. 2.3). See the caption to Fig. 2.5 for more details

in that the Icelandic shields are much smaller and formed in one or a few eruptions, whereas the large shield volcanoes form in numerous eruptions over tens or hundreds of thousands of years. Thus, the large shield volcanoes are central volcanoes whereas the lava shields are not.

In addition to the fractures and lava flows, there is one very noticeable mountain that can be seen from Road 41 so long as it is not pouring down rain. That mountain is **Keilir (3)**, a cone-shaped mountain that stands as a landmark in the surrounding lava field (Fig. 2.7). The mountain rises about 380 m above sea level and is seen from far away. It is made of basalt, but during an eruption in deep water. When hot magma meets the cold water, the resulting explosions change the magma into fine particles or grains. Each explosion forms one layer of particles or ash. As the layers pile up during the eruption, they form a mountain—hence Keilir. Clearly, the area where the mountain stands today is dry—so where did the water come from? Very likely from a lake in a thick glacier that covered Iceland, including the Reykjanes Peninsula, during the **Ice Age** (which consists of many glacial periods) tens of thousands of years ago. Thus, the process of formation of Keilir is similar to that which we observe in Iceland today when there is an eruption under an ice cap (Fig. 1.4). The ice sheet melted and disappeared from the Peninsula some 12–14 thousand years ago, so that Keilir must be older than that. Its age is presumably several tens of thousands of years. Keilir is therefore much older than the surrounding lava flows, all of which are younger than 14 thousand years, and some much younger, such as Arnarseturshraun (Fig. 2.5). The cone shape of Keilir (its Icelandic name means cone) is mainly because the eruption was

Fig. 2.7 The mountain Keilir, a cone-shaped hyaloclastite mountain and a landmark on the Reykjanes Peninsula. View south, the mountain was formed during an eruption in deep water, most likely meltwater during the Ice Age when an ice sheet covered the entire Reykjanes Peninsula

Fig. 2.8 Lava hillock known as tumulus. View south, tumuli are common in pahoehoe lava flows, particularly in the lava flows from lava shields, as seen here next to the road and, like this one, normally several metres in height (compare Fig. 13.39)

through a very short fissure that melted a close-to-circular cavity (in plan view) within the ice sheet.

Much closer to Road 41 you may see many dome-shaped structures or small hills, reaching several metres above the surrounding lava field. These hillocks are named **tumuli** (singular tumulus) and are common in pahoehoe lava flows (Fig. 2.8). They are commonly 2–10 m high and form when pahoehoe lava ponds in a small surface depression. Soon a solidified or frozen surface crust forms on the pond. If lava continues to flow into the pond, the frozen surface lifts or domes so as to make space for the incoming lava, thereby forming a hillock, namely a tumulus. Some of the tumuli seen close to Road 41 are among the largest in Iceland.

There are two comparatively young aa lava flows through which Road 41 passes before it reaches the '**Capital Region**' **(Höfuðborgarsvæðið)**, which includes Reykjavik (the capital) but also the six municipalities adjacent to Reykjavik, namely Hafnarfjördur, Gardabaer, Kopavogur, Mosfellsbaer, Seltjarnarnes, and Kjosahreppur (Hafnarfjörður, Garðabær, Kópavogur, Mosfellsbær, Seltjarnarnes, Kjósahreppur—the main towns are indicated in Fig. 2.1). These are the lava flows **Afstapahraun (4)** and **Kapelluhraun (5)**. The age of Aftstapahraun is probably around 2000 years, whereas that of Kapelluhraun (also called Nyjahraun, Nýjahraun, meaning New Lava) is about 1000 years. The Aluminium Factory in

Straumsvik, to the left of Road 41 just before entering the town of Hafnarfjördur stands on Kapelluhraun, and so does the westernmost part of the town itself.

The Capital Region itself is mostly on rocks as old as several hundred thousand years. These are primarily basaltic lava flows, referred to as ‘Grey Basalt’ and date from periods when Iceland was ice free during the Ice Age. The lava flows can be seen in many road cuts within Reykjavik. Because of the comparatively great age of these rocks, the area occupied by Reykjavik itself is well outside active volcanic systems (Figs. 2.2 and 2.3).

Nevertheless, there are comparatively young lava flows inside the area of the Capital Region. We already mentioned Kapelluhraun, on which part of the town of Hafnarfjördur is built, the rest of the town being on lava flows that are several thousand years old. One very narrow lava flow reaches the coast along the **Ellidaardalur (Elliðaárdalur)** valley in the eastern part of Reykjavik itself. The tiny front is a part of Leitarhraun lava flow, which issued about 5000 years ago (Chaps. 3 and 9).

All the comparatively young lava flows that reach into the Capital Region issue from volcanic fissures in volcanic systems on the Reykjanes Peninsula (Fig. 2.3). In the event of a new eruption in one of these systems resulting in a lava flow heading for the Capital Region (most likely in the systems of Krisuvik (**Krísuvík)** and Blafjöll (**Bláfjöll**)) modern technology allows the flow to be diverted away from that region. Diversion channels and levees (barriers) have been successfully used in Iceland (during the 1974 eruption of Heimaey, the main island of the Vestmannaeyjar archipelago, discussed in Chap. 14) and elsewhere to redirect the flow of lava away from the built environment.

3 Reykjavik

While Reykjavik itself (Fig. 3.1) is not geologically as spectacular as some areas you visit during the later excursions, the city has some remarkable geological features. I have indicated some of those I think of particular interest by number in Fig. 3.1, and describe them briefly below.

The first thing to notice is that Reykjavik is partly located inside an old volcano. This volcano, named the **Videy Volcano** (Viðeyjareldstöðin) or sometimes the Kjalarnes Volcano (Kjalarneseldstöðin), was active about 2.5 million years ago (Fig. 3.1). The volcano is thus totally extinct, with no chances of eruption, but provides much benefit for Reykjavik. The benefit is the **geothermal energy**. In particular, the extinct volcano provides the heat source of the hot water used for space heating of buildings (and to melt snow of some of the streets in the winter time) in a large part of Reykjavik. At great depths the rocks from the Videy Volcano are still comparatively hot—commonly reaching 75–100 °C, and sometimes more. Through drilling wells to these depths, the geothermal water can be used either directly (but at a lower temperature than above), or indirectly (the hot water is used to heat originally cold water) for space heating, swimming pools, melting of snow and ice, and related benefits.

Somewhere close to the centre of this old and extinct volcano is the southern part of the island of **Videy (1)** (Viðey). The island has intrusive (basaltic or mafic)

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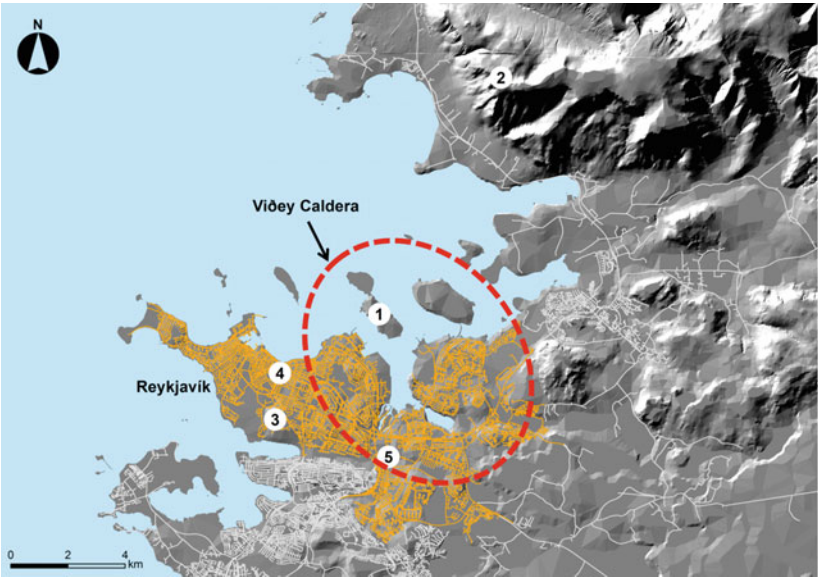

Fig. 3.1 Some places of geological interest in Reykjavik. The street network of Reykjavik itself is in orange; that of the remainder of the Capital Region in white. The geological sites in Reykjavik, or easily seen from Reykjavik, are the island of Videy 1, the western slopes of the mountain Esja 2, the grey-basalt lava hills of Öskjuhlid 3, and Skolavörduholt 4, as well as the valley of Ellidaardalur 5, where a lava flowed into Reykjavik some 5200 years ago. Also indicated, very schematically, is the caldera, the ring-fault, of the Videy (Viðey) Volcano. At its peak activity, the Videy Volcano would have covered a considerably larger area than its subsequently formed collapse caldera

rocks, that is, rocks formed when magma froze or solidified at considerable depth (here about 1 km) below the surface of the volcano (Figs. 3.2 and 3.3). The island is one of the main sights from the north shore of Reykjavik. In addition, there are ferry trips to Videy if you wish to explore its geology in detail.

The Videy Volcano also produced the rocks that constitute the western part of the mountain **Esja (2)** north of Reykjavik (Fig. 3.4). These were formed under different environmental conditions—because during much of the past 2.5 million

Fig. 3.2 Aerial view of the island of Videy as well as part of Reykjavik and the mountain Esja (see also Figs. 3.3 and 3.4). Videy is really two islands connected by a narrow strip of land or sandbar. The northwestern part is named Vesturey (West Island), whereas the main island in named Heimaey (Home Island—not to be confused with the main island of Vestmannaeyjar, which is also named Heimaey; Fig. 14.1). The total length is about 3000 m, the maximum width about 700 m, and the maximum elevation above sea level about 80 m. The island is close to the centre of the collapse caldera of the extinct Videy Volcano (Fig. 3.1). Most of the island is made of one lava flow of the Grey Basalt type (of similar age to the Grey Basalt lava flows elsewhere in Reykavik), but there are also hyaloclastites and intrusions, particularly sills (Fig. 3.3). The intrusions and the hyaloclastites were issued by the Videy Volcano, some 2.5 million years ago, and are thus much older than the Grey Basalt layer which constitutes the bulk of the island of Videy

years there were thick ice sheets, glaciers, covering the entire country. The eastern part of Esja formed much later, and was erupted from a different volcano, namely the **Stardalur Volcano** (Stardalseldstödin) which you see very well from the road to Thingvellir. We thus delay the discussion of that volcano until Chap. 4.

The main part of Reykjavik, however, does not stand on these old rocks but rather on rocks that formed much later. These rocks are referred to as '**Grey Basalt**' (grágrýti) and form most of the hills in Reykjavik, as well as many other

Fig. 3.3 Close-up aerial views of parts of the island of Videy, both of the larger island, Heimaey, which is close to the main harbour in Reykjavik (part of which is seen on the photographs). **a** The northeast part of Heimaey has large intrusions, part of which form the coast and small hills seen here, mostly sills of gabbro, similar to those of Stardalshnjukar (Figs. 4.6, 4.7 and 4.8; Chap. 4). **b** The south part of Heimaey is entirely made of a Grey Basalt lava flow, part of which forms the coast (notice the clear cooling or columnar joints) seen here

Fig. 3.4 Aerial view of the western slopes of the mountain Esja. View northeast, the top parts are composed of close-to horizontal basaltic lava flows. Much of this part of the mountain was generated some 2.5 million years ago, partly through eruptions from the Videy Volcano (Fig. 3.1). Esja is described in detail in Chap. 4. Also seen are part of Reykjavik and the island of Videy (Figs. 3.2 and 3.3)

parts on which the city is located. For example, the hill **Öskjuhlid (3) (Öskjuhlíð)** is made of these rocks (Figs. 3.5 and 3.6). Several large cylindrical hot-water containers stand on the top of the hill, with a half-spherical dome, named Perlan (the Pearl), on top of the cylinders. Similarly, the hill **Skolavörduholt (4)** (**Skólavörðuholt)**, the highest point in central Reykjavik (Fig. 3.6), is also made of grey basalt (Fig. 3.6) and is the site of the well-known church Hallgrimskirkja (Hallgrímskirkja). The grey basalts are almost all basaltic pahoehoe lava flows and many of the hills may have formed as individual lava shields, similar to, but much smaller than, Thrainsskjöldur, which you may have seen on the road from Keflavik (Chap. 2) and will see better examples of at Thingvellir (Chaps. 5 and 6). Some of the grey basalt lava flows, however, originated outside Reykjavik and flowed into

Fig. 3.5 Aerial view of the gently sloping hill of Öskjuhlid. On its top are large containers for geothermal water which, in turn, support the semi-spherical and rotating building Perlan. Öskjuhlid is primarily composed of thick Grey Basalt lava flows, which were formerly used as building materials for the older stone buildings in Reykavik, including the House of Parliament. View north, parts of the mountains (from left to right) Akrafjall, Skardsheidi (Skarðsheiði), both located north of the fjord Hvalfjördur (Chap. 11), and Esja are also seen

the area occupied by the city. Most of the lava flows are thought to be erupted between 100 and 200 thousand years ago. The top parts of the lava flows are seen in many road cuts in Reykjavik. But these show only the uppermost part of the total pile which is commonly tens of metres thick and, occasionally, about 100 m.

While there have been no volcanic eruptions in Reykjavik for about 100 thousand years—meaning that the Reykjavik area has ceased to be a part volcanically active part of Iceland (Figs. 2.2 and 2.3)—lava flows can flow towards and very occasionally into the city from the nearby volcanic systems on the Reykjanes Peninsula (Fig. 2.3). During the past 10 thousand years one lava flow

Fig. 3.6 Aerial view of the hill of Skolavörduholt, the highest point in central Reykjavik and the site of the impressive church Hallgrimskirkja. Like Öskjuhlid (seen here also), Skolavörduholt is primarily composed of thick Grey Basalt lava flows. View south, the Reykjavik Airport and parts of the southern towns of the Capital Region are also seen

has entered the Reykjavik area. This is **Ellidavogshraun (Elliðavogshraun) (5)** which covers part of the valley of Ellidaardalur (Elliðaárdalur) in the eastern part of Reykjavik (Fig. 3.7). The age of the lava flow, which is quite narrow in the valley, is about 5200 years and it is simply the northernmost part of a larger lava flow, named Leitarhraun, from one of the volcanoes on the Reykjanes Peninsula. In the same lava flow there are the rootless craters or pseudocraters **Raudholar (Rauðhólar)** which we discuss in Chap. 9.

Ellidavogshraun erupted from a volcano some 20 km southeast of Reykjavik, located in the Blafjöll Volcanic System (Fig. 2.3). If a similar lava flow were erupted in this or other volcanic systems on the Reykjanes Peninsula, all of which are far from Reykjavik, the flows would simply be stopped or diverted long before they reached Reykjavik. As mentioned earlier (Chap. 2) there are many ways to

Fig. 3.7 Aerial view of the easternmost part of Reykjavik, including the valley Ellidaardalur which hosts Ellidavogshraun, the only lava to have flowed into Reykjavik since the last glacial period, which ended 12–14 thousand years ago. This lava flow is about 5200 years old and issued from a volcanic fissure in the Blafjöll Volcanic System (Fig. 2.3), some 20 km east of Reykjavik

stop or divert a slow-moving lava flow. It is thus highly unlikely that Reykjavik could be in danger from lava flows of the type and size common on the Reykjanes Peninsula.

Reykjavik to Thingvellir (Þingvellir)

4

Once you start on your way from Reykjavik to Thingvellir, you are at last on the Golden Circle (Fig. 4.1). While the main sites of the Golden Circle are normally covered in a one-day driving, there are places on the Circle that are most certainly worth a longer stay—for example Thingvellir and its surroundings. In Fig. 4.1 I assume that you will cover the entire Circle in a single day of driving. I therefore mark the main geological sites, that is, the recommended main stops, in a single sequence, beginning at stop 1 and ending at stop 15. Because the descriptions of some of the sites are rather long, I divide the Circle into six chapters, starting with Chap. 4 and ending with Chap. 9. The excursion begins and ends in Reykjavik. However, some of the geological aspects of the Reykjavik area have already been described in Chap. 3. In this first chapter we focus on geological landmarks on the way from Reykjavik to Thingvellir.

4.1 Esja—Spreading Rate and Structure

The best place to make the **first stop (1)** is the parking place just as you turn from Road 1 (the Ring Road, the main highway) to Road 36, the road to Thingvellir. The parking place is on the right-hand side (south of) the road and marked by the number 1 in Fig. 4.2. From the parking place you can see many interesting geological structures and landscapes features. To the north is the beautifully impressive mountain Esja, rising to a maximum height of 914 m above sea level (Figs. 4.3 and 4.4). Esja, whose east-west length is some 20 km, making it one of the larger mountains in Iceland, is formed through two main processes. One is the piling up of lava flows and other volcanic rocks when the mountain area was part of the West Volcanic Zone (Figs. 2.2 and 2.3). The other process is erosion, primarily through the action of the enormous ice sheets active during the **Ice Age**

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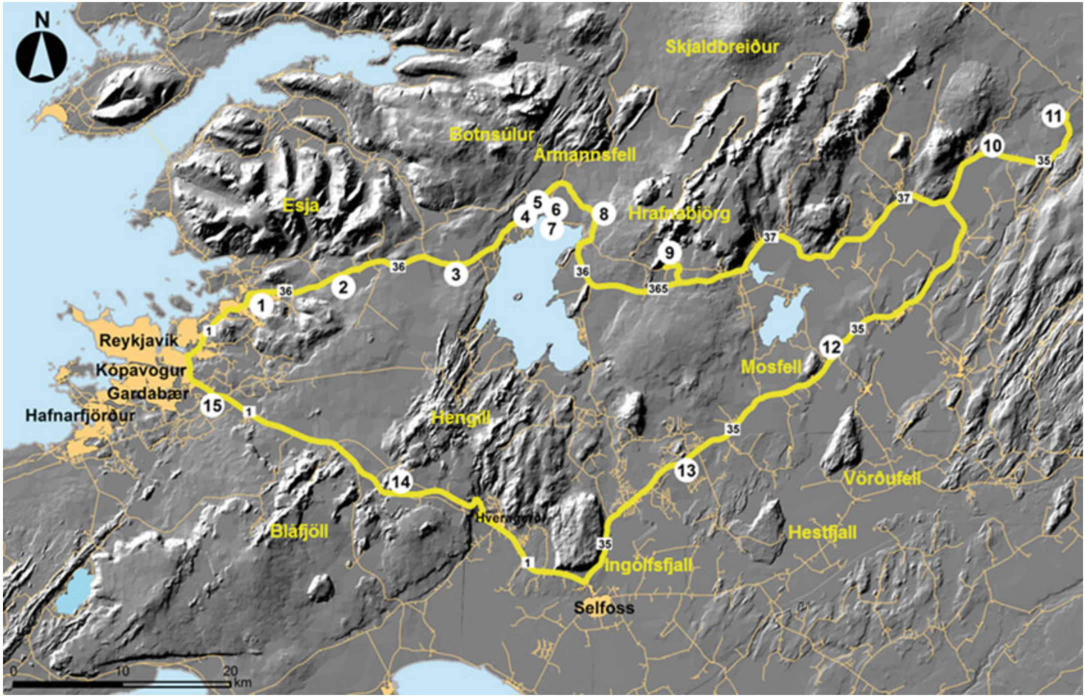


Fig. 4.1 The main stops on the Golden Circle, marked by the numbers 1–15. Also indicated are the main roads that constitute the Circle, clockwise roads 36, 365, 37, 35, and 1. Some of the main geological structures and landscape features visited at stops 1–15 are named here and in similar maps and also identified on the associated photographs presented and discussed at the stops. The main towns are named. For more details on the general topographical names, I refer you to the topographical and geological maps discussed in Chap. 1

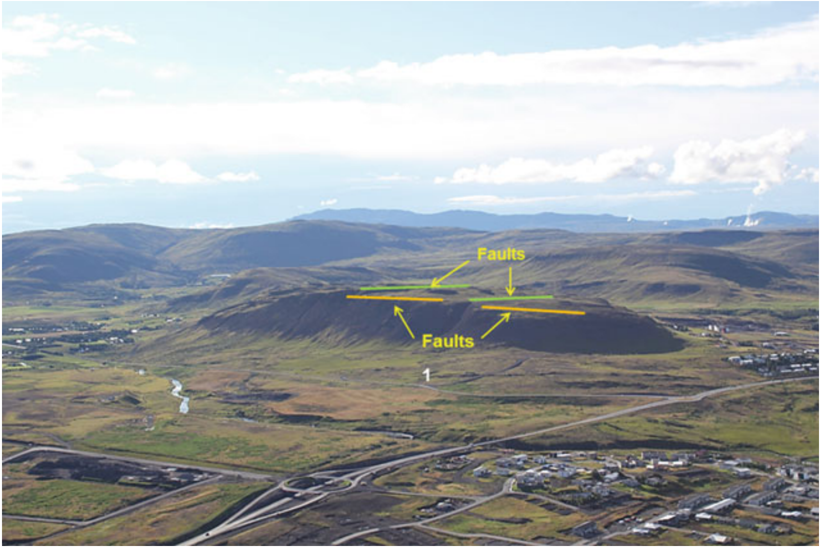
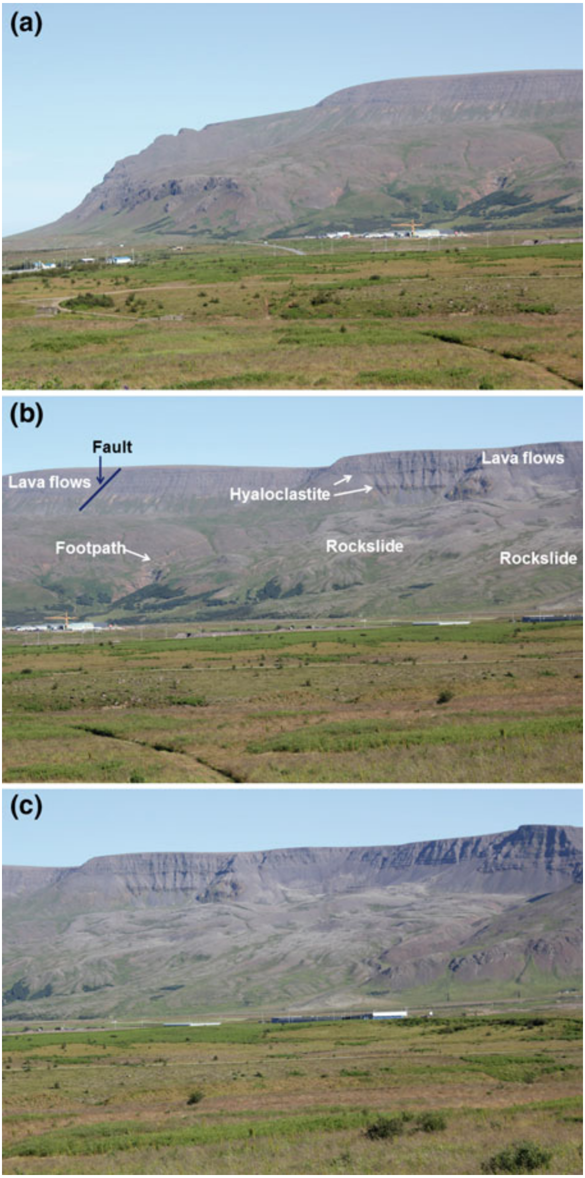


Fig. 4.2 Aerial view of the first stop (marked by 1, the number being at the parking place). View southeast, the tallest mountain in the distance is Hengill (Chap. 12) and the steam further to the south (right) is from the geothermal fields and drill holes of the power plant Hellisheidarvirkjun (both Hengill and Hellisheidarvirkjun are seen later today). From this location, I mainly describe the southern slopes of the mountain Esja (Figs. 4.3 and 4.4). But some additional features are seen on this aerial photograph. First, the mountain itself, Helgafell, is shaped by earthquake fractures, that is, faults, whose main directions, northeast (green) and east-northeast (orange) are indicated. The side of the mountain facing the camera follows an east-southeast trending earthquake fracture, that is, one or more faults (orange), whereas each small depression is the location of a northeast-trending fault. The faults are discussed further in subsequent chapters. Another noticeable feature is that all the layers in Helgafell slope to the east, as is discussed in the main text in connection with the lava flows in Esja (see also Fig. 4.6). The river seen is Kaldakvisl (Kaldakvísl)

(the past 2.8 million years). Thus, the primary reason that Esja is a mountain at all is the work of the Ice-Age glaciers—as, indeed, applies to almost all mountains in Iceland that have formed as **mountains** outside the active volcanic zones. The glaciers erode deep valleys, commonly along weak parts of the crust (often existing fractures), leaving behind the stronger parts which thereby stand as mountains above the valley floors.

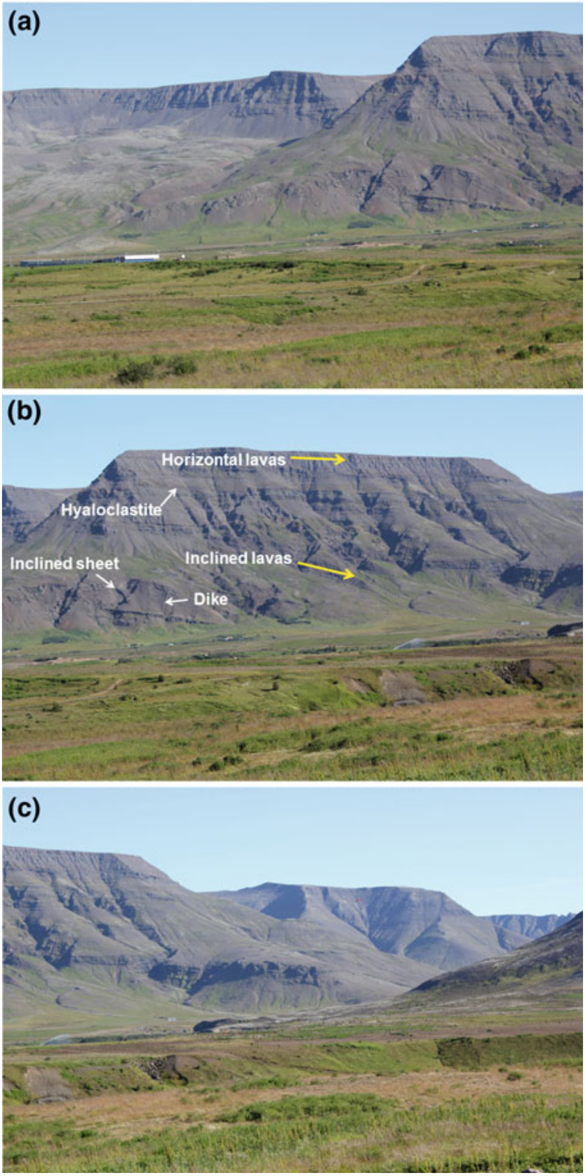
The pile of lava flows that constitutes the main part of Esja (Figs. 4.3 and 4.4) formed during the past three million years. As mentioned in Chap. 3, the Videy


(a)
(b)
Fault
Lava flows
Lava flows
Hyaloclastite
Footpath
Rockslide
Rockslide
(c)

◀ **Fig. 4.3** The western part of the south side of Esja. The mountain is mostly of basaltic lava flows (indicated). In addition to basaltic lava flows, there are thick and thin layers of hyaloclastite (basaltic breccia) and many intrusions, both large ones as well as small ones (dikes and small sills, Fig. 4.4b). **a** The westernmost part. **b** The conspicuous brown layers in between the lava flows are hyaloclastite layers. A large rockslide (landslide) is responsible for the smooth undulating surface in the central part of the picture (indicated). **c** The entire rockslide is seen here, forming a 'scar' in the mountain side

Volcano (Chap. 3) was active 2.8 million years ago, and much of the westernmost part of Esja (Fig. 4.3a) formed at that time (although partly generated by the somewhat older Hvalfjördur Volcano, Chap. 11). Now the eruptions that issued the lava flows that constitute Esja happened at the location of the Thingvellir area or graben (Fig. 4.5), of the Hengill Volcanic System (Figs. 2.2 and 2.3), some 30 km to the east (Figs. 2.3 and 4.1). So how come the lava flows are where they are now? The answer is what used to be called **continental drift**, and is now named **plate tectonics**. There is a horizontal movement, drift or spreading, across the volcanic zones (Fig. 4.5). At the time of the formation of most of Esja, the rate of movement or **spreading rate** to the west from the Thingvellir area was about 1 cm each year. It follows that in 2.8 million years (the oldest rocks in the western cliffs of Esja, Fig. 4.3) the lava pile of Esja would be carried some 28 thousand metres, or 28 km, to the west. And that is exactly where the 2.8 million year-old rocks are found today. They have drifted by some 28 km to the west from the Thingvellir Graben because of plate movement or spreading—a process we will explain in greater detail in the very next Chap. 5.

But the rocks have not only drifted laterally, they have also risen—that is, their elevation above sea level increased as the rocks moved to the west. Why? Because of the erosion during the Ice Age. As the glaciers scraped and carried away the weaker rocks the load or vertical force by the crust on the ductile mantle below became less so that the entire land rose. The elevation increase towards the west is partly along earthquake fractures, that is, normal faults, as indicated in Fig. 4.5. So the lava flows that you now see forming the top of Esja at as much as 914 m above sea level (Figs. 4.3 and 4.4) were formerly at the location of the Thingvellir Graben at only 100–150 m above sea level. They have thus risen by some 800 m in the top parts of Esja. The valley (Mosfellsdalur) closest to you in Figs. 4.3 and 4.4, part of which is also seen in Fig. 4.2, contains weaker rocks than Esja itself, rocks that were mostly eroded away by the glaciers. The rocks that constitute Esja were eroded perhaps to a depth of only a few hundred metres. That means that the top


(a)
(b)
Horizontal lavas
Hyaloclastite
Inclined sheet
Inclined lavas
Dike
(c)

◀ **Fig. 4.4** The central part of the south side of Esja. The lava flows are almost horizontal (subhorizontal) at the top but become gradually more inclined down to the east, that is, towards the West Volcanic Zone (and Thingvellir in particular) with increasing depth (decreasing elevation) in the mountain, as explained in Fig. 4.6. **a** The western part. **b** Inclined layers and dikes and inclined sheets are seen and indicated in the lower parts of the slope. This part of Esja is named Kistufell. It has a very even top part. **c** The inclination or tilting of the lava flows clearly increases downslope, that is, with increasing depth below the original surface

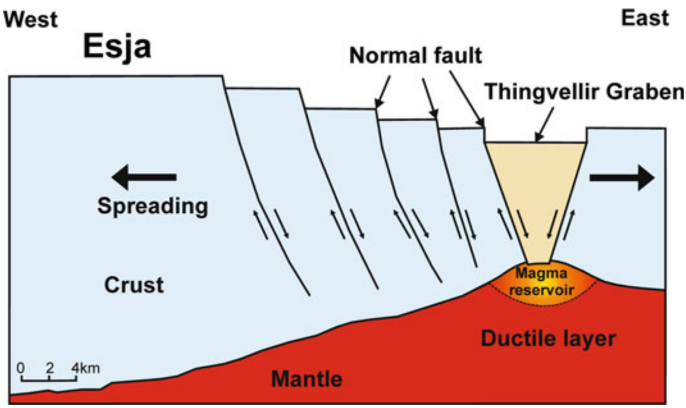


Fig. 4.5 The rocks, primarily lava flows, that constitute the mountain Esja were initially formed in the part of the West Volcanic Zone now occupied by the Thingvellir Graben. Esja itself is largely formed by lateral westward drift of some 28 km from this volcanic zone over the past 2.8 million years combined with deep glacial erosion and uplift. The top layers of Esja have risen by some 800 m as they drifted to the west, partly through normal faulting. The crust (or, strictly, the lithosphere) is floating on the magma or melt-rich ductile mantle. The crust/lithosphere increases in thickness with distance from the volcanic zone (here the Thingvellir Graben) because part of the magma in the mantle solidifies and becomes added to the lithosphere. At the same time, the glacier erosion wears away part of the surface through the generation of numerous valleys and lowlands, lessening the weight or vertical load on the ductile mantle. Both factors contribute—primarily through so-called isostasy—to the remaining parts of the crust rising (partly along faults) to form mountains such as Esja

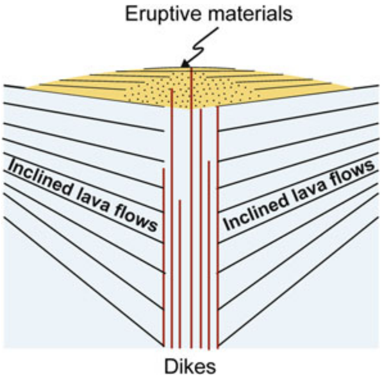


Fig. 4.6 The lava pile in Iceland is normally tilted towards the nearest segment of the volcanic zones. Thus, in East Iceland, the lava pile is tilted to the west (towards the North Volcanic Zone) and to the northwest (towards the East Volcanic Zone—see Fig. 2.2) whereas in West Iceland, such as in Helgafell and Esja (Figs. 4.2, 4.3 and 4.4), the pile is tilted to the southeast, that is, towards the West Volcanic Zone. The tilting or inclination of the lava flows increases with depth in the crust, that is, at less elevation above sea level, in the slopes of the mountains. The reason is that younger lavas continue to pile up at the surface of the active zone and their weight or load presses the lava flows beneath them

layers in Esja that we see today are a few hundred metres below the original top—so a few hundred metres of the lava pile is missing from the Esja area. This top was eroded and the debris carried by glaciers and rivers into the sea. By contrast, the deepest valleys around the mountain were eroded to depths of more than a kilometre, and the deepest fjords close to Esja, such as Hvalfjördur (Chap. 11), eroded to depths of perhaps 1300 m.

Let us now look at some of the interesting features of Esja. The first thing to notice in Fig. 4.3 is that the top of the mountain is almost flat. In fact so flat that small aircrafts have landed on the top. It is partly so flat because of erosion, and partly because the top-most layers have not become much tilted or inclined down to the east (towards the West Volcanic Zone and Thingvellir in particular). As I discuss in a moment, the reason is that the uppermost lava flows have been buried under relatively few flows (now eroded away) and thus not been subject to much vertical load. By contrast the lowermost lava flows have been subject to perhaps one kilometre load of lava flows on top of them, and thus been bent down, that is, tilted down towards the West Volcanic Zone (Fig. 4.6).

The top is made of almost horizontal basaltic lava flows, mostly greyish to bluish. The lava flows formed, one on the top of the other (Fig. 4.6), over a long period of time. Commonly the time between successive flows was many thousand years, occasionally tens of thousands of years. In the time periods between the eruption of each new lava flow, vegetation formed on the top on the last one before, similar to the vegetation you see later today on the young lava flows in the Thingvellir Graben and surrounding areas. Old layers of vegetation and soil commonly become red when buried under younger lava flows. There are also brownish layers in-between some of the lava flows. These are made of **hyaloclastite** or moberg (móberg), that is, basaltic ash, formed in explosive eruptions under glaciers (see Chaps. 2 and 6).

One remarkable feature, seen most clearly in Fig. 4.3c, is a huge pile of rocks that clearly has broken from the main part of the mountain, leaving a scar. Large slides of this type are named landslides or **rockslides**. Small rockslides are common in the mountains of Iceland, but rarely of this size. We do not know when it formed—and it need not have been a single slide—but the greatest instability of the mountain edges or slopes occurred when the last glaciers of the Ice Age melted away. Since the main melting of the ice sheet in this part of Iceland occurred some 12–13 thousand years ago, it is possible that the main rockslide occurred at that time.

In Fig. 4.4 (particularly 4.4c) it becomes clear that the lava flows, horizontal at the top of Esja, are no longer horizontal in the lower slopes of the mountain, but rather inclined or tilted (dipping is the term used in geology) down to the east. That is, the lava flows, and in fact the whole pile that constitutes Esja, is inclined or dipping towards the active volcanic zone, the West Volcanic Zone, at Thingvellir (Fig. 4.1). This is a universal feature in Iceland: the lava pile almost everywhere **tilts** or dips towards the closest part of the active volcanic zone. The reason for the tilting of the lava flows at deeper levels (less elevation above sea level) in the mountains is that younger lavas continue to pile up at the surface of the active zone and their weight or load presses the deeper lava flows (Fig. 4.6).

We also see rock layers, or sheets of rock, that are close to vertical (Fig. 4.4b). These are frozen (solidified) magma paths, named **dikes** (or dykes). We will see many of these close-ups in Chaps. 11 and 13), so that the main discussions about their formation will be in those chapters. Here, however, I can explain dikes briefly as follows. Magma is stored in large cavities, **magma chambers**, at various depths in the crust. These chambers are partly or totally filled with magma. When the pressure of the magma reaches the strength of the rock around the chamber (so-called tensile strength, namely the pressure or stress that rock being pulled apart can tolerate before it breaks or ruptures) the walls of the chamber—or, most commonly, its roof—rupture. As soon as the rupture occurs, a **magma-filled fracture** forms which may or may not reach the surface to erupt. When the hot magma eventually freezes or solidifies in the fracture, the resulting structure is called a **dike**. Most of the dikes we see here never reached the surface of Esja—they stopped on their vertical path to the surface—so that they did not erupt. The word dike is today not only used about the solidified rock in the fracture but also about the fracture when it was filled with fluid, hot magma, and moving (propagating) within the crust and the associated volcano.

The mountain south of the parking place, Helgafell (Fig. 4.2), is of an age similar to that of Esja, and the same applies to the mountains south of Road 36 along **Mosfellsdalur** (the Mosfell Valley, Figs. 4.1 and 4.2). These mountains are partly of hyaloclastite or moberg, and partly of lava flows similar to those seen in Esja. The lava flows all show the same tilting to the east, that is, towards the active West Volcanic Zone as do the lava flows in Esja. The main mountain north of Road 36 closest to the road, Mosfell, is also made primarily of hyaloclastite. This mountain, however, is much younger than the others in its vicinity, or 'only' about 150–200 thousand years old.

4.2 Magma Chamber and Collapse Caldera

After looking at these geological features, we now drive on to the **second stop (2)**, located in Fig. 4.1. This stop is chosen so as to allow you to see the top of an old, that is, a fossil (frozen, solidified) **magma chamber** represented by grey cliffs north of Road 36 (Figs. 4.7, 4.8, and 4.9). Above the grey cliffs are three yellowish peaks. These also tell a remarkable story, but let us begin with the grey cliffs. The cliffs form the top of a fossil magma chamber, namely a chamber that was active (filled with hot magma) and supplied magma to eruptions in a volcano some 1.8 million years ago. This volcano constitutes a part of Esja and was a major contributor to Esja's lava pile. The volcano is named after the valley **Stardalur**, the valley hosting the fossil magma chamber—the cliffs are named **Stardalshnjukar (Stardalshnjúkar)**—and is called the **Stardalur Volcano** (Stardalseldstöðin).

The cliffs are made of basaltic or mafic rock (Fig. 4.9). The crystals in the rock can be seen with the naked eye, but are not very large. This type of rock is commonly referred to as dolerite, but a better name for it is **microgabbro**. Many fossil magma chambers in Iceland (and elsewhere) are composed of micrograbbro. This applies primarily to **shallow chambers**, that is, magma chambers that, when active, were at the shallow depths of 500–2000 m (0.5–2 km) below the surface of the volcano to which they supplied magma. If a fossil magma chamber, a pluton as

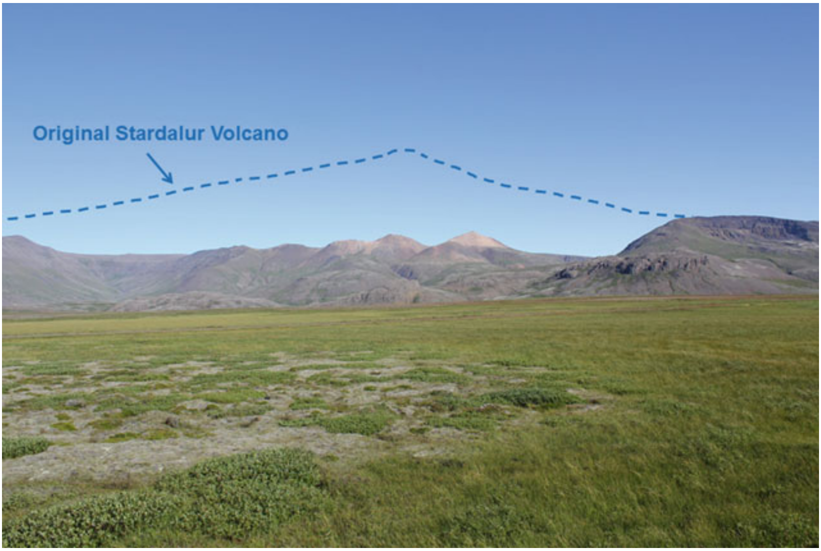


Fig. 4.7 The grey cliffs are Stardalshnjukar, a fossil coin-shaped shallow chamber that supplied magma to the extinct Stardalur Volcano (whose original surface is indicated schematically). View north, the yellow to pinkish peaks are Moskardshnjukar, also a part of the Stardalur Volcano. Stardalshnukar are located close to the centre of a collapse caldera (it cannot really been shown on the photograph and is inferred from geological mapping), about 6 km in diameter, whereas Moskardshnjukar are around 1 km north of the inferred north rim or edge of the caldera fault (the ring fault)

they are normally called, is made of the same type of rock, that is, basaltic (also called mafic), but has solidified more slowly and thus normally at a greater depth, say 2 km depth as is common in Iceland, then the rock is not called microgabbro but rather gabbro.

The fossil magma chamber had a roof or top only some 600–700 m below the surface of the active volcano. This is an unusually shallow depth. Most shallow chambers in Iceland and elsewhere have roofs at depths of **1–5 km** below the surface. Fossil chambers of depths as great as 2 km are seen in other parts of Iceland, particularly in Southeast Iceland. The reason that fossil chambers at greater depths than 2 km are not seen in Iceland is simply because that is the deepest glacier erosion in Iceland. In future ice periods, the glacier erosion will continue and reach greater crustal depths, showing deeper fossil magma chambers. Many active (fluid) chambers at depths from 1 to 5 km have been detected through earthquake studies and surface-deformation studies of active volcanoes in Iceland. For example, I discuss the magma chambers of the volcanoes Eyjafjallajökull and Katla, both in South Iceland, in Chap. 14. Remarkably, one such active fluid magma chamber at a depth of about 2 km below the surface was **drilled into** in the volcano Krafla in North Iceland during drilling for geothermal energy in the year 2009. No eruption happened, or anything dramatic, because the water used to cool the drill-core simply rapidly cooled the magma and changed it into glassy material.

Given the typical depths of magma chambers, it is thus possible that Stardalshnjukar is a part of a deeper and larger chamber either the top of that chamber or a separate 'satellite' chamber of the deeper one. Whichever is the case, the Stardalshnjukar chamber has roughly the shape of a flat (oblate) ellipsoid, or to use a more familiar object, the shape of a typical coin (one euro or one pound sterling, for example) oriented horizontally. Such shapes of magma bodies, where the large dimensions are horizontal, are named **sills**. Their vertical dimension or

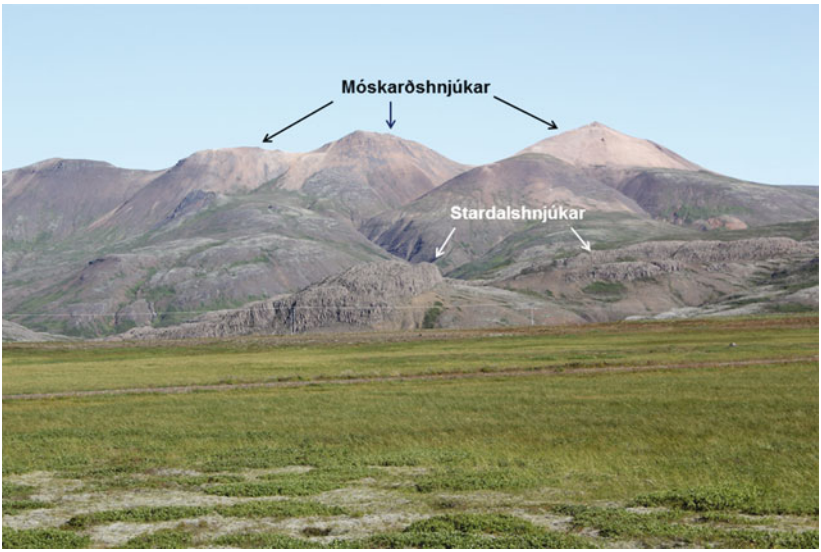


Fig. 4.8 Closer view of a part of Stardalshnjukar as well as the three main peaks that constitute Moskardshnjukar, both indicated. For close-ups see Figs. 4.9 and 4.11

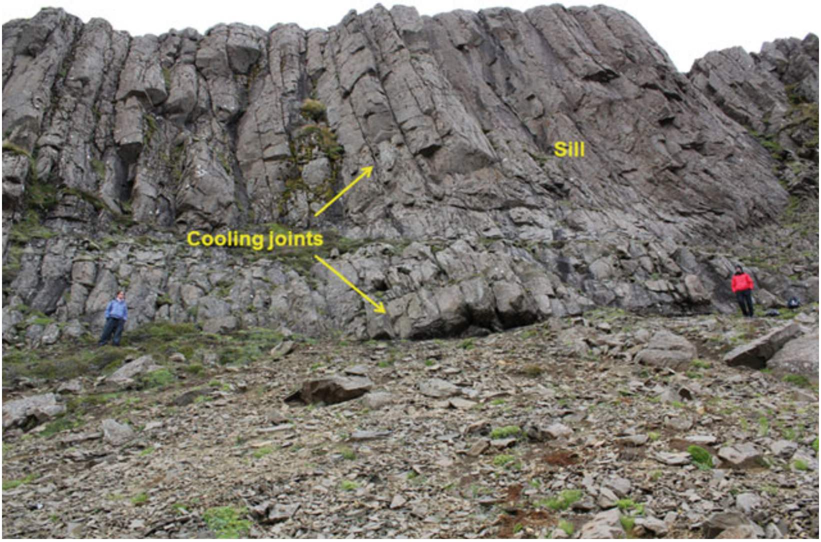


Fig. 4.9 Close-up of a sheet of rock, a sill, that constitutes part of the Stardalshnukar fossil shallow magma chamber. The vertical (and the less noticeable horizontal) fractures are cooling or columnar joints. These form when the magma cools down, that is, freezes or solidifies (discussed in greater detail in Chaps. 11 and 14). Then the magma body shrinks or contracts because solid rock has less volume than fluid magma. Vertical columnar joints, as here, characterise horizontal intrusions, named sills, of which this cliff is a part. The persons provide a scale. Notice: this cliff or outcrop can by reached by car but only if the car is a four-wheel drive (a jeep). The road is a very poor gravel road and is not recommended. Alternatively, the outcrop can be reached on foot from the farm Stardalur

height is thus much less than their lateral dimension; for the fossil magma chamber this means that its thickness is much less than its lateral width or diameter.

Why is the shape important? It is important because the shape of a magma chamber partly determines how often it is likely to rupture, with the potential of supplying magma to a volcanic eruption. If two chambers are located in the same volcanic zone, the one with a flat roof normally ruptures and erupts less often than one with a curved or domed roof, such as a semi-spherical roof. By contrast, a flat-roofed chamber, like the one in Stardalur, is more likely than one with a domed roof to trigger collapse of its volcano, resulting in the formation of a **collapse caldera**.

And that is exactly what happened in the Stardalur Volcano (Fig. 4.10). Just like the Videy Volcano (Chap. 3), the Stardalur Volcano eventually collapsed to form a ring-shaped depression, a collapse caldera. While the ring-fracture along which the subsidence took place is not clear in the landscape today (Figs. 4.7 and 4.8), it is thought to have been roughly circular with a diameter of some 6–7 km. This is a dimension very similar to that of many calderas within the active volcanic zones of Iceland, the most famous of which is the Askja Caldera in central Iceland.

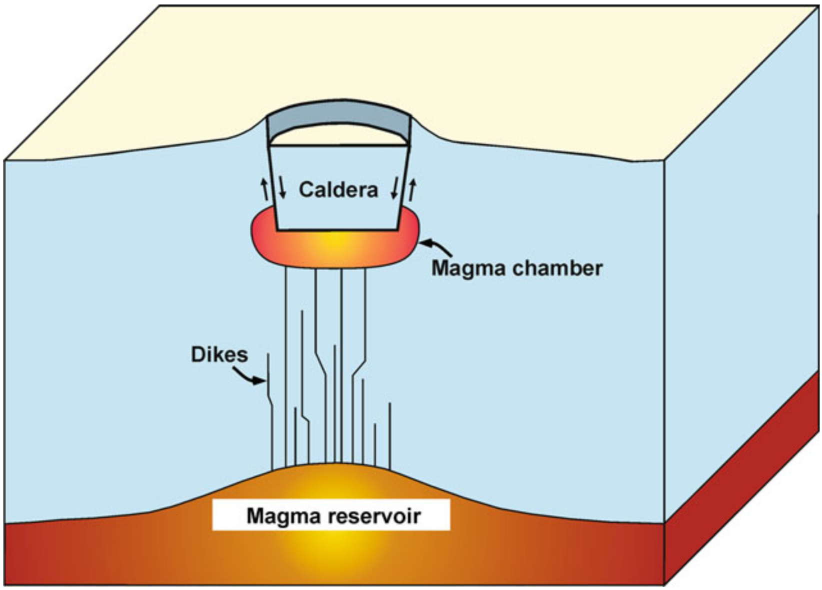


Fig. 4.10 Stardalur Volcano eventually collapsed to form a caldera. Collapse calderas are a common stage in the development of central volcanoes (composite volcanoes, stratovolcanoes) in Iceland and elsewhere. During caldera collapse, a part of the volcano subsides along a ring-fault into a shallow magma chamber. This general illustration indicates the relationship between a collapse caldera and its main shallow magma chamber. Stardalshnukar form only a small part of the shallow chamber. Moskarsdshnukar were injected outside the caldera, most likely from a different compartment of the shallow chamber. The ultimate source of the entire volcanic and intrusive activity, including the formation of the shallow chamber(s), was the deep-seated large reservoir at the depth of 15–20 km

Fig. 4.11 Aerial view of the easternmost rhyolite peak of Moskardshnjukar (Figs. 4.7 and 4.8). View northwest, close to the peak is one of the flat tops of Esja. In the distance, we see part of the fjord Hvalfjördur and the mountains Skardsheidi, to the east or right (also seen in Fig. 3.5) and Hafnarfjall, to the west or left

So from the road you are seeing straight into the core of the volcano; a volcano that is now extinct but was very much active 1.8 million years ago. The structure of the volcano when it was active and had already developed a collapse caldera is illustrated in Fig. 4.10. Thus, only some 15 km west of the active West Volcanic Zone at Thingvellir, where eruptions are still occurring, we can see into the heart of a volcano. The Stardalur Volcano may have been similar to the active **Hengill Volcano**, south of Lake Thingvallavatn (Chap. 12). Later in our excursions we see even deeper into the eroded and inactive or extinct rift zone and associated volcanoes, namely when we drive north to the fjord of Hvalfjördur (Chap. 11).

Before we leave this site, it is worth taking another look at the three yellowish to pinkish peaks (Figs. 4.7, 4.8, and 4.11). These belong to the same volcano, the Stardalur Volcano, and are named **Moskardshnjukar (Móskarðshnjúkar)**. They rise to about 807 m above sea level and are made of rocks with high silica content, that his, rhyolite. The molten rock, the magma, that formed them was at much lower temperature (around 800–900 °C), and with a much greater viscosity, than the basaltic magma that formed the visible part of the shallow magma chamber, the cliffs of Stardalshnjukar, whose basaltic magma was presumably at a temperature of 1100–1300 °C. Moskardshnjukar are not lava flows, but rather are **intrusions**, that is, magma that froze or solidified at some depth below the surface of the volcano. In this sense they are similar to the Stardalshjnukar cliffs—that is, both are intrusions, so that the rocks we see were not emplaced at the Earth's surface—but Moskardshnjukar were very close to the surface. Probably, the peaks we see now were only 200–300 m below the surface (Fig. 4.7).

And why do Moskardshnjukar form these peaks? Like the rest of Esja the peaks are primarily formed by erosions of the Ice Age glaciers. Given that, then the reason they stand above their immediate surroundings—the reason they really are peaks—is that their rock is less easily eroded, is more resistant to erosion, and in that sense stronger, than the rocks in their vicinity.

The rhyolitic magma that formed the peaks of Moskardshnjukar (Figs. 4.7, 4.8, and 4.11) did come from the shallow magma chamber of the Stardalur Volcano. Most likely it did not come from the part presently occupied by the Stardalshnjukar, but rather from another part or compartment of the magma chamber. This compartment may have been just below the peaks, rather than 3–4 km to the south from them, as are the Stardalshnjukar. Several fractures filled with rhyolite, rhyolite dikes, occur close to the peaks, and one or more of them most likely supplied the magma that forms the present peaks.

4.3 Subsiding Crustal Segment—A Graben

We now drive on along Road 36. Before you reach Thingvellir, it is a good idea to make the **third stop (3)** at a parking place for an overview of the landscape of the Thingvellir area, including the beautiful **Lake Thingvallavatn (Þingvallavatn)** and its islands. In addition, this stop provides a good overall view of the mountains around the lake. Another reason for stopping here is to look at the depression or valley **Gagnheidi (Gagnheiði)** between the mountains **Botnsulur (Botnsúlur)** and **Armannsfell (Ármannsfell)**. This valley (Figs. 4.12 and 4.13) is analogous to the valley of Thingvellir itself. The only difference is that the valley of Gagnheidi is located in rocks considerably older, and partly of a different type, than those at Thingvellir. Both valleys are formed by subsidence of the land between two major

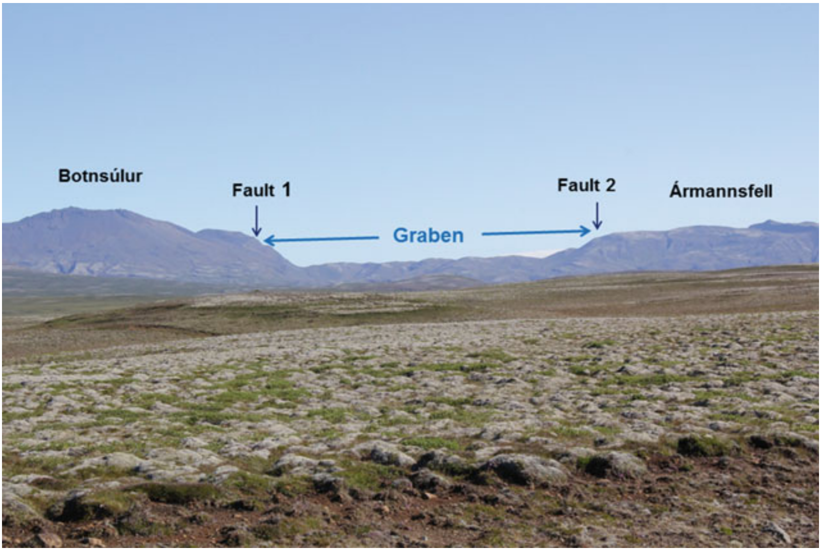


Fig. 4.12 Gagnheidi, the depression between the mountains Botnsulur and Armannsfell, is a graben, that is, a trough formed by the subsidence of land in-between two normal faults (explained in Fig. 4.14). The main boundary faults, Fault 1 and Fault 2, are indicated. Fault 1 (Sulnaberg, Súlnaberg) is one of the largest normal faults in Iceland, with a vertical displacement or subsidence of about 400 m

earthquake fractures, namely normal faults. The technical term for valleys formed in this way is the German word **graben**, whose original meaning is ditch or trench. We have used this word above when discussing the Thingvellir Graben (Fig. 4.5). In geology, graben means a subsided crustal block located between parallel faults, referred to as normal faults (Fig. 4.14).

Grabens characterise areas where the surface is being extended, where the tectonic plates are being separated or pulled apart, as in Iceland in general and in Thingvellir in particular (Fig. 4.5). We will see many grabens on the Golden Circle and in its surroundings, and I explain the forces that drive the pull in the chapter on Thingvellir (Chap. 5). We do not know how old the graben in Gagnheidi is, but we know that it is located in rocks that are tens of thousands to several hundred thousand years old. The earthquake fractures, the normal faults, that form the boundary of the graben (Figs. 4.12 and 4.14) may be much younger, although at least 10–12 thousand years old. One main thing, however, is the sizes of the normal faults. The subsidence or vertical displacement (Figs. 4.13 and 4.14) along the largest one, forming the west boundary of the Gagnheidi graben, is as much as 400 m (Figs. 4.12 and 4.13). As we shall see, this is about 10-times the maximum subsidence along the faults of the Thingvellir Graben itself.

It may thus be that the Gagnheidi graben is 10-times older than the Thingvellir graben. We know that the Thingvellir graben is about 10 thousand years old, which would then make the Gagnheidi graben about 100 thousand years old. But the faults in Gagnheidi may be much younger than this—we simply do not know. Whatever the exact age of the faults, the main thing is that they, like almost all the fractures in this part of Iceland, are formed through the plates being pulled apart. Which brings us to the sizes of the forces that move the plates—the

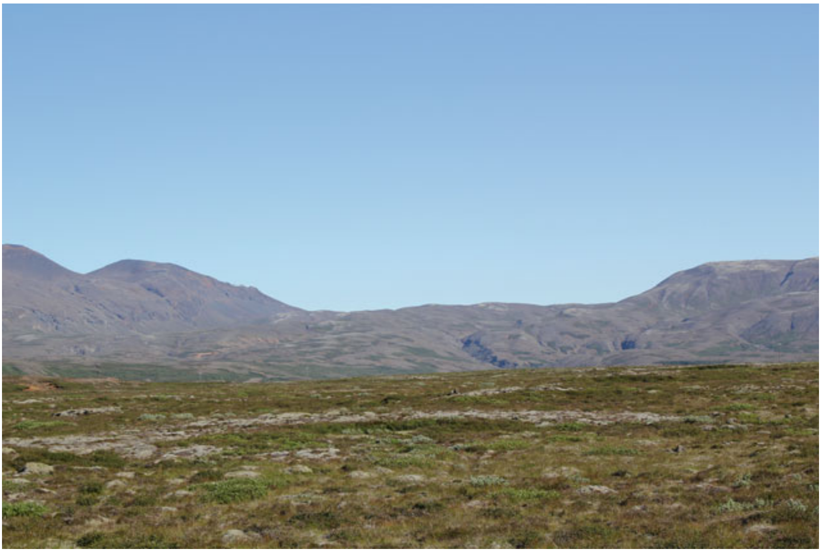


Fig. 4.13 Close-up of the Gagnheidi Graben. The fault on the left-hand side (the western boundary fault) has a displacement (explained in Fig. 4.14) of about 400 m. This is, as mentioned, one of the largest vertical displacements on any normal fault in Iceland

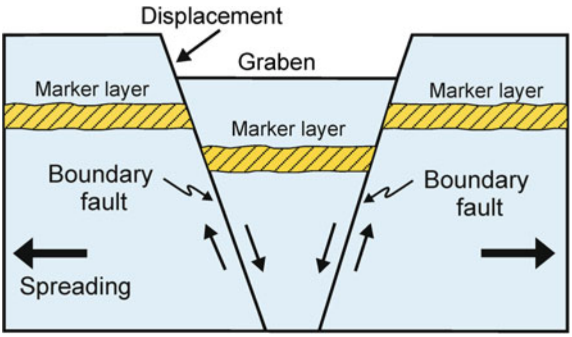


Fig. 4.14 Graben forms through vertical movement or displacement along two boundary faults. Because the boundary faults are inclined or dip towards each other, the subsiding block is wedge shaped, that is, becomes narrower with depth. Here the boundary faults have the same inclination or dip to the surface, but commonly they become vertical close to the surface, particularly if the surface is a thick pahoehoe lava flow such as at Thingvellir (Figs. 4.5 and 5.18). The vertical displacement is partly attributable to absolute subsidence of the graben floor, and partly to absolute rise of the graben flanks, particularly where there is magma or ductile material below the graben. Marker layer is any easily identified rock layer that can be used to measure the displacement along the boundary faults (see Figs. 11.21 and 11.22 for a clear marker layer in a graben)

plate-tectonic forces. And for understanding and analysing these, there are few places on Earth that are as appropriate and educating as Thingvellir itself. So it is time to turn to one of the geological wonders on Earth—Thingvellir.

5 Thingvellir (Þingvellir)

Thingvellir is perhaps the best place on this planet to understand the process of rupturing of the crust in response to the pulling forces of plate movements. You will be driving to, and most likely walking inside, the most spectacular example of the effects of the enormous plate-tectonic forces tearing the crust apart. While it is easy to see the open fractures on the ground—and you will see the large ones while walking in the Thingvellir National Park—it is perhaps easier to explain the processes and forces by looking at the area and some of the sites we visit from aerial photographs (Fig. 5.1; see also Fig. 4.1). I therefore include many aerial photographs in this chapter.

Thingvellir constitutes a graben that forms a part of the West Volcanic Zone (Fig. 2.2). More specifically, the Thingvellir Graben is located in the northern part of the Hengill Volcanic System (Figs. 2.3 and 5.2). Although the area is geologically a wonderland, and all the sites are spectacular, it is worth mentioning that great care is needed while walking among the fractures. There are, as we shall see, numerous small fractures adjacent to the larger ones, and many of the fracture walls are unstable. So my strong recommendation is **never ever go to the edge of a large fracture**.

5.1 Almannagja (Almannagjá)

The **fourth stop (4)** on the Circle is normally at the entrance to the largest fracture of the Thingvellir area, **Almannagja (Almannagjá)**, which means the fracture or fissure for or belonging to the general public. It certainly does so today—and you are likely to see many people walking the path along Almannagja. At this stop the classic view is the one in Fig. 5.3 (cf. Fig. 6.1). For comparison, Figs. 5.1 and 5.2 show a larger part of Almannagja from the air. More detailed aspects of this part of

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GeoGuide, DOI 10.1007/978-3-319-55152-4_5

Fig. 5.1 Aerial photograph showing the location of the four main sites visited at Thingvellir. The numbering refers to the stops indicated in Fig. 4.1. View southwest, the fourth stop (4) is at the entrance to the main fracture in Thingvellir, namely Almannagja. The fifth stop (5) is along the path down Almannagja where the west wall is very high and clear for observations of flow units and related aspects. The sixth stop (6) is at Lögberg, the site for parliamentary meetings when Iceland's parliament was located at Thingvellir, but also geologically an interesting place. The seventh (7) stop has two sites. The first is the popular water-filled fracture Peningagja and its extension Nikulasargja. Both of these, however, are just segments of the main fracture, referred to as Flosagja, which is the second site for the seventh stop. The river Öxara (Öxará) flows to the southwest along part of Almannagja, from the waterfall Öxararfoss (Öxarárfoss)

Almannagja are in Figs. 5.4, 5.5, 5.6, 5.7 and 5.8. In particular, Fig. 5.8 indicates some of the main geological structures associated with Almannagja. The main geological points regarding Almannagja may be summarised as follows (with references to Figs. 5.1, 5.2, 5.3, 5.4, 5.5, 5.6, 5.7 and 5.8):

- The fracture is formed through two main processes: opening and subsidence (vertical displacement). The maximum opening is just over 60 m; the maximum subsidence or vertical displacement is about 40 m (Fig. 5.9). Both processes relate to the plate-tectonic forces that tear the crust apart (discussed further in Sect. 5.3).

- The opening of 60 m by this single fracture gives a spreading rate (Fig. 4.5; Chap. 4) of about 0.6 cm per year. How do we know this? Simply by considering that the fracture is located in a lava flow that is about 10 thousand years old. So if the fracture opens by 60 m in 10,000 years, then we have 60/10,000 or 0.006 m or 0.6 cm per year. When the openings of all the fractures along a line or profile (or section) across the Thingvellir Graben are added up, we obtain 100 m, so that the spreading rate in the past 10,000 years is, on average, about 1 cm per year. This result is generally in good agreement with the spreading rate at Thingvellir measured by other means (such as by satellites) during the past decades.

Fig. 5.2 The fractures of Thingvellir, including Almannagja, are a part of the Hengill Volcanic System, whose central (main) volcano, Hengill, is seen here south of Lake Thingvallavatn. View southwest, this close-up aerial photograph of the southwestern part of Almannagja is taken about a decade later than the one in Fig. 5.1, and the hotel (with a red roof close to the fourth stop) in Fig. 5.1 is no longer seen in Fig. 5.2 (it burned down in 2009). Stops 4, 5 and 6 in Fig. 5.1 can be seen closer here (although not located again). The elevation difference between the western (right) and eastern (left) wall of the normal fault Almannagja is about 40 m, which is the subsidence across the fault in the past 10 thousand years. The last major subsidence was during earthquakes in 1789 when the northern shore of the lake, part of which is seen here, subsided by as much as 2.5 m. The maximum opening or aperture of the fault is about 60 m, close to Lögberg (the sixth stop in Fig. 5.1)

Fig. 5.3 Photograph of Almannagja from the fourth stop in Fig. 5.1. View northeast, the surface of the eastern (right) fault wall is inclined by about 11° to the east, whereas the surface of the western (right) wall is horizontal (see Figs. 5.8 and 5.9 for the geometric details). View northeast, the mountain Armannsfell is seen at the end of Almannagja, as well as part of the lava shield Skjarldbreidur (see also Fig. 6.4 for Armannsfell and Fig. 6.6 for Skjaldbreidur)

- The elevation difference between the top of the western wall to the lowest ground of the eastern wall, is about 40 m (Fig. 5.9). It follows that during the past 10,000 years the average rate of vertical displacement, primarily subsidence, across Almannagja has been about 0.4 cm per year. We see therefore that the rate of vertical displacement is about half the rate of opening or spreading in the Thingvellir area.
- While the plate movements are continuous, opening and vertical displacement across fractures such as Almannagja occur in discrete events. During such events, the eastern (lower) fracture wall of Almannagja suddenly subsides relative to the western (higher) wall (Figs. 4.14 and 5.9). Such abrupt displacements normally give rise to earthquakes. The last major subsidence, by close to 1 m at Almannagja, took place during earthquakes in 1789. The earthquakes lasted many days, during which part of the land on the north shore of the lake subsided beneath the water. In the centre of the Thingvelllir Valley

or Graben, the subsidence may have been greater, or as much as 2.5 m. As a result of this subsidence, the Parliament of Iceland was moved from Thingvellir to the capital, Reykjavik.

- All large fractures such as Almannagja—a large normal fault—are formed of smaller parts or segments (Sect. 5.3). As the tearing apart of the crust continues, that is, the spreading continues, the parts or segments of the fracture link together. But the original segments and the linkage between them are normally marked by offsets (Fig. 5.8; cf. Sect. 5.3). When you walk down the road or path inside Almannagja towards the fifth stop, you start your walk at the south end of one of the main segments of Almannagja. And at that lateral end, the fracture does not reach great depth and is made of pure opening—a tension fracture—so that there is no subsidence (Figs. 5.2, 5.4 and 5.8).

Fig. 5.4 The 'entrance' to Almannagja. This part can be seen on the aerial photographs (Figs. 5.2 and 5.8) as being the end of one of the segments of Almannagja. Where the segments end, as here, they are pure tension fractures. That is, the walls on either side of the fracture are at the same elevation. Tension fractures are best seen in Figs. 5.11, 5.12 and 5.15

Fig. 5.5 The lava flow that constitutes the walls of Almannagja is a pahoehoe lava flow. Such flows are basaltic and composed of numerous flow units, commonly with vertical cooling or columnar joints, and can reach thicknesses of several hundred metres, as does the present lava flow. For a vertical section through a thick pahoehoe lava flow, much older than the Thingvellir flow, see Figs. 11.19, 11.20 and 11.21

Walking down the path along Almannagja, we should make the **fifth stop (5)** so as to take a look at the fracture walls (Figs. 5.5 and 5.6). We see that the walls are made of many layers, each one 0.5–2 m thick. All the layers belong to the same lava flow, which at Thingvellir has a thickness of several hundred metres. In the walls we see only the uppermost twenty metres or so (the maximum height of the western wall is about 28 m). The lava flow is about 10 thousand years old and filled a valley, namely the graben that already existed at the time.

The flow is a thick pahoehoe flow of the type very common in the shield volcanoes of Hawaii and other basaltic edifices. Such flows are composed of numerous thin layers of the kind we see in the walls of Almannagja. The layers are called **flow units** (Figs. 5.5 and 5.6). There may be many tens, sometimes hundreds, of flow units in a single thick pahoehoe flow (in Chap. 11 we see a vertical section thorough a thick pahoehoe lava flow). As mentioned in Chap. 2, pahoehoe lavas are made of magma that is very hot (around 1300 °C). When erupted, the magma forms a flowing lava with a temperature of about 1200 °C at the surface (the lava, at the surface, is about one hundred degrees cooler than the magma in the magma chamber). Pahoehoe lava of this type flows very easily, that is, it has a comparatively low viscosity or, more specifically, viscosity similar to that of tomato ketchup or mustard. Each flow unit normally comes from an underground tunnel or tube, a **lava tube**. When the tubes drain at the end of the eruption, they form **caves**, some of which may reach many kilometres in length.

When looking very closely at the flow units—which is perhaps best to do as you enter Almannagja where the walls are still low and little danger of rock falls (Fig. 5.4)—you see a lot of cavities in them. Most of the cavities are either circular or somewhat elliptical in shape, and with a common diameter between half and one centimetre. The cavities (named **vesicles** by geologists) are initially gas-filled swellings or bladders within the lava. When the gas escapes out of the hot but solidifying lava and into the air, a cavity is left. The shape of the cavity is an

Fig. 5.6 Close-up of some of the flow units and cooling or columnar joints seen in Fig. 5.5. Here we see three main flow units (and part of the fourth one in the top left corner of the photograph). The cooling or columnar joints are best developed in the topmost flow unit. Vertical columnar joints are typical for horizontal flow units and lava flows in general, but are normally much better developed (more beautiful) in intrusions (see Figs. 11.6, 11.8 and 11.10 for horizontal columnar joints in a dike, and Figs. 14.30, 14.32 and 14.33 for vertical columnar joints in a sill)

Fig. 5.7 The surface of the eastern wall (left) of Almannagja is tilted by about 11° to the east. The tilting is most likely because of friction along the fault at depth (illustrated in Fig. 5.9). The tilting of the eastern wall is along the greater part of Almannagja (Figs. 5.1 and 5.8)

indication of the viscosity of the lava—and thereby the temperature of the lava. If the cavity cross-section is circular or somewhat elliptical, as in the walls of Almannagja, the lava flow had a high temperature and low viscosity. If the cross-section of the cavity is highly elongated or angular, the lava flow had a comparatively high viscosity and lower temperature. The lowest temperatures of basaltic lavas are around 1050 °C. These are **aa** lava flows and easily ten times more viscous than the lava flow seen in the walls of Almannagja. (I provide more discussion on vesicles in basaltic rocks, with close-up photographs, in Chap. 13.)

The **sixth stop (6)** is at the site of Lögberg, which was the main site for the parliament meetings while the parliament still met at Thingvellir (Figs. 5.1, 5.2 and 5.8). The reason for the choose of this site soon after the settlement of Iceland is partly that the western wall of Almannagja is ideal for projecting the speakers voice. One thing that is particularly clear at this site is that the surface of the eastern wall of Almannagja is inclined or tilted down to the east by about 11°. This is clearly seen on the ground (Fig. 5.7) and also from the air (Fig. 5.8). We could also see the sloping eastern wall from the first stop (Fig. 5.3) but perhaps less clearly. By contrast, the surface of the western wall is perfectly horizontal (Figs. 5.2 and 5.8).

So why is the eastern wall tilted while the western wall is not? The primary reason is friction between the walls at a certain depth (Fig. 5.9). Almannagja and other fractures in the volcanic zones of Iceland are open or gaping only to shallow depths. At depths of several tens of metres the walls are closed, that is, in contact with each other. Friction is here a measure of the resistance to relative movement of the closed fracture walls of Almannagja. The part of the eastern wall away from the contact with the western wall can subside through bending of the rocks (Fig. 5.9). At the contact between the walls, however, there is so much friction that

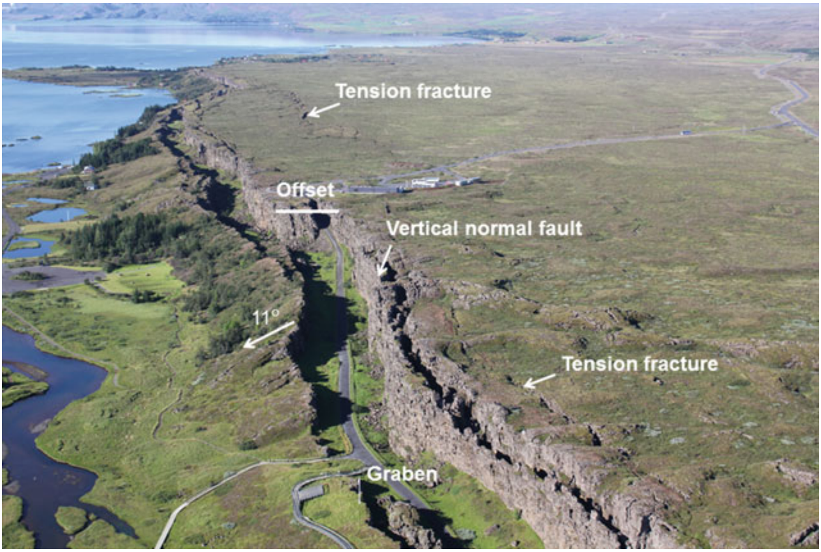

Fig. 5.8 Aerial photograph showing some of the main structures associated with Almannagja. Around Almannagja itself there are many smaller structures. These include small tension fractures (discussed in detail in connection with Figs. 5.11, 5.12 and 5.15) and the inclined eastern fault wall (its surface is inclined by 11° to the east, as shown here). By contrast the surface of the western fault wall is horizontal. Where you enter and start your walk down Almannagja, one segment or part of Almannagja is ending laterally (as a tension fracture). Then there is an east-west offset (indicated) and a new segment takes over and continues to the southwest. While Almannagja is a gaping or open normal fault, its opening is so large (more than 60 m in places) that it resembles a narrow graben (indicated). Compare the vertical section in Fig. 5.9

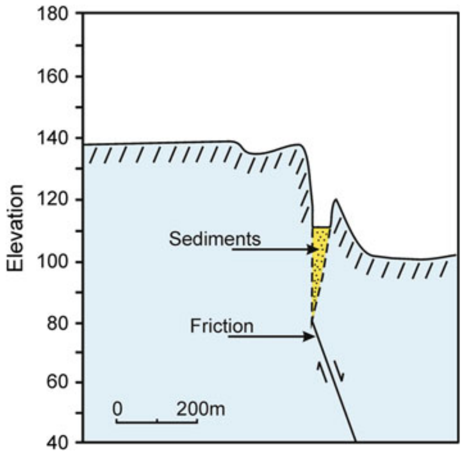

Fig. 5.9 Vertical section through Almannagja (roughly the central part seen in Fig. 5.2). The surface of the western (left) fault wall is at close to 140 m above sea level, whereas the lowest surface on the eastern (right) wall is at about 100 m above sea level. The maximum vertical displacement ('subsidence') across Almannagja is thus about 40 m. The thickness of the 'sediments', which include fractured rocks from the fault walls and gravel, is unknown. Where the fault walls come together at depth, the fault changes from being vertical (originally a tension fracture) to an inclined fault where the eastern wall has moved down relative to the western wall. At this location, the friction between the walls is presumably the reason for the tilting or sloping of the eastern fault wall. The vertical scale is exaggerated about 8-times relative to the horizontal scale

movement can only happen when large forces or stresses make it possible for slip to occur. And when slip occurs, there is an earthquake. The slip along the contact between the walls thus normally lags behind the general subsidence of the Thingvellir Graben, hence the bending or tilting of the eastern wall.

That abrupt slips and earthquakes are comparatively rare on Almannagja is known from recording of earthquakes, and is also indicated by Fig. 5.10. Here we see a stone which is poorly connected to the rest of the wall. In fact, the stone has been in exactly the same position for at least tens of years. During a moderate or strong earthquake the stone is almost certain to fall. But while the entire Thingvellir area is moving or spreading at the rate of close to a centimetre per year, no earthquakes of these sizes have occurred on Almannagja for many decades.

5.2 Peningagja and Flosagja (Peningagjá and Flosagjá)

We now move on to the **seventh stop (7)**. This stop is split in two parts, as explained below, and indicated in Figs. 4.1 and 5.1. It is easy to walk to this stop along the path from stop (6). Part of the path is seen in Figs. 5.1 and 5.2. Alternatively, you may choose to walk back to your car up (south) along Almannagja and then drive to the seventh stop. The seventh stop is one of the most popular in Thingvellir and is at a water-filled fracture named **Peningagja (Peningagjá)**. The name means 'Money Fissure', the money in this case being coins thrown by tourists into the fissure, a tradition established in the early twentieth century. Peningagja is a part of a larger fissure whose name is **Nikulasargja** (**Nikulásargjá**) which, in turn, is a part or segment of a larger fissure whose name is

Fig. 5.10 View west, the uppermost part of the western fault wall of Almannagja at stop 6 (Fig. 5.1). The little stone has been in this position for at least decades, indicating that despite gradual slow crustal movements at Thingvellir, as measured by geodetic instruments, no moderate to strong earthquakes have occurred on Almannagja for many decades and probably not since 1789

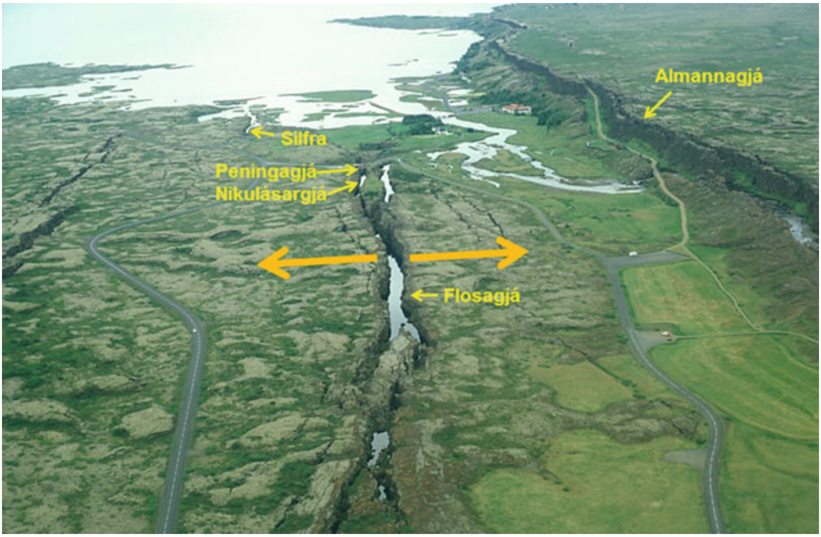

Fig. 5.11 Aerial view of the tension fractures Peningagja (Peningagjá), Nikulasargja (Nikulásargjá), Flosagja (Flosagjá), and Silfra, as well as the normal fault Almanngja (Almannagjá). View southwest, the plate-tectonic forces, indicated schematically by orange arrows at Flosagja, tear the crust apart, rupture it, and form tension fractures and normal faults. Close to the road, at the east (left) margin of the photographs are normal faults

Flosagja (**Flosagjá**). The latter, Flosagja, is the original name of the entire fissure, and Peningagja and Nikularargja are just among its southernmost segments or parts (Fig. 5.11).

Peningagja and Nikulasargja are very impressive structures (Fig. 5.12). But in some ways the main fracture, Flosagja, is the most spectacular of them all (Fig. 5.13). Peningagja/Nikulasargja and Flosagja here count as the **seventh stop (7)** in two parts. Flosagja (Fig. 5.13) can be reached through walking from Peningagja/Nikularsargja. Alternatively, if you drive into the Thingvellir Graben along Road 36 and then take Road 52 to the south to the parking place close to the waterfall **Öxararfoss** (**Öxarárfoss**), seen in Fig. 5.3, and walk from there to the fracture. In Sect. 5.3 I explain how the fractures form, but first let us have a look at the water in the fractures.

The water is very clean and clear—it is a perfect example of high-quality **groundwater**. In fact, the lava fields of Thingvellir and its surroundings are among the largest groundwater aquifer systems in Iceland. The water originally comes from precipitation, either directly from the rain and snow that falls on the area or, more indirectly, from melting of the glaciers in the north—in the highlands of Iceland—particularly the Langjökull ice cap. The groundwater migrates from the highlands surrounding the Thingvellir Graben through the lava flows which act as a sieve or filter for cleaning the water. When the water reaches the fractures it migrates into them and finally into **Lake Thingvallavatn** (Figs. 5.1, 5.11 and 5.14). About 90% of all the water in the lake comes from groundwater springs, the rest being from surface water (rivers). And since much of the water comes from far away, from Langjökull and the surrounding highlands, it takes many years—even tens of years—for the water to migrate the distance of about 50 km from the ice cap to the lake.

The surface elevation of the lake varies somewhat, but is at about 100 m above sea level on average (Fig. 5.14) and is at the same elevation as the water level in the fractures (Figs. 5.11, 5.12 and 5.13). This level is also the general elevation of the surface of the groundwater in the vicinity of the lake, so that the surface of the lake and the surface of water in the fractures close to the lake are at the same elevation (Figs. 5.1 and 5.11), named the **water table**. The lake itself exists because the valley or graben it occupies reaches below the water table—a common

Fig. 5.12 Peningagja, the water-filled fissure with numerous coins at its bottom, is a part of a tension fracture named Nikulasargja, most of which is seen here

Fig. 5.13 Peningagja and Nikulasargja (Fig. 5.11) are a part of a larger tension fracture named Flosagja, part of which is seen here. View northwest, the maximum opening or aperture ('width') of the fracture is about 15 m. Tension fractures form by pure opening, where the forces or stresses causing the opening are directly related to the plate-tectonic forces. Flosagja, as well as Nikulasargja and Peningagja, are seen from the air in Fig. 5.11

reason for the formation of lakes everywhere in the world. In fact, the lake is as deep as 114 m, so that its deepest parts reach below sea level (are at 14 m below sea level, to be accurate). The average or mean depth of the lake, however, is only 34 m. The lake covers an area of about 84 km^2, making it the largest natural lake in Iceland (Fig. 5.14).

The fractures (Figs. 5.11 and 5.12) were not in any way formed by the pressure of the water. All the fractures are formed directly by plate-tectonic forces or stresses (Fig. 5.11) and are just conduits for the groundwater. The groundwater moves or circulates very slowly through the rock before it meets the fractures that conduct the water into the lake. Groundwater has close to the same temperature throughout the year, a temperature which is similar to the average annual temperature in the area. In the fractures the water temperature is mostly 3–4 °C (Figs. 5.12 and 5.13). The water is thus very cold, yet warm enough so that it does not freeze. Thus, even in mid-winter, the water in the fractures does not become covered with ice.

The temperature of the lake (Fig. 5.14), however, changes over the year. It is lowest in the winter months and highest in the summer months. In the winter months of January to March the average temperature is less than 1 °C, but 9–10 °C in July and August. The average temperature of the surface water of the lake itself is somewhat higher than that of the water in the fractures. For the decades before the turn of the century, that is, before the year 2000, the water temperature in the lake was between 4 and 5 °C. In the present century, that is, after the year 2000, however, the temperature has so far been above 5 °C. This is, at least partly, related to the general warming in Iceland (and elsewhere) which has been most noticeable in the past two decades or so. Ice does form on the whole lake during the winter but

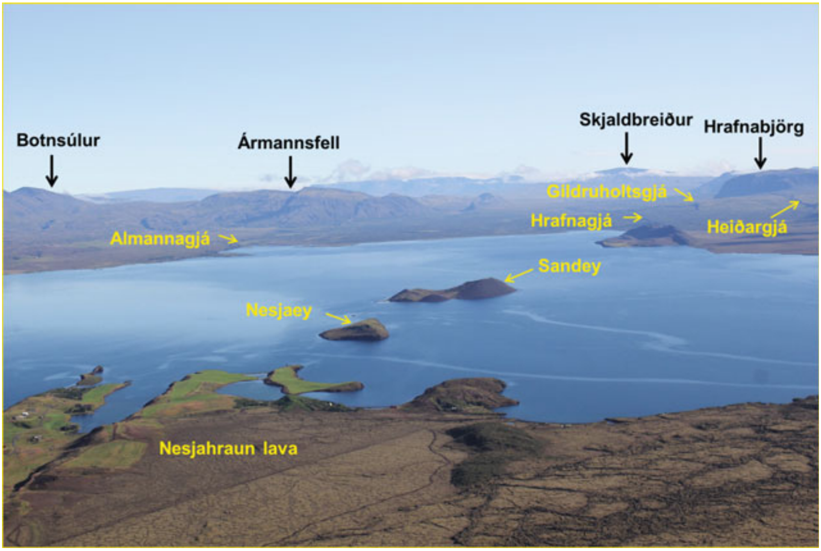

Fig. 5.14 The greater part of Lake Thingvallavatn. This aerial photograph shows the lake from the southwest. Most of the water in the lake originates in groundwater springs. The surface elevation of the lake is about 100 m above sea level, whereas its deepest part reaches a depth of 114 m, so that it extends below sea level. Lake Thingvallavatn, with an area of 84 km^2, is the largest natural lake in Iceland. For discussion of the mountains north of the lake (Armannsfell, Skjaldbreidur, and Hrafnabjörg) see Chap. 6 and for Botnsulur see Chap. 4. The largest faults at Thingvellir are indicated (Almannagja, Hrafnagja, Gildruholtsgja, and Heidargja) and discussed in Chaps. 5 and 6. The lava flow Nesjahraun 2-thousand year-old lava flow Nesjahraun on the south shore of the lake and the island Sandey formed about 2-thousand years ago and are discussed in Chap. 12

the number of days with ice cover has been declining considerably in the past two decades. In fact, in some of the years during the past decade there were no days when the entire lake surface was frozen.

5.3 How Do the Fractures Form?

Coming back to water-filled fractures (Figs. 5.11 and 5.12), how do they form? The general answer is that they form when the tectonic plates on either side of the Thingvellir Valley are being separated or pulled apart, resulting in spreading (Fig. 4.5). As we discussed above, Iceland is being pulled apart across the volcanic rift zones. In Thingvellir the rate of pulling apart, or spreading, is on average over thousands of years, about 1 cm per year. Far away from the volcanic zones (Fig. 2.2), in particular Thingvellir itself, the spreading is continuous, but its effects as regards fracture formation within the Thingvellir Graben is episodic. This means that centuries may pass between major **rifting events** with fracture formation or widening in Thingvellir. Recall that the last main rifting event in Thingvellir was in 1789, so more than two centuries ago.

So why does the rifting or rupture occur in separate events? Why is it not continuous like the spreading or plate movements themselves? The answer to both questions is that the plate-tectonic forces have to build up stress in the crust that is high enough to rupture the crustal rocks, to break the rocks. Gradually, as the forces move the plates apart, the Thingvellir Graben is stretched and its rocks become subject to higher and higher stress. As you know from tearing a sheet of paper, an existing rupture—a 'fracture'—makes it easier to tear the paper asunder. That is because the stress becomes raised or magnified at the rupture ends or tips, and these then propagate to the edges of the paper during the tearing. Similarly, the existing fractures (Figs. 5.11, 5.12 and 5.13) raise or concentrate the plate-tectonic stress at their lateral ends or tips, so that a particular fracture is most likely to lengthen, become longer, when the stress is high enough for a rifting event to take place. Thus, during rifting events, existing fractures become larger, that is, become longer and also deeper (Fig. 5.11). As rifting events continue, small offset fractures propagate and link up into larger fractures (Figs. 5.15 and 5.16).

Flosagja (and Nikulasargja and Peningagja) are clearly different from Almannagja in that the fracture walls in Flosagja on either side of the fracture are at the same elevation (Figs. 5.11, 5.12 and 5.13). By contrast, the eastern wall of Almannagja has subsided by as much as 40 m relative to the western wall (Figs. 5.1, 5.2, 5.7, 5.8 and 5.9). In geological terms, Almannagja is a **fault**, and more specifically a **normal fault**, whereas Flosagja (and Nikulasargja and

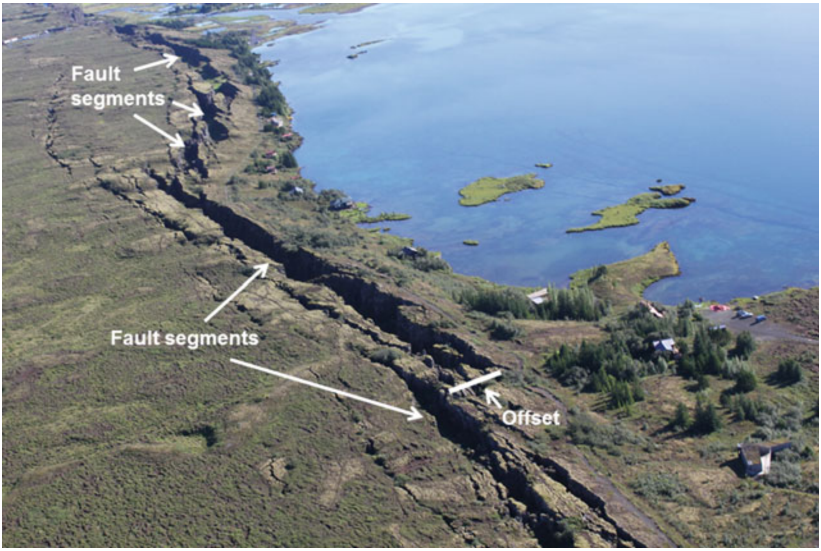

Fig. 5.15 The southwestern part of Almannagja is highly segmented, that is, divided into many smaller fractures. This is partly because the old fault beneath the surface lava flow and which controls where the surface fractures occur is no longer perpendicular to the main plate-tectonic force (as shown in Fig. 5.14). On a local scale, as here, the direction of the plate-tectonic force or spreading vector fluctuates ('wobbles' somewhat) so that a fracture that was initially oriented at right angle to the spreading vector or force may not be so for a while. The length of the offset indicates how much the fault shifts laterally when passing from one segment to another

Peningagja) is a **tension fracture**. In a fault, much of the movement of the rock on either side of the fracture is parallel with the plane of the fracture, either up or down (vertical) the fault plane, or sideways (horizontal). Most **earthquakes** are related to sudden movements of the walls or slip on faults, and all large earthquakes are generated by such movements. On a tension fracture, by contrast, the movement is simple opening, pulling the fracture walls apart (Fig. 5.11). There is thus no fracture-parallel movement during tension-fracture formation, and therefore no friction between the fracture walls (Fig. 5.9). For tension fracture opening, the earthquakes that occur, if any, are normally small.

So how much stress must build up before we will have a new rifting event in Thingvellir? That is easy to calculate and turns out to be about 3 million pascals (3 mega-pascals). Now this may sound as something very great. However, the unit pascal (Pa), which measures stress or pressure as force over area—force per unit area (newtons per square metre)—is tiny. One pascal is equal to the fluid pressure of a film or layer of water that is about 0.1 or one-tenth of a millimetre thick. At the bottom of a 2-m deep swimming pool the pressure due to the water is about 20 thousand pascal. At the bottom of Lake Thingvallavatn, at 114 m, the pressure due to the water is about million pascal. Thus, the stress needed to form Flosagja and other tension fractures at Thingvellir (and in general in rift zones and at ocean ridges worldwide) is of the same magnitude as the pressure at the depth of about 300 m in a lake or the sea. Alternatively, it is of the same magnitude as the pressure or vertical compressive stress (due to the weight of the rocks) at the depth of 120–130 m in the Thingvellir lava flow, the one seen in Almannagja (Fig. 5.5). I say the same magnitude because pressure or compressive stress seeks to compress an object, whereas tension or tensile stress (which may be of the same magnitude as the compressive stress or pressure, but with an opposite sign), responsible for the fracture formation, seeks to expand or extend the object—here the rocks at the

Fig. 5.16 Close to its southernmost end, just as it enters into Lake Thingvallavatn, Almannagja changes into a set of tension fractures. View southwest, this set is seen here. The opening of the fracture to the right (west) of the white car is 12 m. The step-like oblique arrangement of the fractures seen here is known in geology by the French term én echelon

surface of Thingvellir. As regards sign, in geology the sign of tensile stress is normally minus (−) and that of compressive stress plus (+), whereas in physics and engineering the sign convention is exactly the opposite.

The tensile stress needed to form the tension fractures is thus high, but not very high in comparison with the compressive stresses that generally exist in the crust. The compressive stress increases with depth in the crust. For example, in the roofs of many shallow magma chambers in Iceland (Chap. 4), at depths of one to three kilometres, the vertical stress is between about 24 million and 80 million pascal. The magnitude of the vertical stress in the roofs of shallow chambers is thus 8–27 times larger than the tensile stress needed to rupture the crust and form tension fractures at Thingvellir.

5.4 How Deep Are the Fractures?

Now that we know the stresses required to form the impressive tension fractures (and similar stresses are needed for the large faults such as Almannagja), the next question is how deep are the fractures? These are really two questions. One question is: what is the visible depth of the fractures, that is, the part mostly filled with groundwater? The other question is: what is the depth of the fracture as a narrow crack in the crust? As for the first question, Flosagja reaches a maximum visible depth of some 25 m (Figs. 5.11 and 5.13). There are other tension fractures nearby that reach even greater visible depths. The best known is **Silfra**, whose maximum visible depth is around 60 m. Silfra is on the north shore of, and extends into, Lake Thingvallavatn, a few hundred meters to the south of Peningagja (Fig. 5.11). Silfra is very popular for diving.

The second question is to what depths in the crust do the tension fractures reach? Here I mean the depth not as the widely open fractures seen at the surface, or with the openings that people can dive into, but rather the depths to which the fractures continue as narrow cracks down into the crust. You might ask how it is possible to find this depth. The answer is that all large tension fractures, such as Flosagja, Nikulasargja, Peningagja, and Silfra can only reach a certain maximum depth. If (say during a rifting event) they attempt to exceed this maximum depth, they will automatically change into normal faults. That is, one of the fracture walls will then subside relative to the other wall—just like at Almannagja and the other normal faults at Thingvellir. Using this information, and general knowledge of how fractures form (a specific scientific field named **fracture mechanics**), it is possible to calculate the maximum depths of tension fractures such as Flosagja and Silfra as being between 300 and 400 m. They are thus most likely entirely within the thick pahoehoe lava flow that occupies the uppermost part of the Thingvellir Graben—namely the Thingvellir lava flow.

And then, of course, the next question would be: how deep is Almannagja? The answer is that there are no simple methods for calculating accurately the depths of large normal faults such as Almannagja. If Almannagja were highly active seismically—with numerous small earthquakes—then their depths would indicate the depth of the fault. But Almannagja has very little seismic activity. One crude indication of the depth of a fracture, including faults such as Almannagja, is its length at the surface: longer fractures tend to be deeper than shorter fractures.

Almannagja is the longest continuous fracture of the Thingvellir Graben. By continuous fracture I mean that all the fracture segments or parts are physically connected or linked together—there is no strip of land in-between their nearby ends. Its total length as a continuous fracture is about 7.7 km. For comparison the shortest tectonic fracture in the Thingvellir Graben is about 60 m and the average or mean length of all the fractures about 620 m. The longer fractures are generally normal faults and generate earthquakes when they slip, whereas the shorter fractures tend to be tension fractures with little earthquake activity when they grow.

But Almannagja, like all the larger fractures, is composed of parts or segments, many of which, even if comparatively close to each other, are not physically connected—they are disconnected and offset (Figs. 5.8, 5.15 and 5.16). We know that in earthquakes segmented and disconnected faults commonly act as single faults, and the same would apply to all the segments of Almannagja during moderate to strong earthquakes (Almannagja cannot generate really major earthquakes of magnitude 7 or greater). And if all the segments of Almannagja are counted, then its length within Thingvellir is at least 15 km. Similar segments continue into the hyaloclastite mountains north of Thingvellir (Armannsfell, Ármannsfell, Chap. 6), as well as to the southwest along Lake Thingvallavatn and towards the Hengill Volcano (Chap. 12). If all these segments are regarded as parts of Almannagja, then its total length is easily 30–40 km. Similar lengths would be obtained for some other large faults in the area; their lengths may reach several tens of kilometres when all the segments, also in the older rocks, are considered parts of the same faults.

Then we come back to the question: how deep into the crust do Almannagja and the other large faults at Thingvellir extend. The answer is at least 10 km, and more likely about 20 km. Why not more than 20 km? Because at approximately that depth there is magma beneath the West Volcanic Zone (Fig. 2.2), of which the Thingvellir Graben is a part. The large faults of the Thingvellir Graben, and their extensions to the southwest and northeast along the West Volcanic Zone, most likely reach to the bottom of the crust, into the roofs of deep-seated and very large magma reservoirs (Fig. 5.17).

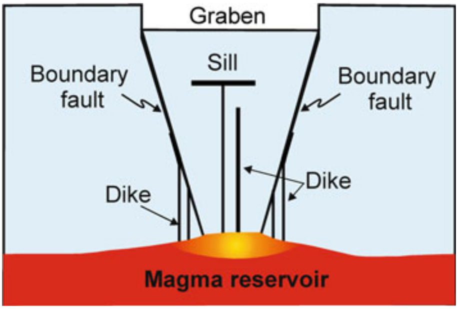

Fig. 5.17 Grabens are common in volcanic rift zones, such as in Iceland. Grabens can often act, temporarily at least, as barriers to dike propagation to the surface, and thus to fissure eruptions. Graben acts as a barrier to vertical dikes when the boundary faults deflect or stop or arrest the dikes and also when temporary compression inside the graben wedge stops or arrests dikes or deflects them into sills. When the graben wedge subsides, it enters into a narrower 'gap', so to speak, and may then become subject to horizontal compression for a while. Horizontal compression arrests vertical dikes or deflects them into sills; in both cases stopping the dike from reaching the surface to erupt

If so, why does the magma then not come up along the faults? The reason is their inclination and the unfavourable stresses generated temporarily in the graben following earthquake slip (Fig. 5.17). In a volcanic rift zone such as Thingvellir, the magma almost always travels to the surface through vertical magma-filled fractures, that is, **dikes** (Chap. 11). The magma very rarely uses existing inclined fractures, such as normal faults, for the simple reason that it requires much more energy to push the inclined fracture walls aside to make room for the magma, the dike, than to use the numerous vertical cooling fractures, columnar joints (Figs. 5.5, 5.6 and 5.10) to generate its own path to the surface. This follows because the plate movements are horizontal, so that it is easier for the magma to push the crust horizontally than in an inclined manner. Additionally, when the Thingvellir graben subsides along the main faults, Almannagja and Hrafnagja (Hrafnagja is discussed in Chap. 6), the effect is to hinder magma movement to the surface. When the wedge-shaped crustal block of the graben subsides, it is forced into a gradually narrower 'gap' in the crust (Figs. 4.14 and 5.17), so that the effect is mechanically similar to pressing a cork into a bottleneck, namely temporary horizontal compression. This compression generates compressive stresses which tend to prevent vertical dike propagation; the dikes either become deflected into horizontal sills or stop altogether (Fig. 5.17). In either case the dike is unable to reach the surface to erupt.

5.5 When Will the Next Eruption Occur?

When can we then expect the next volcanic eruption in the Thingvellir area? The last eruption in the area was not in Thingvellir Graben, but rather at the south end of the lake (Fig. 5.14), close to the volcano Hengill (Fig. 5.2; Chap. 12). This eruption occurred about 2 thousand years ago, during which a lava flow formed as well as the island of Sandey (Fig. 5.14; Chap. 12). Since that time there has not been any eruption in this part of the volcanic zone. The next eruption is in fact more likely to occur in the Hengill Volcano (Chap. 12) than in the Thingvellir Graben. This follows from pure statistics. As we discussed in Chap. 4, outside the main central volcanoes (such as Hengill) there is, at any given locality (such as Thingvellir), one new lava flow erupted every several thousand years—and occasionally there are tens of thousands of years between successive lava flows.

Given that the main lava flow in Thingvellir Graben is about 9 thousand year old (from Skjaldbreidur, Chap. 6), however, we might expect a new flow to come in the geologically near future. In active areas such as Thingvellir, however, the 'near future' commonly means tens or hundreds, even thousands, of years from now. Whether that eruption occurs inside the valley itself, or, as in the eruptions that formed the current lava flows in the graben, outside the graben, we do not know. But when the eruption occurs, it is likely to be much larger than the recent small eruptions in central volcanoes such Grimsvötn (Grímsvötn), Hekla, and Eyjafjallajökull (Chap. 14). In fact, an eruption in the Thingvellir area, when it occurs, is not unlikely to be of the order of several cubic kilometres.

6 Thingvellir (Þingvellir) to Geysir

You can start this part of the journey from several places at Thingvellir but perhaps the best way is to follow Road 36 east through the graben. The **eighth stop (8)**, located in Fig. 4.1, is then made at a parking place close to one of the two main faults on the eastern side of the Thingvellir Graben, namely Hrafnagja (**Hrafnagjá**, The Raven Fissure). Driving along Road 36 from the western part of the graben, you see several large hillocks in the lava. These are **tumuli,** as are described in Chap. 2 (Fig. 2.8). The most noticeable structures from the road, however, are the mountains (Fig. 6.1); from west to east, **Armannsfell** (**Ármannsfell**, the mountain of Armann, a man's name), **Skjaldbreidur** (**Skjaldbreiður**, Broad Shield), and **Hrafnabjörg** (Raven Mountain). Some or all of these mountains can be seen from almost anywhere within the Thingvellir area (Figs. 5.3, 5.14, 6.1 and 6.2). And you have, no doubt, looked at them during your excursion in Thingvellir. However, it is worth while to stop when leaving the Thingvellir Graben (Fig. 6.2) and have a look at these mountains so as to understand their formation. We shall therefore take a look at these mountains from this stop, but first we take a quick look at the main fault seen here and crossed by the road, namely Hrafnagja

6.1 Hrafnagja (Hrafnagjá)

Hrafnagja is one of the two main boundary faults (Figs. 4.14 and 5.17) of the Thingvellir Graben (Figs. 5.14 and 6.2), but there are also faults to the east of Hrafnagja (Gildruholtsgja, Gildruholtsgjá, and Heidargja, Heiðargjá) that may be regarded as the easternmost faults of the active graben (Fig. 5.14).

Hrafnagja is composed of segments (Fig. 6.3), as you see clearly from the eighth stop. The fault walls are less sharp than those of Almannagja, and the

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Fig. 6.1 General view of the mountains north of the Thingvellir Graben (see also Fig. 5.14 for an aerial view of these mountains). View northeast from the fourth stop, Armannsfell is in the upper left (west) corner, followed to the east by Skjaldbreidur, and finally Hrafnabjörg in the upper right (east) corner. The elongated mountain with irregular top between Skaldbreidur and Hrafnabjörg is a hyaloclastiteridge named Tindaskagi

overall impression is of a fracture composed of 'valley like' offset segments, one of the offsets being across the road at the parking place. Hrafnagja extends from Lake Thingvallavatn to the northeast, with a total length of some 11–12 km. The maximum subsidence or vertical displacement is only about 12 m, but the opening is as great as 68 m. Thus, Hrafnagja has a greater opening, is 'wider', but less vertical displacement or subsidence across it than Almannagja. The width of the Thingvellir Graben between Almannagja and Hrafnagja is about 5 km, but the total width to the easternmost fault, Heidargja (Fig. 5.14) is about 7 km. This entire area, 7 km wide, has subsided by a maximum of about 70 m in the past 9 to 10 thousand years. That means an average rate of subsidence for the entire graben of as much as 7 mm per year, which is similar to the rate of spreading or extension of the graben during the same period.

6.2 Botnsulur and Armannsfell (Botnsúlur and Ármannsfell)

Coming back to the mountains, we have already seen **Armannsfell** on Figs. 4.12, 5.3, 5.13 and 6.2. And now we add a close-up (Fig. 6.4). Armannsfell is a typical hyaloclastite mountain, similar to those we saw on the way to Thingvellir (Chap. 4) and on the way to Reykjavik (Chap. 2). Armannsfell was most likely formed during subglacial eruptions (eruptions under the ice sheet that covered Iceland at the time) during the second last glacial period and could be around 150 thousand

Fig. 6.2 Aerial photograph of the Thingvellir Graben (snow-covered in winter). View northeast, the main boundary faults are indicated as 'faults', namely Almannagja (west or left) and Hrafnagja (east or right). The main mountain seen in the left part of the photograph, clearly dissected by faults, is Armannsfell. Skjaldbreidur, indicated, is partly cloud-covered

Fig. 6.3 Aerial photograph of two segments (indicated) of Hrafnagja. View southeast, the segments are offset where the footpath crosses the fracture and form a part of an en echelon system, similar to but much larger than that at the southwest end of Almannagja (Fig. 5.16). The total length of the part of Hrafnagja seen her is about 1 km. The road seen in the upper left corner is part of Road 36

years old. The mountain is then considerably younger than the nearby hyaloclastite mountain Botnsulur (Fig. 5.14), which we discussed briefly in Chap. 4 (Figs. 4.11 and 4.12). **Botnsulur** were probably formed during subglacial eruption in the third last glacial period and may be around 250 thousand years old (Chap. 11).

There are many faults that dissect Armannsfell (Figs. 5.13, 5.14 and 6.4). These are generally the same faults as in the Thingvellir Graben. The faults in Armannsfell are mostly with greater vertical displacements than in Thingvellir simply because in Armannsfell the faults have had longer time to develop. If an earthquake slip occurs once every several hundred or thousand years on such a fault, then a fault dissecting a 9 thousand year-old lava flow, as in Thingvellir, has perhaps had 10–20 slip events. By contrast, a fault dissecting 150 thousand year old rocks, as in Armannsfell, could have had 150–300 slip events. Other things being equal, the latter would normally show a larger displacement.

6.3 Hrafnabjörg Table Mountain

The youngest of the hyaloclastie mountains seen here is Hrafnabjörg (Fig. 6.5), which is from the last glacial period and probably only about 20 thousand years old. Hrafnabjörg is a classical hyaloclastite mountain of the type named **table mountain**. The formation and general structure of table mountains is indicated in Fig. 6.6. Like other hyaloclastite mountains, table mountains mostly form under ice, while some form under water (in the sea or in deep lakes). Most of the hyaloclastite mountains in Iceland formed under ice during the glacial periods of the Ice Age, that is, tens or hundreds of thousands of years ago. Hyaloclastite mountains, however, are still forming today during the many eruptions in the ice sheets of Iceland and in the sea. In particular, the eruptions of **Surtsey** (Chap. 14),

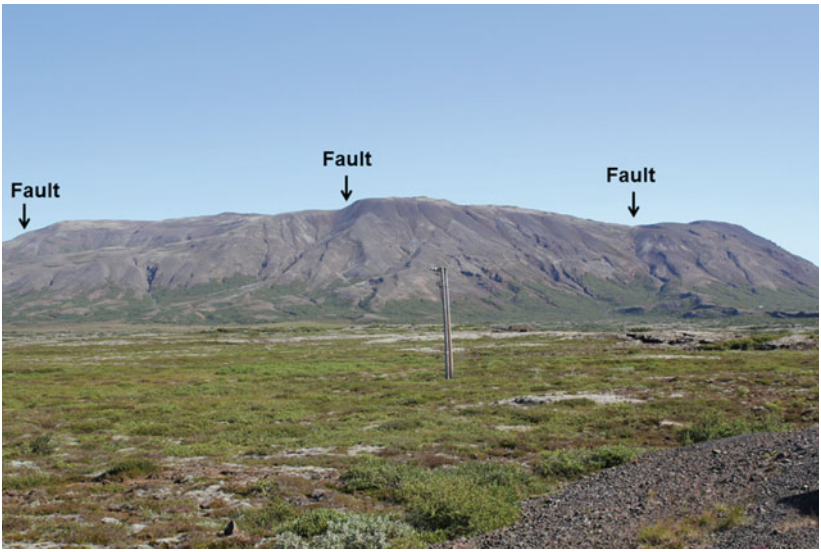

Fig. 6.4 View north, Armannsfell is dissected by many faults (only some are indicated here). These are all normal faults, similar to Almannagja, and some are continuations of Almannagja. In all the faults indicated here the displacement or subsidence is down to the west, that is, down to the left. Other faults dissecting the mountain, however, are with subsidence to the east (right) just like Almannagja

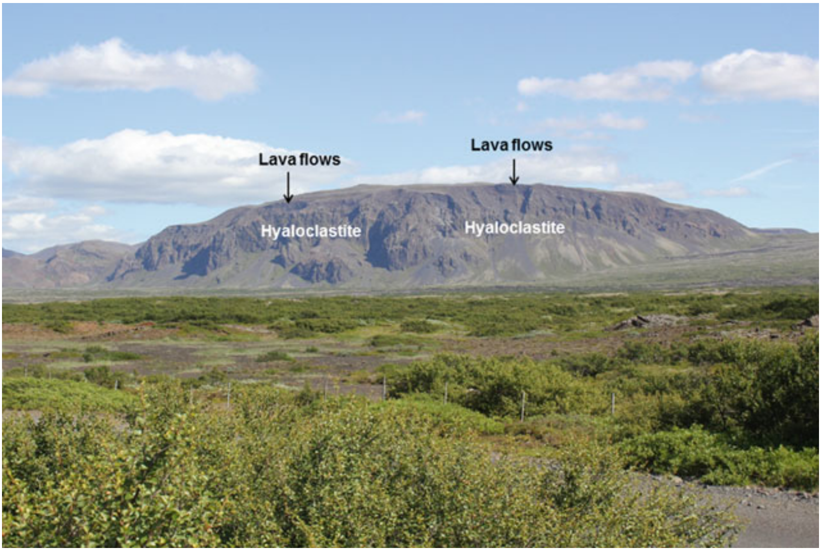

Fig. 6.5 View east, the table mountain Hrafnabjörg. The lava flows are clearly seen in the top part of the mountain, resting on thick layers of hyaloclastite. This photograph is taken from the road close to Armannsfell. Compare Figs. 11.22 and 11.23

an island offshore the south coast of Iceland (formed in 1963–1967), **Gjalp (Gjálp)** in the Vatnajökull ice cap in 1996 (Fig. 1.4, Chap. 1), as well as the many eruptions in the volcano Grimsvötn—which erupts once every 10 years on average—also in the Vatnajökull ice cap (Fig. 2.2), have clarified many aspects of subglacial eruptions and the formation of hyaloclastite mountains.

During such an eruption, the magma gradually melts the ice to form a lake (Fig. 6.6). As the water accumulates and becomes deeper, the magma begins to form a pile of pillow-looking structures. Not surprisingly, these are named **pillow lavas**, and we see a beautiful example of a pillow lava at the ninth stop, where we also discuss its formation in greater detail. As the pile of pillows grows, its top reaches gradually shallower water depths, and when close to the water surface the eruption no longer produces pillows but rather fragmented rocks, named **breccia**. The breccias form as result of explosions when the eruption conduit has come close to the surface of the lake in the ice sheet. Finally, when the breccia has piled up inside the lake to form an island, eruptions on dry land (the island) can take place, in which case the explosions and breccia formation stop, and ordinary lava begins to flow. The lava is normally a typical pahoehoe lava, with the same characteristics as we see in the walls of Almannagja (Figs. 5.4, 5.5, 5.6 and 5.7). The lava flows form the top of the table mountain—the cap (Figs. 6.5 and 6.6).

We discuss the rock types that constitute the table mountains in greater detail at the ninth stop where they can be observed in detail. As mentioned, the best recent eruptive analogy as to table-mountain formation is the eruption that formed the island of Surtsey, offshore the southern coast of Iceland, that is, the southernmost island of the Vestmannaeyjar archipelago (Fig. 14.12b; Chap. 14). Another analogy, a bit more remote since the eruption was entirely on dry land, is the 2014–2015 eruption of Bardarbunga-Holuhraun, north of the Vatnajökull ice sheet (the

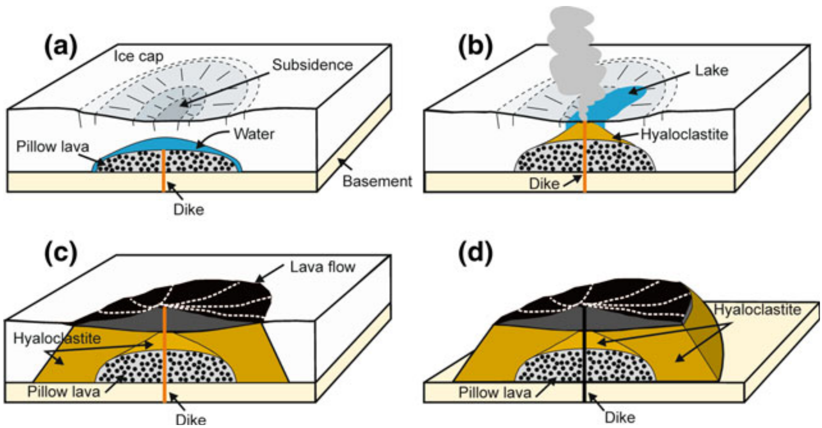

Fig. 6.6 Schematic illustration of the formation of a table mountain during eruption beneath an ice sheet. **a** Eruption begins under an ice sheet or cap. Melting of the ice generates a lake under the ice and subsidence at its surface. The erupted magma forms pillow lava. **b** As the pillow lava piles up, the water pressure decreases and the eruption changes, first, to pillow breccia (broken pillows) and then to hyaloclastite. Eventually, the melting reaches through the ice, and the lake is seen in the depression at the surface. **c** When the eruption conduit and crater reach above the lake, lava starts to flow, most commonly pahoehoe lava. The lava flows build up a lava shield on top of the pile of hyaloclastite and pillow lava. The shape of the mountain, particularly its steep slopes, is much determined by the ice. **d** When the ice sheet/cap has melted, such as at the end of a glacial period, the table mountain stands above its surroundings so that the bottom of its top lava shield is corresponds roughly with the surface of the ice at the time of mountain formation. It follows that the height to the contact between the lava shield and the hyaloclastite foundation indicates crudely the thickness of the ice sheet/cap when the table mountain formed

ice sheet is located in Fig. 2.2). It is true that the eruption was on dry land. However, it was only a few kilometres north of the northern edge of the main ice cap in Iceland, Vatnajökull. If the eruption had occurred some 15 or 20 km further south, it would have happened well within the ice cap, resulting in a subglacial eruption. And what happened early on in the eruption, which lasted close to half a year, is that almost all the lava effusion became concentrated at one eruption site, namely the main crater. If the same would have happened, as is likely, in an otherwise similar subglacial eruption, then, given the duration of the eruption and the volume of magma erupted (close to 2 km^3), a classic table mountain (Figs. 6.5, 6.6, 11.22 and 11.23) would almost certainly have formed.

6.4 Skjaldbreidur (Skjaldbreiður) Lava Shield

The third main mountain seen from the first stop discussed here is **Skjaldbreidur (Skjaldbreiður).** This beautifully shaped volcano is seen from many of the stops in Thingvellir, as well as from the air (Figs. 5.3, 5.14, 6.1 and 6.2). Here we add a

Fig. 6.7 The lava shield Skjaldbreidur (Skjaldbreiður). View east, this photograph is taken from the road just north of Armannsfell. The lava shield is younger than 9 thousand years and the average slope of its flanks is 5°–6°

Fig. 6.8 The lava shield Lyndalsheidi (Lyngdalsheiði) is from the last interglacial period and around 120 thousand years old. The shield is exceptionally gently sloping or flat, with an average flank slope of 2°–3°

close-up taken somewhat north of the Thingvellir Graben (Fig. 6.7). Skjaldbreidur is the youngest of the volcanoes that we have discussed at this stop, less than 9000 years old. It is of the type referred to as **lava shield**. These are generally of the same shape as **shield volcanoes**, the typical volcanoes of the islands of Hawaii (Mauna Loa, Mauna Kea, Kilauea), Galapagos, and many other famous volcanic islands. While the shape is essentially the same—that of a shield with gently sloping (commonly 3°–6°) sides—there is a fundamental difference between lava shields and shield volcanoes. Lava shields, such as Skjaldbreidur, normally form in a single eruption (some show evidence of more than one eruption, but always very few) and have no shallow magma chamber. By contrast, shield volcanoes, such as Mauna Loa, form in numerous eruptions, many of which come from a shallow magma chamber, over periods of hundreds of thousands of years and are much larger than lava shields.

So how long time does it then take for a lava shield such as Skjaldbreidur to form? At least many years, and probably tens of years. This follows because the volume of magma that comes up through the conduit of a lava shield in unit time (effusion rate) is rather small, so it takes a some time, perhaps many tens of years, to form a volcano of the size of Skjaldbreidur—whose estimated volume is about 17 cubic km (17 km^3). By contrast, Mauna Loa may have a volume as great as 75 thousand cubic kilometres and has been erupting (been active) for the past 700 thousand years. The lava flows in the walls of Almannagja are not from Skjaldbreidur but rather from fissure to the east of the Thingvellir Graben, and 9–10 thousand years old. In the northeastern part of the Thingvellir Graben, the lava flows from Skjaldbreidur are on top of the lava flows seen in Almannagja.

There are tens of young lava shields in Iceland. By 'young' I here mean shields formed in the past 11–12 thousand years or so, that is, following the melting of the ice from the last glacial period. Most of these shields are much smaller than Skjaldbreidur, generally about 1 km^3 or less. Skjaldbreidur is the second largest or largest young shield in Iceland (the competitor is the lava shield **Trölladyngja** in the central highlands). You already saw some of the shields while driving from Keflavik to Reykjavik (Chap. 2), and you will see more if you do the excursion in Chap. 13. In addition there are many earlier-formed lava shields buried in the lava pile in the older parts of Iceland. Some of these are described in Chap. 11).

6.5 Laugarvatnshellar and Hyaloclastites

We now continue to the east of Hrafnagja along Road 36 until it meets Road 365, which leads east across a very smooth lava field. This smooth and almost flat field is a part of another lava shield, namely **Lyngdalsheidi (Lyngdalsheiði)**. This is an exceptionally flat (gently sloping) lava shield (Fig. 6.8), and much older than Skjaldbreidur and possibly formed during the last interglacial period around 120 thousand years ago. We continue along Road 365 until it meets a road marked **Laugarvatnsshellar** that takes us to the north, to famous small caves and pillow lavas, which is our **ninth stop** (located in Fig. 4.1).

The caves (Fig. 6.9) are famous primarily because a family used to live there for some years in the early part of the 20th century. In fact, two babies were born in these caves. All this is explained in detail in the information provided at the site of the caves. Our focus, however, is on the rocks of the caves and their surroundings. The rock hosting the caves is hyaloclastite, that is, basaltic breccia. It forms during explosions in the conduit of the volcano when the opening of the conduit is at a comparatively shallow depth below the surface of the lake within the glacier (Fig. 6.6). The explosions continue even when the hyaloclastite mountain has built up to the surface of the lake so long as water can reach into the conduit, that is, the erupting crater. Each explosion in the crater generates one layer. The layers are easily seen in the rock (Fig. 6.10). Some layers are with smaller grains or particles than others. Usually, those with larger particles formed in more powerful explosions, although other factors may also play a part in the grain-size distribution.

Commonly, much larger rock fragments are embedded in the fine-grained layers (Fig. 6.11). This is in accordance with the size distributions—how things arrange themselves according to size—resulting from many natural and human-related processes and activities. They all have the same general distribution: namely, the great majority of the objects or features (here grains) are comparatively small, while a few are very large, with a continuous variation in size between the very small and the very large (Fig. 6.12). Such distributions are named heavy-tailed, of which the subclass of **power laws** is the most important and common. They are

Fig. 6.9 The main entrance to the caves Laugarvatnshellar. The caves are hosted by a typical hyaloclastite, brownish breccia

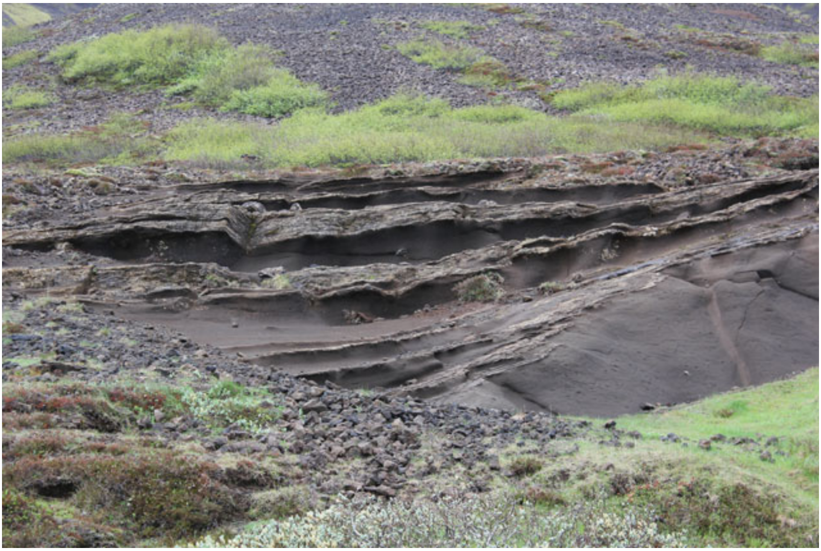

Fig. 6.10 Layering in the hyaloclastite at the caves of Laugarvatnshellar. Each layer normally corresponds to one explosion in the crater during the eruption. The grain size depends on the power of the explosion

named power laws because the shape of the curve which is obtained when the number of objects of a given size depends on the size raised to a certain power or exponent. Power laws apply to the size distributions of the explosions that generated the layers that you see here (Figs. 6.10 and 6.11). More specifically, most of the explosions were comparatively small, a few ones (comparatively speaking) very large. Similarly, most of the grains in the layers are comparatively small, but a few ones (Fig. 6.11) are (always, comparatively speaking) very large.

6.6 The Importance of Power Laws in Natural (and Human) Processes

Given that we have introduced power laws into the discussion, it is perhaps appropriate to mention some of the natural and human processes and objects or structures that follow power laws (Fig. 6.12). These include:

- **Sizes of eruptions**. Most volcanic eruptions are small whereas a few ones are very large. This applies both to the volume of eruptive materials (lava flows, ash or pyroclastic materials) as well as the energy released or transformed in the eruption. For example, the 2010 Eyjafjallajökull eruption in South Iceland was a small eruption. Its total volume was only about 0.1 km^3 (one-tenth of a cubic km)

Fig. 6.11 The grains in the hyaloclastite layers here are generally very small (the rock is referred to as tuff), although with some variation. In addition, there are also some much larger 'grains'. These larger grains are mostly fragments of pillows (see Figs. 6.14, 6.15, 6.16, 6.17 and 6.18). Statistical distributions where objects with small sizes (here the grains) are very common while objects with very large sizes (always in comparison with the small objects in the same distribution) do also occur are named heavy-tailed, of which the most common are power-law distributions

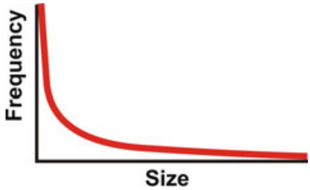

Fig. 6.12 Power laws are characterised by a very high frequency of small objects or features or processes (for example, volumes of lava flows) and a very low frequency of large objects or features of the same type and in the same distribution. Small and large are always relative and refer to the particular distribution being studied. Power law is what is called 'scale free', which implies that there is no size of objects or events that is typical for the distribution. For example, the mean value of volcanic eruption volumes or earthquake sizes are not the typical, in the sense of most common, values for these processes and products. Not all heavy-tailed distributions are power laws, but there exist methods for determining which are. The large sizes of events and objects are of the greatest importance in power laws—the large earthquakes, volcanic eruptions, floods, meteoritic impacts, economic downturns, and so on—and are, as yet, impossible to reliably forecast

of magma. By contrast, during the formation of Skjaldbreidur, as we discussed about, about 17 km^3 of magma was erupted. And in the largest eruptions on Earth, in North America and South-East Asia, about 5000 km^3 were erupted. Such gigantic supereruptions happen very rarely, fortunately for humankind, or, on average, once every several hundred thousand years.

- **Volumes of rock layers**. Most rock layers and units are comparatively small, whereas a few are very large. This applies to all types of rock layers. For example, to volumes of lava flows, which relate to the sizes of eruptions discussed above, and volumes of intrusions(dikes, sills). This relationship is also reflected in the thickness of these structures. If you measure the thicknesses of, say, many dikes or many lava flows, you would normally obtain a power-law size distribution.
- **Sizes of volcanic edifices**. When the volumes of volcanic edifices are estimated, again, they show power-law size distribution. Thus, most volcanoes are very small, as applies to many volcanoes in Iceland, whereas a few ones are very large.
- **Sizes of earthquakes**. There are earthquakes every day in Iceland, and the same applies to all the plate boundaries on Earth. But the great majority are so small (less than magnitude 2) that we do not feel them. There may be 15

earthquakes larger than magnitude 7 on Earth each year, but only one larger than magnitude 8. Similarly, the greatest earthquakes, of magnitude 9 or larger, occur on average only a few times in a century. By contrast, there may be half a million earthquakes, mostly of very small magnitudes, recorded instrumentally every year.

- **Fractures**. Earthquake magnitudes relate directly to the size of the fracture, the fault, generating the earthquake. It is therefore not surprising that fracture sizes also follow power laws. For example, all the main fractures in Thingvellir have been measured. The longest continuous ones are close to 8 km, whereas most of the fractures are shorter than 200 m. And if we took the small fractures, columnar joints, as seen in the walls of Almannagja (Figs. 5.5, 5.6 and 5.10) into account, then the great majority of fractures would have lengths of less than one metre.

There are numerous other geological processes that produce power-law size distributions. These include **landslides**, **floods**, and **meteoric impacts**, in all of which most of the features/structures produced are small while very few are very large.

Power laws are not restricted to natural processes and objects. Many human-generated processes and structures follow power laws. For example, the sizes of **cities** (in terms of area covered, population, or both), the sizes of **buildings**, the lengths of **streets**, the number of visits to **internet sites**, as well as **complex networks** of various types (social, biological, computer, and technological networks) and **economic downturns** all have power-law size distributions. And the list goes on. The question is then: how do these power laws arise?

A related question is: why are these distributions power laws rather than **bell-shaped** or normal (Fig. 6.13)? We know that bell-shaped distributions occur

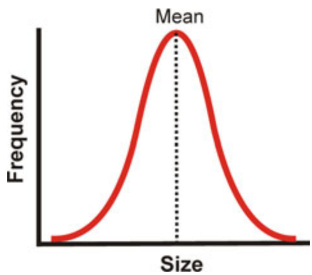

Fig. 6.13 In contrast to power-law distributions, normal or bell-shaped (or Gaussian) distributions have a mean value that characterises the distribution. More specifically, the mean is also the most frequent (mode) value and is thus the typical value—in the sense used here. The mean is thus a very useful concept to characterise a normal distribution, but says rather little about a power-law distribution—where the largest possible values are normally of the greatest interest and concern for the society

when the sizes of events/object or structures are independent. For example, heights and weights of people in a city or a country follow normal distributions because these values or numbers for the individuals are totally independent. By contrast, power laws are known to arise when the sizes of events/objects or structures depend on each other (are interdependent). For example, a long fracture forms because many smaller fractures link together under favourable conditions.

Large volcanic eruptions, earthquakes, floods, landslides, and meteoritic impacts are of great concern for society. In fact, the largest eruptions and meteoritic impacts on Earth are the only known natural disasters with the potential of having devastating effects on all mankind. It is thus of fundamental importance to understand the processes that give rise to these, and power laws in general. There are currently very great research efforts on power laws, particularly in their relation to natural and human-made complex networks of various types. While several mechanisms have been proposed for power laws, it is fair to say that none of them really explains why certain processes and things follow power-law size distributions, whereas others do not. So, briefly, while power-laws are extremely common and important, in natural and human-related processes, at this stage we cannot explain them. One consequence is that we cannot, at this stage, reliably forecast large volcanic eruptions and earthquakes.

6.7 Pillow Lavas

Leaving power laws now for a while, we walk to a little depression, a (normally) dry stream to the right of the caves. While walking along the depression, we first see the layered brownish hyaloclastite on the left side (Fig. 6.11). The brown colour, which we also saw in the rocks at the caves (Figs. 6.9 and 6.10) is due to alteration of glass in the ash. As a result of chemical reaction, the black glass gradually becomes brownish—which is the common colour of the breccia or ash layers in all the hyaloclastite mountains.

After walking along the depression for a few tens of metres, we see an exceptionally beautiful section or outcrop (exposure) of **pillow lava** (Fig. 6.14). You can readily see here why this is called pillow lava, given that the pillow-shaped structures are exceptionally clear as shown in the photograph. On a closer look (Fig. 6.15) you see that while some of the pillows are close to spherical (with circular cross sections, seen where the pillows are broken (Fig. 6.16)), many are comparatively flat (with elliptical cross-sections where the pillows are broken). The flat ones are properly named oblate ellipsoids. They get their flat shape partly because of the weight of the erupted materials above; the weight flattens them and

Fig. 6.14 Pillow lava. Many of the pillows are elongated and elliptical in a vertical cross-section, whereas others are close to spherical. The yellow notebook indicates the sizes of the pillows, which are mostly 20–80 cm in maximum diameter. The flat and very elongated pillow just above the notebook, however, is about 1.5 m in diameter

makes them, in fact, look more like ordinary pillows—which the spherical ones are not.

Another thing you notice, particularly in the central spherical pillow in Fig. 6.15, is that there are cracks or fractures that radiate from its centre to the margin or the pillow surface. These **radial fractures** are formed in exactly the same way as those seen in the walls of Almannagja (Figs. 5.5, 5.6 and 5.10), namely when the very hot magma cooled down from 1200 °C to a few tens of degrees. As the fluid magma changes into solid rock, its volume decreases, it shrinks, and part of the shrinkage results in the formation of these radial fractures. They are radial, rather than vertical as in the flow units in Almannagja, because the body that is cooling is spherical (a pillow) rather than a horizontal layer. In addition to the radial fractures, there are little holes or cavities in the pillows—again of the same type as we saw in the flow units in Almannagja. Here as there these cavities, called **vesicles**, form when gas escapes from the cooling lava.

The pillows normally form when there is high water pressure at the eruption site. It follows that they form mostly at the bottom of the hyaloclastite mountains, that is, during the early part of the eruption where the water pressure in the lake is still high (Fig. 6.6). It is known, however, that an ordinary lava flow may under certain conditions change into pillow lava on entering in a lake or the sea—as has been observed on the coast of Hawaii. In the hyaloclastite mountains of Iceland, the pillow lavas generally form in the lower parts of the mountains and were generated under high water pressure. As the pillows pile up, and the water surface remains the same, the depth from the top of the pillow pile to the water surface gradually decreases (Fig. 6.6).

As the water pressure decreases on the newly formed pillows, the gas inside them tends to expand and rupture the pillows. Thus the pillow lava gradually changes into mixture of pillows and breccia (broken rocks), as we see nicely in Fig. 6.17. As the eruptive materials continue to pile up, the water depth in the lake decreases and explosive activity begins with the formation of pure **breccia**, as seen

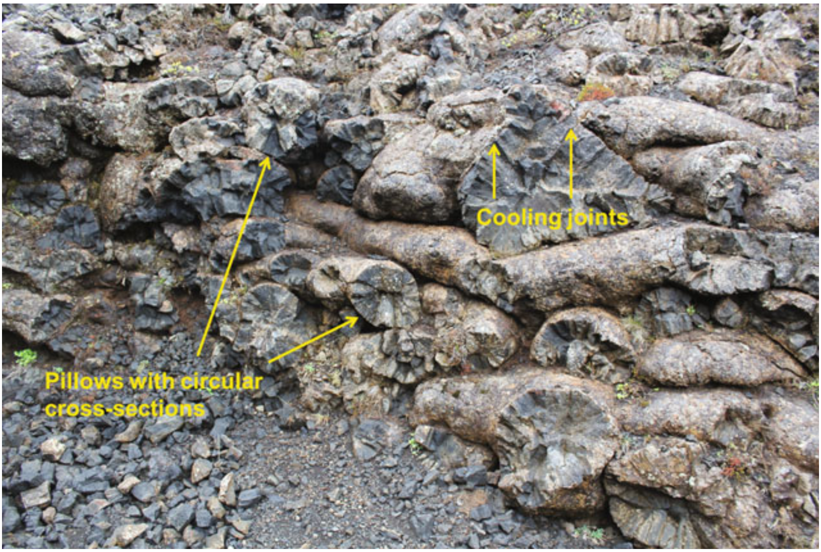

Fig. 6.15 Close-up of part of the pillow lava seen in Fig. 6.14. Here are many pillows with close to circular vertical cross-sections (some indicated). Notice that the cooling joints (some indicated) are everywhere perpendicular to the nearest surface of the pillow (a topic discussed further in Chaps. 11 and 14)

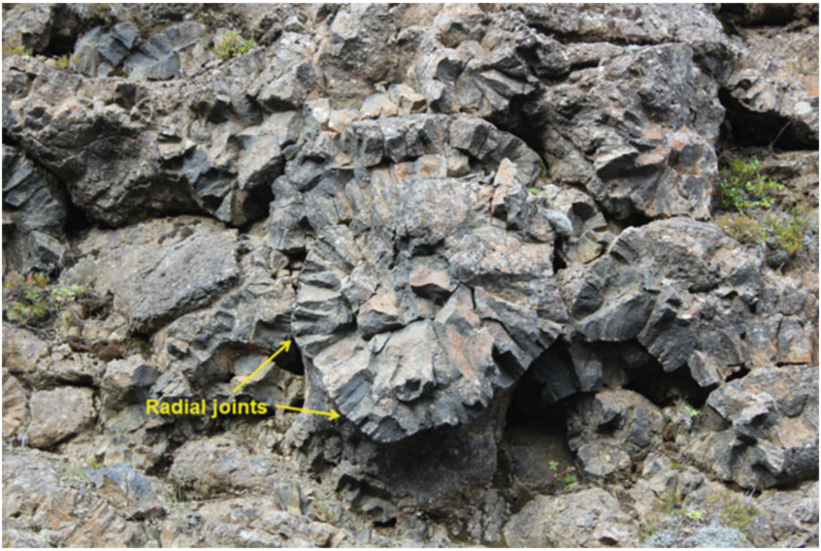

Fig. 6.16 Close to spherical pillow with a circular vertical cross-section showing radial cooling joints (columnar joints). The joints are perpendicular to the cooling surface, the surface of a sphere, and thus with a radial arrangement

in Figs. 6.9, 6.10 and 6.11. That breccia lies on the top of the pillow breccia (and thus on the pillow lava) and is inclined or sloping to the east, that is down the depression or gully, as is clear from the contact between the pillow lava and the breccia in Figs. 6.15 and 6.16.

While pillow lava exists in the lower parts of all hyaloclastite mountains, there are not too many easily accessible localities where it can be seen so clearly and, in particular, where the contact between the pillow lava and the breccia on top if it is as clear as here (Figs. 6.17 and 6.18). All the hyaloclastite mountains you see in Iceland have the same structure: the lowermost part is pillow lava, then pillow breccia, on top of which is breccia (Fig. 6.6). If the mountain peak (and the erupting conduit and crater) do not reach the surface of the lake, that is, if the eruption stops at this stage, then the result is a **hyaloclastite cone** (Fig. 2.7) if the conduit is circular, or a **hyaloclastite ridge** (Chap. 12) if the conduit is a long fissure. If, however, the conduit and its opening, the crater, eventually reach to the surface of the lake and water no longer flows into the crater, then ordinary lava starts to flow and the result is a table mountain (Figs. 6.5, 6.6, 11.22 and 11.23).

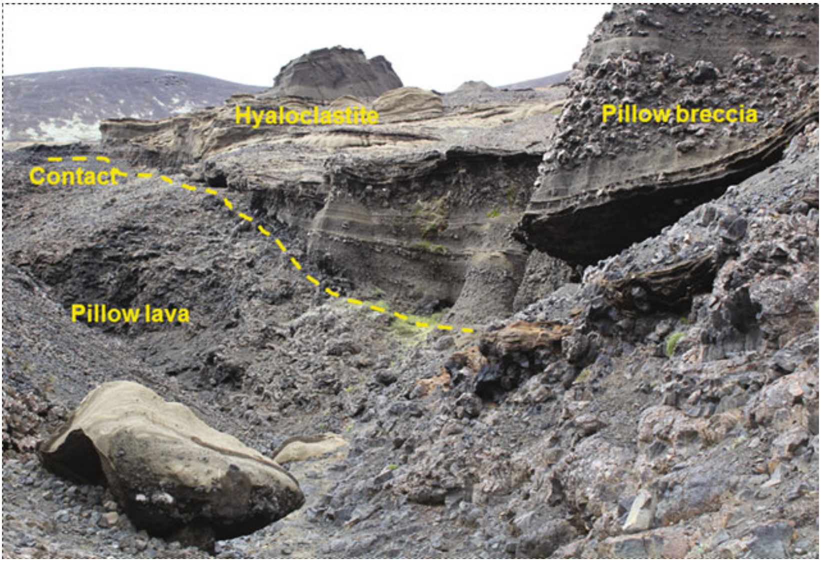

Fig. 6.17 The contact between pillow lava and pillow breccia is indicated. The gradual change from pillow breccia to hyaloclastite is also indicated

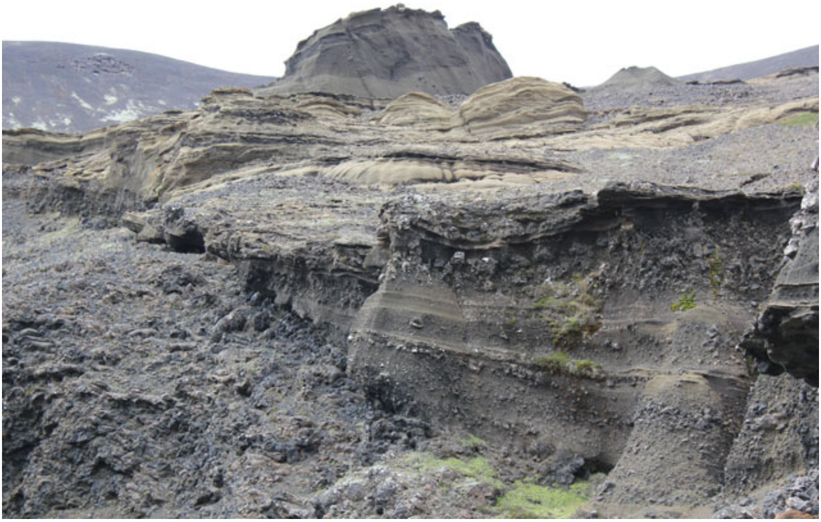

Fig. 6.18 Close-up of part of the contact in Fig. 6.17

From this second stop we drive back to the main road, Road 365, and continue our journey to Geysir. There are many beautiful areas on the way, including the lake **Laugarvatn**, a popular tourist area partly because geothermal water flows into the cold lake and makes its temperature comfortable for swimming. There are, in addition many touristic facilities at Laugarvatn. All the mountains around Laugarvatn and all the way to Geysir are hyaloclastite mountains. But they add nothing new to what we already know about such mountains, so we drive on all the way to Geysir.

7 Geysir

Geysir (the Great Geysir) may be the most famous geyser in the world, and is our **tenth stop (10)** in Fig. 4.1. The Geysir area is also located in Fig. 2.2. Geysir is the nameshake of all erupting (gushing) hot springs: they are referred to as **geysers**. The **Great Geysir** and the nearby **Strokkur** (and occasionally some other hot springs in the same area) are the only erupting hot springs in Europe. The Great Geysir is hardly active now; it erupts very rarely—only a few times each year—and the height of the fountain or water column during eruption is normally less than 10 m. This is considerably less, both as regards fountain height and, in particular, eruption frequency than that of the nearby geysir Strokkur.

Strokkur currently erupts on average once every 5–10 minutes, so if you stay for a while in the geothermal field, you are certain to see it erupting (Fig. 7.1). Each eruption is short, perhaps a few minutes, all counted. The height of the resulting fountain varies much. Occasionally, the fountains are as high as 30–35 m, but most commonly 10–20 m (Figs. 7.1, 7.2, 7.3 and 7.4). By contrast in the late 19th century, there are reports that the fountains of Strokkur occasionally reached the height of 60 m.

That height, however, is less than the height to which the fountains of the Great Geysir were able to reach in earlier times. In the middle of the 19th century it is reported to have occasionally reached the height of 170 m. This may, however, have been an overestimate because about the same time exact measurements by the scientist who first explained, in general terms, how erupting geysers operate (Robert Bunsen) indicated maximum foundation height of only about 54 m. It is well confirmed, however, that the fountain of the Great Geysir commonly reached 60–70 m in the 20th century. Furthermore, following the earthquakes in South Iceland in the year 2000, Geysir became reactivated and for some days reportedly erupted to heights of about 120 m (Sect. 7.2).

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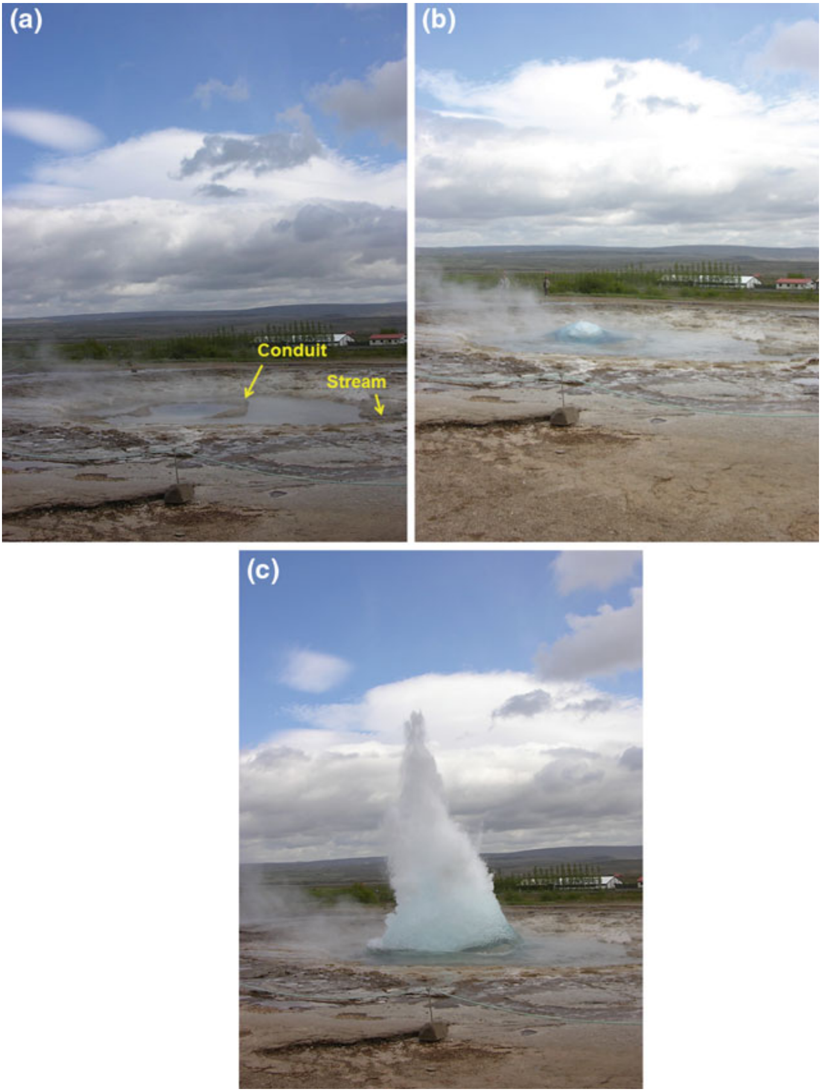

Fig. 7.1 **a**, **b**, **c** Three stages in the eruption of the geyser Strokkur. **a** The main conduit or pipe (indicated) is being filled with water. **b** Swelling of the water surface, so as to form a half-sphere on the top of the conduit, indicating the beginning of the eruption. **c** The eruption itself

7.1 Mechanism of Eruption

This brings us to two questions. First, why is the activity in individual geysers so variable over time—in particular why does it relate to earthquake activity. Second, what is the general reason of the geyser eruptions or gushes. We start with the second question, the one broadly explained by Bunsen during middle of the 19th century.

Eruptions in a geyser are driven by boiling of the geothermal water in the geyser pipe or conduit (Fig. 7.5). The water is everywhere above 100 °C and its temperature gradually increases with depth. However, because the water pressure also increases with depth—as you know from being in a swimming pool or the sea—the temperature at which the water boils (**boiling temperature**) and changes into steam also increases with depth. The boiling temperature is thus well above 100 °C

Fig. 7.2 Some eruptions in Strokkur fail to reach their peak and remain very small. Here is an example of one such 'failed' eruption

Fig. 7.3 Example of a reasonably large eruption in Strokkur. The human-made stream, indicated, helps keep the water level in the conduit comparatively low so as to encourage boiling and eruptions

at deep in the conduit or pipe of a geyser. For example, at the depth of about 20 m in the Great Geysir the normal water temperature is about 120 °C, which is still not enough to cause boiling. Overheating of the geothermal water is needed in order to reach the boiling point. When that happens, for example when water enters the pipe at a certain depth at a higher temperature than the surrounding water at that depth, then boiling begins, which normally results in an eruption.

In detail the process is then as follows. When the boiling begins in the conduit or pipe at a certain depth (Fig. 7.5), the water above that depth must rise somewhat. Why? Because heating the already hot water increases the water volume while at the same time **bubbles** form and grow, resulting in further volume increase of the water.

Fig. 7.4 Large eruption in Strokkur

The volume of the cylindrical pipe or conduit is basically always the same, so the only way that the additional water and steam volume in the pipe can be accommodated is by **lifting the water surface**. And this is what you see happening during preparation for an eruption (Fig. 7.1a, b). If the surface water cools very rapidly, such as when there is a strong, cold wind at the surface, it may be difficult for enough boiling to take place for an eruption to occur. But normally the pressure decrease due to the volume expansion and overflow of water at the surface (Fig. 7.1b) triggers further boiling in the upper part of the pipe, resulting in an eruption (Fig. 7.1c).

As the eruption starts, water is transferred out of the pipe (into the air), which further reduces the water pressure in the pipe. Consequently, boiling extends **deeper into the pipe**, generating more steam, and more water is shot up into the air. Thus, the eruption commonly occurs in several shots or gushes of water into the air in a rapid succession (Figs. 7.3 and 7.4). Following these shots, the water left in the pipe is all overheated so that it boils into steam. The rest of the eruption is therefore primarily a noisy steam eruption. After the eruption is finished, the pipe gradually becomes filled with geothermal water again (water is continuously flowing into the pipe through fractures), and the story repeats itself (Fig. 7.1). Much of the water that makes the fountain falls back into the pipe (Fig. 7.6), or flows back into it form the surrounding bowl, but some leaks away along the tiny stream (Figs. 7.1b and 7.3).

The story described above is the classic course of events from one major eruption to the next. But there are many factors that may affect the eruption scenarios. Thus, many eruptions are small and thus incomplete (Fig. 7.2) and their size distribution presumably follows a power law (Chap. 6). Then the pipe does not become anywhere close to being empty, so that the time to refill it so as to be ready for the next eruption is often very short. There are several other factors that affect eruption size and frequency. One, indicated above, is the weather. Rapid cooling of the surface water by wind may prevent the eruption from happening for

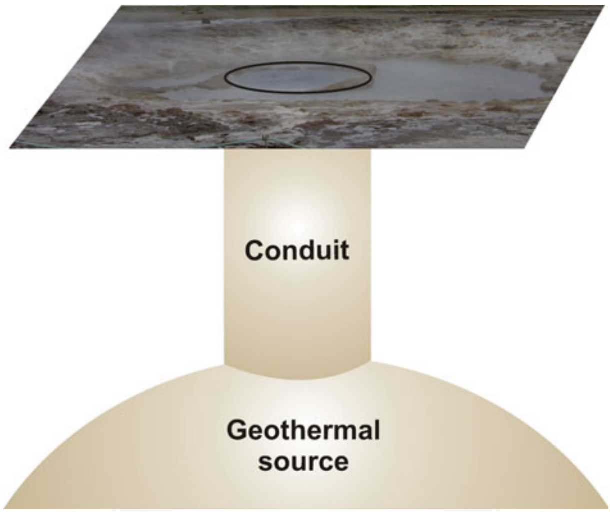

Fig. 7.5 Very schematic illustration of the conduit, the pipe, of a geyser. The pipe is normally a crude cylinder, into which hot water flows. Here we use the rim of the conduit of Strokkur as a model (Fig. 7.1a). The geothermal source is shown in a generalised way. Water can flow into the pipe from all directions, not only from below, but also through the walls of the conduit, and mostly through narrow fractures. When conduit is subject to extra loading, such as during earthquakes, stresses concentrate, that is, become raised at and around the conduit/pipe resulting in forming or reopening of fractures, thereby, commonly, increasing the flow of hot water into the conduit at various depths

a while. Another factor is the rate of inflow of water into the pipe (Fig. 7.5), which is variable, and also the water temperature. Generally, geothermal fields, such as the Geysir area (the formal name is **Haukadalur**, the Haukar Valley), are continuously changing. For example, the inflow of water into the pipe depends on the openings or apertures of the fractures through which the water flows. The apertures gradually change because of mineralisation, that is, particles or minerals from the hot water gradually fill and seal the fractures, thereby making them narrower and partly closed. Even a small change in the aperture or opening of a fracture has very large effects of its ability to transmit or conduct water. Which brings us to the question: why are there geothermal fields in Iceland, and why in the Geysir area or Haukadalur in particular?

Fig. 7.6 Much of the water that goes into an eruption of Strokkur falls again into the bowl around the conduit, as is seen here. Comparatively small amount of water flows from the geyser, mostly along a human-made stream (Figs. 7.1a and 7.3)

7.2 Geothermal Fields

All geothermal fields, whether they have geysers or not, are related to rain (and in Iceland also snow) migrating to great depths in the crust, becoming hot, and then rising as hot water to the surface. (The general term for condensed atmospheric water—including drizzle, rain, sleet, snow, and hail—falling on Earth's surface is **precipitation**.) Part of the rain and snow that falls on the ground runs off in streams and rivers, but part migrates into the soil and the solid rock below the soil. The water that remains in the soil and uppermost few tens of metres of the solid rock (bedrock) is referred to as groundwater, the water which we see in the open fissures and the lake at Thingvellir (Chap. 5). Some water, however, goes deeper, and in areas of volcanic activity, or recent volcanic activity, this water, if it migrates to great depths, becomes geothermal water.

How does the water migrate to great depths? Partly through numerous columnar or cooling fractures (joints), cavities from expanding gas (vesicles), as we saw in the lava in the walls of Almannagja (Figs. 5.5 and 5.6, Chap. 5) and in the pillow lavas (Figs. 6.14–6.18, Chap. 6). Much of the water, however, migrates to great depths through earthquake fractures, namely faults. So here is the connection between earthquakes and geysers, as is clear in the Geysir area. Earthquakes generate, or reopen, fractures that increase the flow of geothermal water towards the geysers. Not only that, but new or reopened fractures are very likely to occur exactly at the main geysers. Why? Because these are fed by pipes, cylindrical conduits (Figs. 7.1a and 7.5), and all such holes or cavities tend to **magnify stresses**, that is, concentrate stresses. It follows that when earthquakes occur, stresses become magnified at the geyser pipes (Fig. 7.5), and new fractures form or older ones become reopened at and around the pipes.

And it does not matter if the earthquake fractures themselves do not reach to the Geysir area. All earthquakes, the quakes themselves, carry stresses (and strains) and these become magnified at the geyser pipes. The new and reopened or reactivated fractures at and around the pipes then contribute to the activities of the geysers in two main ways. First, they normally (but not always) allow more geothermal water to flow into the pipes, thereby refilling them more quickly. Second, they change the fluid-flow paths and commonly make it possible for hotter or warmer water to enter the pipe at a shallower depth, thereby encouraging eruptions in the way we discussed above (Sect. 7.1). The fractures that supply the water into the pipes of erupting geysers, and of hot springs in general, are normally tiny. For example, the volumetric flow rate from Geysir (in other words, the volume of geothermal water flowing from the Great Geysir) is about 1.5 litre per second. A **single fracture** some tens of centimetres long and with an opening or aperture of about one millimetre could theoretically conduct all the water needed for the eruption activity of the Great Geysir.

Fracture reactivation normally results in increased activity of the geysers, but not always. While reactivation generally increases the ease of fluid transport through the rock, that is, the **permeability**, the fluid-flow paths also commonly change. This means that cooler water may be injected into the pipe than before. Alternatively, the hot water necessary to trigger eruptions may no longer be injected at a suitable depth into the pipe to cause boiling but at a greater depth where the pressure is too high for a water of the given temperature to boil. And there are various other scenarios possible whereby earthquakes can change the situation at and around the geysers so as to hinder, rather than help, eruptions to occur. In most cases, however, earthquakes increase the chances of eruption in geysers, as is indicated by the following brief account.

7.3 Geysers and Earthquakes

The connection between earthquakes and the activities of the erupting springs in the Geysir field is well established. Based on written accounts, the Great Geysir seems to have **become active** following large earthquakes in South Iceland **in the year 1294**. It may of course have been active much earlier, but these are the oldest written accounts (annals) describing its activity. Even if it was active earlier, it may have been dormant for a long period before the large earthquakes in 1294 and thus not mentioned in the annals.

Later strong and major earthquakes in South Iceland have affected the activity of the Great Geysir. For example, Geysir was essentially dormant before the large earthquakes in South Iceland in 1896, but following those earthquakes it produced long-lasting eruptions many times each day. The activity then declined over the next decades, while channels (for lowering the water level in the pipe) and the addition of soap to the water in the pipe (to make it easier for the water surface to rupture) where made to keep its eruptions going. However, Geysir had been essentially dormant for decades prior to the 2000 earthquakes in South Iceland (Chap. 14) when, for a while, it became very active. Measurements suggest that in the days following the June 2000 earthquakes, the fountains may have reached the height of over 120 m. The activity soon declined, however, and, as indicated above, eruptions are currently very rare and small.

A similar story applies to Strokkur. Like the Great Geysir it is unknown when its eruption activity began. However, following large earthquakes in 1789—the largest historical earthquakes in South Iceland—its eruptions became very noticeable after having been dormant for a considerable time. Following these earthquakes, Strokkur erupted frequently and with great force; in fact, its eruptions were more spectacular at that time than those of the Great Geysir. The 1896 earthquakes, which renewed the eruption activity of the Great Geysir, had the opposite effect on Strokkur, which became dormant. Strokkur remained largely dormant until a hole, tens of metres deep, was drilled into the bottom of its conduit in 1963. Since then Strokkur has been erupting on average once every 5–10 minutes.

7.4 Heat Sources

How is the geothermal water generated in the first place—why does the water become hot? The basic answer is that the **temperature** of the rocks in the Earth's crust everywhere **increases with depth**. On average, worldwide, the temperature increases with depth by about 25 °C for every kilometre. But the temperature increase with depth in the crust is much faster in active volcanic areas, and at plate boundaries in general. For example, in parts of Germany, in the Rhine Graben, which is volcanically active (although not very active), the temperature at the depth of 1 km may be as high as 40 °C. In areas with great volcanic activity, such as Iceland, the temperature in the volcanic zones is commonly at or above 200 °C at the depth of one kilometre. In fact, that is the definition of a **high-temperature geothermal field** in Iceland: the water temperature is above 200 °C at the depth of 1000 m. By contrast, if the temperature at the depth of 1000 m is below 150 °C, the geothermal field is referred to as a **low-temperature geothermal field**. While there are over 250 low temperature areas all over Iceland, there are only 32 well-defined high-temperature fields in the country, one of which is the Geysir area.

So why does the Earth become hotter at depths—what are the heat sources? For the Earth as a whole, the heat source is mainly **radioactive decay** of elements, accumulated when the Earth formed. Heat is generated in the crust through radioactive decay, but flows also from the **outer core**, which is molten, through the mantle (partly as **mantle plumes**, cylindrical conduits of partly molten material, one of which forms Iceland) and the crust to the surface. For volcanic areas such as Iceland, however, the main heat sources are much more local, and mostly **shallow magma chambers** and associated intrusions of the type we saw in Stardalur in Esja in Chap. 4. Thus, most of the high-temperature fields in Iceland are directly related to, and occur inside or close to, active **central volcanoes**—an excellent example being Hengill (Chap. 12).

The Geysir or Haukadalur high-temperature field, however, is somewhat special in the sense that no eruption has occurred in the area for the past 10 thousand years, so that it is not regarded as volcanically active. Rather, it is located at the **margin** of the active volcanic zone (Fig. 2.2). It presumably was an active volcano, some tens of thousands of years ago, but appears to be either dormant or totally extinct by now. There may of course be some intrusions, cooling magma bodies, at depth below the Geysir area even if no eruption has occurred for the past 10 thousand years. It is well known that most magma-filled fractures, dikes (Chap. 11), never reach the surface to supply magma to eruptions, and some may have propagated under the Geysir area without erupting. It is, however, more likely that the water in the Geysir area becomes heated up while circulating through deep fractures, many of which may be related to earthquakes. The water can reach depths of several kilometres in such fractures (Figs. 4.5 and 5.17), and then migrate to the surface, where it forms hot springs and erupting geysers. When the water finally reaches the surface in the hot springs and geysers it has been migrating through the rocks for a long time—some geothermal waters in Iceland circulate through the crust for **thousands of years**, while others circulate only for tens to hundreds of years, before they reach the surface as hot springs.

As a final thought on the Geysir area (Great Geysir), it is worth emphasising again the fracture network that allows the flow of geothermal water through rocks in most geothermal fields worldwide is **maintained through earthquakes**. When there are no earthquakes in a geothermal field for some time, the fractures and cavities and contacts (between rock layers and units) and tend to become filled with secondary minerals (zeolites, calcite, quartz, etc.) and block the flow. Thus, for a geothermal field to maintain its fluid transport, to maintain its permeability, earthquakes are needed from time to time, and that is exactly what has been observed over centuries in the Geysir area.

8 Gullfoss

The drive from Geysir to the waterfall Gullfoss along Road 35 is short. The waterfall, which constitutes the **eleventh stop (11)**, is located in Fig. 4.1. The main features to see on the way are the hyaloclastite mountains north of Geysir and, if the visibility is good, the southern part of the ice cap **Langjökul**l (the Long Ice Sheet). However, to really enjoy the glaciers or ice caps, one needs to approach them, and such a trip is beyond the present excursion. We therefore drive on to Gullfoss, the most famous waterfall in Iceland (Fig. 8.1).

8.1 Why Has Gullfoss Two Oblique Steps?

Gullfoss (The Golden Waterfall) is in the glacier river **Hvita** (**Hvítá, White River**). Part of the beauty of Gullfoss lies in the **two main steps** that constitute the waterfall whose total drop (waterfall height) is 32 m (Fig. 8.1). These steps make an (acute) angle of about 60°—and thus a larger (obtuse) angle of about 120° (Fig. 8.2). More specifically, the upper waterfall or step has a direction (trend, strike, azimuth) of about 75° (the angle is always referred to the geographical north), whereas the lower waterfall or step (into the main channel or gorge) has a direction of about 15°. Then main river channel to the southwest of Gullfoss has a general trend or azimuth of about 40°. All these fracture directions are indicated schematically in Fig. 8.2.

The same main directions are seen at other locations southwest along the main river channel, that is, the main channel itself trends about 40° (always referring to azimuth, that is, east of north) and is dissected by fractures with the two other trends, the one at about 15° and the other at about 75°. These fracture trends are also seen in some of the nearby river channels and all over South Iceland. These directions clearly mark systems or sets of earthquake fractures and are easy to explain.

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Fig. 8.1 General overview of the waterfall Gullfoss. View east, the waterfall consists of two main steps, with a total drop (waterfall height) of 32 m. The waterfall is located in the river of Hvita (Fig. 8.2)

Fractures with the trend of about 40°, that is, trending northeast-southwest, characterise all the volcanic systems in the southern half of Iceland, as well as the West and East Volcanic Zones (Fig. 2.2). As we already know from the Reykjanes Peninsula and Thingvellir (Chaps. 2, 5 and 6) the larger fractures with this trend are mostly **normal faults**, and that applies to the fracture forming the main channel southwest of Gullfoss (marked by B in Fig. 8.2). Thus, the main canyon presumably developed along a normal fault zone, containing also tension fractures, which may originally have been partly similar to Almannagja (Figs. 5.1, 5.2 and 5.8).

The other two directions, 15° (marked by C in Fig. 8.2) and 75° (marked by A in Fig. 8.2) coincide with well-known fracture systems that produce earthquakes in entire South Iceland. These fracture systems are faults that are generated (slip, move) in the same stress field as controls the presently active **South Iceland Seismic Zone** (Chaps. 9 and 14). Everywhere in South Iceland there are faults with these two directions: one is north-northeast (about 15° at Gullfoss but somewhat variable), and the other one is east-northeast (about 75° at Gullfoss but also somewhat variable). In contrast to the normal faults at Thingvellir, such as

Almannagja, where the movement of the fault walls is primarily vertical (up and down the fault plane; Figs. 4.14 and 5.9), the movements of the walls of the faults that characterise South Iceland Seismic Zone are primarily horizontal. Such faults are named **strike-slip faults**—San Andreas in California (the United States) is among the most famous examples of such a fault. Beautiful examples of these types of faults can be seen in some of the hyaloclastite mountains in South Iceland, perhaps the best example being in **Vördufell**, which we will discuss at the twelfth stop later today (Chap. 9).

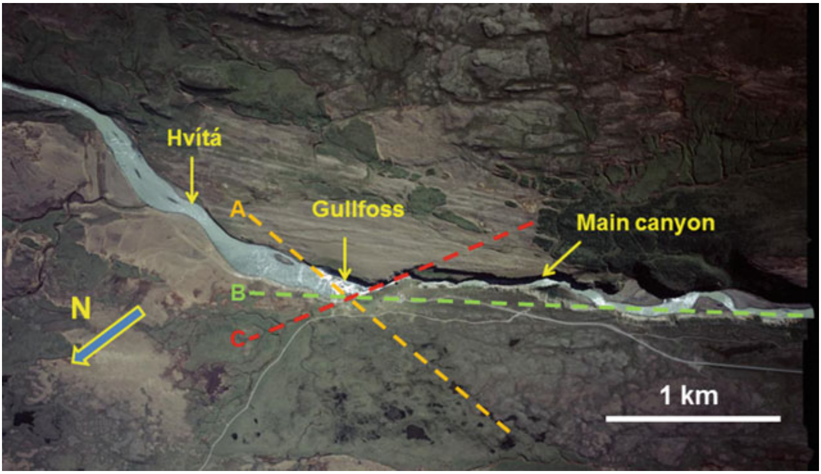

Fig. 8.2 Aerial view of the canyon of the river Hvita (Hvítá), namely Hvitargljufur (Hvítárgljúfur), and Gullfoss. The main steps that constitute Gullfoss have very different orientations, and are also different in orientation from the trend of the main canyon itself. All the three orientations are related to earthquake fractures, that is, faults. The main canyon, running parallel with the broken green line B, relates to a normal fault zone, perhaps originally similar to Almannagja (Chap. 5), and trends about 40°. The upper step, running roughly parallel with the broken orange line A, is related to a fault, and so is the lower step, which runs parallel with the broken red line C. All the faults A, B, and C are typical for South Iceland. A is called a sinistral or left-lateral strike-slip fault, whereas C is called a dextral or right-lateral strike-slip fault. These technicalities need not concern us here, but are mentioned in case you wanted to explore the fault pattern in greater detail—here and in later chapters. The direction of geographic north (N) is indicated with a thick arrow, and so is the scale, that is, the length of 1 km

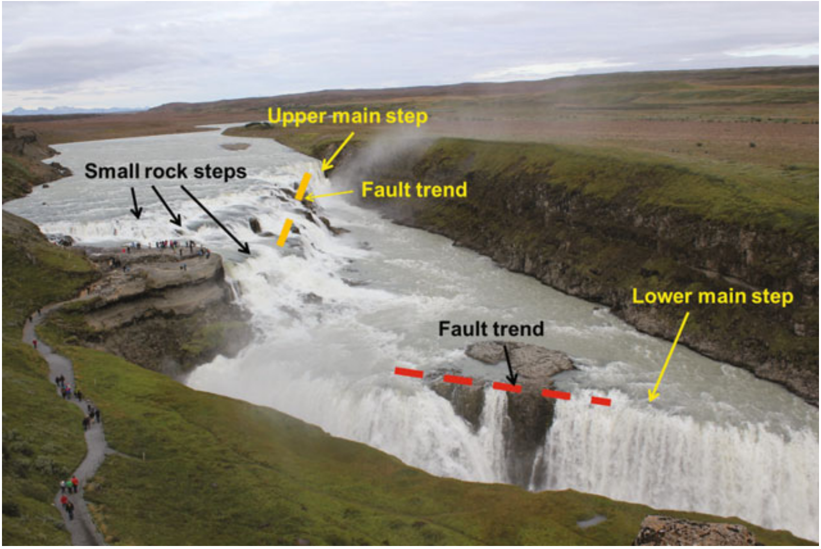

Fig. 8.3 Details of Gullfoss. View east, the main upper step is composed of several smaller rock steps (forming a series of cascades). By contrast, the lower main step is a single one. The upper step is about 11 m in total, whereas the lower step is about 21 m. Also indicated, crudely, are the two fault trends that contribute to the formation of the oblique steps (Fig. 8.2)

So the earthquake fractures offer zones of weakness in the rocks that the water of the river Hvitá can easily erode, forming a major gorge or canyon (Figs. 8.2, 8.3, 8.4 and 8.5). The main canyon (Fig. 8.2) follows the more fractured and thus more easily eroded normal fault zone. It is named for the river as Hvitargljufur (**Hvítárgljúfur**, White River Canyon) has a maximum depth of about 70 m and a length of some 2500 m. The **zig-zag geometry** of the river channel (Fig. 8.2), particularly close to and at Gullfoss, is, as discussed, the consequence of the two other main earthquake fault (strike-slip fault) directions (A and C in Fig. 8.2) that make for fractured rocks and easy erosion.

Fig. 8.4 The part of the canyon that is parallel with the strike-slip fault, indicated by red broken line in Figs. 8.2 and 8.3. View south, at its south end, this canyon joins the main canyon at an acute angle of about 35°

8.2 How Did the Canyon Evolve?

How easily the river **erodes** the rocks and expands the canyon depends on the properties of the rock layers themselves (Figs. 8.1, 8.3 and 8.6). The lower step in the waterfall, and the associated canyon, is primarily composed of a thick basaltic lava flow with numerous columnar or cooling fractures (Figs. 8.4 and 8.5). The upper step, however, is composed of various rock layers. The layers are primarily of two types: basaltic lava flows with columnar joints, and sedimentary layers. The **lava flows** were formed during interglacial (ice free) periods, whereas the **sediments** (rocks formed through erosion and transport of rock particles) were mostly formed during the glacial (ice) periods of the past several hundred thousand years (Chaps. 3 and 4).

How the rock layers respond to the flowing water and its pressure depends on many factors and is not always easy to forecast. We might think that the 'strong' basaltic lava flows would be very resistant to erosion, but that is not necessarily so. This follows because the vertical cooling fractures, **columnar or cooling joints**, make the lava flows weak in response to water pressure from above. The lava flow resistance to erosion also depends on the layer thickness: thin layers with numerous vertical fractures are generally more easily eroded by flowing water than thick layers.

Some sedimentary rock layers are comparatively strong, whereas others are weak; the strength depends on their grain size and other properties. What we see is that the top lava flow has been largely eroded whereas the topmost sedimentary layer (the one the people are standing on in Fig. 8.6) is comparatively strong—and thus forming an overhang. The sedimentary layer below is easily eroded, whereas the lowermost sedimentary layer is comparatively strong. Below that layer is again a lava flow that is comparatively resistant to erosion. This layering of the upper

Fig. 8.5 The same part of the canyon as in Fig. 8.4 but from a different perspective. The canyon is primarily composed of basaltic lava flows, indicated, with numerous vertical columnar joints. These are aa lava flows, and thus formed of a single unit, in contrast with the pahoehoe lava flows seen in the walls of Almannagja, which are composed of many thin flow units (Figs. 5.5 and 5.6). The columnar joints are very well developed in these lava flows and made them comparatively easy to erode by the river

Fig. 8.6 Gullfoss gradually moves inland as the erosion of the steps that constitute the waterfall continues. View east, the upper step, seen here, is composed of rock layers of different composition and strength. 'Strength' here means resistance to erosion. The lava flows are only moderately resistant to river erosion because they contain numerous fractures, columnar or cooling joints (Fig. 8.5), that make the rocks more easily eroded, or 'weaker'. The sedimentary layers are of several types. Depending on the grain size and other factors, some of the layers are comparatively strong or resistant to river erosion, whereas others are weaker or less resistant to erosion, as indicated. The different layers are reflected in the small rock steps that characterise the upper step (Fig. 8.3)

step can be seen in the geometry of the waterfall itself. The layering results in the upper step not being a single one, but rather composed of four to five **small rock steps**, **cascades** (Figs. 8.1 and 8.3), each of which corresponds to one of the layers in Fig. 8.6.

Gradually, however, the layers become eroded and the canyons become longer. The main canyon, Hvitargljufur (Figs. 8.4 and 8.5) and all the structural features associated with Gullfoss itself as seen today must be formed since the ice caps of the last ice period disappeared. This follows because glaciers always tend to change narrow river canyons and valleys into larger U-shaped valleys. This has not happened here—the canyon walls are clearly vertical (Figs. 8.4 and 8.5)—so that the canyon and the entire associated landscape must be younger than the last glaciers in this part of Iceland. This part of Iceland became permanently ice free some 8–9 thousand years ago. If, as is likely, the entire 2500 m long Hvitargljufur was formed in the past 8–9 thousand years, then the rate of growth or expansion of the canyon must have been, on average, about 30 cm per year. And this is the same rate as Gullfoss itself is moving up the canyon. So every year, on average, the waterfall itself moves some 30 cm to the northeast, that is, further inland.

9 Gullfoss-Kerid (Kerið)-Reykjavik

We now drive on towards Reykjavik. There are several roads that can be taken, but it is best to stay on Road 35, that is, to go back to Geysir and then follow Road 35 towards the next main stops (Fig. 4.1). On the way you pass through some of the main agricultural areas of Iceland and, in particular, several geothermal fields. The fields are not open for observation, but the geothermal water is used for space heating—both for buildings in general, and for greenhouses in particular. Almost all these geothermal fields are related to earthquake fractures. Many people like to visit Skalholt **(Skálholt)**—a site of cathedral that in medaeval times was one of the two bishop sites in Iceland (the other one being at Holar in North Iceland), including a school and variuos activities. Skalholt is located at the glacier river Hvita, the river of Gullfoss.

Whether you decide to drive to Skalholt or not, it is worth stopping, as the **twelfth stop** (**12**), somewhere in this area (close to Road 31 which leads to Skalholt) to take a look at two prominent, although not very tall mountains. The one to the north of the road is a small hyaloclastite mountain named **Mosfell** (Figs. 4.1 and 9.1). This moutain is similar to those we saw at Thingvellir and on the way from Thingvellir to Geysir. Mosfell is several hundred thousand years old. The mountain south of the road, and south of Skalholt, is much larger and one of the more proniment moutains in South Iceland. Its name is **Vördufell** (**Vörðufell**), and it is located in Fig. 4.1. Vördufell is considerably older than Mosfell. More specifically, the age of Vördufell is at least 800 thousand years and may be a few million years. The mountain is composed partly of hyaloclastites, and partly of lava flows (Figs. 9.1 and 9.2).

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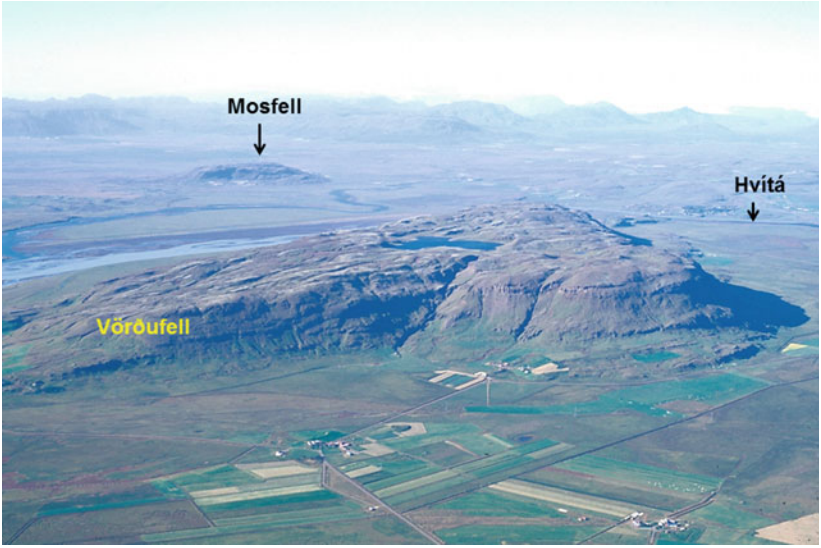

Fig. 9.1 Aerial view of the mountains Vördufell and Mosfell, with the river Hvita in between them. View north, both mountains are primarily hyaloclastite mountains that, to a large degree, are shaped by earthquake faults that offer weakness zones for easy glacier erosion

9.1 Earthquake Faults and Mountain Shapes

Why are these mountains of any particular interest? The answer is that they are dissected by exactly the same types of fractures, earthquake faults, as we saw control the geometry of Gullfoss and associated canyons. As mentioned above, one set of faults is the normal faults of the type we saw at Thingvellir, but the other two different from those at Thingvellir in one fundamental way: the movements of the fault walls are not vertical (up and down along the fault) but rather horizontal and are referred to technically as **strike-slip faults**. For a given fault-wall movement, or **slip**, on a strike-slip fault the earthquake generated (its magnitude) is normally larger than for the same wall movement or slip on a normal fault as in Thingvellir. Indeed, the **largest earthquakes** in Iceland all occur on strike-slip faults, and many occur in the South Iceland Seismic Zone, primarily in the area south of Vördufell.

Not only are the mountains dissected by earthquake faults, but the faults, to a large degree, control the **shapes** of the mountains. This is particularly clear for Vördufell. The mountain shape is that of a triangle or, perhaps more accurately, that of the heart symbol. This is seen in Figs. 9.1 and 9.2, and also in Fig. 4.2. This shape is primarily the result of the interaction of the main fault trends in South Iceland, roughly the same as shaped Gullfoss (Figs. 8.2 and 8.3). I have indicated these main trends schematically in Fig. 9.2. As for Gullfoss, I denote east-northeast-trending faults by A, northeast-trending faults by B, and north-trending faults by C. The faults offer weakesses that are easily eroded through subsequent **glacial erosion** (during the Ice Age).

The direction of faults with these general trends is somewhat variable. For example, the northeast-trending faults B have a direction of about 30° at Vördufell

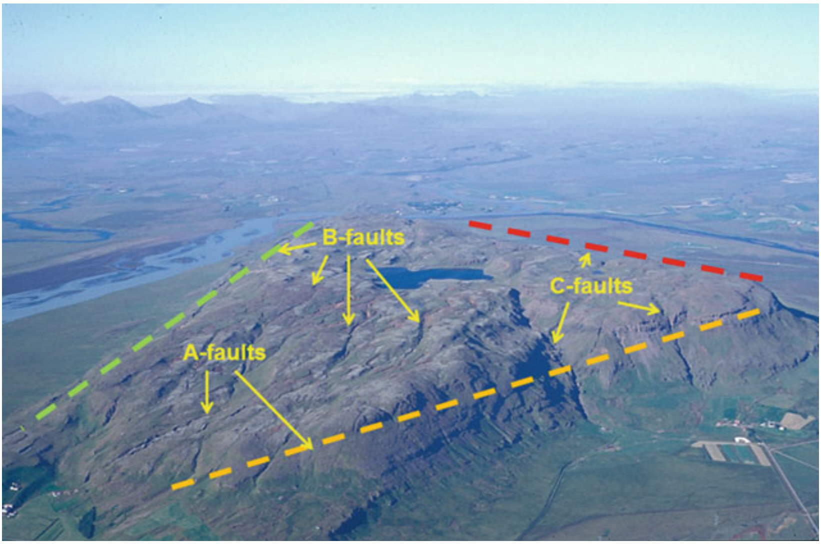

Fig. 9.2 Aerial view of Vördufell showing some of the main earthquake faults that have shaped the mountain. The slopes of the mountain coincide crudely with the three main trends of faults which, when combined with glacial erosion, have given the mountain its triangular or heart-symbol shape. View northeast, the A-faults are east-northeast trending strike-slip faults, B-faults are northeast trending normal faults, and C-faults are north-trending strike-slip faults. The same faults occur throughout South Iceland. The letters used here to identify the faults are the same as used for similar faults at Gullfoss (Fig. 8.2). The A and C faults are the main sets of faults that constitute the South Iceland Seismic Zone (located in Fig. 2.2) and were active during recent strong earthquakes in that zone (in the years 2000 and 2008)

(and the nearby **Hestfjall**, located in Fig. 4.1) but 40° at Gullfoss. The reason is that at Vördufell the direction is controlled by the stress field in the West Volcanic Zone, so that the direction coincides with faults at Thingvellir (trending about 30°) rather than that of the East Volcanic Zone (trending 40°–45° as at Gullfoss). Similarly, the north-trending faults can vary from about 15°, as at Gullfoss, to 0° (true north) or even several degrees west of north, as some of the C-faults at Vördufell. As for the A-faults, they trend mostly between 60° and 70°. There are, in addition, some northwest-trending fractures in Vördufell and South and Southwest Iceland in general, but we leave these out here (they are discussed and illustrated in Chaps. 13 and 14). The same fracture sets occur in Mosfell but are not seen in Fig. 9.1. Other mountains in the vicinity of Vördufell whose shape or geometry is controlled by earthquake faults include Hestfjall and Ingolfsfjall (Fig. 4.1). We discuss Ingolfsfjall further later today.

The north-northeast (or north) trending and the east-northeast trending strike-slip faults are the main directions of earthquake fractures in the young lava flows in large parts of the South Iceland Seismic Zone (located in Fig. 2.2). The main discussion of this seismic zone is in Chap. 14, but here we mention briefly how large the earthquakes on the faults in Vördufell and surrounding areas in South Iceland can be. Based on the lengths of the faults (earthquake magnitudes normally increase with fault-rupture length) and measurements of the slips and descriptions of earthquakes in historical accounts (annals) they can exceed **magnitude 7** (M7, where M stands for magnitude). Earthquakes reaching magnitude 7 and above are classified as **major earthquakes**. About 20 such earthquakes occur every year on Earth. But they are not great earthquakes, whose magnitude is 8 and above. Great earthquakes cause the greatest damage and loss of life and properties, but fortunately occur only once every 5–10 years on Earth as a whole. The seismic zones in Iceland are presumably too small to be able to generate great earthquakes.

Earthquakes of magnitude 7 generally cause much damage of buildings, and commonly loss of life. Collapse of buildings and loss of life has occurred in many of the earthquakes in South Iceland during historical time (the past 1100 years). Since we have the Cathedral at Skalholt in front of us—whether or not you decide to drive up to the Skalholt itself—it is worth mentioning that buildings have been damaged and collapsed at Skalholt in earthquakes. In many of these earthquakes some of the faults that we see in Vördufell (Fig. 9.2) were active and caused damage. In fact, the earthquakes in 1784, estimated to have reached M7.1, generated so much damage at Skalholt and its surroundings, that the episcopal see (the official seat of the bishop) at Skalholt was moved to Reykjavik. We come back to the earthquake zone in South Iceland in Chap. 14.

9.2 Pit Crater

We now drive on towards the west along Road 34 until we come to the **thirteenth stop** (**13**), namely the crater **Kerid** (**Kerið**). Kerid is a beautiful crater (Figs. 9.3 and 9.4) of scoria and spatter, the most common materials to constitute volcanic craters. The rock is basalt. **Scoria** is broken, stiff lava fragments, mostly from millimetres to centimetres in size (diameter). They derive from the spray of the **lava fountains** during fissure eruptions, such as the one that formed Kerid. When the scoria fragments hit the ground the fragments are already solidified and stiff. **Spatter** is also derived from the spray of the lava fountain, but the difference is that when the clots hit the ground, they are still hot and plastic. Consequently, the spatter clots commonly combine into larger flattened 'pancake-like' masses—which, eventually, may flow for a while. The scoria is either black or red (Figs. 9.3 and 9.4). The black colour is the most common. The red colour is due to oxidation whereby oxygen in the atmosphere combines with iron in the rock and changes the colour to red.

Fig. 9.3 The pit crater Kerid. Red scoria, black spatter, and a welded layer are indicated. The depth of the lake (groundwater) is about 10 m. People in the lower right corner provide a scale

Fig. 9.4 Kerid and its surroundings. View northeast, the maximum diameter of Kerid is elliptical in plan view. The long axis of the ellipical crater is in the northeast direction and is about 300 m, while the short axis is about 170 m and the depth some 50 m

As a crater cone builds up rapidly due to the rain of scoria and spatter from the lava fountains, much of the heat (from the spatter and scoria) is trapped inside the cone. Heat conduction in rocks is very slow, so the heat cannot escape rapidly and the temperature rises and may reach the melting temperature of the rock. When the melting temperature is reached, part of the scoria/spatter may re-melt and flow as lava, or rather like an intrusion, inside the crater cone. The process is referred to as **welding** and results in lava-like (or intrusion-like, sill-like) welded layers inside crater cones. Many welded layers are seen in Kerid (Figs. 9.3 and 9.4; cf. Fig. 9.9).

Kerid is a part of a larger (but still very small) volcanic field, referred to as the **Grimsnes (Grímsnes) Volcanic Field**. Some of the crater cones that constitute that field are seen as small hills in Fig. 9.4. The entire field was generated in many eruptions that occurred some 8–10 thousand years ago. All the crater cones where formed by fissure eruptions. As is common for monogenic (single-eruption) crater cones, the fissures supplying magma to the eruptions are comparatively short. The longest fissure is about one kilometre, and most are with lengths of several hundred metres. Some of the crater cones are large, standing many tens of metres above their surroundings, in contrast with Kerid, which is primarily a depression, formed in an eruption some 9 thousand years ago. Kerid, together with several nearby crater cones, was supplied with magma from a north-northeast trending volcanic fissure, about 900 m long. This fissure has the same direction (exactly 29°) as many of the volcanic fissures in the nearby Hengill Volcanic System, a system we discuss in greater detail in Chap. 12.

So how did Kerid form? Why is it so different from many of the nearby craters? Let me first say what **it is not**. Kerid is definitely not an explosion crater, that is, a depression formed by explosion when hot magma comes into contact with groundwater. Explosion craters of this type are also called **maars**, but they produce very little if any eruptive materials (which, by contrast, Kerid did) and have an overall form different from that of Kerid. Maars are common in Iceland: we will see good examples in Chap. 13. Kerid is also definitely not a collapse **caldera**, as it is commonly thought to be. Collapse calderas are very common in Iceland, and we have already discussed some in Chap. 4. Most collapse calderas are circular to somewhat elliptical in plan view. By definition, the diameter of the depression must reach at least 1 km so as to be classified as a collapse caldera. Calderas are also known to be related to pressure changes in shallow magma chambers (Chap. 4). By contrast, Kerid is elliptical in plan view, with a major axis (the long axis) of the ellipse of about 300 m and the minor axis (the short axis) of about 170 m, and a depth of about 50 m. And there is certainly no evidence that Kerid is related to a shallow magma chamber. Kerid is, however, clearly generated by a volcanic fissure and its feeder-dike (the dike that fed the eruption), and the formation of the Kerid relates to events in the feeder-dike.

How? During the end stages of an eruption, the magmatic overpressure (the pressure driving the magma up to the surface) in the feeder-dike normally decreases. This follows because there is less magma coming from the source (here a deep reservoir, Fig. 2.4). When the overpressure in the feeder-dike decreses gradually, nothing much happens, and this is the most common style of fissure eruption. The crater cones end their eruptive activities one after the other, until usually one is left which then finally fades away as well. When the overpressure decreases abruptly, however, one or more of the crater cones may **collapse**. Abrupt decrease in overpressure may happen when the magma finds a new path, commonly to a crater located at a lower surface elevation. Abrupt decrease may also happen when the dike connection with the magma reservoir at depth closes. Either way, the abrupt decrease in the magmatic overpressure leads to a sudden decrease in the height of the magma column below the crater, and that is most likely the reason for the depression that characterises Kerid (Figs. 9.3 and 9.4).

More specifically, there was presumably a standing magma column in the conduit that supplied magma to Kerid. Such a standing magma column, when small, is referred to as a **lava pond** (when large it is referred to as a **lava lake**). When the overpressure in the dike feeding the magma column in the crater suddenly dropped, then the pond dried out, leaving the depression that we know as Kerid. The process is thus essentially that of the formation of a **pit crater**, a very common feature along volcanic fissures. Since the depression extends below the **water table**, that is, the level where the rock is satuated with groundwater (Chap. 5), the lower part of the depression is filled with groundwater (Figs. 9.3 and 9.4). The water level fluctuates somewhat but the lake is normally about 10 m deep (the range is from 7 to 14 m). The reason for the existence of the lake within Kerid is thus basically the same as the reason for water filling the fractures at Thingvellir and for Lake Thingvellir itself (Chap. 5): the water level in both marks the water table, the surface of the groundwater.

9.3 Rockfalls and Earthquakes

From Kerid we continue our drive along Road 35 towards the mountain **Ingolfsfjall** (**Ingólfsfjall**), located in Fig. 4.1. This is a hyaloclastite mountain (Fig. 9.5), basically a table mountain, similar to the one we described in Chap. 6 (Hrafnabjörg). There is evidence, however, that part of Ingolfsfjall was formed during eruption in the sea, rather than beneath an ice cap. Its age is several hundred thousand years. The most remarkable features seen on the eastern slopes of the mountain are an enormous number of stones, that is, **boulders** of various sizes (a small summerhouse is located in-between some of the largest boulders). In fact, the eastern slope itself, trending roughly north-south (Fig. 4.1), is partly controlled by an **earthquake fault** of type C (Fig. 9.2).

Many, and perhaps most, of these boulders are the consequences of **rockfalls** and **landslides** generated during earthquakes. Many of the strong to major earthquakes in South Iceland have occurred close to Ingolfsfjall. The shape or geometry of the mountain is controlled by faults. Thus, not only the east slope but also the west slope of the mountain coincide with north-trending **C-faults** (Fig. 4.1). C-faults have also propagated through the mountain; some of these are indicated in Fig. 9.5. The other main types of earthquake faults (A and B) also dissect Ingolfsfjall and have similar directions as in Vördufell and elsewhere in South Iceland. The most recent earthquakes to cause rockfalls in Ingolfsfjall were the

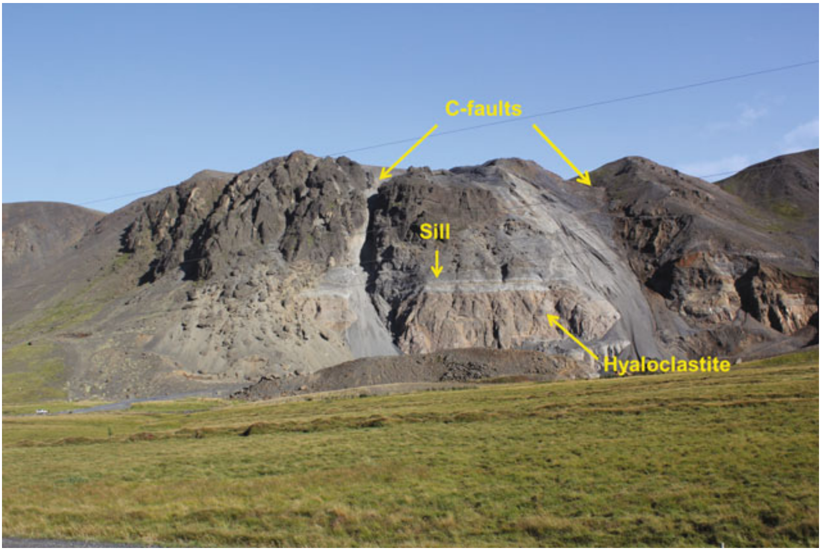

Fig. 9.5 View north of the south slopes of the mountain Ingolfsfjall (located in Fig. 4.1). The mountain is primarily of hyaloclastite, with many intrusions, such as sills. Ingolfsfjall is dissected by earthquake faults, particularly north-trending C-faults, two of which are indicated here

M6.5 earthquakes in June 2000 (Chap. 14), and the M6.3 earthquake in May 2008, both of which were primarily on C-faults, but also with rupture along A-faults. In fact, the north and south slopes of Ingolfsfjall have east-northeast trends and are partly controlled by **A-faults**. Earthquakes are also very common in the nearby central volcano Hengill, as discussed in Chap. 12.

We now drive along Road 35 to the south until it meets Road 1, the Ringroad, which we follow to Reykjavik. In the south slopes of Ingolsfjall there are many depressions and small gullies that are originally earthquake faults. Some of the C-faults can be seen above the quarry (Fig. 9.5). The quarry also shows the internal structure of the mountain, which is primarily of hyaloclastite (tuff) but also with many intrusions, such as sills. As we drive along Road 1, we soon have the town of Hveragerdi (located in Fig. 4.1) on the right hand side (to the north of the road), where there are many geothermal fields which are used for space heating of greenhouses. Hveragerdi is also the town in Iceland where earthquakes are probably most common. Partly because of its location within the South Iceland Seismic Zone, and partly because earthquakes are so common in the nearby volcano, Hengill (Chap. 12).

9.4 Lava Erupted in the Year 1000

As Road 1 climbs up the so-called Kambar, we have the Hengill Volcano to the north (located in Fig. 4.1). The highland area that we enter now is the central part of the West Volcanic Zone, and more specifically the Hengill Volcanic System (Fig. 2.3)—the same system to which the Thingvellir Graben belongs. The normal faults and grabens extend all the way from the northern part of Thingvellir south to Road 1 here, some 60–70 km. The normal faults you see on the right-hand side (north of the road) all belong to the Hengill Volcanic system (Fig. 2.3). However, we see these and similar faults much better when we visit Hengill itself (Chap. 12), so that we do not make a specific stop to look at the faults here. The highland area itself is named **Hellisheidi** (**Hellisheiði**), and the reason for its being a highland is the high volcanic productivity in this part of the Hengill Volcanic System. Most of the lava flows on the highland itself are several thousand years old, and some of the volcanic fissures are easily seen to the right-hand side or north of the road.

As we drive down from the highland of Hellisheidi, we see geothermal fields to the north or right of the road. Since we have already seen the Geysir area, and can see larger geothermal fields later (Chaps. 12 and 13), we do not make a stop here to look at these fields. In front of us, however, is a famous lava field, with a very clear contact between a younger lava flow on the top and an older one beneath (Figs. 9.6 and 9.7). It is worth making a stop here, **the fourteenth stop** (14), at an appropriate parking place. The older lava flow is named Svinahraun (**Svínahraun**, **Pig Lava**) and is about 5200 years old (Fig. 9.7). It is a part of a larger lava flow named **Leitarhraun** and which flowed all the way into Reykjavik (Chap. 3), and which we will discuss again at the last stop before we come to Reykjavik.

The younger lava flow, on the top (Figs. 9.6 and 9.7), has also two names and is historically famous. It is know as **Svinahraunsbruni** (**Svínahraunsbruni**, **The Burning Pig Lava**) but also as **Kristnitökuhraun** (literally 'The lava (formed) during the taking of (adopting) Christianity'). Icelanders became officially Christianised in the year 1000. Before that they believed in Norse (Nordic, Skandinavian) gods, such as Odin (Óðinn) and Thor (Þór). While the case was being debated in the parliament at Thingvellir (sixth location in Fig. 5.1; Chap. 5), as to whether or not Christianity should become the official religion of the nation instead of the old religion (also known as Asatru or Odinism), a man came running to Thingvellir to tell people that there was an eruption in Svinahraun. Those who did not want to adopt Christianity argued that the eruption indicated that the pagan gods were angry because of the proposal of changing the official religion to Christianity. But one of the chieftains then asked the question: what made the gods angry when the lava flow on which we stand was formed? He was there referring to a 9–10 thousand-year-old lava flow, the Thingvellir lava (Chap. 5). The remarkable thing is that Icelanders had by then understood that all these comparatively old rocks were lava flows formed in eruptions of the type they had seen since the settlement in 874.

Thus the argument as to the wrath of the Norse gods being reflected in an eruption was regarded as invalid and, by law, Christianity became the official religion of Icelanders. Svinahraunsbruni or Kristnitökuhraun has been dated to the year 1000 which, according to the written documents, is exactly the year when the debate about the different religions took place at Thingvellir.

Fig. 9.6 View west, Road 1 passes here into the lava flow Svinahraunsbruni, which is also known as Kristnitökuhraun. The margin of the aa lava flow is seen here. The lava erupted in the year 1000, at the time when, at Thingvellir, it was decided that Christianity should become the official religion of Icelanders

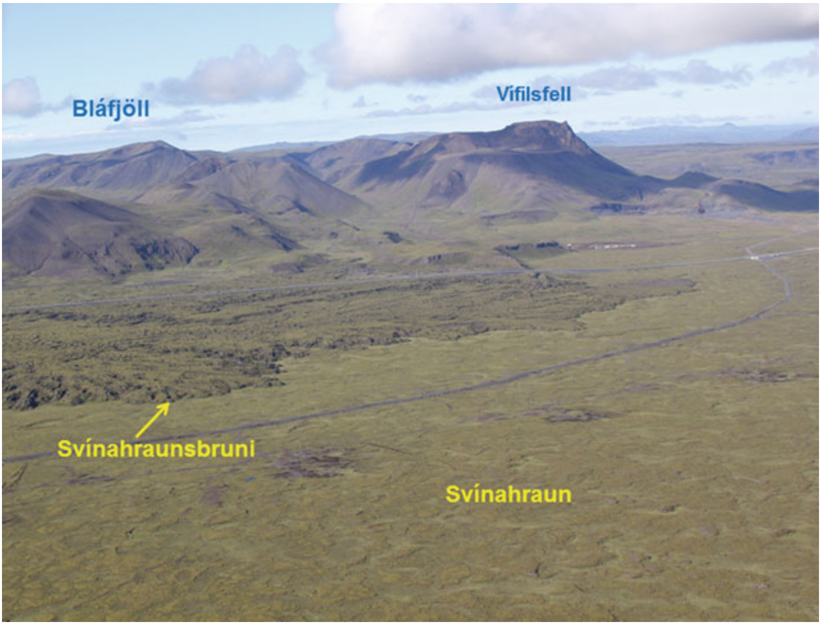

Fig. 9.7 Aerial view of part of Svinahraunsbruni, also known as Kristnitökuhraun, partly covering the older lava flow Svinahraun. View southwest, Road 1 is close to the mountains, whose general name is Blafjöll, a part of which is the mountain Vifilsfell

If you are interested in the use of geothermal energy, then it is worth your while to drive the short distance along Road 378 north to the **Hellisheidi** (**Hellisheiði**) **geothermal power plant**. The power plant is one of the largest of its kind in the world, with a capacity for electricity production of about 300 MW (mega-watts). The power plant is open to tourists and there are free video and slide shows on the power plant itself, how it operates, on geothermal energy, and on the general geology of Iceland.

9.5 The Youngest Lava Flow to Have Entered Reykjavik

Back to Road 1, we continue on the way to Reykjavik. The road crosses the 1000-year-old Svinahraunsbruni/Kristnitökuhraun, with hyaloclastite mountains to the south of the road, and the old lava shield Mosfellsheidi (Mosfellsheiði)—which we crossed on our way to Thingvellir early in the tip—in the distant north. Further to the north is the mountain Esja, which was described in detail in Chap. 4. Road 1 is inside the young Svinahraunsbruni/Kristnitökuhraun only for a short while, and for the rest of the drive all the way to the last stop before entering Reykjavik, the road is along a very narrow **lava stream** of the **Leitarhraun** lava flow. A very narrow lava stream, mostly 500–1000 m wide, but in places as narrow as 100 m, from Leitarhraun flowed to the northwest all the way to Reykjavik (Ellidaardalur) some 5200 years ago (Fig. 3.7; Chap. 3). In this area, that lava stream is used as a foundation for Road 1 all the way to Reykjavik.

Our final stop today, the **fifteenth stop** (15), is where that narrow lava stream expanded somewhat as it met with a shallow lake or wetland and formed crater cones referred to as **Raudholar** (**Rauðhólar**, **Red Hillocks**). Raudholar (Figs. 9.8 and 9.9) are the type of crater cones referred to as **pseudocraters** or **rootless craters**. They are different from ordinary crater cones (spatter and scoria or cinder cones) in that no magma is issued from them—their material is entirely derived

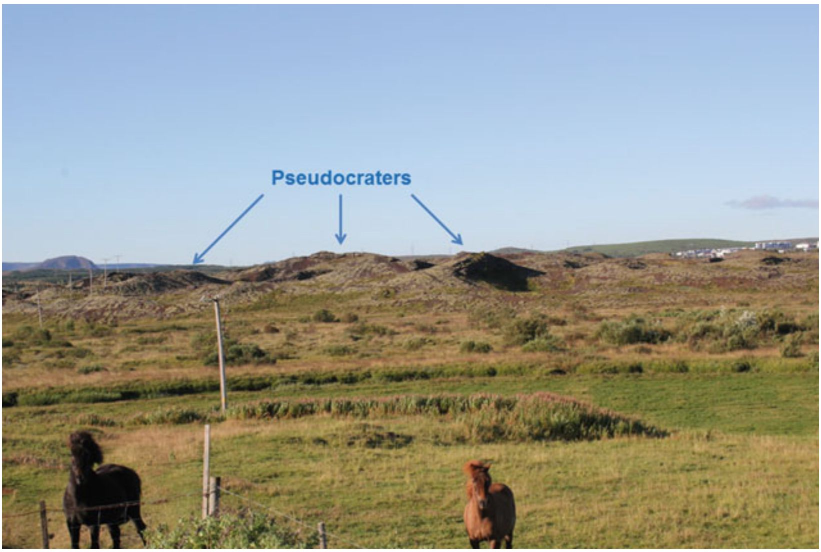

Fig. 9.8 View south, part of the field of pseudocraters (rootless craters) known as Raudholar. The lava flow within which the craters are located is about 5200 years old and flowed into the present Reykjavik through the valley of Ellidaardalur (Fig. 3.7). The horses provide a scale

Fig. 9.9 View west, one of the pseudocraters of Raudholar. Red scoria and welded layers characterise the craters. The scoria has been used for many years as foundation material for road construction, including scoria from the present crater. What you see is thus just what remains of the original crater

from the lava flow in which they occur. More specifically, on contact between the lava, at a temperature of 1100–1200 °C, and the cold water, the water is transformed at an instant into rapidly expanding vapour, resulting in hydromagmatic explosions. The resulting hillocks are what we call pseudocraters or rootless craters, of which Raudholar are fine examples. As we discussed in connection with Kerid, the red colour from which the Raudholar derive their name is attributable to **oxidation**, the result of oxygen in the atmosphere combining with iron in the rock.

On arriving in Reykjavik we have completed the classical ‘Golden Circle’. We have seen many aspects of the geological processes that operate continuously in Iceland and the resulting landforms and structures. With reference to the Golden Circle, as well as the trip from Keflavik to Reykjavik and Reykjavik itself, the landforms and structures we have seen so far include as follows:

- The internal structures of old and deeply eroded volcanoes, including the tops of their fossil magma chambers (Chap. 4)
- Volcanoes formed in eruptions beneath ice sheets/caps (Chaps. 2, 4–6)
- Young lava flows, both those with rough surface (aa) and smooth surface (pahoehoe) as well as pillow lavas (Chaps. 2, 3, 5–7)
- Large (and small) fractures where the gigantic tectonic plates that constitute the surface of the Earth are being pulled apart (Chaps. 2, 5, and 6)
- One of the largest groundwater reservoirs in Iceland (Chap. 5)
- Erupting hot springs or geysers (Chap. 7)
- The second most powerful waterfall in Iceland and its canyon, whose direction is partly controlled by earthquake fractures (Chap. 8)
- Mountains whose geometries are largely controlled by earthquake fractures (Chap. 9)
- A remarkable collapsed crater, with a lake inside (Chap. 9)
- Pseudocraters or root-less craters (Chap. 9).

The Golden Circle has such a great wealth of beautiful geological structures and form, reflecting nature's forces and physical processes, that it may contain as much of geology as you could wish to explore in Iceland. For short visits to Reykjavik, the Golden Circle is an ideal excursion for exploring geological processes and landforms. The Circle is normally done in one day, as I have assumed here. But if you stay longer in the country, or visit the Reykjavik area often and like to explore other geological and landscape beauties, there are many additional one-day trips that can be made from Reykjavik. These trips show exciting features and geological processes in action, and offer great photo opportunities.

Other One-Day Geological Excursions from Reykjavik 10

In addition to those seen on the Golden Circle, there are many geological and landscape wonders and beauties to explore in the vicinity of Reykjavik. Where you go depends on what you want to see. If you want to see beautiful mountains and deep valleys, or go hiking, then the fjord **Hvalfjördur** (**Hvalfjörður)** just north of Reykjavik would be ideal. If you want to explore exotic landscape of narrow valleys and ridges with young lava flows and crater cones (some forming islands in a lake), then the volcano **Hengill** east of Reykjavik should be of interest. If you like lakes, geothermal fields, explosion craters, lava fields, and sea cliffs, then the lake **Kleifarvatn** and its surroundings as well as **Reykjanesta (Reykjanestá)**, the southwestern 'toe' or tip of the Reykjanes Peninsula, have much to offer you. And if you would like to see the most famous volcanoes in Iceland, as well as the most beautiful rock columns in Iceland, then that is exactly what a trip to **Eyjafjallajökull** (**Eyjafjallajökull**) and **Reynisfjara** offers.

I have selected 4 trips to give an indication of landscapes and geological processes and features in the vicinity of Reykjavik that cover the above topics and are most certainly worth visiting (Fig. 10.1). Some of the highlights of these trips are indicated in Fig. 1.6. These are all one-day trips, and some can be made in half a day, depending on how much you want to see and explore and how often you like to stop. As indicated above, these excursions cover a variety of geological processes, structures, and landscape forms and features. The described excursions are the following:

- Reykjavik-Hvalfjördur
- Reykjavik-Hengill
- Reykjavik-Kleifarvatn-Reykjanes
- Reykjavik-Eyjafjallajökull-Reynisfjara.

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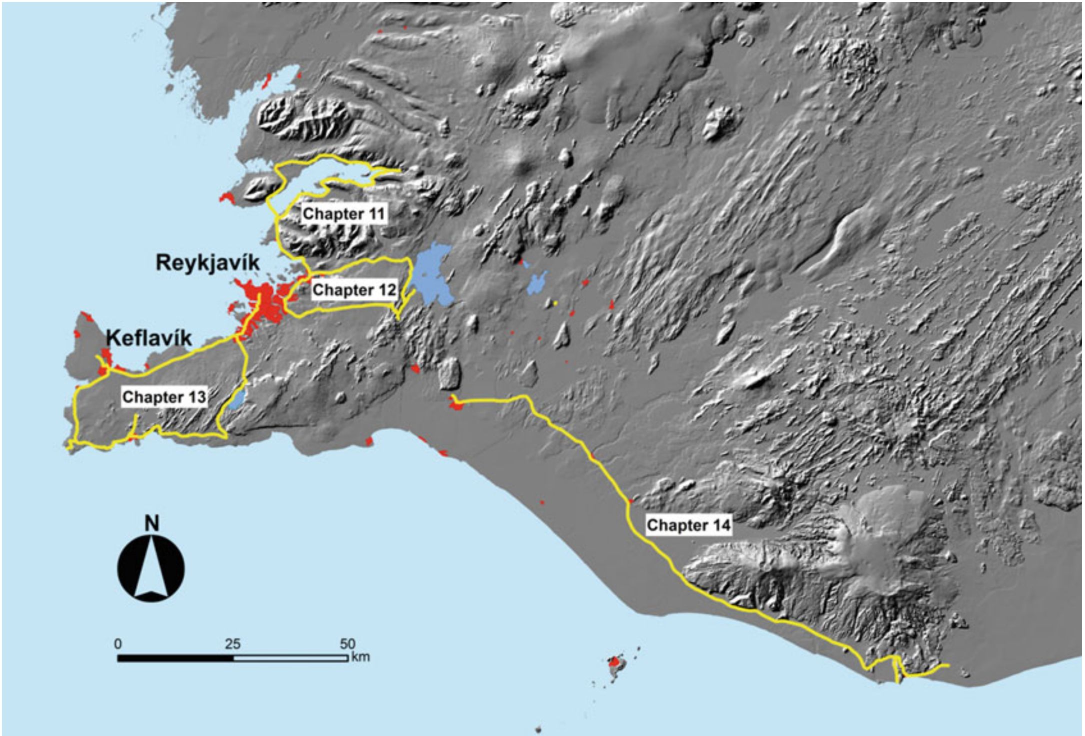

Fig. 10.1 Four excursions, in addition to the Golden Circle, are discussed in Chaps. 11–14. These are indicated here and are to Hvalfjördur (Chap. 11), Hengill (Chap. 12). Kleifarvatn and Reykjanesta (Chap. 13), and Eyjafjallajökull and Reynisfjara (Chap. 14)

I shall now describe some of the main geological structures, landforms, and processes that can be seen on these excursions. These are all much less travelled than the Golden Circle. In particular, Hvalfjördur and Hengill do not receive anything like the same number of visitors as the Golden Circle. Thus, these sites, especially Hvalfjördur, offer the chance of observing some of the beauties of Iceland's nature in the tranquillity that can only be reached away from the main traffic. I should mention that Chaps. 11 and 13, that is, Reykjavik-Hvalfjördur and Reykjavik-Kleifarvatn-Reykjanes, are somewhat **more** '**geological**' than other chapters in the book. This is because I discuss in greater detail some geological processes for which more technical terms and concepts are needed than in the other chapters.

Reykjavik-Hvalfjördur (Hvalfjörður) 11

During the first part of the excursion, we drive the same way as we did during the beginning of the Golden Circle drive (Fig. 11.1). That is, we drive from Reykjavik through the town of Mosfellsbaer (Mosfellsbær). Where Road 1 meets Road 36 (east to Thingvellir) we continue to the north along Road 1 towards the mountain Esja. We have already discussed the main geological and landscape features of the south slopes of Esja, including the flat top and clear rockslide (Figs. 4.3 and 4.4). The best views of the slopes in front of you now are those seen in Figs. 4.3a, b. An aerial view of the part of Road 1 that you drive towards Esja is seen in Fig. 4.2.

As the road turns to the west (nearby the Forestry Research Institute), it passes through rocks with various colours. These are mainly intrusive rocks, that is, rocks formed at great depths in the **Videy** (**Viðey**) **Volcano**, which was discussed in Chap. 3, and partly related to the fossil shallow magma chamber of that volcano. (Notice that the caldera shown in Fig. 3.1 includes only part of the volcano—the diameter of the volcano was larger than that of its caldera.) The variety in colour is partly because there used to be geothermal activity associated with the volcano—an activity whose heat sources were the shallow chamber and associated dikes and sills—and the associated geothermal water has altered the rock (changed its composition) so as to change its colour. A similar alteration of the rock colour is seen in the Geysir geothermal field (Chap. 7), in geothermal fields in general (Chap. 13), and also at the first stop today (Fig. 11.2). Since we have already discussed a shallow magma chamber and related intrusions in Chap. 4 (Stardals-hnjukar), we drive on and continue to the north along the western slopes of Esja.

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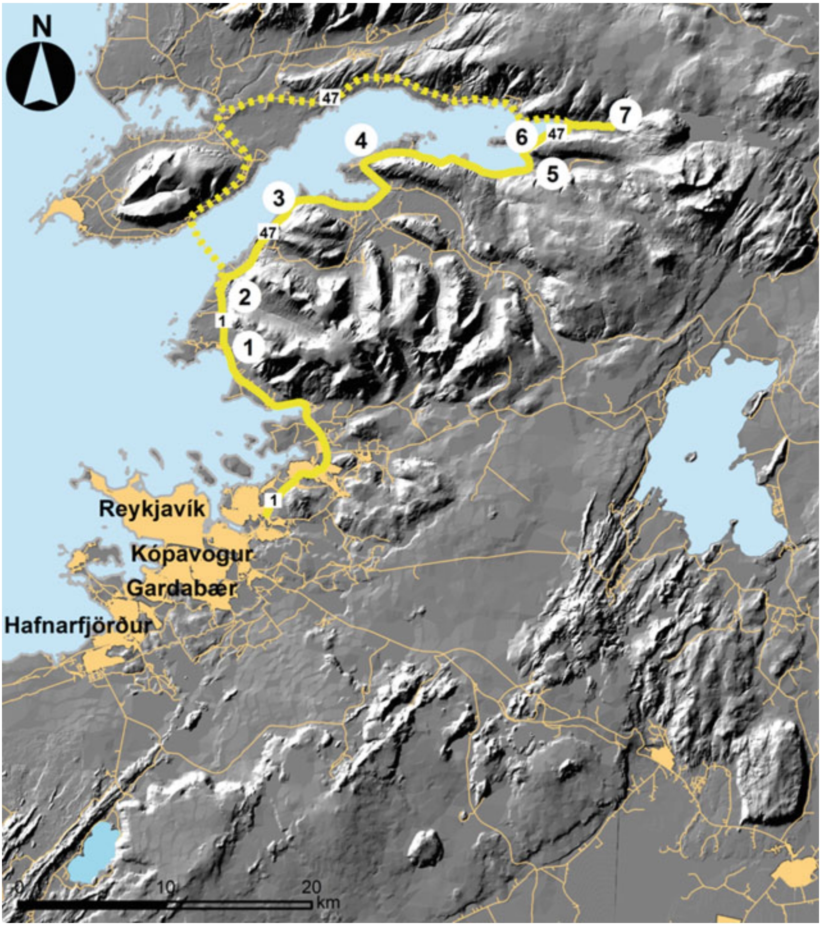

Fig. 11.1 Map showing the excursion to the fjord Hvalfjördur. The suggested main stops are indicated by encircled numbers, 1–7

11.1 Magma Transport Inside a Volcano

The **first stop** (**1**) is to indicate how magma is transported inside volcanoes. The type of transport seen here will be compared and contrasted with that seen at the next two stops. Because you have to cross the road to park in a safe place, you may decide not to stop here and drive straight to the second stop. Also the features I want to show you here are seen really well only in suitable sunshine. But if you have the time, and the visibility is good, it is worthwhile to have a look at these structures (Fig. 11.2).

What we see in the slopes of Esja are: (1) Horizontal layers on the top, (2) light-coloured rocks (light grey and bluish) in-between inclined layers, and (3) inclined rock layers within the light-coloured rocks. The structures and rocks we see here typical for the interiors of volcanoes. More specifically, what we are seeing here is the uppermost part of one of the old volcanoes that formed Esja. So let us now explain these three items with reference to Fig. 11.2.

First, the horizontal layers at the top are all basaltic lava flows. These are the same types of lava flows as we saw in the south side of Esja from a distance in Chap. 4. They are not hundreds of metres thick, as the pahoehoe (smooth surface) lava flows at Thingvellir, but rather from several metres to, maximum, ten or twenty metres thick aa lava flows (with rough surface) and thus similar to the lava flow of Svinafellsbruni (Figs. 9.6 and 9.7).

Second, the light-coloured rocks are mostly hyaloclastites, similar to those we have seen in numerous mountains before. In particular, we had a close-up view of

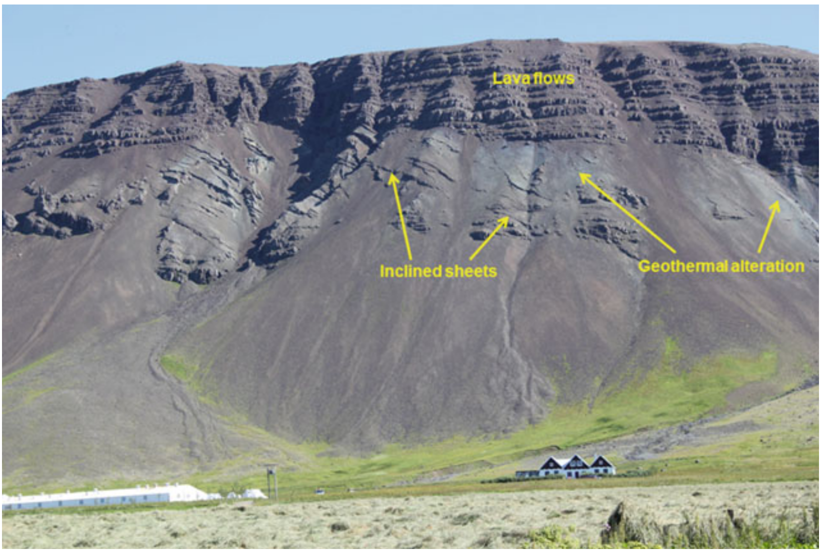

Fig. 11.2 Lava flows, inclined sheets, and geothermal (hydrothermal) alteration on the west slopes of the mountain Esja. View east, the rock layers seen here are a part of the interior of an extinct central volcano who contributed to the formation of Esja

the hyaloclastites in Chap. 6 at the caves (Figs. 6.9, 6.10 and 6.11). So why are the hyaloclastites here then not brown or black in colour? Because the rocks have been altered or chemically changed through the circulation of geothermal water. So we are here seeing at depth what we see at the surface in the Geysir area (Chap. 7) and in even greater detail at Kleifarvatn (Chap. 13). The hot geothermal water gradually changes the chemistry, and hence the colour, of the rocks through which it circulates. And the heat sources for the geothermal fluids are partly the inclined layers, to which we come now.

Third, the inclined layers are not lava flows at all. Rather, they are intrusions, that is, magma-filled fractures that subsequently became solid, solidified or 'froze', as the magma cooled down. Magma-filled fractures, or sheet-like intrusions, of this kind are of three basic types based primarily on their inclination (Fig. 11.3):

- **dikes**, which tend to be vertical, or close to vertical (as you will see in a moment)
- **sills**, which tend to be horizontal, or close to horizontal
- **inclined sheets** (sometimes called cone sheets) which, as the name implies, are neither vertical nor horizontal but inclined (the inclination varies but is commonly 30°–60°).

The inclined layers you are seeing are **inclined sheets**. They tend to be thin, commonly around one metre, and thus much thinner than the horizontal lava flows in the top of the mountain (we saw one sheet from a distance in Fig. 4.4b). One of the remarkable things about the inclined sheets is that they are found only close to shallow magma chambers (Fig. 11.3). The reason is that the shallow chamber, which is basically a cavity filled with hot magma under pressure, modifies or changes the normal stress field—namely, the stress field found elsewhere in the rift zone, in the regional dike swarms far away from magma chambers (Fig. 11.3). The modification is rather simple: the paths along which the magma-filled fractures propagate must meet the boundary of the chamber at right angles (90°). This means that for spherical or flat ellipsoidal magma chambers, most injected magma-filled fractures (except from the very top of the chamber) are not vertical but rather inclined (Fig. 11.3). In other words, they are inclined sheets of the type we saw so clearly in Fig. 11.2.

Where are then the horizontal sills and the vertical dikes? Both are seen in many places in Iceland. In fact, many and perhaps most of the shallow magma chambers, such as we saw in Stardalshnukar (Fig. 4.9; Chap. 4) are sills, either single thick ones, or formed by many thinner sills that accumulate to form the chamber. As for

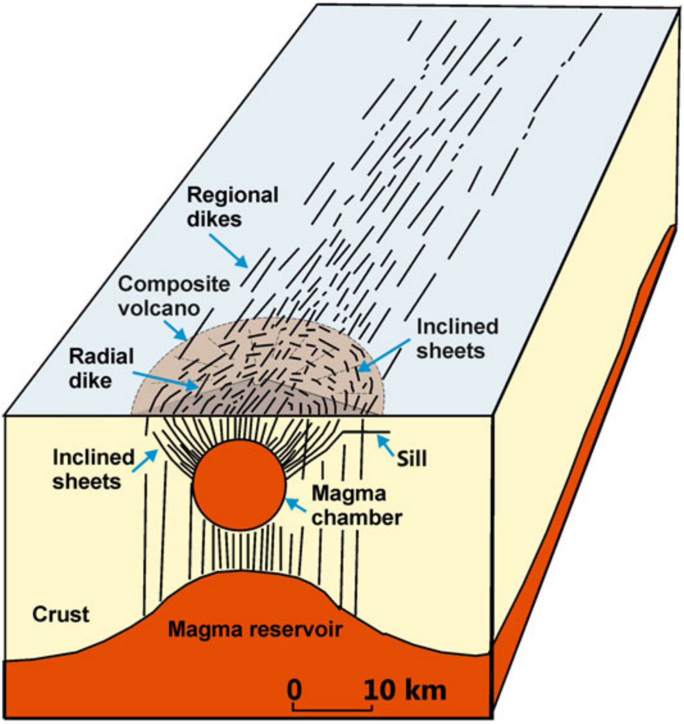

Fig. 11.3 Internal structure of a typical rift-zone volcanic system in Iceland (see also Figs. 2.2, 2.3 and 2.4). The volcanic system shown here has, in addition to a deep-seated reservoir, a shallow magma chamber which supplies magma to the central volcano (here a composite volcano/stratovolcano). The composite volcano is mainly supplied with magma from thin inclined sheets and radial dikes injected from the shallow chamber, whereas the eruptions outside the central volcano are primarily supplied with magma through much thicker regional dikes. Most dikes and inclined sheets do not reach the surface to erupt but stop, become arrested, at contacts between dissimilar layers at some depth—some deflecting into sills at these contacts. Figure 11.2 shows the shallow part of the crustal segment close to and within the composite volcano. Figures 11.4 and 11.5 show the shallow part of the volcanic system that is just outside the composite volcano, that is, a part of the swarm of regional dikes

the vertical dikes—they are the structures we see at the next stops where, again, we refer to the details of Fig. 11.3.

After the first stop, we drive again along Road 1 to the north until we come to next cross-roads, namely where Road 1 meets Road 47. Road 1 continues to the undersea Hvalfjördur tunnel. However, we take Road 47 into the fjord, but first

Fig. 11.4 Regional dikes, like here, tend to be close to vertical (subvertical) and much thicker than inclined sheets (Fig. 11.2). View northeast, regional dikes

make a stop, the **second stop** (**2**) at the parking place to look at the very conspicuous vertical dikes seen there in the cliffs (Fig. 11.4). Some of the dikes are injected radially from the shallow chambers, but most, particularly the thick ones, some of which we see later (third stop), are injected from very large and deep-seated reservoirs (Fig. 11.3). The dikes from the deep-seated reservoirs, such as occur below large parts of the rift zone in Iceland (Figs. 2.4 and 4.10), follow the stress field of the rift zone. Since that stress field is primarily related to horizontal opening or spreading across the rift zone (Figs. 4.5 and 5.11), the dikes tend to be oriented perpendicular or at right angles to the spreading direction, and are therefore vertical.

11.2 Volcanic Systems and Dike Swarms

These dikes, referred to as **regional dikes**, form elongated swarms of similar dimensions as the **volcanic systems** of the rift zones (Figs. 2.2, 2.3, 2.4 and 11.3). As indicated earlier, **volcanic systems** are the sites of the main volcanic and tectonic (fracture formation) activity in the volcanic zones. Generally, in Iceland, the volcanic systems are from 20 to 190 km long and from 7 to 38 km wide (Figs. 2.2 and 2.3). Active volcanic systems have tension fractures and normal faults at the surface, as well as some volcanic fissures (supplied with magma by feeder-dikes). Some have also lava shields. You have already seen these structures in the earlier chapters. What you see now in Hvalfjördur is what these volcanic systems—including the fissures and fractures and lava shields—look at crustal depths from a few hundred metres (the highest peaks of Esja are only a few hundred metres below the original, now eroded, top of the lava pile) to about 1200 m below the original top of the active volcanic systems. At sea level in Hvalfjördur a 1200-metres-thick part of the crust, the lava pile, has been **eroded away** by the glaciers of the last Ice Age. And it is this erosion that allows us to really see the internal structure of the volcanic systems—see them in three dimensions—and thus how they operate.

So how do they operate? Basically, the magma-filled frozen fractures, the **regional dikes**, we see here come straight from the deep-seated reservoirs—perhaps from depths of 15–20 km—whereas the **inclined sheets** we saw in Fig. 11.2 come from shallow magma chambers at the depths of 1–5 km. This is illustrated in Fig. 11.3. Of course these depths may vary—for example, in central Iceland the reservoirs may be at greater depths than 20 km because there the crust is very thick—but the indicated depths are the most common ones. So the regional dikes come from great depths and bring very hot magma (commonly at 1200–1300 °C) to the surface to form nearly all the pahoehoe lava flows, including all the lava shields, as well as many of the basaltic aa lava flows—particularly the large ones such as the famous Laki lava flow which erupted in Southeast Iceland in 1783. The inclined sheets, by contrast, bring magma from the shallow chambers, both basaltic or mafic magma as well as felsic magma (more silica rich, such as rhyolite).

To bring the differences between regional dikes and inclined in recent eruptions into focus, I mention here two recent eruptions that are known to many. One is the Eyjafjallajökull eruption of 2010 (Chap. 14), which put a halt to air traffic (because of ash clouds) in a large part of Western Europe for many days. The main part of this eruption was supplied with magma by an **inclined sheet** injected from a shallow magma chamber—more specifically a chamber at the depth of about 5 km. The eruption caused a lot of problems because of ash clouds that spread to Europe, but the volume of magma erupted was small (about 0.1 km^3). By contrast, the eruption in Bardarbunga-Holuhraun (Bárðarbunga-Holuhraun) in central Iceland, just north of the main ice cap, Vatnajökull (cf. Fig. 2.2), in the years 2014–2015 was supplied with magma by a **regional dike** whose length is nearly 50 km. The dike is vertical and many metres thick. The lava flow erupted is about 1.5 km^3, or 15-times larger than the magma erupted in Eyjafjallajökull. And if the volume of the regional dike is included then the total volume flowing into the crust from the deep-seated reservoir and partly up to the surface was probably 3–4 km^3.

These types of intrusions are easily distinguished in the field, as you have already seen. The inclined sheets are normally very thin (around one metre, many thinner) and, of course, inclined (Figs. 4.4b and 11.2). They also tend to have somewhat different chemistry—they are more silica rich or 'evolved' as this is called—from that typical of the regional dikes. The regional dikes are generally close to vertical and much thicker than the inclined sheets. In many parts of Iceland, the regional dikes are, on average, with a thickness of 3–5 m. So here we are really seeing how the volcanic systems, how the rift zones, operate at depth. Later today I shall show you what the earthquake fractures look like at depth in the crust, but let us first have a close-up view of the dikes.

11.3 Sandpit, Secondary Minerals, and Dike Propagation

That view of dikes is offered by the **third stop** (**3**) at the coast starting from **Hvalfjardareyri (Hvalfjarðareyri)** and walking to the west. Hvalfjardareyri is a sand-and gravel 'peninsula' or what is called a **sandspit** in geology (Fig. 11.5). Sandpits of this type form as sediments (e.g. sand) carried by the sea are deposited as the sea loses its carrying power. As the sediments accumulate, the sandpit gradually extends into the fjord (Fig. 11.5). Please notice that you can only look at the dikes when the **tide is low**, so it is a good idea to check the tides before you go on the excursion. Also, please notice, that some of the thick dikes project into the sea. Thus, you should make sure that you can **pass them** at any time, so as to be able to get back to the car before the tide comes in. If you decide to go down to the coast, then you can park your car close to Road 47 and walk a few hundred metres along a gravel road to the coast. Alternatively, if you have a 4-wheel drive car, you can drive along the gravel road to the parking place that is on the top of the thickest dike in the area.

While you walk to the west, in the direction of the opening towards the sea (towards the mouth of the fjord), there are many things to see and observe on the coast. Not only are there impressive dikes, but also numerous beautiful minerals, so-called **secondary minerals** or **amygdales** which many people like to collect (Fig. 11.6). There are beautiful minerals on many other parts of the coast of Hvalfjördur. The most noticeable minerals found here are **zeolites**, some forming beautiful fibres. The secondary minerals do not only form amygdales, that is, crystals within the vesicles (the holes or cavities formed as gas escapes from the

Fig. 11.5 View northeast showing part of Hvalfjardareyri, a sand-and gravel 'peninsula', that is, a sandspit. Sandpits are generated when sediments carried by the sea are deposited as the sea loses is carrying power

lava—see Chap. 5) but also thin white veins (Fig. 11.7). Some veins, mostly with thicknesses of millimetres or less, occur inside the dikes, others at the contacts between the dikes and the surrounding rock—the lava pile—and still others in the lava pile itself.

The reason for the amygdales and the veins is that this is an **old geothermal field**. At the coast we are some 1200 m below the original surface of the volcanic system (Fig. 11.3). As we have seen earlier today (Fig. 11.2), and in other areas (Chaps. 7, 12 and 13), there are geothermal fields associated with active volcanic systems. Here we are far away from the shallow magma chamber and the dense swarm of inclined sheets, so the geothermal field was not so intense as to alter the colour of the rocks much or transform them into clay (Fig. 11.2). Nevertheless, the amygdales and the mineral veins show that this was a geothermal field, and the heat sources are obvious—the dikes. It is thus about time to take a look at the dikes, which are exceptionally well exposed (easily observed and studied) on this coast.

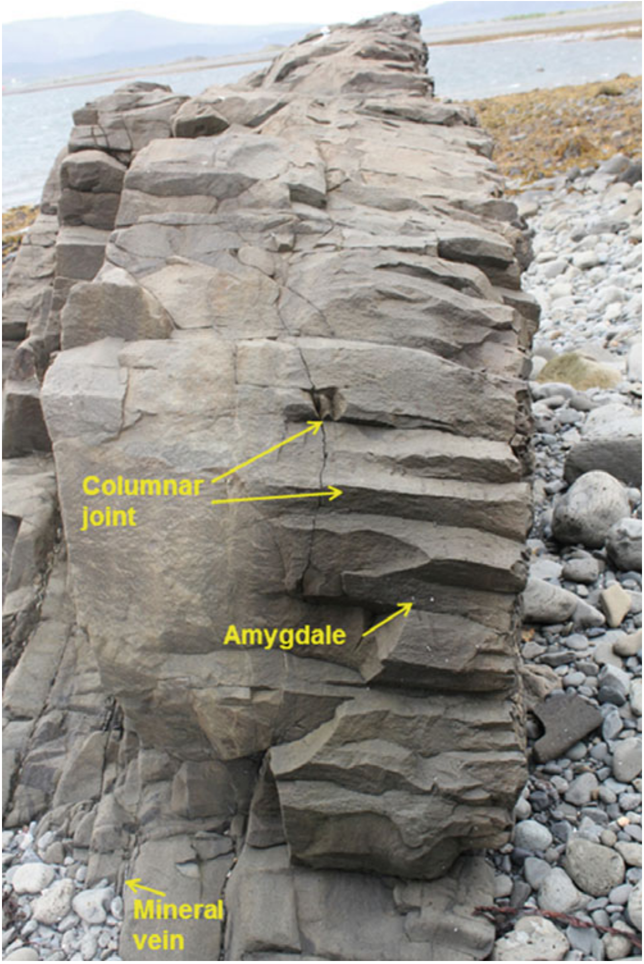

Fig. 11.6 Regional basaltic dike on the south coast of Hvalfjördur, close to Hvalfjardareyri. Columnar joints form when the hot magma that forms the dike cools and solidifies. They form perpendicular to the cooling surface, the surrounding hosting rock (here eroded away) of the dike, and are therefore horizontal in a vertical dike. Amygdales and mineral veins are generated through the flow of geothermal water along the dike. The dike is southwest-northeast trending or striking and about 1.5 m thick

The thickest dike in this coastal area, and the one that you need to pass—hence the need for low tide—is many metres thick (Figs. 11.7 and 11.8). Like nearly all the dikes in this coastal section, it is a basaltic regional dike with a northeast

Fig. 11.7 Regional basaltic dike on the south coast of Hvalfjördur, close to Hvalfjardareyri. The dike varies in thickness between 3 and 4 m. View southwest, the dike trends or strikes southwest-northeast and is steeply inclined (by about 80°) to the northwest. The dike has crude columnar joints in the centre, but finer and better developed near its margins (see Fig. 11.8)

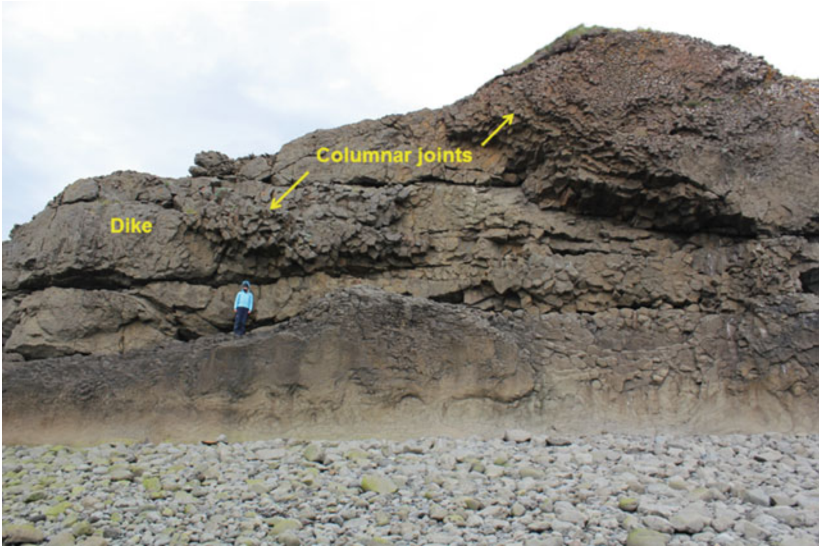

Fig. 11.8 View east, showing the west margin of the dike in Fig. 11.7. Well-developed, horizontal columnar or cooling joints are indicated. The horizontal fractures that cut through the dike relate to stresses generated when the rocks, including the dike, were uplifted following erosion during the Ice Age. The person provides a scale

orientation. The close to horizontal fractures in the dikes are cooling fractures or **cooling or columnar joints** of the type we discussed in Chaps. 4 and 5. But you notice the contrast in the fracture attitude: in Stardalshnjukar (Figs. 4.8 and 4.9) and at Thingvellir (Figs. 5.5 and 5.6) the fractures are mostly vertical whereas here they are mostly horizontal (Figs. 11.7 and 11.8, and see also the joints in the dike in Fig. 11.6). The reason is that most of the cooling joints are oriented at right angles or perpendicular to the surfaces, fluids or rocks, into which the heat from the cooling magma is conducted. Here the heat from the dike-magma flows into the rock that forms the walls of the dike-fracture—seen in Fig. 11.7. The heat flow in dikes is mostly horizontal. This follows because heat always flows primarily in the direction of greatest temperature difference, which here is about 1000 °C, that is, between the dike-magma, at about 1200 °C, and the nearby rock, the host rock, at perhaps 200 °C when the magma was injected. The cooling joints in the lava flows at Thingvellir are vertical because the flow units are horizontal and in contact with the atmosphere, so that the heat is conducted mainly vertically—into the air.

Similarly, the cooling joints in the horizontal sills of Stardalshnjukar are vertical because the flow of heat is vertical, up and down, from the magma that forms the sill. Ideally, cooling joints form hexagonal columns, but squares and pentagons, for example, are also common. The cooling or columnar joints are better developed in some parts of the dike than others, as we see in Fig. 11.8. In Chap. 14, at the coast of Reynisfjara, we will see some of the most beautiful columnar joints known in Iceland.

You notice in Fig. 11.7 that the dike is really **not quite vertical** but rather inclined at about 80° to the right (west). But did I not say that the regional dikes were generally vertical? Yes, and they are—initially. So how come this one and many others along the coast here are inclined—mostly steeply to the west? The reason is the tilting or inclination of the entire lava pile, as we discussed in Chap. 4 in connection with the south slopes of Esja (Figs. 4.4 and 4.6). The entire crust, including all the lava flows and all the dikes, has gradually become tilted towards the active volcanic zone, that is, to the east. The tilting increases with depth in the crust. So at sea level in Hvalfjördur the tilting to the east is about 10°. It follows that the dikes, originally vertical, have become tilted 80° to the west.

There are many dikes to explore and observe along this coast. Depending on your time and interests, you can walk along the coast and observe them as well as beautiful minerals such as zeolites. Here, however, we go only a short distance to the west, namely to what is perhaps the most striking structure on the entire cost. Again, it is a dike, with well-developed columnar joints (some in fact being bent or close to vertical in parts of the dike presumably because of 'plastic' or semi-ductile deformation in the dike centre while the columns were developing). But the most remarkable thing about the dike is that it is so much more **resistant to erosion**—'stronger'—than the surrounding rock that it forms a **ridge or wall** that reaches many metres above the surroundings (Figs. 11.9 and 11.10). Many dikes are more resistant to erosionthan the host rock, some forming small islands or skerries, but this one is unusually clear.

So why are dikes commonly harder or more resistant than the host rock? Primarily because the dike rock is more uniform in mechanical properties, has less internal variation and weaknesses, than lava flows and other host rocks. Of course the dike rock has numerous columnar joints, but apart from them there are few weaknesses in the rock. **Weaknesses** in rocks include cavities (vesicles from gas expansion), contacts, and irregular fracture networks. All such weaknesses tend to cause the rock to break when it is subject to force or stress, for example from a weight and movement of a glacier and by the sea waves as they hit the rocks. The dike rock has only a regular fracture system, the columnar joints, normally very

Fig. 11.9 Regional basaltic dike nearby the dikes in Figs. 11.6, 11.7 and 11.8. The dike makes a noticeable wall on the coast because its rock is harder, more resistant to erosion, than the surrounding host rock. The dike direction (indicated) or trend or strike is north-northeast, its inclination or dip is 82° west-northwest (indicated), and its thickness is 3 m. The person provides an additional scale

small and regular (elliptical or circular) vesicles, and generally no weak layers in-between stronger layers. By contrast, as we have seen, for example at Gullfoss (Fig. 8.3; Chap. 8), there are often weak hyaloclastite or sedimentary layers in-between the lava flows, making the pile more easily eroded. In addition, aa lava flows have commonly very irregular fracture systems and numerous large, irregular vesicles that concentrate stresses when subject to force (from glaciers, from the sea waves, from the tides), and more easily broken down and eroded.

You can also find the opposite, namely that the dike is more easily eroded than the host rocks. This happens, for example, when the host rock is very hard—very resistant to erosion. For instance, many plutonic rocks, such as fossil magma chambers (Chap. 4) are very resistant to erosion and basaltic dikes that dissect them, or cut through them, are often more easily eroded than the host rock and thus form depressions—**gorges and gullies**.

Fig. 11.10 Close-up of a part of the dike wall in Fig. 11.9. The columnar joints are inclined or dipping by about 8° to the east-southeast simply because the dike itself is inclined by 82° to the west-northwest. The person provides a scale

11.4 Difference Between Eruptive Fractures (Dikes) and Earthquake Fractures (Faults)

Before we leave the dikes here, it is perhaps worth putting their formation in context of general fracture formation in the Earth's crust. So what is the main difference between a dike and an earthquake fracture, such as you have seen in Thingvellir (Chap. 5)? The differences are perhaps best summarised as follows:

- A fracture that subsequently becomes a dike is generated by the pressure of the magma. The **magma pressure** breaks or ruptures the rock. There are several analogies. An everyday analogy is a container of a fluid. If the fluid pressure becomes too large, the container ruptures, and the fluid forms a hydrofracture through which it leaks out. A technical analogy is hydraulic fracturing in the oil industry. Water-driven fractures are propagated, sometimes for a kilometre or more, to allow gas or oil to flow more easily along the fracture(s) to the production well.
- All fluid-driven fractures open up against the least stress or pressure in the crust. More specifically, the total fluid pressure must overcome the least stress plus the tensile strength of the rock. The main movement of the fracture walls is thus pure opening, as in tension fractures (Figs. 5.11, 5.12 and 5.13)—the walls become separated by the fluid pressure, here by the magma pressure. Now, rocks are **weak in tension**—their tensile strengths are low (Chap. 5)—so it is easy for the fluid pressure to rupture the rock. By contrast rocks are **strong in compression**. There is an interesting analogy here. In all the large buildings from the medieval period (roughly from the 5th to the 15th century), castles, cathedrals, mosques, palaces, that stand today the stones that constitute the buildings are everywhere in compression. Those that were built so that significant parts of the buildings were subject to tension, simply collapsed, and are thus not seen today.
- Earthquake fractures, by contrast, are basically closed (except some normal faults close to and at the surface, Chap. 5). The main movement of the fracture walls is not perpendicular (opening) but rather parallel to the fracture plane (Figs. 4.14 and 5.9). Such parallel movement is named **shear movement** and is similar to the movement of the blades of scissors or, even more homely, the movement when rubbing the palms of your hands together when you are cold. While fluids are involved in earthquakes—all earthquake ruptures occur in zones of high fluid (usually groundwater) pressure—the earthquake fracture itself is related to shear movement and thus to shear forces or shear stresses.

- The fault movement may be horizontal, as happens on so-called **strike-slip faults** such as the San Andreas Fault in Western United States (part of the fault being close to San Francisco) and the North Anatolian Fault in Northern Turkey (part of the fault being close to Istanbul). The largest earthquakes in Iceland also occur on strike-slip faults (Chaps. 9 and 14). Alternatively, the movement may be vertical, up and down the fault plane, as in the normal faults of Thingvellir (Chaps. 4 and 5). While earthquakes on **normal faults**, such as in Thingvellir, are usually not very large—rarely reaching magnitude 7—the largest earthquakes on the planet are on faults where the movement is up and down, namely on **thrust faults**, such as offshore Chile, Japan, and Sumatra (Indonesia). The difference is that during normal-fault movement (slip or displacement) there is horizontal extension (Figs. 4.5 and 4.14), that is, the crust is dilating horizontally, so that the fault walls are not pressed very hard together. By contrast, in thrust faults there is contraction, that is, the crust is being shortened horizontally, so that the fault walls are pressed very hard together and when they finally slip, very great energy can be released, hence large earthquakes.

So how quickly do these fractures move or propagate? Could we escape running from a propagating fracture front? The **rupture velocity**—how quickly the tip or front of the fracture moves and ruptures the crust—is well known for earthquake fractures. The rupture velocity is **several kilometres per second**. Remembering that 1 km/s is equal to 3600 km/h (kilometres per hour), namely the speed of sound in common rocks. For comparison, the speed of sound in the air is three-times lower than in common solid rocks, or about 1236 km/h—the reason being that the speed of sound increases with increasing density of the medium through which the sound propagates (and rock is much denser than air). Thus, at 3600 km/h, there is clearly no way that you can outrun a propagating earthquake rupture—it happens effectively **in an instant**.

The rupture velocity is very different, though, for propagating dikes and other magma-filled fractures (inclined sheets, for example). The velocity of propagation of many dikes has been measured. The dikes generate small earthquakes which are easily located and can be used to monitor the dike-fracture propagation. Most dikes propagate at an average velocity of **half to one metre per second**. Now one metre per second is equal to 3.6 km/h, which is a typical **walking speed**, pacing, of a person on a smooth path or road. So, yes, surely a reasonably fit person should be able to outrun a propagating dike. In fact, in several cases in Hawaii the observing volcanologist has followed the approaching dike (approaching the surface close to where the geologist was making his/her observations) and then run away when the tremors and steam (generated when the hot magma boils groundwater at shallow depth) and fractures at the surface indicated that the dike had only minutes to go before reaching the surface to erupt.

So why is the movement of the front of a magma-filled fracture, a dike, so comparatively slow? The reason is that it takes time for the magma to flow into the tip of the fracture. When the magmatic pressure in the dike-fracture reaches a certain level, the fracture advances or propagates very quickly for a certain distance at its ends and then stops. Because of its comparatively high viscosity, the magma cannot move or flow as quickly as the fracture ends propagate, so that there will be a temporary empty 'cavity' or front between the magma front and the fracture front. Magma has then to flow into this cavity, fill it, and build up a pressure so high that the fracture tip can advance or propagate again. This process, filling the cavity and building up the magma pressure for further rupture takes some time. In earthquakes, the fracture front does not have to wait for a fluid to follow—so there is no 'time lag' between the fracture front at any time and a fluid front. Thus, earthquake fractures propagate at the **speed of sound**, whereas dike fractures, on average, at the **pace of a walking person**.

If we look again at Figs. 11.9 and 11.10 we could now answer the question: how long time did it take for this part or segment of the dike to form? Since the segment is a couple of tens of metres long, it would have taken some tens of seconds to form. The entire dike is much longer. The one in Figs. 11.9 and 11.10 was most likely between one and three kilometres long at the depth in the crust where it is seen now (about 1200 m below the surface when the dike formed), and presumably much longer at greater depth. By contrast, the dike in Figs. 11.7 and 11.8 was most likely somewhere between three and nine kilometres long at the depth where we see it now, and much longer at greater depth. If these dikes had reached the surface, that is, if they fed lava flows at the surface and were thus **feeder-dikes**—which we do not know—then the lengths seen here would be similar to those at the surface and similar to those of volcanic fissures seen today on the Reykjanes Peninsula (Chap. 13). Most of these have lengths from a few hundred metres to about three kilometres, and thus corresponding to the estimated length of the dike in Fig. 11.9. Some, however, are as long as eight to ten kilometres, which would agree with the upper estimate for the dike in Fig. 11.8.

You might now ask: How can we know the lengths of dikes when we do not really see the entire dikes? The answers are, first, through comparison with observations and, second, using well-established theory that applies to all fractures. The first answer relates to actual dikes measured along their lengths in Iceland and elsewhere. All the results show that there are certain relations between the dike thickness and the dike length. The relations vary somewhat depending on the magma pressure when the dike formed and the mechanical properties of the rocks within which the dike is located. But these can be assessed. All the results show that thicker dikes tend to be longer. From measurements of dikes in Iceland at similar crustal depths as these, we can say that the dikes are normally **500–1500 times** longer than they are thick. Thus, a dike that has the thickness of about 2 m should, at this depth, be between 1000 and 3000 m long. Similarly, the length of a dike that has a thickness of 6 m, could, at this depth, be anywhere between 3000 and 9000 m. Not only are these direct measurements and observations, but they are also in perfect agreement with the basic theory dealing with fractures of any kind, whether the fractures are in rocks or other types of solids (such as wood, concrete, brick, ceramic, steel, glass, and composite materials). The results of this theory, referred to as **fracture mechanics** and initially derived from the laws of thermodynamics, show clearly that longer fractures tend to have larger displacements, whether wall-parallel displacements, such as in faults, or wall-perpendicular displacements, such as openings or thicknesses of dikes.

How long do dikes like these stay fluid? That is: how long time does it take for the magma **to solidify** and the dike to become solid rock? The answer depends on several factors, such as the original temperature of the magma and the surrounding rock, and also on what we define as solid dike-rock. If we take, for example, the original magma temperature as 1200 °C and assume the magma to be solid at 1100°, then a 4-metre thick dike (similar to the one in Figs. 11.7 and 11.8) would solidify in about 50 days. A dike at 1100° is, however, definitely plastic, ductile, and partly fluid. Perhaps it is better to ask: how long time does it take for the dike to cool down so much that cooling joints or columnar joints begin to form? Cooling joints begin to form at about 60% or the original magma temperature. So for dike-magma originally at 1200–1300 °C, cooling joints may be assumed to begin to form at temperatures of about 800 °C. For the magma in a dike 1.5–2 m thick (Fig. 11.6) to cool from 1200°–1300° down to 800° would take about 270 days, or about **9 months**.

The time for solidification, the cooling time, however depends on the thickness of the dike in second power. This means that it takes 2 × 2 = 4-times longer for a dike of the thickness of 2 m to cool down than one of the thickness of 1 m. Thus, it would take the dike in Figs. 11.7 and 11.8, which is close to 4 m in thickness, about 1060 days or nearly 3 years to cool down to the temperature of 800° so as to begin to develop columnar joints. The thickest dike on the coast of Hvalfjördur is about 25 m. If that dike was formed in a single magma injection—which we do not know—then it would have taken it about 42,000 days, that is, about 115 years or

over a century to cool down to 800°. For any of the dikes to cool down to the temperature of the surrounding rocks takes much longer time, from hundreds to thousands of years, and all that time they act as heat sources for geothermal fields.

11.5 Landscapes, Landforms, and the Hvalfjördur Volcano

We have now discussed dikes and magma-filled fractures in detail—perhaps in greater detail than you like. So it is time to move on to other areas of Hvalfjördur. We therefore get back to the car and follow Road 47 to the east to the inner part of the fjord. There are many options for driving around in this part of Hvalfjördur and many beautiful features to see. For example, you may like to drive to the east along Road 461 to see the beautiful lake Medalfellsvatn (Meðalfellsvatn). If you do so, you can either drive back the same way to Road 47, or go around the mountain Medalfell. In the latter case you drive along Road 48 to the west until it meets our main Road 47 at the beautiful river Laxa in Kjos (Laxá í Kjós). All the mountains here are similar, composed mainly of basaltic lava flows, with horizontal sills in-between some of the flows (the sills can normally be recognised by their very intense and well-developed vertical columnar jointing, as you saw earlier). There are also some layers or units of hyaloclastite in-between the lava flows. Furthermore, many earthquake fractures, normal faults, dissect the lava pile, most of the fractures being directed northeast-southwest just like the fractures at Thingvellir (Chap. 5)—where, indeed, they originated as the entire lava pile seen here.

In this excursion, however, we do not drive to Lake Medalfellsvatn, but rather drive on towards the inner part of the fjord. Our **fourth stop** (**4**) is to get an overview of the great variety of landscapes that can be seen in the inner part of Hvalfjördur. You should, here and everywhere, be careful to select a proper parking place for the car and yourself. Once that is done, the view from here, when the visibility is good, offers a unique opportunity to observe and understand the processes that take place inside the volcanic rift zones of Iceland, and the eruptive and intrusive materials these processes produce.

If we first look southwest (Fig. 11.11), we see the mountain **Eyrarfjall** across the fjord and small hills on the point of **Halsnes (Hálsnes)**. The gently inclined layers in Eyrarfjall are basaltic lava flows. They are all inclined or tilted down to the southeast, that is, towards the West Volcanic Zone at Thingvellir. The same tilting of the lava pile is seen in all the mountains here and is explained in Chap. 4 (Fig. 4.6). The hills on the point of Halsnes are primarily formed by an intrusion,

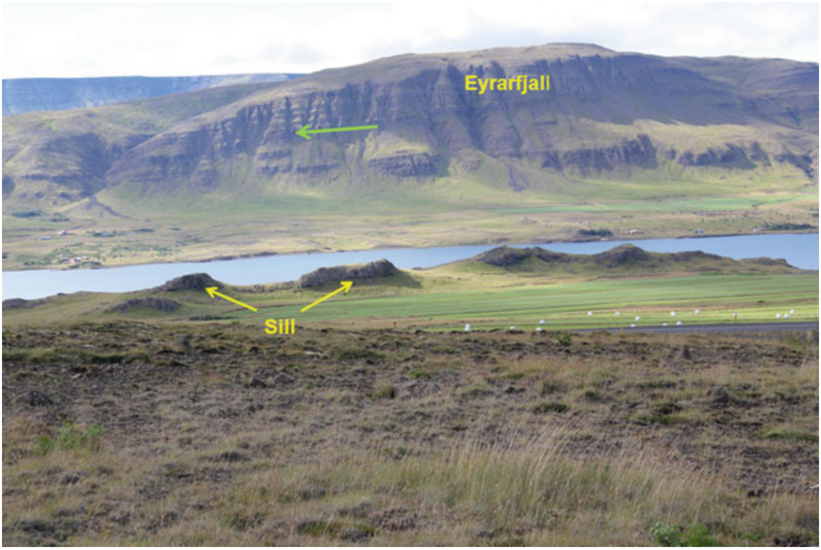

Fig. 11.11 Gently inclined or dipping basaltic lava flows in the mountain Eyrarfjall. View southwest, the hillocks on the peninsula Halsanes constitute a part of an intrusion, a sill

a sill, which forms a part of the main volcano in the inner part of Hvalfjördur, the **Hvalfjördur Volcano** (**Hvalfjörður Volcano**). Associated with that sill is the thickest dike in Hvalfjördur, 25 m. The dike crosses the cove of Laxarvogur (Laxárvogur) that we just passed and is of a rather coarse-grained basalt, named dolerite or microgabbro—the same as in the sill. The grains in the rock become larger or coarse—become visible—when the body is thick and solidifies or freezes slowly. There are also acid rocks, including dikes, on the coasts of Laxarvogur and in the mountain on whose slopes we have made this stop, namely **Reynivallahals (Reynivallaháls)**, all of which form part of the Hvalfjördur Volcanic System. Based on our discussion at the third stop as to the ratio between dike thickness and dike length, a dike with a thickness of 25 m could have a length anywhere between about 12 and 40 km. In the latter case, the dike's length would approach that of the dike formed in the 2014–2015 Bardarbunga-Holuhraun eruption north of Vatnajökull, where the dike length was about 45 km.

If we now look north, we see two remarkable mountains. To the west is Thufufjall and to the east Brekkukambur, both of which belong to the Hvalfjördur Volcano. **Thufufjall (Þúfufjall)** is remarkable for having a prominent volcanic

plug or neck in its centre (Fig. 11.12). Plugs form in volcanoes with a central ‘pipe’ or conduit. They normally develop where eruptions are comparatively common and are usually composed of a mixture of thin dikes, a dike cluster, and fragmented rocks, that is, pyroclastic rocks. That is also the type of rock we find in the neck of Thufufjall. (Perhaps the most famous volcanic plug or neck in the world is Ship Rock in New Mexico, United States.) The plug of Thufufjall reaches an elevation of about 540 m above sea level. Since sea level is some 1100–1200 m below the surface of the volcano when it was active, this means that the plug we see is the old central conduit at 500–600 m depth inside the volcano (indicated schematically in Fig. 11.12). The conduit is elongated (elliptical in plan view) in a northeast direction, that is, parallel with the direction of dikes and normal faults in the area. The maximum diameter is only several tens of metres. So we are seeing here an additional part of the shallow magma-transport system in a volcano, the Hvalfjördur Volcano, which was active just over 3 million years ago.

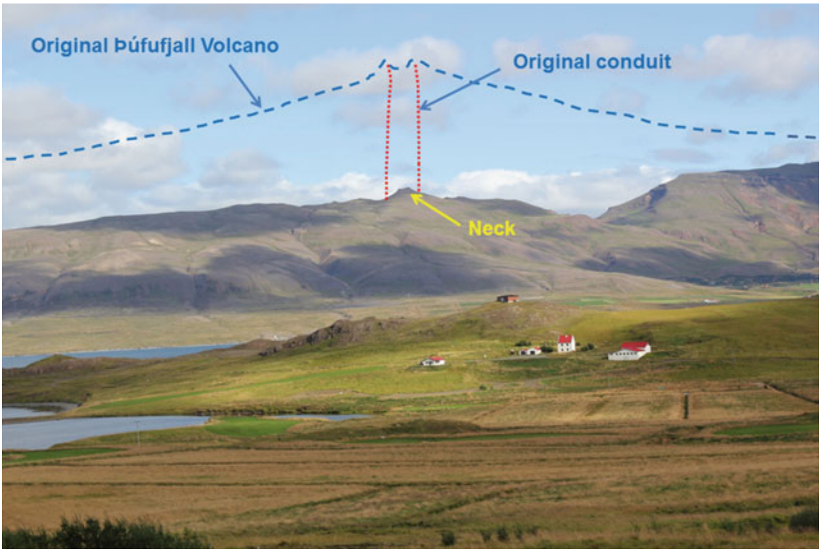

Fig. 11.12 Thufufjall (Þúfufjall) forms a part of the extinct Hvalfjördur Volcano. When Thufufjall was active it was a much larger and taller volcano than the eroded remnants seen today, since several hundred metres of the upper part of the volcano have been eroded away, primarily by glaciers. View northeast, the neck or plug is a solidified part of an old conduit in the centre of the volcano. I indicate here, very schematically, what the profile of the volcano may have looked like when it was active some 3 million years ago

Fig. 11.13 Mountain Brekkukambur in Hvalfjördur. View northeast, the mountain is partly composed of lake sediments, with many intrusions (Fig. 11.15). The sediments accumulated within the caldera lake of the Hvalfjördur Volcano some 2–3 million years ago

The mountain east of Thufufjall, **Brekkukambur**, is in some ways even more interesting (Fig. 11.13). What we are seeing here is a sort of cross-section through the centres of volcanoes such as Askja (Central Iceland), Grimsvötn (Grímsvötn, in the Vatnajökull ice sheet), and, somewhat differently, Eyjafjallajökull (South Iceland). How? Because we are looking into the collapsed part of the Hvalfjördur Volcano, namely a circular depression of the type called **collapse caldera** (Fig. 11.14, see also Fig. 4.10). Collapse calderas are very common in volcanoes worldwide. They relate to subsidence or collapse of a part of the volcano—a piston —into an underlying shallow magma chamber often, but not always, during a large eruption. Famous collapse calderas include those on Hawaii, many of the volcanoes of the Galapagos Islands, and the beautiful Crater Lake in Oregon, United States. The best-exposed active collapse caldera in Iceland is Askja in the mountain Dyngjufjöll in central Iceland—which is, in fact, not a single caldera but rather a triple (three) caldera or even, according to some interpretations, a quadruple (four) caldera. Most of the other calderas in Iceland are within the ice caps, and thus difficult to see. The small caldera in the top of Eyjafjallajökull (Chap. 14), and even more the Grimsvötn Caldera in the Vatnajökull ice sheet, are perhaps among the easiest to see in the ice caps—and then primarily from the air. One we discuss further in this book, however, is the Katla Caldera in South Iceland (Chap. 14).

But a collapse caldera is a depression—it forms when part of the volcano collapses or subsides into the associated shallow magma chamber (Figs. 4.10 and 11.14). Now Brekkukambur is anything but a depression; it is obviously a positive topography, a mountain, rising to about 650 m above sea level. So how can a tall mountain be a caldera? To answer that question, we must look into the history of the area where Brekkukambur currently stands.

We begin the story when a central volcano, the Hvalfjördur Volcano, had already formed. For given—fortunately very rare—stress conditions in a volcano, a ring-fracture, a ring-fault, forms (Fig. 11.14) and then part of the volcano subsides into the shallow magma chamber. Because shallow magma chambers in Iceland are commonly one or two kilometres thick, the subsidence can easily reach one

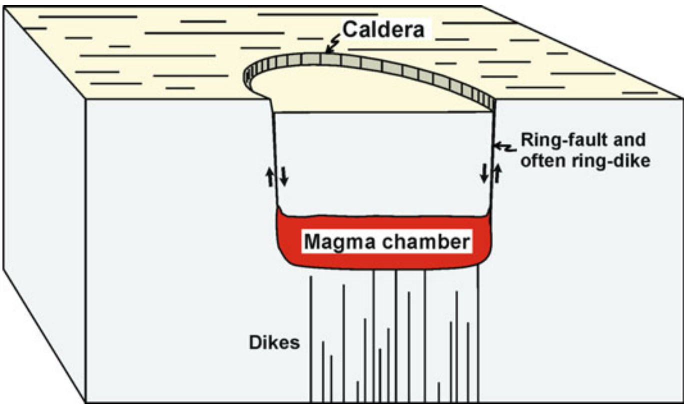

Fig. 11.14 Collapse caldera is a common stage in the evolution of many central volcanoes, including those in the volcanic zones of Iceland. Calderas form when a piston-like central part of the volcano subsides, collapses, into the associated shallow magma chamber, as indicated here. The subsidence occurs along a fault that is circular or somewhat elliptical in plan view, and referred to as a ring-fault. Commonly, the ring-fault is partly, or entirely, injected by magma to form a ring-dike. The Hvalfjördur Volcano developed a collapse caldera which formed a deep caldera lake that subsequently became partly filled with sediments which, in turn, were injected by various intrusions (dikes, sills, and inclined sheets). The lake sediments and the intrusions today constitute the bulk of the mountain Brekkukambur (Figs. 11.13 and 11.15). Only half of the caldera is seen in this illustration. Compare Figs. 3.1 and 4.10)

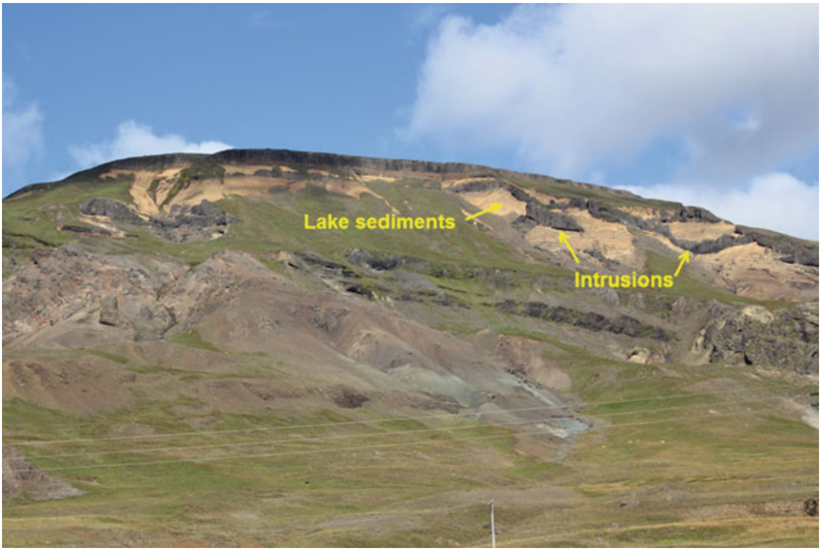

Fig. 11.15 Close-up of part of the mountain Brekkukambur. View northeast, lake sediments and intrusions are indicated. Compare Fig. 11.13

kilometre. In fact, directly observed caldera collapses in the past decades worldwide have commonly had subsidences or displacements from several hundred metres to about one kilometre. After the subsidence, the caldera becomes a lake. Most calderas in the world become lakes for the simple reason that their bottoms are far below the water table (Chap. 5 on Lake Thingvallavatn discusses this point)—witness Crater Lake and many other calderas. The lake gradually collects loose rocks, sediments, from the walls and its surroundings, which eventually reach a thickness of hundreds of metres. While the volcano is active, it injects dikes and inclined sheets into the sediments, as is seen in close-ups (Fig. 11.15). Eventually, the volcano becomes inactive (extinct), the lake dries out, and the lake sediments, together with intrusions and lava flows, become solid, stiff rock (Figs. 11.13 and 11.15). Because there were many plants living in the lake and its vicinity, the lake sediments of Brekkukambur contain numerous plant **fossils**. Subsequently, during the past two million years or so, the Ice-Age ice sheets and glaciers have eroded and sculptured the land, leaving the hardest cores of the volcanoes—those with many stiff or hard intrusions—standing as mountains in-between the deeply eroded valleys and fjords, one of which is Brekkukambur.

This, in a nutshell, is the story of the landscape evolution everywhere in Iceland outside the volcanic zones and, in particular, the story of the mountain formation (although few mountains are of former caldera lake sediments; most are of lava flows and hyaloclastites). Mountains in Iceland (and most mountains elsewhere as well) are formed through **two basic processes**: (1) **erosion and uplift** and (2) construction through **eruptions**. Many mountains in a volcanic area such as Iceland are, of course, the result of both processes, although normally one is the dominant process. Hvalfjördur offers text-book examples of both types (Fig. 11.16). We have already seen that Thufufjall (Fig. 11.12) and Brekkukambur (Figs. 11.13 and 11.15) are mountains today primarily because of erosion and uplift. Looking east to the innermost (easternmost) part of the fjord (Fig. 11.16) we see more examples of mountains formed by these two basic processes. These include the mountains Thyrill (Þyrill), Hvalfell, Mulafjall (Múlafjall), and Botnsulur (Botnsúlur). Of these, Thyrill and Mulafjall are primarily formed

Fig. 11.16 Overview of the mountains in the inner part of Hvalfjördur. View east, Thyrill (Þyrill) and Mulafjall (Múlafjall) are almost entirely generated through glacial erosion and uplift, Botnsulur (Botnsúlur) is largly generated as a result of eruptions, but has also been shaped by erosion. Hvalfell, a table mountain, is mostly pure construction through a volcanic eruption

through **erosion** and subsequent crustal uplift. By contrast, Botnsulur is mostly formed by construction through **eruptions**, although its present shape, particularly the peaks, are partly related to glacial erosion. You already saw Botnsulur on the way to Thingvellir (Fig. 4.12), although the three peaks were not so clear there as here. Hvalfell, by contrast, is almost purely a constructive mountain formed in an eruption. More specifically, Hvalfell is one of the better examples of table mountains in Iceland. As we drive on to the innermost part of Hvalfjördur, we will have close-ups of all these mountains (least so of Botnsulur, though).

11.6 What Does the Thingvellir Graben Look like at Great Depth?

We now drive to the east until we come to the **fifth stop** (**5**), at the head of the valley of Brynjudalur and facing the mountain Mulafjall (Múlafjall) to the north (Figs. 11.17 and 11.18). We see how Mulafjall is primarily formed by lava flows and layers of breccia (broken rocks, also called pyroclastics) in-between. We also see some intrusions, including sills, characterised by their clear and well-developed vertical columnar joints. The mountain exists primarily because of the erosion and subsquent uplift of the crust, the erosion here being best demonstrated in the valley of Brynjudalur and, of course, its extension into Hvalfjördur in the form of the cove Brynjudalsvogur. The main reason why we stop here, however, is not the landscape, the mountain and the valley, but rather the striking structure that can bee seen in the slopes of Mulafjall.

The structure I refer to here is an exceptionally clear 170-metre-wide **graben**, that is, a palaeovalley formed between earthquake fractures, namely normal faults (Figs. 11.17 and 11.18). I indicate the boundary faults in Fig. 11.17 and all the faults associated with the graben in Fig. 11.18. The faults as seen here are some 900 m below the original top of the lava pile. There are several remarkable features about these faults in relation to the faults in Thingvellir that is worth drawing your attention to:

- The faults are all **normal faults**, of exactly the same type in Thingvellir and, generally, in the volcanic zones of Iceland (Fig. 2.2). I have indicated a ‘**marker layer**’, namely a breccia layer, used to measure the displacements along the faults—in geology such layers, used to mark or measure displacements, are called ‘marker layers‘. I have also indicated all the main faults.

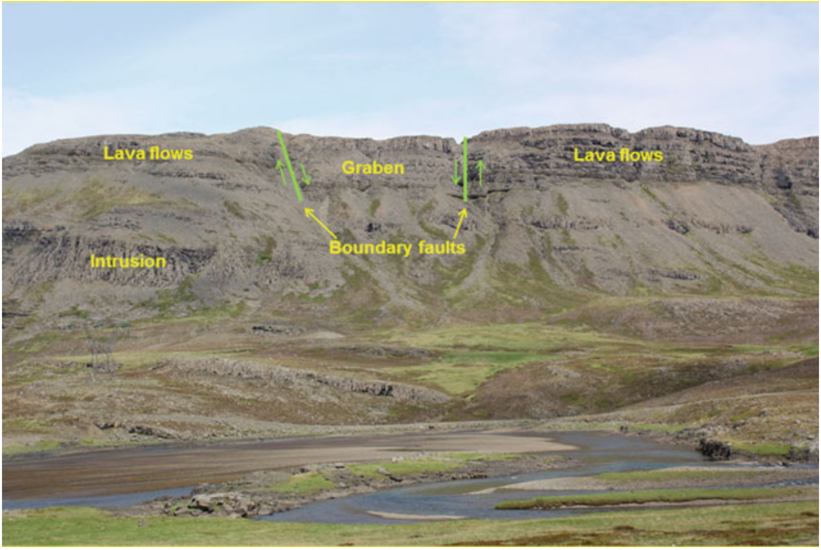

Fig. 11.17 Graben in Mulafjall (the mountain is seen in Fig. 11.16). View northeast, the two boundary faults are indicated. The mountain is primarily composed of basaltic lava flows, with some intrusions, primarily sills (a thick intrusion is indicated) and a few dikes. The overall subsidence or vertical displacement associated with the graben is about 30 m, which is similar to that measured at the boundary faults of the Thingvellir Graben (Chaps. 5 and 6)

- The faults are only 17–18 km northwest of the main faults at the western boundary of Thingvellir Graben, namely Almannagja and nearby faults (Figs. 5.1, 5.2, 5.3, 5.8, 5.11 and 6.2).
- Not only are the faults in Mulafjall of the same type as in Thingvellir, they also have the essentially the same movements or displacements. For example, the western boundary fault, the one on the left (Fig. 11.17), has a vertical displacement or subsidence of 30 m, similar to the maximum displacement on Almannagja (40 m). And the other three main faults have displacements of, from west to east (Fig. 11.18), 6 m, 15 m, and 8 m. Together these add up to 29 m or effectively a cumulative displacement equal to that on the western boundary fault.
- The total widh of the graben is 170 m. This is very similar to the distance from Almannagja to some of its nearby faults (Figs. 5.1 and 5.11). And the main fault is inclined down to the east (just like Almannagja, Fig. 5.9) whereas the

other three are inclined down to the west, again like the faults nearby Almannagja.

- So what we are seeing here in a vertical cross-section is exactly what we can expect Almannagja and other faults at Thingvellir to look at depths of some 900 m below the surface. You notice that (a) all the faults are **closed** (not open as in Thingvellir), and (b) there are **no** large **tension fractures** (such as are common at Thingvellir). This is exactly as expected from fracture-mechanics theory. As we discussed at Thingvellir (Chap. 5) large tension fractures should normally not extend deeper than a few hundred metres into the crust. They are therefore not expected here at depths of about 900 m below the surface of the rift zone as it was (before the erosion) when the fractures formed.

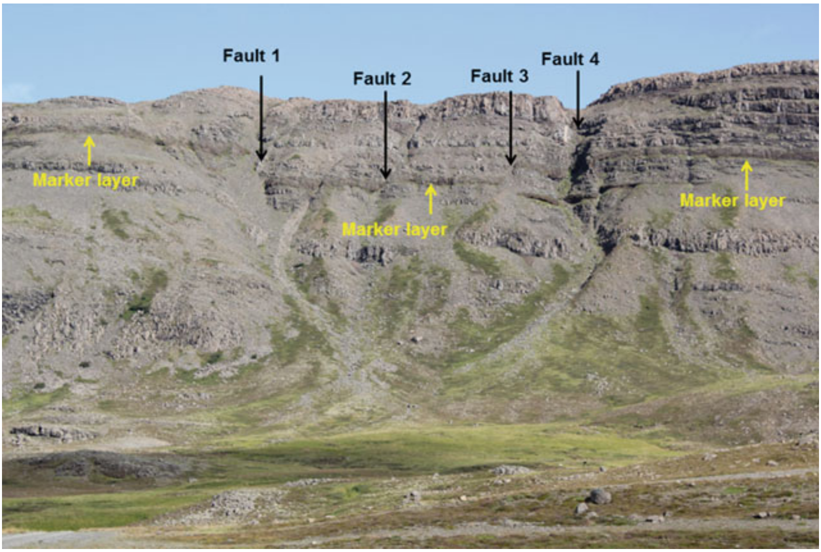

Fig. 11.18 Close-up of the graben in Mulafjall (Fig. 11.17). The 170-metres-wide graben is composed of four main faults. Fault 1 is the west boundary fault with a vertical displacement or subsidence of 30 m. Fault 2 has a displacement of 6 m, Fault 3 a displacement of 15 m, and Fault 4 a displacement of 8 m. The average inclination or dip of the faults is 71°. Fault 1 is inclined to the east, the other three to the west. The marker layer, an easily identified layer and used as a reference layer to measure the vertical displacements across the faults, is a brownish breccia layer. Compare Fig. 4.14 for the general illustration of a graben and a marker layer

- We also see that the faults are no longer vertical, as at Thingvellir, but **inclined**—dipping as it is called in geology. Again, this is in perfect agreement with fracture-mechanics theory, which states that the vertical parts of normal faults occur only, if at all, in the uppermost tens or metres or, at most, the uppermost few hundred metres, namely to the same maximum depths as tension fractures can reach.
- We also notice, perhaps less easily, that there are no major dikes anywhere within the graben. This is what is normally found in Iceland. Where the plate movements are accommodated by dike intrusions, there are rather few faults. By contrast, where the plate movements are accommodated mainly by comparatively large normal faults, there are few dikes. At depths of many kilometres, however, most of the plate movements in Iceland, as well as at mid-ocean ridges in general, is accommodated almost exclusively by dikes—not by faults.

Now, why is what we see here remarkable and interesting? What is so special about these faults and the graben? My first answer is that this section provides a unique opportunity for studying the structure of a plate boundary, a volcanic rift zone, in three dimensions. That is to say, observing the slopes of the mountain Mulafjall and its vicinity we get a very clear view what active rift zones, where plate movements are currently taking place and eruptions occur from time to time, look like below the surface—here at depths of 900–1200 m. Thus combining the observations here with the surface observations in the active areas we can really understand the **processes** that generate **earthquakes**, volcanic **eruptions**, **groundwater** reservoirs, and **geothermal** reservoirs and fields (occasionally with geysers) in the volcanic rift zones. Such processes cannot be understood if we restrict our attention to the surface. Because then we are trying to infer three-dimensional structures (and processes), such as earthquake fracture, dikes, and magma chambers, from two-dimensional (surface) measurements, and there are no unique solutions in such an undertaking.

My second answer is that this part of Hvalfjördur is remarkable because it is so close to the active volcanic rift zone. A distance of 17 or 18 km to Almannagja is very short—we are almost at the active rift zone itself. So this—perhaps **unique combination**—of being able to observe to considerable crustal depths the three-dimensional structures of exactly the same type, and generated by exactly the same processes, as in the nearby active volcanic rift zone allows for a really deep understanding of how a divergent plate boundary, such as in Iceland, really operates. And at the next the corner—the next turn on the road—we will get another unique three-dimensional view of common volcanic rift-zone structures.

11.7 Vertical Section Through Two Lava Shields

For the **sixth stop** (**6**) you need to find a good parking place; such a place is not so easily found as at the previous stop. Provided you find a good parking place for your car, the view to the north shows something really remarkable. You recall that there are many **lava shields** in the active volcanic zones of Iceland. In particular, you saw the lava shield Thrainsskjöldur on the way from the Keflavik to Reykjavik on the first day (Chap. 2) and the large Skjaldbreidur lava shield north of Thingvellir on the Golden Circle (Figs. 5.14, 6.1, 6.2 and 6.7; Chaps. 5 and 6). If you make the excursion to the Reykjanes Peninsula (Chap. 13) you will see more lava shields. The shields are all composed of pahoehoe lava flows. Exceptional section through young, that is, 9000-year-old, pahoehoe lava flow is provided by the west wall of Almannagja (Figs. 5.5 and 5.6).

However, looking north at the sixth stop you see the mountain **Thyrill (Þyrill)** which provides a vertical section through not one but through parts of **two lava shields** (Figs. 11.19, 11.20 and 11.21). The contact between the two shields is very sharp and easily seen, partly because the lower shield is much darker in colour than the upper shield (Figs. 11.20 and 11.21). The flow units can be distinguished in close-up (Fig. 11.21). The flow units of the younger shield, the light-grey one on the top (the upper shield), are clearly thicker than those that constitute the lower and older shield. The flow units are here much more altered than those in the walls of Almannagja, simply because the units in Thyrill have been the channels of groundwater and geothermal water for one or two million years whereas the Almannagja units are only 9–10 thousand years old. Because of the circulation of geothermal water through the rocks, there are numerous secondary minerals, particularly zeolites, in the rocks. In fact, there are **secondary minerals** in the rocks on both sides of the innermost part of Hvalfjördur, and you can even see them as white spots in the rocks at the present stop as well as when you drive on after this stop.

The name of the mountain, Thyrill, refers to its well-known gusty winds. The reason for the gusty winds is partly the shape of the mountain. In plan view (seen from the air) the mountain has a wedge shape, the thin end of the wedge pointing to southwest. In fact, the mountain slope facing northwest and thus trending northeast follows the general direction of normal faults in the area (the same is in Mulafjall, Figs. 11.17 and 11.18) and that of the entire West Volcanic Zone (Fig. 2.2). North-easterly winds prevail in the Hvalfjördur area. They thus become magnified along the edges and slopes as well as some of the northeast oriented fault-generated gullies of Thyrill.

Fig. 11.19 Thyrill (Þyrill) is a mountain composed primarily of two lava shields. The rocks of the upper one are grey, whereas those of the lower one are dark grey. In addition, there are some different layers, including tuff and breccia layers (hyaloclastites), in the uppermost part of the mountain. View northeast, the contact between the main lava shields is shifted up and down across normal faults (compare Figs. 11.20 and 11.21)

At Thingvellir we estimated that the pahoehoe lava flow at the surface may reach a thickness of several hundred metres. In the cliffs of Thyrill, rising to a maximum elevation of about 390 m above sea level, you see similar thicknesses for the upper and younger lava shield, and tens of metres are seen of the older and lower shield as well (Figs. 11.19 and 11.20). We do not know if the section in Thyrill is close to the margins of the shields, where the shields, and thus the lava flows, are thinner, or close to their centres, where the flows are thicker. But the Thyrill section shows clearly that pahoehoe lava flows can reach thicknesses of tens or **hundreds of metres**.

If you follow the contact between the older and the younger shield by your eyes, you notice two things (Figs. 11.19 and 11.20). First, that the contact is gently **inclined** to the east and, second, that it abruptly **changes elevation**, that is, changes height in the mountain. The first observation is one you have now made many times, namely that the entire lava pile in all Hvalfjördur area is inclined down to the southeast, that is, towards the active volcanic rift zone at Thingvellir. The second observation relates to earthquake fractures or, more specifically, to numerous **normal faults** that cut through the two lava shields. The frequency (number per kilometre), distribution, and displacements of the faults here is similar to that of the Vogar Fissure Swarm you saw on the way from Keflavik to Reykjavik (Figs. 2.5 and 2.6; Chap. 2). And here, like in the Vogar Swarm, the faults cut through thick pahoehoe lava flows.

In fact, the extension or crustal dilation (**spreading** of the crust) is here dominated by normal faults. For example, in a 2.5-kilometre-long horizontal section or profile along the cliffs seen in Fig. 11.19—and a bit further to the east—there are only 2 dikes but 20 normal faults. There are also some tens of potential normal faults, that is, columnar joints that have started to link together, with some centimetres of opening, but not as yet developed into faults. These are most likely related to (1) subsequent uplift of the lava pile, during the development of the mountain following the erosion that formed the main valleys and the fjord and (2) **compression** from the West Volcanic Zone. This follows because the direction of half of the joints is to the **northwest** (whereas the direction of almost all the

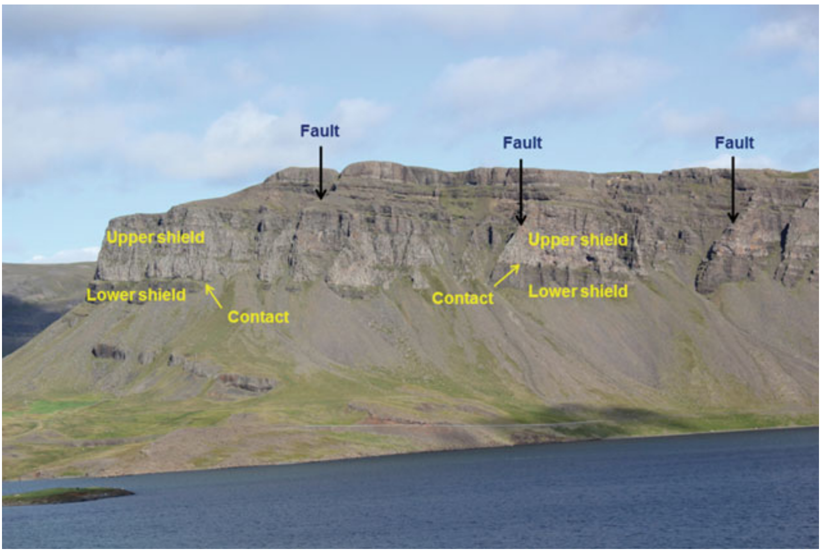

Fig. 11.20 Close-up of part of Thyrill (Þyrill). Parts of the upper and lower shields are seen here (indicated) as are many normal faults. View northeast, the contact between the upper and lower shield is indicated

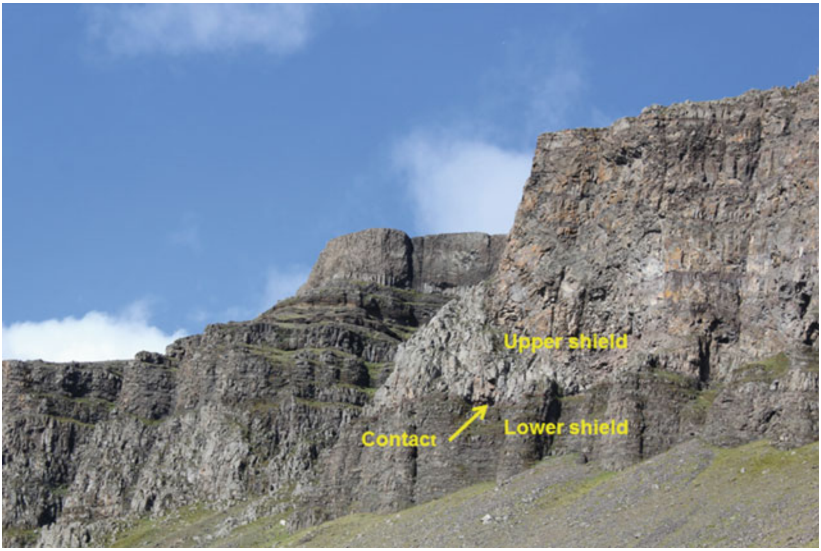

Fig. 11.21 View northwest, close-up of the contact between the upper and lower lava shields in the mountain Thyrill (Þyrill) in Hvalfjördur. Compare Figs. 11.19 and 11.20

normal faults is to the northeast) and are thus not formed within the volcanic rift zone (which strikes northeast and has most of its fractures directed northeast as you remember from Thingvellir, Chaps. 5 and 6). Some of the joints may actually be potential strike-slip faults because most are close to vertical, while others are most likely related to related to horizontal northwest-trending **compressive stress** related to the pressure of northeast directed dikes in the volcanic rift zone (this mechanism is discussed further in Chap. 13). This latter is supported by many of the northwest directed joints being at nearly right angles (90°) to the normal faults (and the few dikes) in the area. The cumulative spreading due to the joints, however, is very small, and none of them can be classified as large, open tension fractures of the type seen in Thingvellir and in the rift zone in general.

As in Mulafjall (last stop), the faults here are also generally closed. This is to be expected since, again as in Mulafjall, we are looking at the lava flows at depth of 800–900 m below the elevation that the surface had when most of the normal faults formed. The small openings seen on the joints and some of the faults, of the order of centimetres on most of these, are primarily related to events that happened once the rocks and the fractures had drifted out of the rift zone through spreading.

In the present case, these small openings are, as we have discussed, primarily related to uplift and associated stress changes following the erosion that formed the valleys and the fjord.

The message to take away here, as regards rifting at divergent plate boundaries such as in this part of Iceland, is that **spreading** is partly accommodated through the formation of **normal faults** and partly through the formation of **dikes**. Where dikes are common, as at stops 2 and 3 today, faults are rare and small. By contrast, where dikes are rare, as at stops 5 and 6, normal faults, some with comparatively large displacements, are more common. It is easy to understand why these relations exist when we consider that the long-term spreading rate associated with the plate movements in a given area—a volcanic zone or a rift zone—is close to constant. That means that, over tens of thousands or hundreds of thousands of years, the spreading-related rate of dilation or extension of the crust at a given locality, such as Southwest Iceland, is generally the same. It follows if large part of that spreading is taken up by dikes, there is very little left for the normal faults to contribute, and vice versa.

11.8 Table Mountain Hvalfell

We now drive on to the east and into the valley of **Botnsdalur** where we make our **seventh stop** (**7**). From Road 47 you drive to the east on to the gravel road that leads into the valley. There are many parking places here and I suggest you make at least two stops: one at the beginning of the gravel road and another where the road ends. I suggest this so that you get a very good overview of the main attraction here, namely the beautiful table mountain **Hvalfell** (**Whale Mountain**). This is a close-to-perfect example of a classical table mountain (Figs. 11.22 and 11.23), which, as you recall, are generated in basaltic eruptions within a thick ice sheet or cap. Hvalfell is somewhat larger than Hrafnabjörg which we saw at Thingvellir (Fig. 6.5) and has more the classical shape of a table mountain. However, Hvalfell is presumably from the second last ice period, or possibly around 150 thousand years old (similar to Arnmannsfell at Thingvellir), whereas Hrafnabjörg formed during the last ice period and is about 20 thousand years old.

Hvalfell has the ideal shape of a table mountain. It rises to an elevation of about 850 m above sea level. When it formed, it partially blocked the river Botnsa (Botnsá), an event that contributed, together with accumlation of meltwater from glaciers at the end of the Ice Age, to the formation of one of the deepest lakes in Iceland on the east side of the mountain (not seen from the present stops). The maximum depth of the lake is about 160 m, making it the third deepest lake in

Fig. 11.22 View east, the table mountain Hvalfell in the valley of Botnsdalur. Located in Fig. 11.16. A typical table mountain. Compare Fig. 11.23

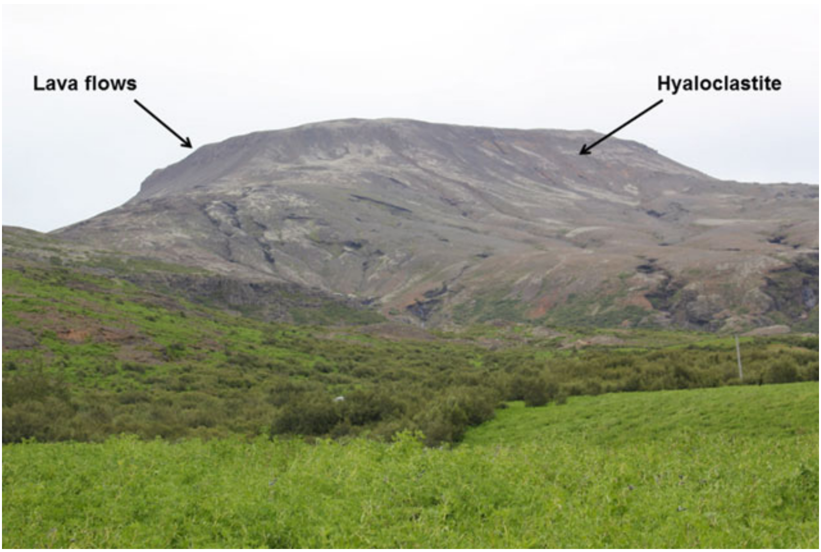

Fig. 11.23 Close-up of the table mountain Hvalfell. View east, like other table mountains it is primarily composed of hyaloclastite with a cap of basaltic lava flows (indicated)

Iceland. It follows in depth Jökulsarlon a Breidamerkursandi (Jökulsárlón á Breiðamerkursandi)—a famous lagoon in Southeast Iceland, a popular tourist area, and used for some scenes in films, e.g. James Bond—which has a maximum depth of about 260 m and Öskjuvatn (Lake Öskjuvant) in the Askja Caldera, Central Iceland, whose maximum depth is about 220 m. In the river Botnsa, which originates in Lake Hvalvatn and follows the northwest slopes of Hvalfell, is the second highest waterfall in Iceland (second only to Mosarfoss, Mosárfoss, in Southeast Iceland, which has a fall of about 240 m). This is the waterfall **Glymur**. The total drop is about 198 m into a narrow canyon. It is popular to hike to the waterfall from the second and final parking at our sevent stop (the hiking distance is about 5 km).

We do not hike to Glymur, but rather enjoy a close-up of **Botnsulur (Botnsúlur)**, the mountain we had seen earlier today and also on the way to Thingvellir. Botnsulur is noticeable for its high peaks (Fig. 11.24), all of which exceed 1000 m above sea level, the highest one being 1093 m. Botnsulur is a somewhat eroded hyaloclastite mountain and was, therefore, presumably taller at its formation, which may have occurred in a subglacial eruption some 250 thousand years ago.

Fig. 11.24 Botnsulur, an eroded hyaloclastite mountain with several noticeable peaks. View east, the mountain is also seen in Figs. 11.16 and 4.12

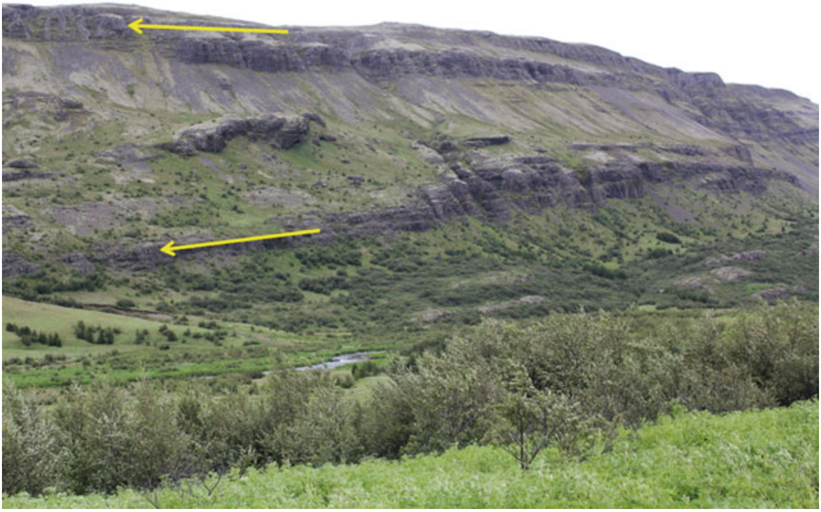

Fig. 11.25 Gently inclined (dipping) lava flows on the north slopes of the mountain Mulafjall (Múlafjall), located in Fig. 11.16. View southwest, the inclination or dip of the lava flows increases with greater depth in the lava pile, that is, deeper in the crust, as is explained in Fig. 4.6 and discussed in Chap. 4. The inclination or dip is to southeast, towards the West Volcanic Zone

The valley of **Botnsdalur** is an area of great natural beauty, with the river Botnsa (Botnsá) flowing between mostly green banks (Fig. 11.25). Above the river, on the northern slope of the mountain Mulafjall (in our sixth stop we observed the graben seen in the southern slopes of Mulafjall) you see gently inclined lava flows, sloping down to the east towards Botnsulur and the volcanic rift zone at Thingvellir.

We have now completed the main part of the Hvalfjördur excursion. There are two options as to the way back. One is to drive back the same way as you came, that is, along the southern side of Hvalfjördur to Road 1 and then to Reykjavik. The other option is to continue from the seventh stop to Road 47 but drive along the northern side of Hvalfjördur until Road 47 meets Road 1. From there you drive along Road 1, south and then southwest, using the undersea Hvalfjördur Tunnel to reach the south side of Hvalfjördur and then to Reykjavik. If you follow the second option, then you will be driving through the centre of the Hvalfjördur Volcano. We have already described the main mountains you drive along. These are, from east to west, Thyrill, Brekkukambur, and Thufufjall. There are many things to see on the way, but as regards the general geology, these have already been discussed. As you approach the tunnel, the lava pile in mountain Akrafjall is seen sloping towards you (to the east), while the much taller and larger mountain Skardsheidi (Skarðsheidi) is in the north (cf. Fig. 3.5). Both mountains are primarily composed of lava flows and formed through erosion and uplift in the way we have already discussed today.

12 Reykjavik-Hengill

There are many good reasons to make an excursion to the Hengill Volcano (Fig. 12.1). It is an active volcano—indeed the centre of the Hengill Volcanic System (Figs. 2.2 and 2.3) to which the Thingvellir Graben belongs—whose most recent eruption was only 2000 years ago. It has many active geothermal fields and a geothermal power plant. But the main reason to visit Hengill is its unique and, depending on your taste, **beautiful landscape**. The landscape is characterised by narrow valleys separated by even narrower ridges. Some of the ridges are constructional, that is, hyaloclastite ridges formed in eruptions under a glacier. Many of the ridges and valleys in-between them, however, are mostly generated by large normal faults—including the largest ones you are likely to see close-up in Iceland. In addition, one of the proposed stops on the excursion offers a excellent and easily accessible view of Lake Thingvallavatn. Of course much of what Hengill has to offer depends on making the excursion when the visibility is good.

We drive out of Reykjavik along Road 1, heading to the east. Just after we pass the pseudocraters Raudholar (Figs. 9.8 and 9.9), we turn to the left (to the northeast) onto Road 431, which is the road we follow to Hengill. The road is for the main part along more or less a completely flat ground. This ground is the surface of the lava shield **Mosfellsheidi (Mosfellsheiði)**, which formed during glacial-free periods (interglacials) some hundreds of thousands of years ago. In fact, Mosfellsheidi may be of an age similar to that of many of the shields that form part of the landscape in Reykjavik (Chap. 3). On our way to Thingvellir, the road is, for a while, along the same lava shield (Chap. 4). Running parallel with the road is a large pipe conducting hot water from the Nesjavellir geothermal power plant at Hengill.

The landscape of Hengill is unique and spectacular. Parts of it are best seen in context from the air, because many of the structures are so large that it is difficult to appreciate them and fathom on the ground. So for many of the structures, I show

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GeoGuide, DOI 10.1007/978-3-319-55152-4_12

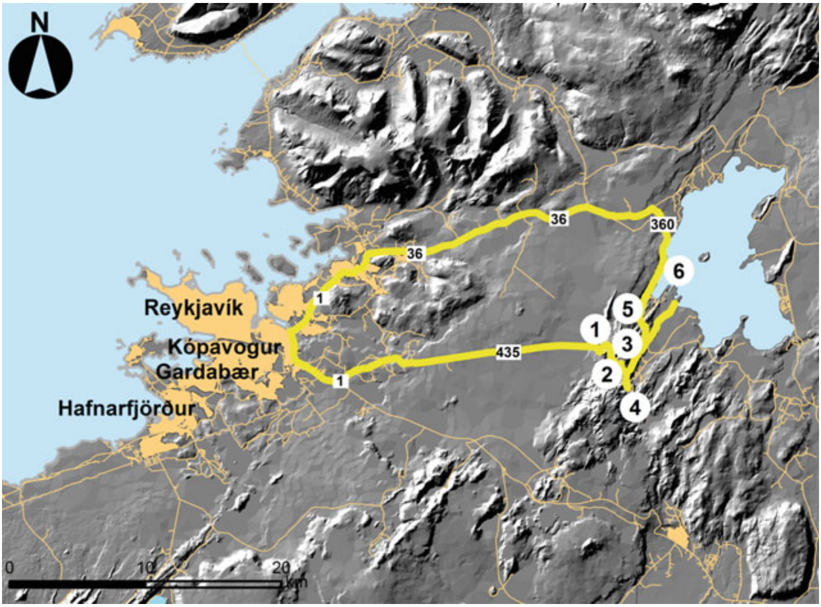

Fig. 12.1 Map showing the excursion to the volcano Hengill. The suggested main stops are indicated by encircled numbers, 1–6

them on the ground, as you see them in the excursion, and then put them into a wider context with other structures, as seen on aerial photographs.

12.1 Hyaloclastite Ridge

We make the **first stop (1)** as soon as we enter the hyaloclastites of Hengill proper. Here is a good parking place from which we view a small part of a much larger hyaloclastite ridge (Fig. 12.2). We mentioned hyaloclastite ridges briefly in Chap. 6, but here we can discuss them in greater detail, particularly with reference to Fig. 12.3, which is an aerial view of a large part of the hyaloclastite ridge **Sköflungur**. There are two main differences between a hyaloclastite ridge and a table mountain (Figs. 6.5, 6.6, 11.22 and 11.23). The first is that during the ridge formation the eruption conduit never reaches the surface of the lake (formed through melting of the ice sheet, Fig. 6.6), so that no lava cap forms on the top of the hyaloclastite. This is in contrast to a table mountain where the conduit eventually reaches above the surface of the lake so that subaerial (meaning under or in air, that is, on dry land rather than under lake water, subaquatic, or sea water, submarine) lava flows can form. The second difference is that when generating a ridge the eruption does not become noticeably confined to one or few conduits; the ridge is primarily formed when a large part, or even the entire, conduit (the volcanic fissure) is erupting. Again, this contrasts with table mountains where the eruption becomes confined to a short part or segment of the volcanic fissure, often in fact a single part (a conduit or crater) and continues from there during most of the eruption.

There are recent eruptive analogies to the hyaloclastite ridge we see here. Perhaps the most relevant is the **Gjalp (Gjálp) eruption** (Fig. 1.4) in Vatnajökull ice cap (located in Fig. 2.2) in 1996. This eruption, mentioned in Chap. 1, produced a hyaloclastite ridge (in fact a ridge apparently on top of an older ridge) but the conduit did not really reach well above the water level in the melted and highly elongated lake, so that no subaerial lava was erupted. The resulting ridge was about 6000–7000 m long and over 200 m high. The Gjalp ridge is thus somewhat larger than the Sköflungur ridge which is about 4500 m long and rises to a maximum of some 180 m above its surroundings—to an elevation of about 420 m above sea level (the adjacent surface is mostly between 200 and 300 m above sea level).

The Sköflungur ridge is geologically young and most likely formed during the late stages of the last glacial period. That is, Sköflungur is of an age similar to that

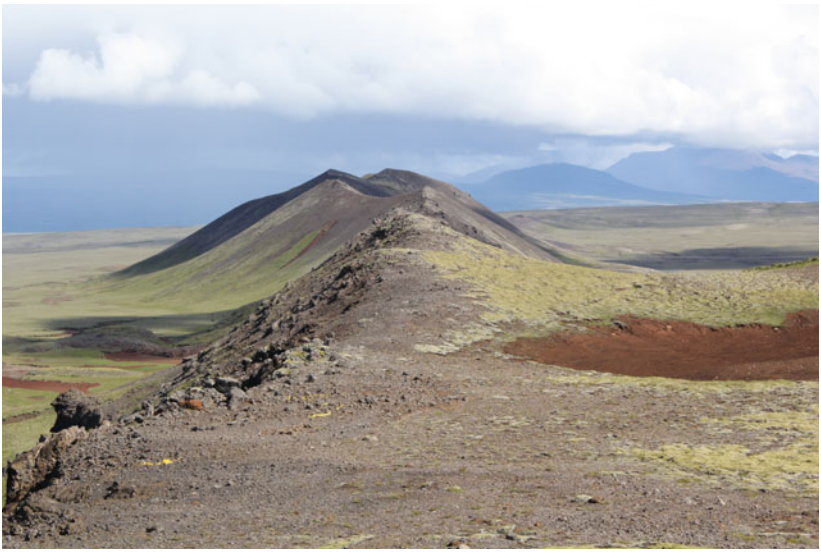

Fig. 12.2 Part of the hyaloclastite ridge Sköflungur. View north, compare Fig. 12.3

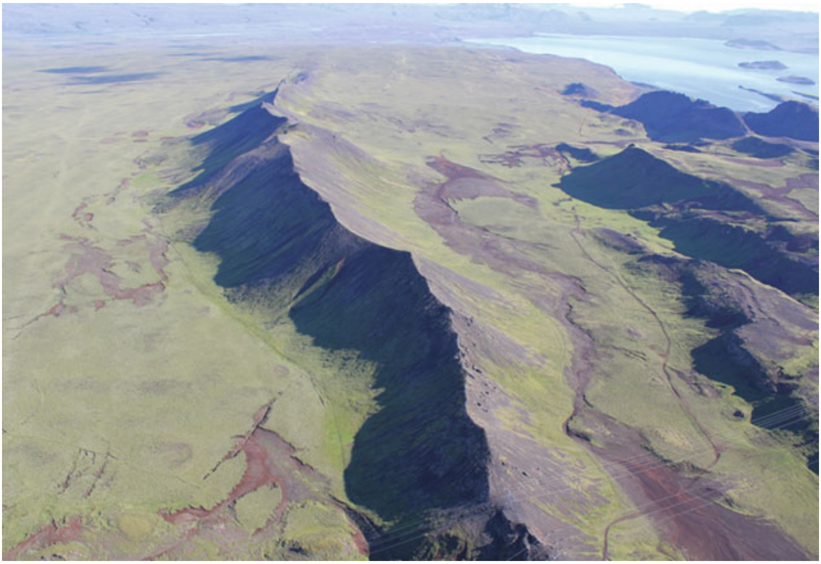

Fig. 12.3 Aerial view of the northern part of the hyaloclastite ridge Sköflungur. View north, part of Lake Thingvallavatn is also seen

of the table mountain Hrafnabjörg at Thingvellir, or about 20 thousand years old. Its direction is remarkable in that the southern half of the ridge trends north, whereas the northern half trends northeast (Figs. 12.2 and 12.3). Thus the direction of the northern part is entirely parallel with that of most faults (and ridges) in the Hengill Volcanic System (Fig. 2.3), whereas the direction of the southern part, close to the Hengill Volcano itself, is different. This change in direction with distance from a central volcano is commonly seen in volcanic systems. The central volcanoes, in particular their shallow magma chambers, **change** the regional **stress field** that controls the formation of normal faults, tension fractures, and regional dikes away from the volcano, so as to form a local stress field around the volcano itself. The local field is normally much different from the regional one—in particular, the local field is more complex—and gives rise to the inclined sheets and radial dikes we discussed during the Hvalfjördur excursion (Fig. 11.3; Chap. 11). The local stress field is, for example, responsible for the formation of collapse calderas (Figs. 4.10 and 11.14). In fact, there are many north-directed earthquake fractures in the vicinity of Hengill, as well as east-west directed, all of which relate to the local stress field within the volcano.

12.2 Feeder-Dike

The **second stop (2)** consists of two parts. The first part is a brief stop in a valley to get an overview of its main structure on the ground (Fig. 12.4). The second part is a stop at a feeder-dike to a hyaloclastite ridge, the dike being eroded on one side of the road, so as to leaving an open cavity, a sort of cave (Fig. 12.5), while being un-eroded on the other side of the road where it is connected to a small intrusion in the hyaloclastite (Fig. 12.6). The dike is clearly a **feeder (supplier of magma)** to part of the hyaloclastite ridge. It is eroded where it turned out to be weaker than the surrounding rock, hence the cave. On the other (north) side of the road, however, the dike and the host rock are of similar resistance to erosion, so that the dike is neither a little ridge or wall (standing above its surroundings) as many of the dikes in Hvalfjördur (Figs. 11.8, 11.9 and 11.10) nor a depression or cave.

To put these ground observations into perspective, let us view the second stop and its surroundings from the air (Figs. 12.7 and 12.8). Here we see the ridges that

Fig. 12.4 Close-to vertical fault plane in the Hengill area. View northeast, the normal fault cuts through hyaloclastite. Compare Figs. 12.7 and 12.8

Fig. 12.5 View south, part of a feeder-dike. The dike material has largely disappeared, presumably mainly through erosion, resulting in a close-to empty dike-fracture, a cave, with an opening of about 1.5 m. The central part of the dike may also have 'drained away' at the end of the eruption, as is common during the formation of pit craters and similar structures (compare the discussion about Kerid, Sect. 9.2). The surrounding rock, the host rock, is hyaloclastite

characterise the landscape very clearly. In particular, we see the ridge that is cut through by the road and how the dike has formed a cave on the south side of the road. All the ridges have basically the same direction, that is, northeast, as do most faults and volcanic fissures in the Hengill Volcanic System. The landscape is controlled partly by hyaloclastite **ridges** and partly by **normal faults** (Figs. 12.3 and 12.4), as is also indicated in Figs. 12.7 and 12.8. More specifically, the hyaloclastites are all formed in eruptions under the ice sheet of the last ice period, and thus mostly between 20 thousand and 100 thousand years old. While the ridges are formed in eruptions and are thus constructional landscape features, it is clear that many have too steep walls to be formed in that way. We saw this already for the wall in Fig. 12.4, and even better in Figs. 12.7 and 12.8. These close-to-vertical walls are earthquake fractures or, more specifically, normal faults, which have to a

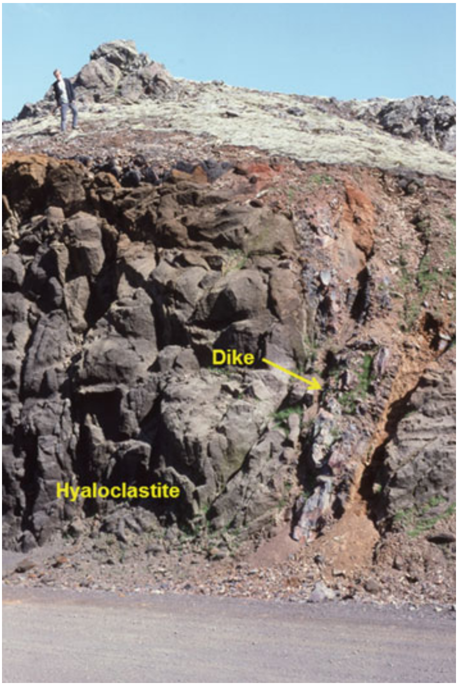

Fig. 12.6 View north, the same feeder-dike as in Fig. 12.5, but here uneroded. The dike varies in thickness, but is on average around 1.5 m. The host rock is mostly hyaloclastite

large degree shaped the landscape of Hengill and its surroundings. I indicate some of the fault walls, fault scarps, in Fig. 12.7. This part of the Hengill area is named **Dyrafjöll (Door Mountains)**, a name presumably derived from the many 'door-looking' fault walls. As we discuss later in the excursion, the geothermal springs are also related to the faults and dikes, that is, the springs tend to follow faults and dikes.

12.3 Geology of Hengill—Overview

We now drive on to the **third stop (3)**. Here you can either park the car next to the two white cylinders and walk a few tens or metres to the north along the gravel path or, in case you have a four-wheel drive car, drive the path up the hill. Once on the hill, and assuming good visibility, the view is spectacular. Here we first discuss the features that can be seen on the ground and then, for a wider perspective, as seen from the air. Looking first to the northeast (Fig. 12.9), you see almost the entire **Lake Thingvallavatn** with its islands, and in the distance the mountains Armannsfell, Skjaldbreidur, Hrafnabjörg, and others (Fig. 5.14, Chaps. 5 and 6). Closer by, on the south shore of the lake, is the youngest lava flow in the entire system, the 2000-year-old pahoehoe lava flow **Nesjahraun**. Essentially the same view from the air gives a different perspective (Fig. 12.10), but shows basically the same structures. The mountains on the east side of Lake Thingvallavatn, namely **Arnarfell** (at the lake) and **Midfell (Miðfell)** close to the lake, both of which are hyaloclastite ridges, are more easily seen on Fig. 12.10 than on Fig. 12.9.

Looking next to the east, we see part of the **Nesjavellir Power Plant** located in a graben, whose main eastern normal faults are very clear (Fig. 12.11). The movements or displacements on these faults reach tens of metres. To see these large faults in perspective, I also provide aerial views (Figs. 12.12, 12.13 and 12.14). Figure 12.12 gives a view to the southwest and the faults seen in

Fig. 12.7 Aerial view showing the location of the feeder-dike in Figs. 12.5 and 12.6, and some close-to vertical walls, forming parts of normal faults (see also Fig. 12.4). Part of the mountains Blafjöll (Bláfjöll) are also seen (compare Fig. 9.7 of Blafjöll)

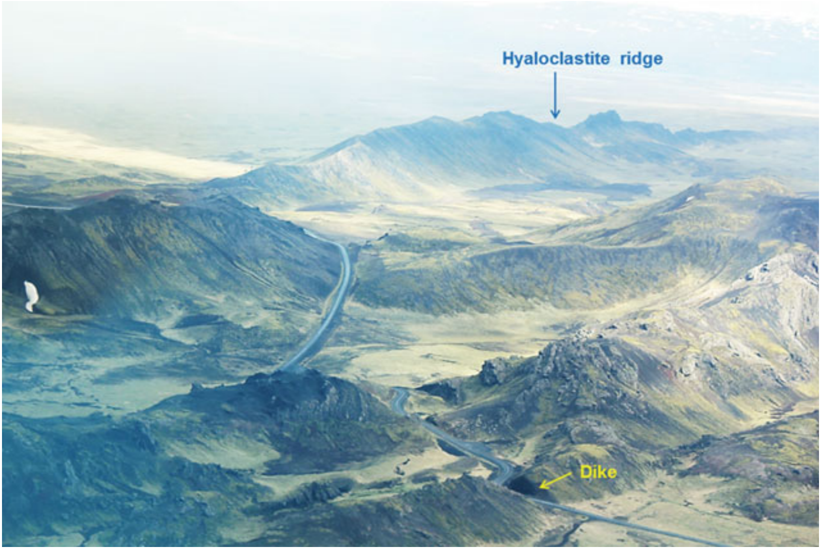

Fig. 12.8 Aerial view, from a different perspective, of the location of the dike in Figs. 12.5, 12.6 and 12.7. View northwest, the hyaloclastite ridge Sköflungur (Figs. 12.2 and 12.3) is also seen

Fig. 12.11, as well as the third stop from which Fig. 12.11 is taken. On Fig. 12.12 it is clear the Nesjavellir Power Plant is located inside a narrow graben, the **Hengill Graben**. Several of the main faults seen on Fig. 12.11 are indicated on Fig. 12.12 which, however, is taken late in the day so that large part of Hengill itself and the faults on the western side of the graben are in a shadow. The geothermal fields tend to follow the faults—as can be seen from the vapour coming up from the south-western parts of some of the faults, and discussed in more detail below.

To see the faults in greater detail and how they develop over time, I provide two additional aerial photographs, namely Figs. 12.13 and 12.14. In both of these the view is to the northeast. On Fig. 12.13 the fault walls in the shadows are those seen closest to the power plant on Figs. 12.11 and 12.12 and form the deepest part of the Hengill Graben. The Hengill area produces many earthquakes, but most are so small (less than magnitude 2) that you would not feel them. For example, in the period from 1994 to 2000 there was an exceptional activity in Hengill, with uplift or doming of its surface of the volcano (Fig. 12.15) by several centimetres. During this period of about six years around 100 thousand earthquakes occurred in Hengill and the nearby areas. This was exceptional, however, and as said most of the earthquakes are so small that they are not felt by people.

Nevertheless, the faults are active, as is easily demonstrated by tracing them into the youngest lava flows in the area, particularly Nesjahraun. This we do in Fig. 12.14. Here we view the continuation of the large faults in the hyaloclastite layers into the 2000-year-old Nesjahraun as well as the 5700-year-old **Hagavikurhraun (Hagavíkurhraun)**. What you notice immediately, particularly when you compare this figure with Figs. 12.12 and 12.13, is how small the fault movement, or displacement of the fracture walls, is in these younger lava flows. The displacement changes abruptly from many tens of metres to, at most, a few metres as the faults enter the young lava flows.

Why is the change in displacement so abrupt? The answer is that normal faults in the volcanic zones of Iceland are active over, commonly, tens or hundreds of thousands of years (For comparison, the largest faults on Earth, may be active over many millions of years.) So when a new lava flow is erupted that partly or totally

Fig. 12.9 View northeast over Nesjahraun, Lake Thingvallavatn, its islands, and the surrounding mountains. For names of the islands and mountains, see Figs. 5.14 and 12.10

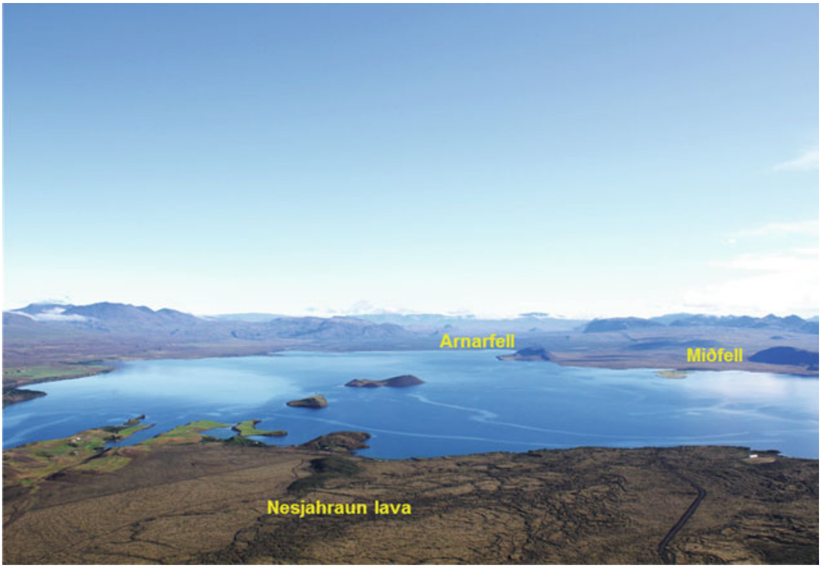

Fig. 12.10 Aerial view of Nesjahraun (compare Figs. 12.14 and 12.21). Also indicated are the main hyaloclastite mountains east of Lake Thingvallavatn, namely Arnarfell and Midfell (Miðfell). For the names of other mountains and islands seen here see Fig. 5.14

covers an existing and still active fault segment then, following the lava emplacement and depending on its thickness, for a while no fracture may be seen at the surface where the lava has covered the fault walls. But as spreading and earthquakes continue, the old fault walls break through the new lava flow. The movements of the walls as seen in the young lava flow, however, are then only those that have happened after the lava was formed. It follows that the fault displacements we see in Nesjahraun and Hagavikurhraun are only those that have happened during the past 5700 years for Hagavikurhraun and the past 2000 years for Nesjahraun since the lavas formed. Because the spreading rate is low, less than a centimetre per year, the rate of fault movement in associated earthquakes is also low, hence the **small displacements** on the active faults.

Another thing you may notice is that there are no large wide-open pure tension fractures in the young lava flows, as are so common in the Thingvellir Graben (Figs. 5.1, 5.11, 5.12 and 5.13). In Nesjahraun and Hagavikurhraun all the fractures are clear normal faults. The primary reason for this difference in fracture geometry is that the pure tension fractures at Thingvellir are just that—pure tension fractures. Many, perhaps most, are not related to any normal faults at depth—those that are, may form en echelon systems (Fig. 5.16). By contrast, the fractures in the Nesjahraun and Hagavikurhraun are clearly all the surface expressions of buried normal faults—as is clear from their continuation to the south into the faults in the hyaloclastite layers (Figs. 12.13 and 12.14).

A second reason for the difference in appearance relates to the properties of the lava flows. At Thingvellir the fractures dissect a comparatively stiff ('hard') and very thick pahoehoe lava flow. Tension fractures easily form along the existing columnar (cooling) joints that extend to the surface of the flow. The Nesjahraun and Hagavikurhraun lavas, however, are partly aa lava flows, with a comparatively 'soft' breccia, that is, loose, broken rock surface layer. This soft layer does not easily develop pure tension fractures, but can clearly develop faults. The analogy would be wet and dry sand. We all know from childhood that suitably moist sand

Fig. 12.11 View east, showing part of the geothermal power plant and many of the normal faults forming the eastern part of the Hengill Volcanic System—in particular, the Hengill Graben. For aerial views of these faults and the surrounding areas, see Figs. 12.12, 12.13 and 12.14

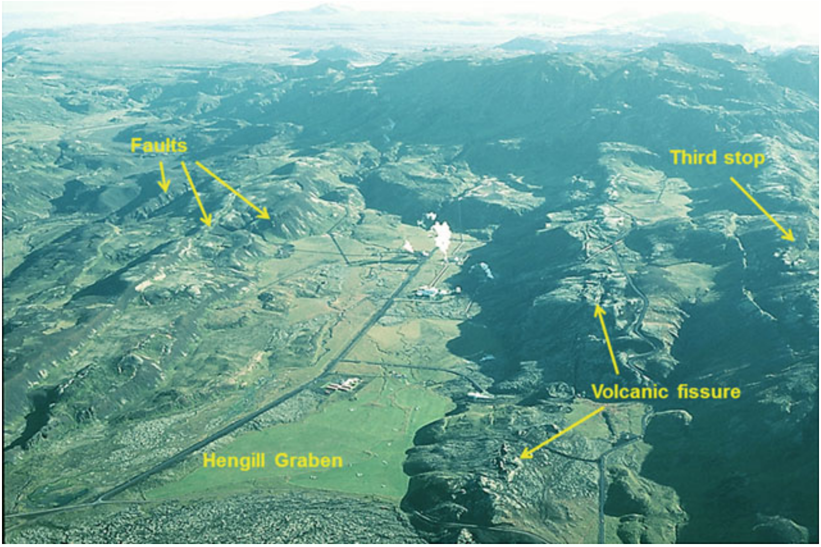

Fig. 12.12 The deepest part of the Hengill Graben (indicated). Aerial view to the south, this part is mostly from about 700 m to about 1800 m in width. The photo is taken late in the day, hence the shadows. The craters on the ridge to the right of centre (to the right of the grass field and the power plant) are part of the Nesjahraun volcanic fissure. Thus, the fissure, and its feeder dike, hit the surface not on the lowest ground but by as much as 150 m higher, showing that there was probably an existing weakness, a fault or a fracture, that guided the magma-dike in this part. Also indicated is the location of the third stop

can fracture—form tension fractures—even if we do not know the name of the fractures. That is because the water gives the sand cohesion, which in geological terms is the same as tensile strength. That, again is the reason that we can build sand castles from moist or wet sand. Dry sand, by contrast, can only fail it wall-parallel movement, that is, in geological terms, through faulting. Dry sand has very little cohesion in tension, or tensile strength, which is the main reason we cannot build sand castles from dry sand.

We also notice how winding or curved some of the faults are, both in the older rocks (Fig. 12.13) as well as in the young lava flow (Fig. 12.14). This wavy geometry is typical of normal faults in rift zones (Figs. 2.5 and 2.6). It is partly related to how they grow through linking up of, originally offset, segments or parts. Partly, however, it relates to the variations in the strengths and stresses in the rock. Stresses vary through the crustal rocks depending on their properties and the loading (here the plate-tectonic forces). Furthermore, the fractures eventually form, for a given stress, where the rock is weakest. The weakness or strength of rocks varies considerably, even within a single lava flow. As times goes on, the small faults in the Nesjahraun and Hagavikurhraun lava flows will gradually **become larger**, that is, the wall movements will continue and make the displacements larger. This process continues so long as these faults are favourably located so as to concentrate stress and generate earthquakes. Eventually, however, the faults drift out of the main rift zone and become **inactive**. In the Hengill Volcanic System that drifting takes several hundred thousand years.

Turning back to our parking place, namely the third stop, we have already looked to the north and to the east. Now, we look to the south, that is, to **Hengill** itself (Fig. 12.15). Here the main graben cutting through the top of the mountain is clear. Furthermore, the shape of the mountain reminds us of that of table mountains (Figs. 6.5, 11.22 and 11.23). Hengill, however, is not a table mountain in the normal sense of being formed in a single eruption—or, at most, a few eruptions. By contrast, Hengill is a **central volcano** which has formed in numerous eruptions

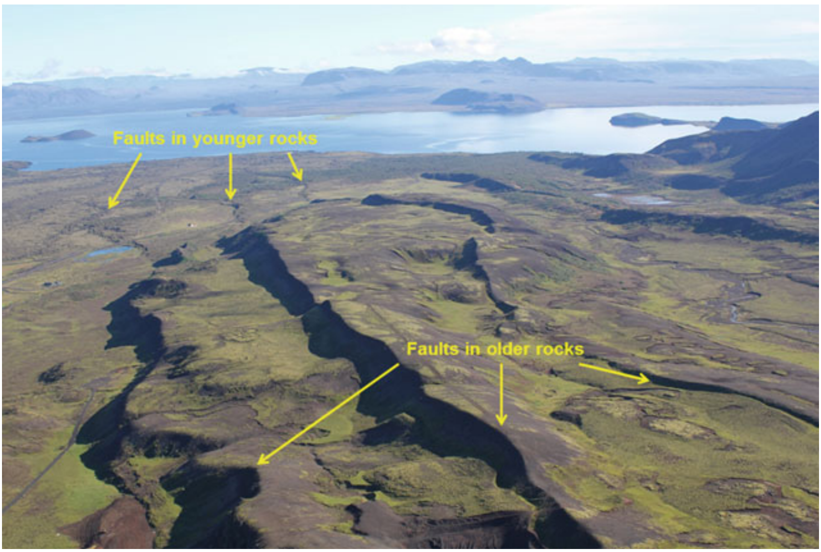

Fig. 12.13 Aerial view of some of the faults seen from the ground in Fig. 12.11. These are the main faults on the east side of the Hengill Graben. Some of them are also seen in Figs. 12.12 and 12.14

Fig. 12.14 Fractures extending from hyaloclatite into Holocene lava flows. There is an abrupt decrease in the displacement or subsidence along the normal faults as they pass from the old hyaloclastite layers into the young lava flows. Nesjahraun is about 2000 years old, whereas Hagavikurhraun (Hagavíkurhraun) is about 5700 years old

during the past several hundred thousand years. In fact, the oldest rocks in the Hengill area have an age of about 800 thousand years. Hengill itself rises to a maximum of just over 800 m above sea level, and is thus close to 700 m above the deepest part of the Hengill Graben and the surface of Lake Thingvallavatn.

We have already seen Hengill and the eastern part of the fault and graben system from the air (Figs. 12.12, 12.13 and 12.14). Here we can add the relation between the largest faults on the western side of the fault system and Hengill (Fig. 12.16). The graben in the top-area of Hengill is, again, very clear. The main structure, however, is the large fault at the western boundary of the entire fissure swarm. This fault, whose name is **Jorukleif (Jórukleif)**, is a continuation of Almannagja at Thingvellir and forms the western boundary of the entire Hengill-Thingvellir graben. We discuss the fault in detail at a later stop, as we drive to the northeast from the Hengill area. Now, however, we move on to take a better look at the geothermal field and the power plant.

12.4 Geothermy

So we drive on to the next place, the **fourth stop (4)**. Here is a good, small parking place and several paths that you can walk both for a great view of the power plant itself and the graben within which it is located as well as to the geothermal fields and wells. Taking first the path to the east, we view the power plant and the nearby normal faults (Fig. 12.17). The power plant produces some 120 MW (120 million watts) of **electricity**, in addition to supplying part of the Capital Region (Chap. 2) with about 1100 l/s (litres per second) at a temperature of about 84 °C for **space heating** (primarily the heating of buildings). In fact, the big pipe we saw on our way to Hengill transports that hot water to the Capital Region. The Nesjavellir Power Plant is the second largest geothermal plant in Iceland, second only to the nearby Hellisheidi (Hellisheiði) Power Plant, whose electricity production is just above 300 MW of electricity (Chap. 9).

Fig. 12.15 View south, showing the boundary faults of the central part of the Hengill Graben where it passes through the Hengill Volcano itself. This is the view from the third stop. The same boundary faults are also seen on the aerial view in Fig. 12.16

Fig. 12.16 Aerial view of the fault Jorukleif (Jórukleif) and the Hengill Volcano. View south, part of cove of Hestvik (Hestvík) is also seen (see also Fig. 12.23 of Hestvik)

We now walk back to the small parking place and drive further 300 m to the south to the main parking place from which there is a footpath to many of the wells and geothermal fields. There are many **wells** to be seen. Those currently producing steam can be very noisy (Fig. 12.18). The **geothermal field** in the Hengill area is one of the largest in Iceland—perhaps in total somewhere around 100 km^2, or larger than the entire Lake Thingvallavatn (82 km^2). Most of the hot springs are clearly related to faults (Fig. 12.19). This is as expected because active faults transport fluids, including geothermal fluids, more easily than the surrounding rocks. One way that the geothermal fields at Hengill are maintained is through **earthquake activity**. Recall that the fields and erupting geysers of the Geysir area are maintained in the same way. In fact, and perhaps surprisingly given the large distance between Hengill and Geysir, both relate to the stresses and earthquakes associated with the main earthquake zone in this part of Iceland, namely the South Iceland Seismic Zone, discussed in Chaps. 9 and 14.

Now what are the **heat sources** for the large geothermal field in Hengill and its surroundings? The answer is magmatic intrusions of the types we have discussed in the earlier chapters. First, there are swarms of dikes and inclined sheets, as were observed in Esja and on the coast and in the mountains of Hvalfjördur (Chap. 11). In addition, we have seen at least one dike on this excursion to Hengill, namely at the second stop (Figs. 12.5 and 12.6). During exploration of the Hengill geothermal system, many drill holes or wells (Fig. 12.18) dissect dikes and inclined sheets, similar to those in Figs. 11.2, 11.3, 11.6, 11.7, 11.8, 11.9, 11.10 and 11.15. The deepest drill holes reach depths of close to 2300 m, and they show that the number of dikes and inclined sheets increases as we drill deeper into the volcano. In fact, at depths between 1500 and 2000 m below the surface, almost the **entire rock** is made of **dikes and inclined sheets**. This is in accordance with studies in the most deeply eroded extinct central volcanoes in Iceland, such as in Southeast Iceland. There the erosion by the glaciers allows us to study the cores or the hearts of the old volcanoes to depths of 2000 m. What we normally see at the core are shallow magma chambers (plutons) and at their margins nearly all the rock is composed of inclined sheets and radial dikes. It is on these observations that Fig. 11.3 is based, and this general picture of Icelandic central volcanoes is confirmed by the drill holes in Hengill and studies of many other active and extinct volcanoes.

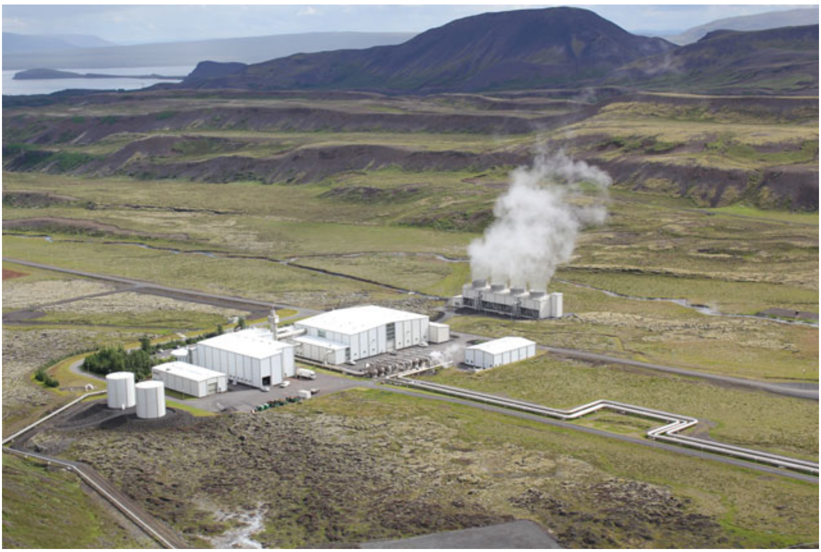

Fig. 12.17 View northeast the geothermal power plant at Nesjavellir. Also seen are the normal faults on the east side of the Hengill Graben and parts of the Nesjahraun lava, close to the power plant. Compare Figs. 12.11 and 12.12

Fig. 12.18 One of the geothermal wells in the Hengill area

Thus, dikes and inclined sheets are heat sources for geothermal fields. But they are not the only ones. A **shallow magma chamber** is another one. We have already seen a fossil shallow chamber on our way to Thingvellir, namely Stardalshnjukar (Stardalshnjúkar), a thick gabbro body (Figs. 4.7, 4.8 and 4.9), and it is likely that a similar magma chamber exists in Hengill. Such a chamber in Hengill, however, is most likely deeper in the crust than Stardalshnjukar, most likely at the depth of about 2 km. How do we know that? Simply because at that depth nearly all the rock is composed of inclined sheets, and these intrusions invariably derive from a shallow magma chamber. Thus, when nearly all the rock consists of intrusions, the source magma chamber cannot be far below the intrusions. Although the chamber has not been found (by indirect geophysical methods such as earthquake locations and geodetic studies), it is almost certainly with a top somewhere at the depth of 2–3 km.

It is also almost certain that the shallow magma chamber is supplied with magma from a deeper reservoir, perhaps at a depth of 15 or 20 km (Figs. 4.1 and 11.3). We have already discussed such reservoirs in connection with the volcanic systems on the Reykjanes Peninsula (Fig. 2.4), and many geochemical and geophysical studies in recent years have indicated such reservoirs beneath many volcanic systems in Iceland. The deep-seated reservoir provides magma for most of the eruptions that occur in the Hengill Volcanic System far from the Hengill Volcano itself. The volcanic fissure issuing the 2000-year-old Nesjahraun lava, for example, may have originated, partly at least, from the deep-seated reservoir.

12.5 The Nesjahraun Lava

Which brings us to the **fifth stop (5)**, focusing on the **Nesjahraun lava**. The lava field is large, and the island of **Sandey** formed in the same eruption. In fact, the volcanic fissure feeding the eruption is extremely long, or about **30 km**. It extends from southwest of Hengill to the northeast to Sandey (Fig. 12.20). However, the

Fig. 12.19 Geothermal springs, hot springs, are aligned along one of the faults in the Hengill area. Most hot springs, including erupting ones, are related to active (earthquake) faults, whose slips maintain the permeability for the geothermal fluids (see Chap. 7)

Fig. 12.20 Close-up aerial view of the island of Sandey, formed in the same eruption as Nesjahraun, about 2000 years ago

fissure and its feeder-dike are discontinuous, with many kilometres between some of the fissure parts or segments. For example, the feeder-dike did not cut through the Hengill mountain itself so that the fissure does not cross Hengill. Given the length and location of the volcanic fissure, it is likely that its magma derived partly, perhaps entirely, for the deep-seated reservoir. The feeder-dike may have hit the shallow magma chamber beneath Hengill (compare Fig. 11.3), and a radial dike or an inclined sheet may have been injected from the shallow chamber during the eruption. Clearly, however, that dike/sheet never reached the surface but rather became arrested (stopped on its way to the surface) at some depth in Hengill.

It is, in fact, common to have **simultaneous eruptions** from a shallow chamber and a deeper source reservoir, where magma from the reservoir goes partly straight to the surface, and partly enters the shallow chamber and triggers eruption from the chamber (Fig. 11.3). It is even more common, however, that the dikes/sheets become arrested and never reach the surface to feed eruptions. It is estimated that only some 10–20% of all injected dikes/sheets in Icelandic volcanoes ever reach the surface to erupt. This means that 80–90% of the injected dikes/sheets become **arrested**, stop, on their way to the surface, usually at contacts between layers in the volcano. We discuss this topic of dike/sheet arrest again in relation to the events that resulted in the 2010 the Eyjafjallajökull eruption (Chap. 14). As for Nesjahraun the evidence is that it was supplied with magma from the deep reservoir through vertical dike movement or propagation to the surface. Given the great length of the volcanic fissure, the dike is likely to have been comparatively thick. Notice, however, that the dike is discontinuous and composed of widely separated segments. For such dikes, the dimensions of individual segments control the thickness, rather than the dimensions of the entire dike. The segments, however, are many kilometres long, so that the dike is most likely many metres thick—perhaps similar to some of the thicker dikes in Hvalfjördur (Chap. 11).

As for the lava flow itself, we have already seen the Nesjahraun lava on several aerial photographs (Figs. 5.14, 12.10 and 12.14). The next aerial view is to the southwest and shows the Nesjahraun lava flow as well as its volcanic fissure (Fig. 12.21). Here we see clearly that part of the volcanic fissure is on the ridge or

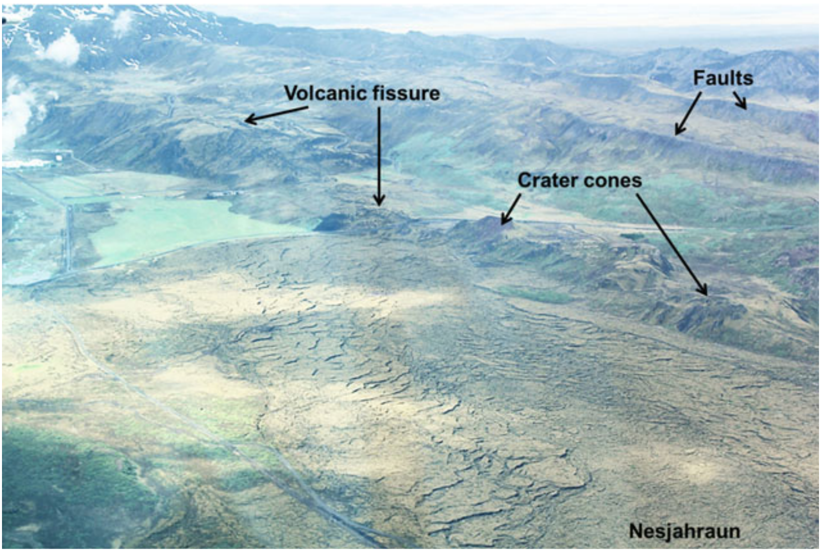

Fig. 12.21 Aerial view southwest showing the main volcanic fissure that issued Nesjahraun (identified) some 2000 years ago. Some of the crater cones are also identified. The photograph shows that parts of the volcanic fissure, and its feeder-dike, reached the surface some 150 m above the lowest ground in the Hengill Graben (the site of the green grassfield and the power plant). Some of the western faults of the Hengill Graben are also indicated

plateau to the west of the deepest part of the graben. The ridge is in part as much as 150 m above the bottom of the valley (the green grass field), so that the feeder-dike of Nesjahraun eruption decided to propagate some further 100–150 m in places above the surrounding valley floor. This may sound strange. We might, on first throught, consider it most likely that the magma, like any fluid, would look for the 'easiest' way to the surface, where by easiest we mean the **path** that requires **minimum energy** used by nature to bring the magma to the surface. And, indeed, that is what nature does. The dike path, like most other physical processes, follows the 'principle of least (strictly, stationary) action'. Here least 'action' means that the path chosen is the one that makes the energy used times the time it takes for the magma to reach the surface a minimum: more specifically **energy** × **time is a minimum**. You can view this principle, depending on your taste, so that nature is either lazy or, alternatively, simply very efficient. She does most thing, she operates most (but not all) of her processes, through minimum of effort.

This is one of the most beautiful principles in physics (also named **Hamilton's principle** after the famous Irish mathematician) and can be shown to control much of mechanics, including quantum mechanics. The main point for us, however, is that the energy required for the feeder-dike is not just related to the length of the dike path—a longer path normally requires the use of larger energy to rupture a longer fracture, the dike fracture. But if, as is likely here, close to the surface the magma found a very weak zone, say a vertical active normal fault, then hardly any energy was needed to rupture the rock, because the fault plane is itself was already a fracture, a rupture. Thus the total energy needed for the magma to propagate the extra 100 m or so to the surface, using and existing fault or weakness zone as a channel to the surface of the ridge instead of the surface of the valley floor, very likely was less than that of forming a new fresh fracture, rupturing, the path to the valley floor. In fact, such magma paths are very common—to the top of ridges or mountains rather than to the surroundings at lower elevations. If they were not common, no large **volcanic edifices** would ever form. A particularly striking example of similar dike-path formation is the Hekla Volcano in South Iceland where most of the fissure eruptions occur—most of the feeder-dikes reach the surface—at the top of the volcano (Chap. 14).

What else do we see of interest in Nesjahraun? One interesting feature is that during the early stages of the eruption, the flow was primarily of a comparatively **smooth pahoehoe** type. Later, however, it changed into a more **rough aa** lava flow. This change is quite common, namely that a lava flow changes from pahoehoe to aa, but the opposite never happens. The change from pahoehoe to aa may be related to change in chemistry, lowering of the temperature, or increase in the volumetric flow rate during the eruption.

The total area of the Nesjahraun lava flow is estimated at about 14 km^2, of which about half is on the bottom of Lake Thingvallavatn. The average thickness is not known, but could be around 10 m, in which case the volume of the lava flow would be about 0.14 km^3, or similar to that which erupted in Hekla in 1991 and in Eyjafjallajökull in 2010. Such an eruptive volume is common in Icelandic volcanoes, but this volume estimate applies only to Nesjahraun; it does not consider the volume that was issued at the same time from the volcanic fissure southwest of Hengill. Also, the volume of the feeder-dike is not considered here. But given the length of the volcanic fissure, the likely volume of the feeder-dike was several cubic kilometres, so that the overall magma flow into the crust during the Nesjahraun eruption may have been similar to that of the 2014–2015 Bardarbunga-Holuhraun eruption in central Iceland.

12.6 Major Faults

We drive on to the crater cones in Fig. 12.21. You may wish to take a closer look at them. They are typical scoria cones. We have already looked at scoria close-ups before (Figs. 9.4 and 9.9), so I do not make this an official stop in this excursion. Rather, I suggest we drive on towards the last stop in this trip, namely the **sixth stop (6)**. Here, again, there are several places where you could stop to view the largest normal fault that we discuss in detail in this book, namely **Jorukleif (Jórukleif)** (Figs. 12.16 and 12.22). It is certainly worthwhile to drive to the cove of **Hestvik (Hestvík)** before you come to the main wall of Jorukleif. Not only is Hestvik beautiful, but you also have a very good overview of Jorukleif from the eastern boundary of the cove (Fig. 12.23). I come to the view from Hestvik in a moment, but first assume that you do not have time to go there, in which case the main ground view of the fault wall of Jorukleif could be to the southwest (Fig. 12.22). The main wall, in places, reaches well over 100 m—for example, 120–130 m at several localities where this photograph is taken (Fig. 12.24). The flow units, similar to those in the walls of Almannagja at Thingvellir (Figs. 5.5 and 5.6), show that the Jorukleif cuts through a very thick pahoehoe lava (Fig. 12.24). In fact, that pahoehoe lava flow is the one that forms the lava shield of Mosfellsheidi, the shield we drew along on our way to Hengill in the beginning of this excursion. We see that the walls, however, look older, more **weathered**, than those of Almannagja—although not as old-looking as those of Thyrill (Figs. 11.19, 11.20 and 11.21). The noticeable weathering of the flow units of Jorukleif is because the age of Mosfellsheidi is several hundred thousand years, whereas the lava flow constituting the walls of Almannagja are about 9000 years old.

While the visible displacement or subsidence along the main wall of Jorukleif is as much as 130 m (Fig. 12.24), the real displacement along this part of the Hengill Graben is over **200 m**. This is because the fault is split in two step-like parts (Figs. 12.23 and 12.24) the road being along the plateau between the two steps. The step-like parallel fault walls can be seen on the ground along the road, but, as indicated above, are perhaps better observed from the Nesjar peninsula, forming the eastern boundary (a fault also) of the cove of Hestvik (Hestvík). In case you drive to Hestvik, either before or after the stop seen in Fig. 12.22, you will, I am sure, appreciate the beautiful view of the more than 200 m high step-like fault wall as well as that of the cove itself (Fig. 12.23). Part of the cove as well as the walls of Jorukleif, are seen from the air in Fig. 12.16. The fault walls of Jorukleif, however, are seen in more detail on Figs. 12.23 and 12.24.

Jorukleif is really a continuation of Almannagja at Thingvellir or, perhaps more correctly, Almannagja is a continuation of Jorukleif. Jorukleif has been active as an earthquake fault for at least many tens of thousands of years, and most likely hundreds of thousands of years. Eventually, as the spreading in the Hengill Graben

Fig. 12.22 Part of Jorukleif (Jórukleif). View south along the fault at Hestvik (Hestvík), this location coincides roughly with the one indicating the subsidence or displacement in Fig. 12.24

Fig. 12.23 Jorukleif (Jórukleif) as seen from Hestvik. View west, the two steps together constitute a vertical displacement or subsidence of about 200 m. Compare Figs. 12.16, 12.22 and 12.24

continues, Jorukleif drifts to the west, out of the main active graben, and becomes inactive (Fig. 4.5). Normal faults of similar displacements, 150–200 m, occur in the Hvalfjördur area, but are not easily seen as clear fault planes—clear fractures—which is the reason I do not show them to you in our Hvalfjördur excursion (Chap. 11). Rather, these large faults in Hvalfjördur are inferred from mapping of the lava pile and observing that the pile on one side of a depression does not fit with the pile on the other side. Similar faults, and some in fact larger—with displacements of some 300–400 m—also occur in Botnsulur (Figs. 4.12 and 4.13), but we do not observed them in a close-up. The Hengill area is clearly **enormously active**, particularly in terms of earthquakes and faulting. Volcanic eruptions can most certainly be expected in the area and there are strong indications, from the doming and earthquakes that occurred in the years 1994–2000, that dike/sheet intrusions occur from time to time in the volcano. As indicated above, most dikes in any volcano in Iceland become arrested or stop on their paths to the surface. As we will discuss in connection with Eyjafjallajökull (Chap. 14), an **unrest period** in a volcano with many dike injections may last tens of years before, eventually, one of the dikes is able to reach the surface and supply magma to an eruption.

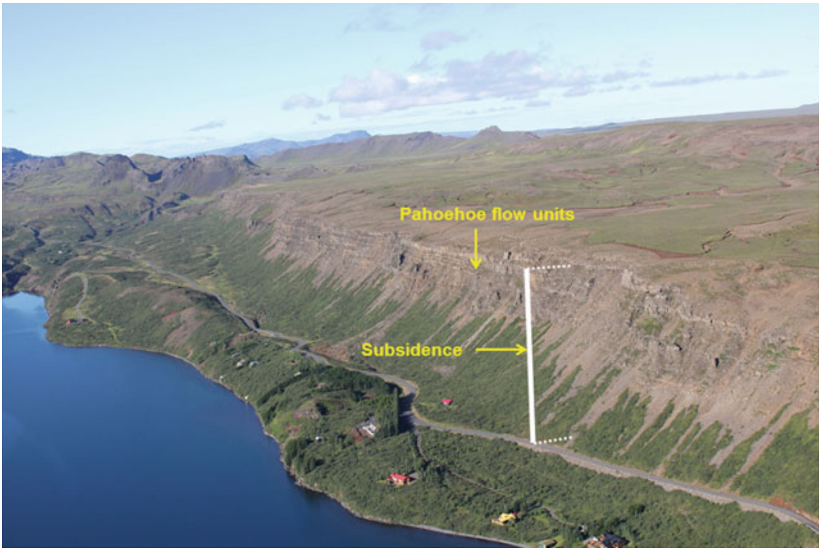

Fig. 12.24 Aerial view of Jorukleif (Jórukleif). View south, the maximum displacement or subsidence along the upper step of the fault (Fig. 12.23) is about 130 m. Also indicated are pahoehoe flow units of the Mosfellsheidi (Mosfellsheiði) lava shield which the Jorukleif fault dissects. Compare Figs. 12.16, 12.22 and 12.23

This is the last stop in our excursion to Hengill and its surroundings. As we drive on to the north along Road 360, we soon meet Road 36 which takes us back to Reykjavik. All the geological structures and processes seen on the way from the cross-road between Road 360 and Road 36 are already described in Chap. 4, in case you would like to observe and explore them further.

13 Reykjavik-Kleifarvatn-Reykjanes

To reach Reykjanes, the southwesternmost part of the Reykjanes Peninsula, you first drive either Road 40 or 41 to the town of Hafnarfjördur. When you come to the southwest part of the town, there are two roads that can take us to Reykjanes. One is Road 41, that is, the one you are on, which is the same as you drove from the Keflavik Airport to Reykjavik when you arrived. The other one is to turn left (to the south) and drive along Road 42 to the geothermal field and explosion craters close to the lake Kleifarvatn, and then follow roads 427 and 425 to Reykjanes. In this excursion, we take the second option, Road 42 (Fig. 13.1).

There are numerous things to observe on the way, and depending on your taste you may wish to stop at many more places than I suggest here. In particular, some people find Lake **Kleifarvatn** beautiful, or at least impressive. We have, however, already seen Lake Thingvallavatn from various points of view—from the air and on the ground (Chaps. 5, 6 and 12), so I shall not make a stop particularly to view the entire Kleifarvatn, but mention sites where a good view of it is possible. Also, the hyaloclastite ridge **Sveifluhals (Sveifluháls)** is a very impressive a hyaloclastite ridge. We shall look at some geological features associated with the ridge, but we have already seen many similar (though mostly smaller) hyaloclastite mountains (Chaps. 2, 4–6, 11 and 12), so I shall not make a particular stop to view the entire ridge (which is difficult, because of its size, except from an aircraft). I shall, however, show an aerial photograph where Kleifarvatn and Sveifluhals are seen in relation to the local earthquake fractures and volcanic fissures. The focus in this excursion is on some geological features and processes that are mostly different, in nature or in context, from those you have seen in the other excursions. The volcanic systems on the Reykjanes Peninsula in relation to the oblique plate boundary are shown on Fig. 2.3.

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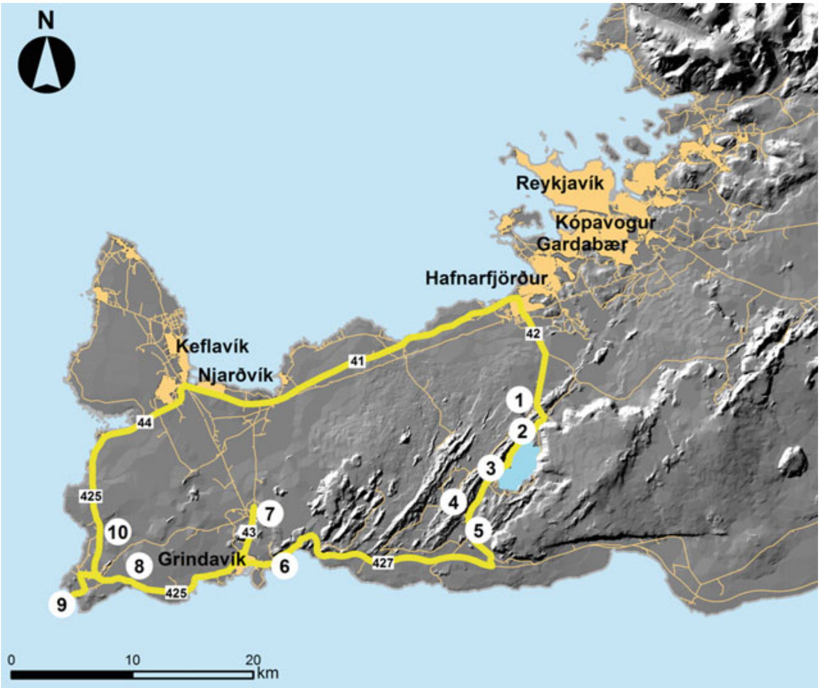


Fig. 13.1 Map showing the excursion to the fjord Hvalfjördur. The suggested main stops are indicated by encircled numbers, 1–10

13.1 Lava Channels

At the cross-road between 41 and 42, we take Road 42 and drive to the south through young lava flows. For several kilometres, the road is along **Obrynnisholahraun (Óbrynnishólahraun)** and then passes on to another lava flow, **Kapelluhraun** (also known as Háibruni). These are both aa lava flows. Obrynnisholahraun is just over 2000 years old, whereas Kapelluhraun is around 750 years old, that is, it formed at around the year 1170 of the current era. Our **first stop (1)** is to take a look at a structure that is characteristic for aa lava flows, namely a **lava channel** (Fig. 13.2). The channel floor is gently sloping downstream. The channel has walls or banks, referred to as levees in geology. The depth of the lava river changes during the eruption and the surface may rise, from time to time, to the edges of the walls, so as to generate overflows on the tops of the walls. The walls thereby build up and become gradually higher. When flow of magma out of the crater that feeds the channel stops, much of the lava that is left in the channels flows downslope and leaves an essentially empty channel—as you see here.

The channel in front of you was formed some 750 years ago. However, to better understand the process, it is worthwhile to take a look at similar channels while active (Figs. 13.3 and 13.4). For this we take a look at one of the channels of the **1991 eruption of Hekla** in South Iceland. (We shall see Hekla from a distance and discuss this eruption again during our excursion to Eyjafjallajökull in Chap. 14.) The photographs show the channels close to the lava source—an underground path, a tube from the main crater (seen in Fig. 14.9)—but two different channels and at different times. The one in Fig. 13.3 is from earlier in the eruption and closer to the main crater, so that the lava flow is hotter, as indicated by the orange colour. The one in Fig. 13.4 is taken later, at a greater distance from the main crater, so the

Fig. 13.2 Lava channel in Kapelluhraun formed around 750 years ago. As seen here the channel is mostly 40–50 m wide (persons provide a scale) but expands to a width of 200–300 m downstream

Fig. 13.3 Part of a lava channel active during the early part of the Hekla 1991 eruption. Close to the source, the channel is 5 m wide, about 2 m deep, and inclined (dipping) downstream about 8°. The maximum flow velocity of the lava in the channel is about 40 cm per second—similar to the pace of a walking person or the rate of propagation of a dike in the crust (Chap. 11)

lava flow is cooler and dark red rather than orange at the surface and already with considerable solid but very thin lava crust at the surface. Many similar channels formed, for example, in the earlier one of the two eruptions in Eyjafjallajökull in 2010 (Chap. 14).

The channel in Fig. 13.2 originates in a depression in the walls of the source crater, as is very common. The lava channel is mostly parallel to Road 42 and you drove through it a bit earlier. But there it was so wide and, in comparison, shallow that you probably did not notice it. The channel west of the road, namely the part which Fig. 13.2 shows, is 40–50 m wide. However, immediately east of the road, the channel widens to about 90 m, and then runs parallel with the road, mostly 200–300 m wide, for at least 3 km.

Aerial photographs of a nearby lava channel from a volcano named Eldborg undir Geitahlid (**Eldborg undir Geitahlíð**) put the lava channel here and in Hekla in a wider perspective (Fig. 13.5). The Eldborg channel is easily reachable from Reykjavik first in car to Heidmörk (Heiðmörk at Vifilstadahlid, Vífilstaðahlíð) where you drive along the wider part of the channel (Fig. 13.5). If you want to go all the way to its source, the crater cone Eldborg, then that is a nice hike (walk). At its source, at the wall of the crater Eldborg, the channel is only about 20 m wide, and 30–50 m further away from the crater. But then it becomes much wider, and where it is parallel to the road it is as wide as 200–250 m, or similar to the channel width at the first stop. Lava channels can be much larger, both on Earth as well as on the other terrestrial planets. Some lava channels on Marsh and Venus have widths of tens of kilometres and reach lengths of thousands of kilometres.

Fig. 13.4 Part of another lava channel active during a later part of the Hekla 1991 eruption. This channel is of similar but somewhat smaller dimensions than the one in Fig. 13.3. Here the lava has been transported a longer distance (in an underground tube) from the main crater and is thus cooler than the lava in Fig. 13.3. Thus, the lava is here dark read and its surface quickly forms a thin, solid crust. The person provides a scale

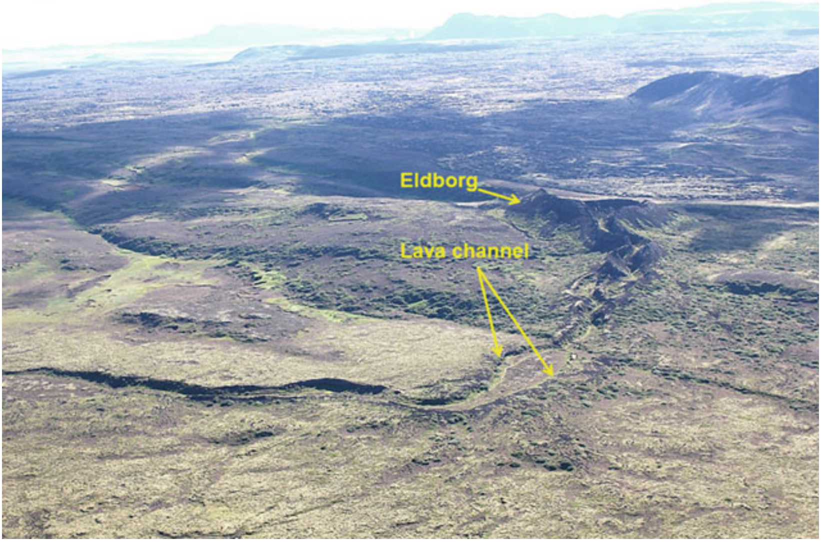


Fig. 13.5 The lava channel from the crater Eldborg undir Geitahlid. At the crater, the source, the channel is about 20 m wide but becomes wider away from the source. The widest part on this photograph is where the arrows mark the channel; there it is about 50 m. However, further downslope, not seen here, the channel reaches a width similar to that of the channel in Fig. 13.2, namely 200–250 m

Continuing our excursion, we drive on passing by a quarry (to the left of the road) where part of the internal structure of the hyaloclastite ridge **Undirhlidar (Undirhlíðar)** can be seen. If you are allowed to go into the quarry, it may be worthwhile to take a look at the walls (showing various rock types), but this may not be possible while the staff is working in the quarry, so I do not include this as a stop.

13.2 Sveifluhals (Sveifluháls) and Kleifarvatn

Driving along Road 42 we pass through a depression between the two hyaloclastite ridges, namely Undirhlidar and **Sveifluhals (Sveifluháls)**, from which we can see **Lake Kleifarvatn**. There is no particular spot where the entire lake is easily seen on the ground—except of course if you climb up to the top of Sveifluhals. Our **second stop (2)** is thus somewhat arbitrary, but is selected so as to discuss the two main types of processes involved in deformation of the crust as a result of applied forces and stresses. One process is **ductile deformation**, the other is **brittle deformation**. But before I discuss these processes, and their relation to Kleifarvatn, a brief overview of the lake itself is in order.

Kleifarvatn is primarily the result of groundwater flowing into the basin that constitutes the lake. There are **no rivers** or streams that flow into the lake, or out of it. The basin thus cuts through, and reaches far below, the **water table**—the surface of the groundwater—as we saw also in the fractures at Thingvellir and Lake Thingvallavatn itself. It is not surprising that the lake goes below the water table because it is quite deep—reaching a maximum depth of about 97 m (its surface elevation fluctuates, however, and so does the maximum depth). Its maximum length and width of the lake vary depending on the fluctuations in the water surface. The maximum length can reach about 5200 m and the width about 2300 m. Similarly, the area of the lake varies, but is commonly about 10 km. Following the **earthquake** in June 2000, discussed below, there was considerable leakage from the lake, presumably through earthquake fractures, which resulted in a **lower lake surface**, by several metres, and **reduced the lake area**. These effects lasted for many years, but the lake surface elevation and area are now back to normal.

Fig. 13.6 Example of ductile deformation in the hyaloclastite ridge Sveifluhals (Sveifluháls). This fold, formed through gravity gliding of soft layers down the slopes of the ridge, is a very small example of an anticline

The ductile structure is a stack of curved layers, referred to as a **fold** (Fig. 13.6). By ductile I mean that the layers were able to flow. Now how can rock layers flow? In this case the answer is very simple: they flow in the same way as sand or mud flows. These layers are formed in subglacial (or, perhaps, submarine) eruptions, in the way we discussed in Chap. 6 for hyaloclastite mountains in general. Here the flow was simply driven by gravity, that is, it was down the slope of the mountain while it was forming. In geological terms, this type of fold is named **anticline** or antiform. Anticlines are dome-shaped, they form ridges, are convex, whereas the other main form, **synclines** or synforms, are valley-shaped, they form depressions, are concave.

This anticline is very small—the total thickness of the layers seen here is about one metre. Much larger anticlines and synclines, and also much smaller, occur in the main mountain belts of the world, such as the Alps and the Himalayas. There, however, the fold formation is not driven directly by gravity, but rather by **plate-tectonic forces**. The plate-tectonic forces responsible for these mountain belts are compressional forces—the lithospheric plates collide head on. For example the plate carrying India is colliding with the Eurasian plates, and the resulting 'crumpling' and folding and uplift is the Himalayas mountain belt.

The ductile deformation or flow you see in the ash in front of you occurs, perhaps surprisingly, **also on large faults**, referred to as fault zones, that extend to great crustal depths; faults of the type we have seen dissecting the rocks in Botnsulur (Fig. 4.12), Thingvellir (Figs. 5.1, 5.2, 5.3, 5.14 and 6.2), and Hengill (Figs. 12.16 and 12.24). The upper parts of faults are **brittle**, meaning that when the stress reaches a certain level the rock breaks, rupture occurs and (commonly, but not always) an earthquake. At a great depth, the exact value of which depends on temperature, rock type, and rate of movement or displacement of the rocks on either side of the fault—so-called **strain rate**—the fault rocks no longer break or rupture when the stress becomes high but flow. At that depth no earthquakes are generated; the rocks in the fault zone just flow. In Iceland flow in large fault zones begins at different crustal depths, but commonly at depths of 10–15 km. The depth is less in active central volcanoes, simply because at several kilometres depth there are often fluid or semi-fluid magma chambers, as we have discussed earlier, where the deformation must be ductile, that is, by flow. Some parts of central Iceland show brittle behaviour to much greater depths—as deep as 25 km in places—but generally the fault zones (and the crust in general) in Iceland begin to flow rather than break or rupture at high stresses at depths of 10–15 km.

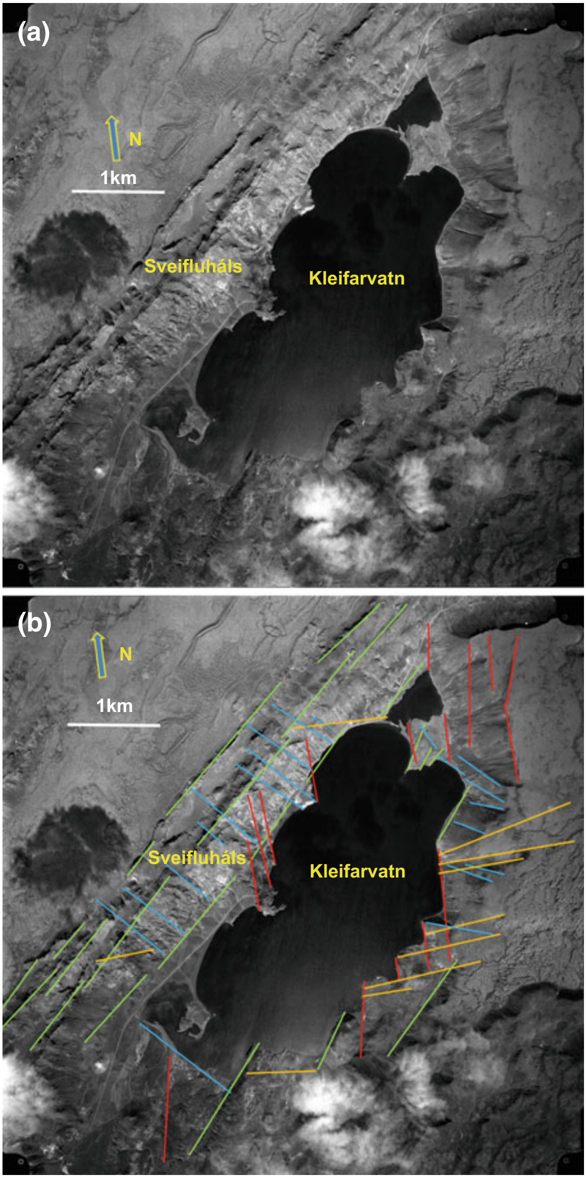
How do we know this depth? Simply by locating the depths of earthquakes in Iceland. Most earthquakes in Iceland are **shallower than 15 km**. This means that brittle behaviour—or earthquake fracturing—does not normally occur at greater depths than this. Below this depth the crustal rocks behave as ductile, like the rocks in Fig. 13.6, that is, they flow. Do we have any analogy to this behaviour closer to the surface? Yes, indeed we do: **glaciers**. Perhaps you have already been at or on some of the glaciers in Iceland. If so, you have seen that they have fractures—and dangerous ones—at the surface. But these fractures, even if widely open at the surface, extend to depths of only several tens of metres. Commonly the glacier is flowing at depths greater than about 50 m. Similarly, the crust in Iceland is flowing, and all the faults are ductile, below depths of 10–15 km. But above this depth earthquakes and fractures are common in a layer that is referred to in geology as the **seismogenic layer**—the layer in which earthquakes are generated.

Fractures to a large degree control the **landscape**, particularly in an area like Iceland where there is active faulting. This control is obvious when we look at valleys such as the Thingvellir Graben and the mountains Armannsfell and Botnsulur (Chaps. 4 and 5). We have also discussed this control in connection with the trends of the valleys in Hvalfjördur, as well as that of the fjord itself, and related mountains (such as Esja, Chap. 4). But the earthquake fractures have large control on other landscape features, including lakes. And one of the most striking examples of this is the shape or geometry of Lake Kleifarvatn.

13.3 Kleifarvatn—Controlled by Fractures

Kleifarvatn has a strange shape (Fig. 13.7). While its main direction, or long axis, is northeast-southwest, there are unusually many small irregularities along its shore or boarder (Fig. 13.7a). How unusual this shape is can be seen on comparing the geometries of Thingvallavatn (Figs. 5.14, 6.2 and 12.10) and Kleifarvatn (Fig. 13.7a). Clearly, Thingvallavatn has much more regular shape which is, indeed, almost entirely controlled by the northeast-trending faults of the grabens of Thingvellir and Hengill. So the question arises: why is the geometry, the shape, of Kleifarvatn so complex?

The answer is: the geometry or shape of Kleifarvatn is largely controlled by **several sets of fractures**. I illustrate the main fracture directions or trends (called strike in geology) around the lake in Fig. 13.7b. We see four main directions of fractures around the lake, namely northeast, north-south, east-northeast, and northwest. The shore, and therefore the shape, of the lake is to a very large extent controlled by these fractures. So why, in contrast to Thingvellir, are there so many


(a)
N
1km
Sveifluháls
Kleifarvatn
(b)
N
1km
Sveifluháls
Kleifarvatn

◀ **Fig. 13.7** Black-and-white aerial photograph of the hyaloclastite ridge Sveifluhals (Sveifluháls) and the lake Kleifarvatn. **a** Sveifuhals is a northeast-trending hyaloclastite ridge, formed in many eruptions and over perhaps some 200 thousand years. It follows the trend of most volcanic fissures on the Reykjanes Peninsula. Kleifarvatn is a lake that is unusual in that (1) there are no rivers or streams that either flow into or out of the lake—it is thus entirely the result of groundwater filling the basin where it is located—and (2) its shape is highly irregular and unlike that of most lakes in Iceland. **b** Schematic presentation of the main tectonic lineaments or fractures—mainly faults—that control the shape or geometry of Kleifarvatn. Green are northeast trending and mostly normal faults and some volcanic fissures. Red are north trending and mostly strike-slip faults. Orange are east-northeast trending and also mostly strike-slip faults. Blue are northwest trending and mostly partly normal faults and partly joint systems. The same fracture sets occur in South Iceland and part of West Iceland

different fracture directions in this area? How do these fractures originate—why are they here?

The **northeast-trending fractures** are the easiest to explain. They are directly related to plate-tectonic spreading or rifting, just like in the Vogar Fissure Swarm (Figs. 2.5 and 2.6) and in the Hengill Volcanic System, including Thingvellir (Chaps. 5, 6 and 12). The general direction of the northeast-trending fractures is very similar to that in Hengill and Thingvellir, namely about 33° east of north. The fractures in this area, around Kleifarvatn, are mostly normal faults, of exactly the same type as at Vogar, Thingvellir, and Hengill. Also, we see that these have the same direction as the hyaloclastite ridge Sveifluhals itself, which is clearly one reason for the existence of the lake. A few of these fractures are thus dikes, some feeder-dikes, related to volcanism in the area. The northeast-trending fractures thus relate directly to the spreading of the tectonic plates.

But what about the **north-south-trending fractures**? How do they come about, and why are they here? The north-south directed fractures are earthquake fractures of the same type and direction as are very common in the South Iceland Seismic Zone (Chaps. 9 and 14). In the South Iceland Seismic Zone north-south trending fractures give rise to large earthquakes (magnitude 7 or larger) about once in a century. The surprising thing is that the earthquakes in South Iceland and around Kleifarvatn act in **harmony**. This was demonstrated during the largest earthquakes in Iceland in the past decades, namely the two magnitude 6.6 earthquakes in South Iceland 17 and 21 June 2000. Within half a minute after the 17 June earthquake, there were 4 smaller earthquakes to the west of and triggered by that earthquake. One of these earthquakes occurred at Kerid (Chap. 9). Another one at Hengill (Chap. 12). And then there were two earthquakes on the Reykjanes Peninsula, one in its southeastern part (Selvogur), and the other one beneath the eastern shore of

Kleifarvatn. It is remarkable that the earthquake at Kleifarvatn occurred **77 km** to the west of the 17 June earthquake which was the trigger, and yet the Kleifarvatn earthquake reached magnitude 5.5 and ruptured the surface.

And earthquakes continue to occur at Kleifarvatn. Many earthquakes of magnitude 4–5 have occurred there following the magnitude 5.5 earthquake in the year 2000. For example, earthquakes of these magnitudes, that is, 4–5, occurred at Kleifarvatn in the years 2005, 2011, 2015 and 2016. These earthquakes occur mostly on north-south trending fractures which, as we see (Fig. 13.7b), have great effects on the shape of the lake. It is thus clear that the same types of earthquake fractures that occur in South Iceland, occur also in the central part of the Reykjanes Peninsula, and in particular at Kleifarvatn.

The **east-northeast-trending fractures** then follow because they relate to the same stress field that controls the north-south fractures. We have already discussed their formation in Chap. 9 and discuss it further in Chap. 14. The east-northeast fractures are earthquake fractures, formed together with the north-south fractures and in the same stress field (Figs. 8.2 and 9.2). Like the north-south fractures, the east-northeast fractures are strike-slip faults.

So we have now explained the three main types of fractures controlling the shape of Kleifarvatn. But what about the **northwest-trending fractures**; where and how do they fit in? The northwest-trending fractures are in fact very common in the southern part of Iceland. We have seen them in the hyaloclastite mountains in the vicinity of Thingvellir (Chaps. 5 and 6), in Hvalfjördur (Chap. 11) and in Hengill (Chap. 14). They occur also inside the South Iceland Seismic Zone (Chaps. 9 and 14), while not as common there as the north-south and the east-northeast trending fractures. How the northwest-trending fractures form has been something of a mystery. Clearly, they are not generated by ordinary rifting within the volcanic zones. And while they occur within the South Iceland Seismic Zone, they are not easily explained by the dominating stress fields that generate the north-south and the east-northeast trending fractures. So how do the northwest-trending fractures form and why are they so widespread?

One thing we notice is that the northwest fractures are right angles, are **orthogonal**, to the northeast-trending fractures (Fig. 13.7b). It follows that the northwest fractures most likely form exactly like many other orthogonal fractures, namely by fluid pressure that shifts or flips the stress that generates fractures by 90°. Here is how this happens. When, say a dike, is injected into the crust, it is able to break the rock because its magma has high pressure. This pressure pushes the dike walls apart. This push, in turn, bends the walls and the nearby crustal segment slightly and, like when you bend a pile of cards or a stick, generates **tension** at right angles to the dike (Fig. 13.8). So if the dike is directed northeast, then the

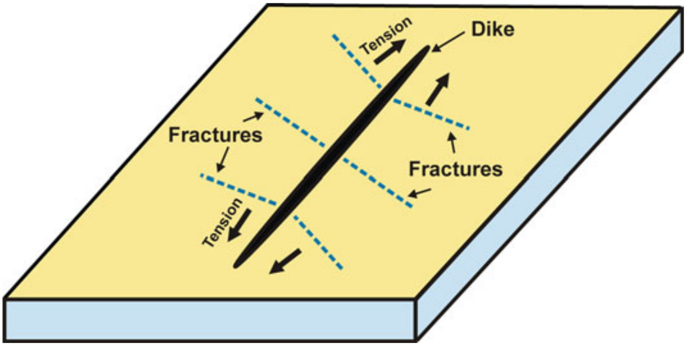


Fig. 13.8 Dike-fracture forms when the magmatic pressure breaks the rock and pushes the fracture walls apart. The magmatic pressure inside the dike generates tension parallel with the dike walls. This tension may result in tension fractures, particularly joints, and normal faults (and occasionally strike-slip faults), in a direction roughly orthogonal (perpendicular) to the dike walls. The trend of the fractures changes somewhat along the length of the dike. For a northeast trending dike, the induced fractures would be northwest trending, as commonly observed in the landscape of South Iceland and part of West Iceland

tension fractures, largely sets of joints, and are directed northwest, that is, at roughly 90° to the dike. This is the most likely explanation for the northwest-trending fractures that control the shape of Lake Kleifarvatn.

But we can generalise this idea further, so as to explain major aspects of the beautiful **landscape** in South and West Iceland that we have been discussing and looking at in the earlier chapters. We mentioned how the northwest-trending fractures dissect, and determine the slopes of, mountains such a Hrafnabjörg, Armannsfell, Botnsulur, Reynivallahals (Hvalfjördur), and Esja, as well as controlling the directions of the fjord and valleys in much of West Iceland. In addition, northwest-trending fractures are common in the older surface rocks everywhere in South Iceland. Now we can explain all these fractures and why they give rise to the landscape forms.

The northwest-trending fractures that control the landscapes in the older rocks of West and South Iceland form not inside, but rather **outside**, the active volcanic zones (Fig. 13.9). That is, these fractures form long after the crustal rocks, composed of the lava flows and hyaloclasites, have drifted out of the active volcanic zones so as to become essentially inactive in terms of earthquakes and faulting. Most of the northwest trending fractures are not faults, and not dikes. So what are they? Most are **joints**—that is, very narrow fractures, similar to the columnar joints you have seen in many places, but formed in a different way and much larger, particularly longer, structures.

So the northwest-trending fractures that control the landscape in these parts of Iceland are largely the result of tension resulting from lateral bending of the crust. On a regional scale such as here the tension is not generated by the magmatic pressure of a single dike, such as was the case for Kleifarvatn and inside the volcanic zones in general, but by the cumulative pressure of many dikes within a large part of, or an entire, volcanic zone (Fig. 13.9). We know from measurements, such as when dikes were emplaced in the volcanic zone in North Iceland in the so-called Krafla Fires in the years 1975–1984, that the stresses or forces in the crust change out to a distance of more than 100 km from the dike. Similarly, in the more recent and much better documented (measured) dike emplacement in the Bardarbunga Volcanic System in Central Iceland, 2014–2015, the stress and force effects of the 45-km-long dike were recorded out to a distance of more than 100 km from the dike. Thus over time, when many dikes are emplaced within a volcanic zone, the cumulative effects is as if the entire volcanic zone was a big intrusion under magmatic pressure. The stress effects reach to distances of many tens to one or two hundred kilometres from the volcanic zone. Generally, the main stress effects reach out to a distance similar to the length of the volcanic zone. Thus, great effects are

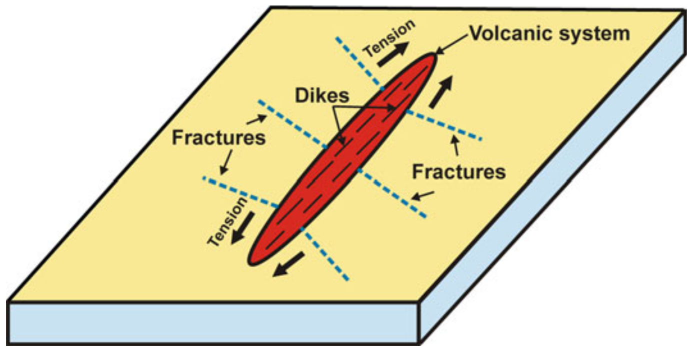


Fig. 13.9 Volcanic system acts essentially an elliptical hole (strictly an elastic inclusion) filled with material under pressure. The pressure is primarily derived from dike injections. Similarly to an individual dike under pressure (Fig. 13.8), an entire volcanic system, some of which reach lengths of 190 km, may act as a pressured elliptical hole and generate tension. The induced tensile stresses, in turn, commonly result in fracture formation, both tension fractures and faults, with great effects on landscape evolution, including the shapes of mountains, valleys, fjords, and lakes such as Kleifarvatn (Fig. 13.7b)

felt to at least 100 km west of the West Volcanic zone, and to as much as 200 km on either side of the North and East Volcanic Zones.

The effects are not confined to West and South Iceland: many of the fjords in East Iceland relate to similarly generated stress field, partly by dikes in the North Volcanic Zone, and partly by dikes in the East Volcanic Zone. In any case, we have now explained all the main fracture directions on that control the geometry of Kleifarvatn, so we turn our discussion to volcanic activity in relation to fracture trends in the area.

13.4 Fracture Sets and Volcanic Eruptions

I have illustrated in Fig. 13.8b how the lake geometry is controlled by the main fracture directions. But what is the connection between these fracture sets and volcanism in this part of the Reykjanes Pensinsula? Starting with the **northeast-trending fractures**, these clearly have the same direction as most of the feeder-dikes and, therefore, most of the volcanic fissures in the area. Not only has the hyaloclastite ridge itself, Sveifluhals, this direction but also most of the volcanic fissures that have issued the young (Holocene) lava flows in the surroundings. This is not surprising because we have already established that the great majority of volcanic fissures and feeder-dikes, and dikes in general, in South and West Iceland (and, indeed Southeast Iceland as well, even if that part is outside the scope of our excursions) are directed northeast-southwest. We also know why that is. It is simply because all these fractures, and the associated normal faults, are directed at roughly right angles or perpendicular to the plate-tectonic spreading direction, the so-called **spreading vector**, as they should be (Fig. 2.2).

What about the **northwest trend**? Are there any volcanic fissures with the same direction? In general in Southwest Iceland, yes. There are northwest-trending intrusions in Hengill, for example (Chap. 12). And there are many dikes, some of which may have been feeders to eruptions, with a northwest trend in the Hvalfjördur area (Chap. 11). As regards volcanic fissures younger than the last ice age, that is, younger than about 12 thousand years, there do not appear to be any with a northwest trend on the Reykjanes Peninsula. Thus, in geologically recent time, there are, apparently, no northwest-trending feeder-dikes on the peninsula.

Are there any volcanic fissure/feeder-dikes with a **north-south trend**? Yes, indeed there are. There are several young (less than 12 thousand years old) volcanic fissures whose direction is roughly north-south. Perhaps the most striking one is the fissure that issued the lava flow **Ögmundarhraun**, which you pass through later today on your way to Reykjanes. That fissure, located at the south end of Sveifluhals, is remarkable for the fact that its direction changes abruptly from being northeast-southwest in its southern part to north-south in its northern part. Similar changes are seen in some other volcanic fissures on the peninsula. In addition, and perhaps most relevant to our excursion, is the fact that the explosion craters (maars) located at the south end of Kleifarvatn and where we stop later today, are apparently also related to roughly north-south trending volcanic fissure.

Are the north-south fissures necessarily earthquake fractures that are then used as channels for the feeder-dike, for the magma, to the surface? Not necessarily. Local stresses are often very different from the regional (plate-tectonic) ones, particularly if there is a shallow magma chamber nearby (Fig. 11.3; Chaps. 4 and 11) or if the rocks have very widely different mechanical properties (some being soft, others stiff). Most dikes form their own fractures, as we discussed in Chap. 11. However, it is also common to see dikes use existing weaknesses in the crust, such as earthquake fractures, for parts of their paths, particularly close to the surface. And that is, most likely, what we are seeing on the Reykjanes Peninsula. Some of the north-south earthquake fractures where used as channels for magma, that is, as parts of dikes.

Since the north-south earthquake fractures on the Reykjanes Peninsula, are of an origin similar to the north-south fractures in the South Iceland Seismic Zone (Chap. 14), you might ask: why are there no eruptions, no feeder-dikes, using such fractures or the associated east-southeast trending fractures in that zone? The answer is: because there is no available magma beneath the greater part of the South Iceland Seismic Zone. A dike can form, and eruption can occur, only if magma is available, and there is no available magma beneath the South Iceland Seismic Zone—that is, until we come to its east and west extensions into the volcanic zones. In the west, the extension is into the Hengill Area, which we have already discussed (Chap. 12). In the east, the zone extends into the East Volcanic Zone. And there is a beautiful example, or a dramatic one—depending on how you view these processes—of an **earthquake fracture being used by magma**, that is, by feeder-dikes. That example is the most famous volcano in Iceland—**Hekla**. We come again to this point in Chap. 14. But here it is sufficient to say that the Hekla Volcanic Fissure is an earthquake fracture, related to the stresses in the South Iceland Seismic Zone. It is not, however, a north-south earthquake fracture, but rather an east-southeast trending earthquake fracture, of the type we saw at Kleifarvatn (Fig. 13.7b) and Gullfoss and Vördufell in Chap. 9, and will see and discuss again in Chap. 14.

13.5 The Structure of Sveifluhals (Sveifluháls)

We have now had a long discussion about the shape of Keifarvatn and what it tells us about the types of fractures that occur in this area and, indeed, in the entire South (and part of West) Iceland. Let us now drive on along the road to the next stop. There is no particularly good stop for an overview of the lake, nor is there a good stoop where we see the entire hyaloclastite ridge, Sveifluhals. The **third stop (3)** gives as good an overview of the lake as you are likely to get on the ground, save for climbing the nearby mountains, and allows us also to discuss two other topics, namely (1) the formation of the hyaloclastite ridge itself and (2) potential, or **'baby' fractures**.

On the way to the third stop you see the internal structure of Sveifluhals (Figs. 13.7 and 13.10), and how variable it is and complex. This complexity is in contrast to many other smaller hyaloclastite ridges and is partly because Sveifluhals is not formed in one or two eruptions but in **many eruptions**. This is perhaps not surprising given that the ridge is rather large. If we include the adjacent smaller

Fig. 13.10 View north from the south shore of Lake Graenavatn (Grænavatn) showing part of the hyaloclastite ridge Sveifluhals (Sveifluháls). Also seen is part (light-coloured rocks) of the geothermal fields of Seltun (Seltún)

ridges and hyaloclastite mountains (Sveifuhals itself is only the largest of these, but we need not discuss all the different names) then the total length is about 25 km and the maximum width about 1.5 km. This is considerably greater length than the mountains that constitute Esja (Chap. 4) and Hengill (Chap. 12) and many times larger than the small ridges we discussed earlier, such as Sköflungur at Hengill (Chap. 12). Sveifluhals and adjacent mountains may be composed of some 20 different rock units and thus, perhaps, formed in equally many eruptions.

Sveifluhals is formed over a considerable period of time, probably some **200 thousand years**. It is not entirely formed in subglacial eruptions because eruptions occurred in the area occupied by the ridge during the last interglacial (ice-free) period. Thus there are some subaerial (under air rather than under water) eruptive materials in the ridge, but these do not form conspicuous and continuous lava flows, such as those formed during the past 12 thousand years. Sveifuhals shows that, even if we assume most hyaloclastite mountains to be formed in a single or, at most, a few eruptions, that assumption no longer holds in detail, and particularly not when the hyaloclastite ridge is large.

13.6 Deformation Bands and Fluid Flow

One very interesting type of structures to observe here are **potential earthquake fractures**, or what may be called 'baby faults' (Figs. 13.11 and 13.12). These potential fractures are named **deformation bands** in geology. As the name implies, they are bands or zones where the rock has become deformed, but not yet developed real earthquake fractures, that is, faults. Some of the deformation bands evolve later into faults, but many do not. So let me then tell you briefly why deformation bands are important before I explain, even more briefly, how they form.

Deformation bands are important primarily for two reasons. First, they give an indication as to how faults in certain types of rocks form. Second, they largely control fluid circulation in the rocks where they occur. Consider first fault formation. Fault slips generate almost all earthquakes, yet we still do not understand well how faults form and develop. Deformation bands are confined to special types of rocks, namely very granular rocks, that is, rocks composed of visible grains or particles. Most granular rocks are **sedimentary rocks**, that is, formed when particles of silt, sand, and larger boulders are transported by wind, rivers, and the sea. When the particles settle on the solid surface (e.g., lake or sea bottom) they form soft layers that, over long periods of time, may transform into hard solid rocks. Well-known examples are siltstones and sandstones, as well as limestones, shales, and other common types of rocks found in many countries.

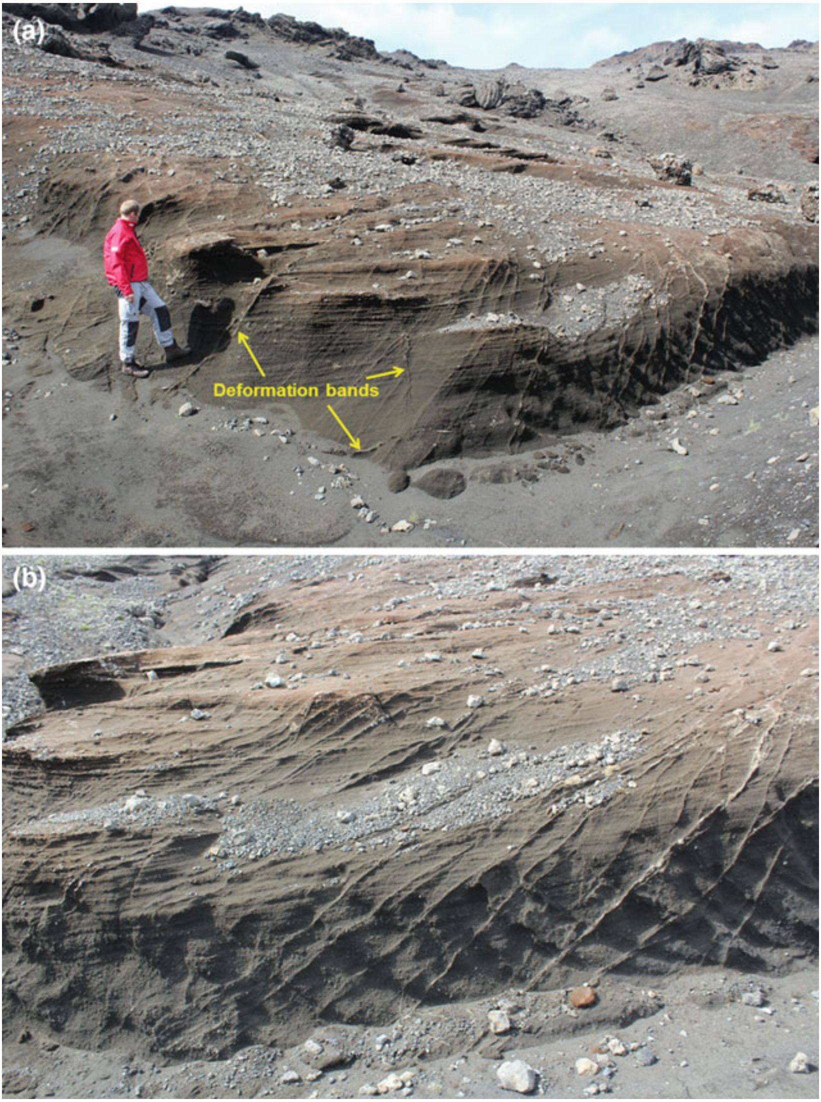


Fig. 13.11 Deformation bands in the hyaloclastite of Sveifluhals (Sveifluháls) as seen **a** from a distance and **b** close-up. Deformation bands are potential faults and particularly common in sedimentary rocks. Deformation bands have great effects on fluid transport in the rocks within which they occur. The person provides a scale

Fig. 13.12 Close-up of deformation bands in hyaloclastite in Sveifluhals (Sveifluháls). The bands are potential faults and here form systems composed of two sets of bands intersecting at an angle of 70°–80°. This is a common angle between faults that are formed in a given stress field. For example, some of the red and orange faults at Kleifarvatn—which formed in the same stress field—meet at this angle (Fig. 13.7)

The hyaloclastites containing the deformation bands in Figs. 13.11 and 13.12 are only to a degree sedimentary rocks. They are, as we discussed in Chap. 6, primarily igneous rocks made of particles of various sizes, that is, **pyroclastic rocks**, formed in eruptions beneath ice sheets or in the sea. Since they form in water, such as a lake within an ice sheet (Fig. 6.6), it follows that they also settle as sediments on the bottom of the lake. In addition, when the ice sheet has melted, and also during the formation of the hyaloclastite mountain when the layers are still soft and weak, movement under gravity, **gravity gliding**, on the slopes of the mountains—sort of landslides—are common. There are therefore several processes affecting the hyaloclastite layers and their development that may be regarded as analogous to those playing a role in the formation of sedimentary rocks. So, in a sense, hyaloclastites are similar to sedimentary rocks.

Furthermore, hyaloclastites are clearly **granular rocks**—you can easily see the particles—and contain many small cavities and holes, that is, are **porous**. Hyaloclastites thus behave in response to forces and stresses in a very similar way to that of many sedimentary rocks. It follows that the deformation bands we see here are similar to those formed in many sedimentary rocks worldwide, and thus give an indication of the general properties of the bands and their relation to faults.

The second point mentioned above is that deformation bands have great effects on **fluid circulation** or transport in the rocks within which they occur. How? Primarily through acting as **barriers to fluid transport**. Deformation bands are partly formed through **crushing and compaction** of the rock, resulting in smaller grains and less space between them which, in turn, means that fluid flow through the bands becomes difficult. In other words, deformation bands have **low permeability**, where permeability is simply a measure of the ease of flow of fluids in rocks. It follows that deformation bands can hinder fluid flow through groundwater aquifers, geothermal reservoirs, oil and gas reservoirs, and, most importantly for the geology of Iceland, magma chambers and reservoirs. The last one may be surprising because most people think of magma chambers and reservoirs as totally fluid, as completely molten. But they are not: they are normally mostly made of solids, a crystal mush or network, with magma in the pores or cavities in-between the crystals. The details of how deformation bands affect magma chambers and reservoirs, as well as the other fluid-containing rock bodies, is beyond the scope of the book. However, the main effects are largely the result of how the bands form—to which we come now.

Deformation bands form in several ways. Here, however, we focus only on the most common mechanism, as seen in Sveifluhals (Figs. 13.11 and 13.12). Most of the deformation bands we see here are formed by stress from the surrounding rocks (primarily through gravity loading during and soon after the formation of the hyaloclastite ridge) that results in **crushing** of the weak glassy grains (in hyaloclastite the grains are mostly glass rather than crystals). The stress is partly shear stress as in faulting, so that some relative movement of the walls of the deformation bands (in the opposite directions on either side of the band) as in faults can be seen. Most of the movement is similar to that of normal faults (Chaps. 5 and 11), but generally much smaller than in most faults, that is, on a millimetre or centimetre scale, although occasionally larger. The crushing of the grains results in their becoming much smaller, less angular, and more **pressed together**. This means that the space between the grains becomes less in the deformation bands than in the surrounding rock, with the result that the bands are much **less permeable** (allow much less through-flow of fluids) than in the adjacent rocks. This reduction in permeability (and porosity) is the primary reason why deformation bands act as barriers to fluid flow in rocks where they occur.

Many, perhaps most, of the deformation bands form during and soon after the eruption of the hyaloclastite (breccia) layers. The bands thus tell the story of the deformation of the rock layers while they were emplaced, and soon after. The forces, mainly gravity, and stresses are, for the present bands, of local origin. Other deformation bands, however, formed later and and due to plate-tectonic forces. This applies in particular to Sveifluhals because it has formed over a comparatively long period of time and is composed of many rock units of different age. It follows that some of the deformation bands are related to regional tectonic forces that control earthquakes and eruptions on the Reykjanes Peninsula, and reflected in the faults and fissures in Fig. 13.7. These bands, which can often be identified based on their orientation, thus give indications of how, perhaps later, large earthquake faults will develop.

13.7 Geothermal Fields

We now drive on to the next stop, the **fourth stop (4)**, namely at the geothermal fields of **Krisuvik (Krísuvík)**. This is the common name given to these fields, but they are also known as **Seltun (Seltún)** or Krisuvik-Seltun. The geothermal fields are well marked, with easy paths, and are surely well worth a visit, even if you have already seen the geothermal field at Geysir. This is partly because the geothermal fields at Krisuvik-Seltun are very different from those at Geysir. In particular, Krisuvik-Seltun has a rich variety of different types of geothermal vents, including boiling hot springs and mud pots, as well as fumaroles and yellowish sulfuric deposits (Figs. 13.13, 13.14, 13.15, 13.16 and 13.17).

Why are there geothermal fields here at this location? Is there any clear reason for them to be here and not, say, some hundreds of metres to the east of Seltun? The most likely reason is that north-trending (some north-northwest trending) fractures dissect Sveifluhals exactly where the main geothermal fields at Krisuvik-Seltun are. These fractures are entirely similar to the ones we see in Fig. 13.7–but Krisuvik-Seltun is south of Fig. 13.7–and are most likely earthquake fractures. Thus, once again it is clear that earthquake fractures control geothermal fields—as they do, in fact, worldwide. One reason for this control is that earthquake fractures, many of which slip (move) repeatedly, maintain the permeability—make it possible for the geothermal fluids to flow to the surface. Another reason is that large earthquake fractures are normally very deep, easily reaching to crustal depths of many kilometres, and some of the northerly trending fractures reach depths of 10 km. At depths of many kilometres the crust in Iceland is very hot—hundreds of degrees Celsius. So, even if there is no known shallow magma chamber beneath Sveifluhals, the fact that the earthquake fractures reach great depths and a very hot crust, means that the hot fractured crust can act as a heat source for the geothermal fluids. Thus, **circulation of water** from rain and snow to **depths of many kilometres** in an **earthquake fracture** is the primarily

Fig. 13.13 Overview of part of the Krisuvik-Seltun (Krísuvík-Seltún) geothermal fields. View east, the light colours are due to the geothermal fluids. Compare Fig. 11.2

Fig. 13.14 Close-up of the steam-vents in Fig. 13.13. The people provide a scale

Fig. 13.15 Yellowish sulphuric deposits in the Krisuvik-Seltun geothermal fields

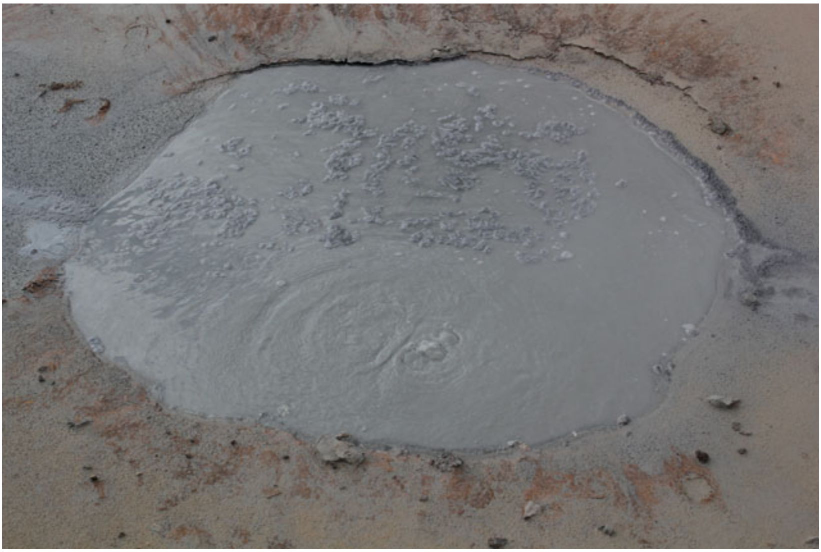


Fig. 13.16 Muddy hot spring in the Krisuvik-Seltun geothermal fields

Fig. 13.17 Mudpots and deposits in the Krisuvik-Seltun geothermal fields

explanation of the geothermal fields at Kristuvik-Seltun. And the same applies to the nearby fields southwest of Seltun (in the Krisuvik area, proper). These fields are also related to north-trending earthquake fractures.

13.8 Explosion Craters—Maars

Some of the earthquake fractures associated with the geothermal fields at Krisuvik-Seltun meet with a lake in the south. This lake is famous for its green colour and most appropriately has the name the Green Lake **(Graenavatn**, **Grænavatn)**, which is our next stop. So we drive on from the geothermal fields at Krisuvik-Seltun to Graenavatn, our **fifth stop (5)**. Both Graenavatn, east of the road, and a somewhat smaller lake, Gestsstadavatn **(Gestsstaðavatn)**, west of the road, are **explosion craters**—which in geology are named **maars**. There are several other explosion craters in the area, mentioned below, but here I focus on Graenavatn, which is the most striking one, not the least because of its impressive green colour (Figs. 13.10 and 13.18).

Graenavatn is somewhat elongated in the east-west direction, reaching a diameter of about 360 m compared with a diameter of 300 m in the north-south direction. Its maximum depth is about 45 m. The **sea-green colour** of the lake is primarily related to minerals of silica and sulphur, both of which enter the lake through geothermal springs, all of which are presumably related to the north-trending earthquake fractures discussed above.

During its formation Graenavatn produced some eruptive materials, both lava and **pyroclastics** (fragmented rock, where the fragments are generated in explosions in the associated crater). These eruptive materials are well known for containing and unusually high proportion of gabbro **xenoliths**. In geology a xenolith, which literally means 'foreign rock', is used for rock fragments, usually from the surrounding crust, which the magma in a dike or other types of conduits picks up on its way to the surface. **Gabbro** forms when basaltic magma solidifies at considerable depth in the crust. You recognise the gabbro xenoliths as being somewhat light grey within the much darker basaltic fragments around the lake. And in contrast to the basalt, where the crystals or grains that constitute the rock can hardly be seen (they are too small for the unaided eye), the crystals in the gabbro are easily seen by the naked eye.

Fig. 13.18 Aerial view of the explosion craters (maars) of Graenavatn (Grænavatn) and Geststadavatn (Gestsstaðavatn) south of Lake Kleifarvatn. Graenavatn has a maximum diameter of 360 m and a maximum depth of 45 m. Its sea-green colour is primarily related to minerals of silica and sulfur

Now, why are these gabbro xenoliths important? The reason is that they tell us that somewhere below Graenavatn there is a gabbro body that may be developing into a shallow **magma chamber**. We know what such a shallow chamber would most likely look like, namely a sill-like body similar to what we saw in Stardalshnukar (Figs. 4.7–4.9; Chap. 4). So far, however, we have no evidence of any shallow magma chamber in any of the volcanic systems on the Reykjanes Peninsula, (Fig. 2.3); the magma seems to come straight from a deep-seated magma reservoir (Fig. 2.4). Gabbro xenoliths, by contrast, indicate thick intrusions, sills, at a shallow depth beneath Graenavatn, and, as we know, sills are 'baby' magma chambers. Thus, the xenoliths may indicate that a magma chamber is developing below Graenavatn and, presumably, the surrounding areas, including the geothermal fields at Krisuvik-Seltun and the southern part of Kleifarvatn.

In addition to the explosion craters of Graenavatn and Gestsstadavatn, there are several small explosion craters about half a kilometre south of Graenavatn. These craters are named **Augun (The Eyes)**. Augun are much smaller craters than Graenavatn and Gestsstadavatn, and are not related to north-trending fractures, but rather follow the common volcanic fissure trend in the area, namely northeast-southwest (Fig. 13.7). Each of these explosion craters produced comparatively little eruptive material (lava and pyroclastic materials). They are thought to have formed in the past 10 thousand years; they may be as young as 6–7 thousand years, but their ages are really not well known.

What is well known, however, is how explosion craters of the maar type form. They are the result of **explosions** which occur when the **hot magma** in the feeder-dike **meets groundwater**. Recall that the common temperatures of basaltic magmas are 1100–1300 °C. The sudden vaporisation of the groundwater results in explosions that generate the craters. Just like Lake Kleifarvatn itself, these craters extend below the water table (the surface of the groundwater), so that they become mostly filled with groundwater—hence the beautiful lakes (Figs. 13.10 and 13.18). While the largest maar here, Graenavatn, is 360 m in diameter, some maars in other areas reach many kilometres in diameter, and may be anywhere between one and two hundred metres deep. They reach below the water table and are thus lakes.

13.9 Table Mountain Cut in Half

We now drive on along Road 42 until it meets Road 427, where we turn right, that is, to the west, on our way to the edge or toe of the Reykjanes Peninsula. On Road 427 we first pass through interglacial lava flows (erupted in-between two glacial periods; perhaps around 100 thousand years old) but soon enter the young lava flow of **Ögmundarhraun**. This aa lava flow formed at around the year 1170 and is thus about 850 years old. It is one of the youngest lava flows on the Reykjanes Peninsula, and is of a very similar age as Kapelluhraun—which forms part of the site of the town Hafnarfjördur (Fig. 2.1; Chap. 2)—in the northern part of the Reykjanes Peninsula. In fact, it is probable that these two lava flows may have formed in the same 'fires'. In geological context, **fires** refer to extensive volcanotectonic episodes, lasting many years, sometimes tens of years, the most recent one of which was the Krafla Fires in North Iceland from 1975 to 1984. We do not make a formal stop in Ögmundarhraun, but rather drive further to the west through the lava fields until we come to the sixth stop.

The **sixth stop (6)** is at **Festarfjall**. This hyaloclastite mountain is remarkable for three things: it is cut in half by sea erosion, it has an exceptionally well exposed feeder-dike, and it was most likely not formed in a subglacial eruption but rather in a submarine eruption. Festarfjall (Fig. 13.19) makes it possible to study the interior of a young hyaloclastite mountain. Basically, it has the shape of a table mountain, similar to the ones we have seen before, namely Hrafnabjörg at Thingvellir (Fig. 6.5; Chap. 6) and Hvalfell in Hvalfjördur (Fig. 11.23; Chap. 11). Festarfjall, however, is much smaller than either of these table mountains. The name Festarfjall, which means the mountain with the chain (or rope or anchor, possibly necklace), derives from the feeder-dike, which looks like a chain/rope/necklace from a distance (Fig. 13.19). Please notice that the cliffs are very steep. The beach is beautiful, but **rockfalls** are common, so that the beach is sometimes closed and in general great caution is in order if you want to go down on the beach. In fact, most of the geology can be seen very well from a distance.

The mountain itself stands on an interglacial basaltic lava flow, perhaps 100 thousands years old, which in turn rests on hyaloclastites. Thus the hyaloclastites and breccias beneath the lava flow (Fig. 13.19) are older than Festarfjall itself, which is only about 90 m tall when measured from the lava flow on which it stands. The mountain is also mainly composed of hyaloclastites with several dikes and inclined sheets. The **feeder-dike** (Fig. 13.19) clearly dissects the entire mountain and reaches to the top, where it supplied magma to a small lava flow. In this sense, therefore, Festarfjall is a very small table mountain, as mentioned. I emphasise, however, that the lower part of the cliff section, that is, the section below the lava flow in the middle part, is older. This means that the cliff section is formed in at two and possibly three eruptions. The lowermost **breccia** (black) changes abruptly to a very fine brownish hyaloclastite of the type which geologists call **tuff**. On top of the tuff is basaltic **lava flow**, with clear brecciated layers (scoria), at its top and bottom.

This lava flow is the foundation on which Festarfjall itself stands. Again, it is likely that Festarfjall formed in an **eruption in the sea** rather than in an ice sheet.

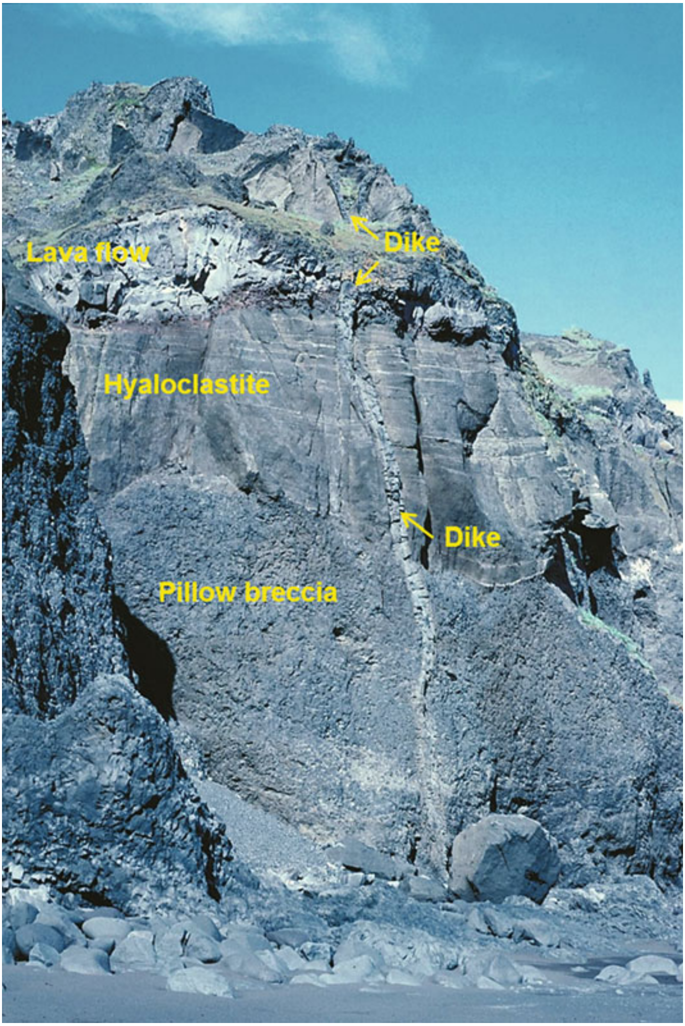


Fig. 13.19 Hyaloclastite mountain Festarfjall, a sort of small table mountain cut in half by sea erosion and presumably formed in submarine eruption. View northeast, the mountain is largely made of pillow breccia and hyaloclastite tuff. Its upper part rests on a lava flow. A feeder-dike (indicated) dissects the mountain to feed a small lava flow at its top

This follows because the total height of the mountain is only 190 m, and its height above its foundation, the basaltic lava flow, only 90 m. The mountain is dissected by several intrusions. The most notable intrusion is, of course, the **feeder-dike** to the top lava flow (Fig. 13.19). It is not common to see a feeder-dike connected clearly to its lava flow. The rarity of such an observation is partly because the magma level in the feeder commonly sinks at the end of the eruption, sometimes resulting in the formation of a pit crater. Partly, however, the reason is that feeder-dikes are connected to their lava flows in a clear way primarily in the main crater cones that form along the volcanic fissure, and the crater cones constitute only a fraction of the length of the fissure. Thus, to see the feeder-dike well connected, the section, here the sea cliffs, has to be exactly where the feeder was connected to a cone. In addition, **most dikes are non-feeders**; they never reached the surface to feed an eruption, and therefore end without being connected to either lava or pyroclastic layers. Later today we will see exceptional examples of both types of dikes, namely a clear feeder-dike and, nearby, an exceptionally clear non-feeder, that is, an arrested dike, both of which are seen only a few metres below the surface.

But how did Festarfjall form during an eruption in the sea? Clearly, its top is now at 190 m above the sea level. The answer is that during the ice periods the enormous weight of the ice sheet, perhaps 2000 m thick, pressed Iceland downwards so that the land under the ice sheet was everywhere much lower than it is today. Even today, the land below the largest ice cap in Iceland, Vatnajökull, has a lower elevation (is bent down) than it would have if Vatnajökull disappeared through melting—perhaps by as much as many tens of metres. At the end of each glacial period, melting of the ice-age ice sheets is comparatively rapid, resulting in the amount of water in the sea increasing and the sea level rising **rapidly**. When the weight of the ice sheets decreases during the melting, the land also rises, but much **slower** than the sea level rises.

In some areas, such as Scandinavia, the land is still rising some 10 thousand years after the main ice sheets melted. This slow rise is because the mantle below Scandinavia is essentially solid and responses very slowly to pressure or stress changes, such as result from the ice load vanishing. By contrast, the mantle below Iceland contains a lot of magma or melt and responds much more quickly to load changes than in Scandinavia. Yet, it took as long as 1000 years for Iceland to **rise fully** after the ice sheets melted away. The **sea level** was much lower during the **ice periods** than it is today (considerable water, that would otherwise be a part of the sea, constituted the ice sheets in the Northern Hemisphere, extending far south in North America and Eurasia). Therefore the ice sheet covered not only the present Iceland, but out to considerable distances from the present coast. In fact, at its maximum during the last glacial period the ice sheet on Iceland covered an area more than **double the present size** or area of Iceland. There was, therefore, thick ice on the Reykjanes Peninsula, and Iceland in general, suppressing the land.

As the sea level rose quickly during the melting of the ice sheet, the rise of the land **lagged behind** (came later), so that, for a while, the sea level was **much higher** than it is today. In West Iceland the shoreline was, at one period, **100–150 m higher** than today. It follows that part of the area where Festarfjall is today was most likely in the sea when the eruption that formed the mountain happened—hence the **submarine eruption**. As indicated above, many of the smaller hyaloclastite mountains on the Reykjanes Peninsula may have formed, partly at least, in submarine eruptions.

Do we have any analogies to such eruptions? Yes, indeed, we do. One of the most famous eruptions in Iceland in recent decades is the Surtsey Eruption. Surtsey, and island offshore South Iceland and a part of the Vestmannaeyjar Archipelago (Fig. 14.12, Chap. 14), formed in an eruption in the sea, a submarine eruption that lasted from 1963 to 1967. The eruption reached the sea level, at which stage a lava cap formed on the top of the hyaloclastite island, just like in a table mountain (Figs. 6.5 and 6.6). Although Festarfjall is smaller than Surtsey, their formation may have been very similar: eruption in a shallow sea.

13.10 Young Lava Flow and a Graben

We now drive on to the west along Road 427, until we arrive at the town of **Grindavik (Grindavík)**. The entire town is located on lava flows ranging in age from about 2000 years to perhaps 8000 years. Grindavik is one of the very few towns in Iceland that is located entirely within the active volcanic zone. The evidence of this activity is not only the fact that the town stands on young lava flows, but also the nearby volcanic fissures, seen to the northeast of Road 43, on which we drive out of the town, and in the large tension fractures, normal faults and grabens, in the vicinity of the town.

From Grindavik we drive to the north along Road 43 to the next stop, the **seventh stop (7)**. To the right (east) of the road is part of the volcanic fissure that generated one of the lava flows on which the town of Grindavik is built, namely Sundhnukahraun (**Sundhnúkahraun**, or Sundhnúkshraun), formed some 2400 years ago. I show you part of Sundhnukahraun, as well as its volcanic fissure and associated lava channel, from the top of the hyaloclastite mountain **Thorbjörn (Þorbjörn)**. I am not recommending that you climb up the mountain to get a similar view (the gravel road/path up the mountain is normally closed and not recommended). However, the panorama view in Fig. 13.20 gives you a good overview of the volcanic fissure, whose total length is about 8.5 km, its surroundings and related structures.

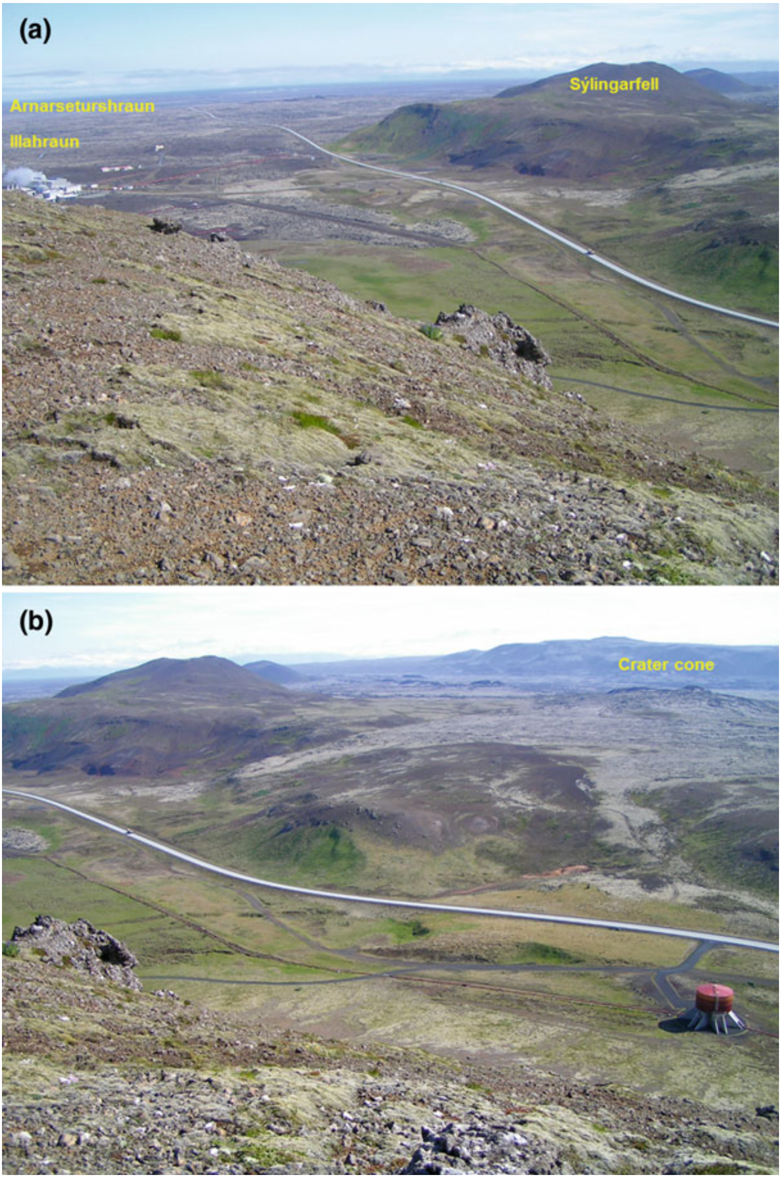
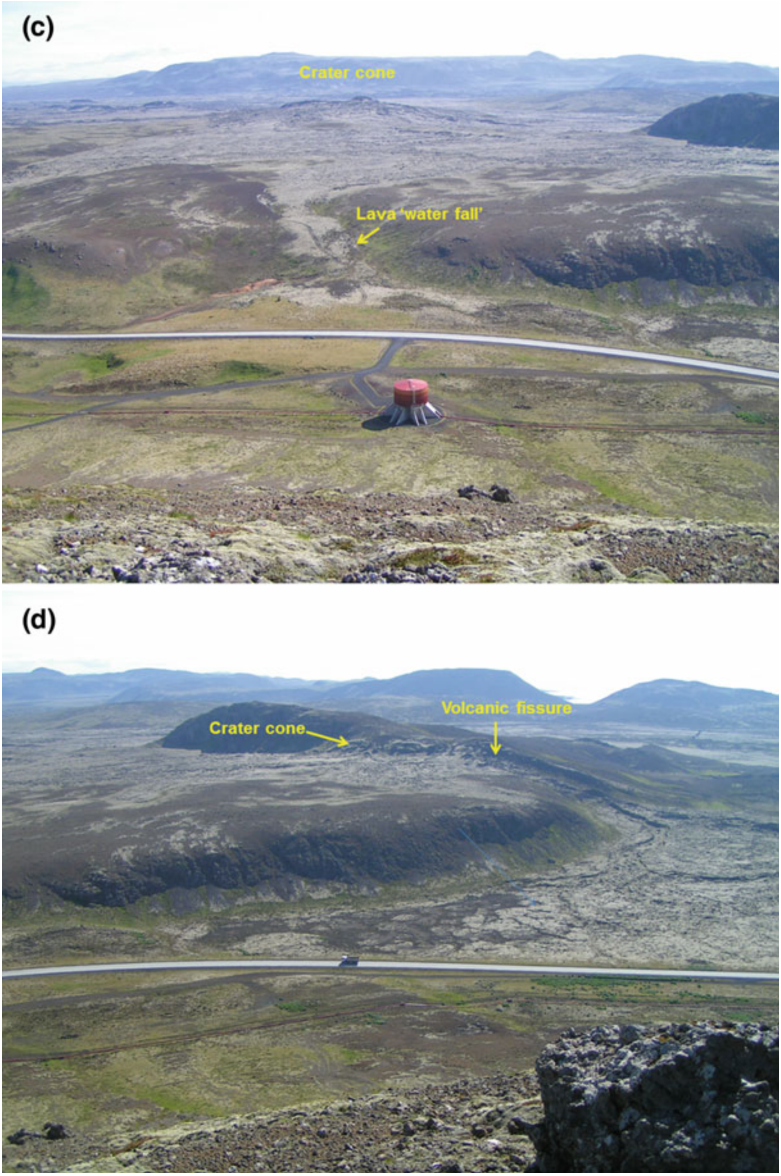


Fig. 13.20 (continued)



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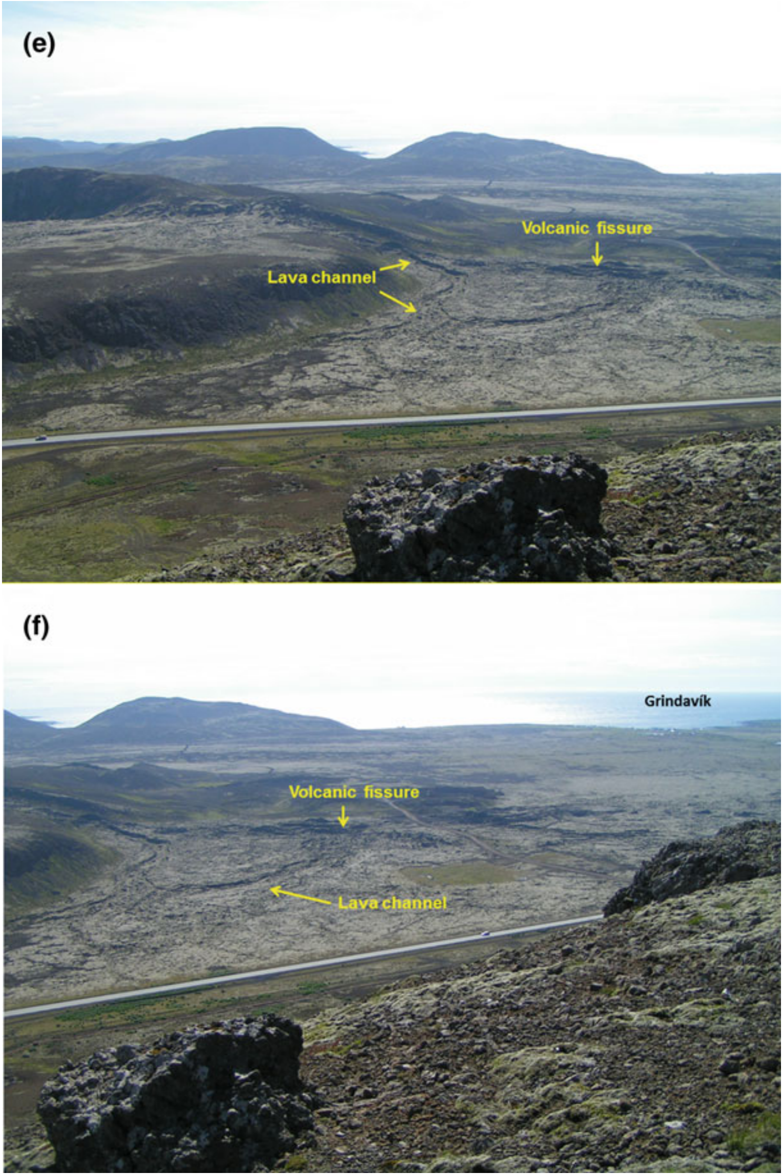

Fig. 13.20 (continued)

◀ **Fig. 13.20** View on the lava flows, volcanic crater cones, and volcanic fissures as seen from close to the top of the mountain Thorbjörn (Þorbjörn). **a** View north, Arnarseturshraun and Illahraun formed during the Reykjanes Fires, some 800 years ago. Part of the Svartsengi geothermal power plant is also seen. The Blue Lagoon is located in Illahraun (Fig. 13.24). Sylingarfell (Sýlingarfell) is a hyaloclastite mountain. **b** View northeast, crater cone and lava streams associated with Sundhnukahraun (Sundhnúkahraun), about 2400 years old, and related lava flows. **c** View east, close-up of the crater cone and lava stream or 'water fall' in Fig. 13.20b. **d** View east, crater cone and volcanic fissure associate with Sundhnukahraun. **e** View southeast, lava channel and volcanic fissure associated with Sundhnukahraun. **f** View southeast, volcanic fissure, curved lava channel, and continuation of the lava flow to the coast, where part of the town of Grindavik (Grindavík) is seen standing on the lava flow

This stop is also to take a look at the very noticeable **graben** through the mountain **Thorbjörn (Þorbjörn)**. This is one of the most striking evidences of rifting in this area (Figs. 13.21 and 13.22), again showing that the town of

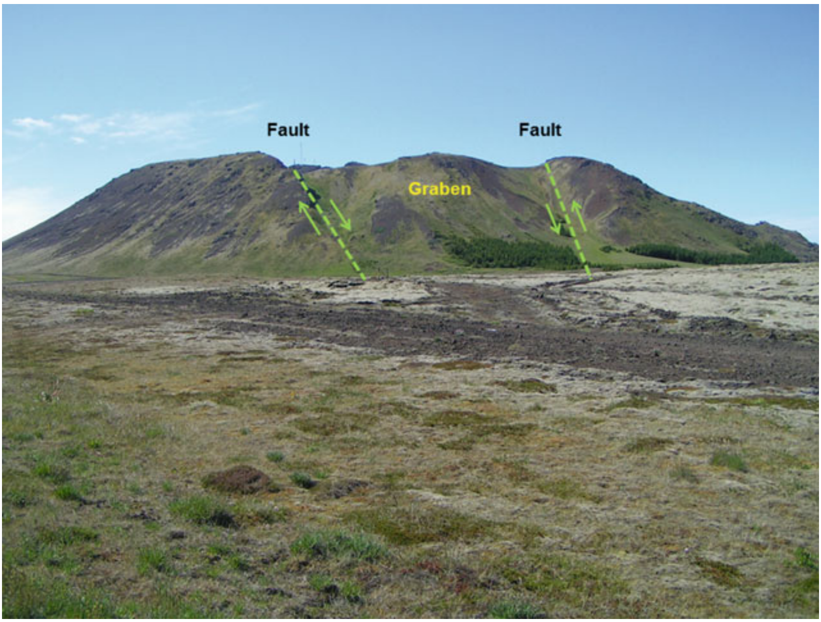

Fig. 13.21 View southwest, part of the graben of Thorbjörn (Þorbjörn), close to the Blue Lagoon. The green arrows indicate the relative movement along the normal faults that constitute the boundary faults of the graben. Thus, the rock segment constituting the graben itself has subsided, moved down, relative to the part of the mountain outside the faults. The graben is about 400 m wide (compare Figs. 11.17 and 11.18)

Fig. 13.22 View southwest, close-up of the graben dissecting the mountain Thorbjörn (Þorbjörn). The subsidence or vertical displacement reaches tens of metres

Grindavik is within a very active area of rifting. The faults have vertical displacements, throws or subsidences, of tens of metres and the width of the graben is about 400 m. The graben is in many ways very similar to the graben in Brynjudalur in Hvalfjördur (Figs. 11.17 and 11.18; Chap. 11), although the graben in Brynjudalur is narrower than the one in Thorbjörn. When we drive back, we shall see the same graben **from the south** (Fig. 13.23). It should be noted that even if we classify Thorbjörn as a hyaloclastite mountain—as a moberg (móberg) mountain in Icelandic terminology—it is largely composed of **pillow lava**, such as we have seen earlier (Chap. 6) and see again later today. The height of the mountain is 243 m above sea level and it was most likely formed in a subglacial eruption.

Grabens of this type characterise rifting everywhere in the world. At a width of about 400 m, this graben is rather narrow, although more than double the width of the graben in Brynjudalur. We recall that the deepest, central part of the Hengill Graben is mostly 700–1800 m wide, and the Thingvellir Graben is 5000–7000 m wide. However, the vertical displacements or subsidences of the Thingvellir Graben (Figs. 5.8, 5.9 and 6.3; Chaps. 5 and 6), the Brynjudalur Graben (Figs. 11.21 and 11.22; Chap. 11), and many of the faults associated with Hengill (Figs. 12.12 and 12.13) are, in fact, similar to those of the Thorbjörn Graben.

How large earthquakes would have been generated during the formation of the Thorbjörn Graben? The answer is that it depends much on how the graben formed; more specifically, on whether the total subsidence occurred in one or two seismic events, or in many such events. If the subsidence occurred in a few events or fault slips, then the earthquake during each slip could have reached magnitude 7. These are strong to major earthquakes, such as are felt widely when they occur. More

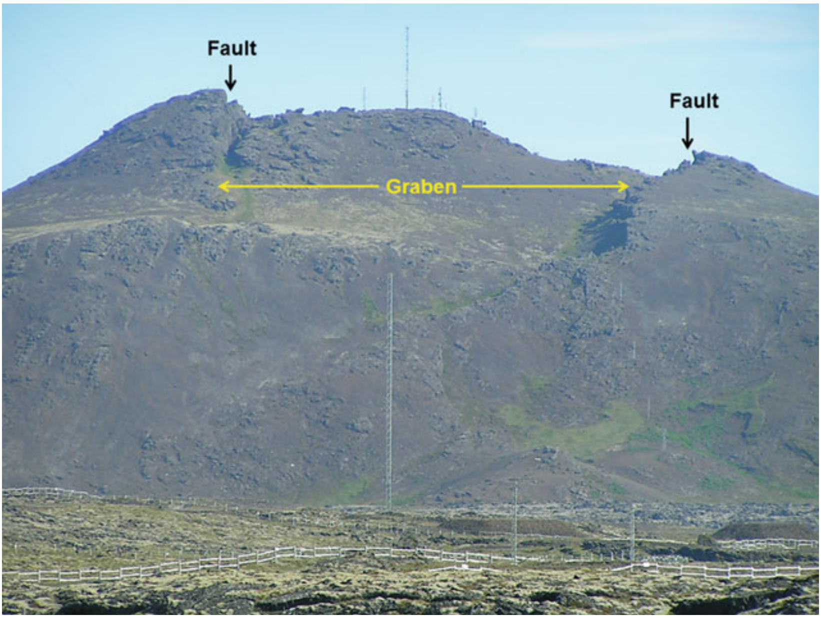

Fig. 13.23 View northeast, the graben of Thorbjörn (Þorbjörn) as seen from the southwest. Grabens of this type are very common in areas undergoing horizontal extension, rifting, such as in the West Volcanic Zone. Compare Figs. 4.14, 13.21 and 13.22

Fig. 13.24 View northwest, Illahraun and the Blue Lagoon as seen from the graben at the top of the mountain Thorbjörn (Þorbjörn). Compare Fig. 13.20a

likely, based on analogy with subsidence events in the Thingvellir Graben and elsewhere in the rift zone of Iceland, the total subsidence was reached in many comparatively small displacements, perhaps one to three metres each. These would normally give rise to earthquakes of the order of magnitude 6, which are regarded as strong but not major earthquakes. Earthquakes of magnitude 6 do occur on the Reykjanes Peninsula from time to time and are felt over large areas in Southwest Iceland.

At the seventh stop we are also close to the **Blue Lagoon (Bláa Lónið)**, which has become one of the most popular touristic place in Iceland (Fig. 13.24). The geothermal water of the Blue Lagoon derives from deep wells that pass through the aa lava flow **Illahraun** (Fig. 13.20a), which is one of the youngest lava flows on the Reykjanes Peninsula and erupted some 800 years ago (erupted sometime in the years 1210–1240). Stopping at the Blue Lagoon may be interesting, but is not a part of the present excursion whose focus is on the geology.

13.11 Volcanic Fissure and Flow Channelling

We now turn, and drive back, to the south along Road 43 until it meets Road 425. We drive onto Road 425, to the right, that is, to the west, on our way to the southwest edge or the toe of Reykjanes. We may make a short stop, then regarded as part of the seventh stop, just to view the Thorbjörn Graben from the southwest (Fig. 13.23), but then move on to the west along the road. On Road 425 we see many tectonic fractures, both tension fractures and normal faults, as well as volcanic fissures. One volcanic fissure erupted at about the same time—in the same 'fires'—those that generated Illahraun (at the Blue Lagoon), and we make a short stop where the road passes through the associated lava flow.

This is the **Eldvarpahraun (Eldvörp Lava)**, the **eighth stop**, which appears to have formed at a very similar time to that of Illahraun and Arnersetarshraun (Fig. 13.20a), and perhaps other lava flows. They are all formed at around 800 years ago probably during a volcanotectonic episode, fires, which lasted many years. At the eighth stop, the road passes through a broad lava stream that extends all the way to the coast. The volcanic fissure, also named Eldvörp, that issued this lava is discontinuous and composed of many offset segments, but its total length is close to **11 km**, one of the longest one on the Reykjanes Peninsula. Because of its length and location, the volcanic fissure and the associated crater cones are best seen from the air (Fig. 13.25).

The Eldvörp lava varies in character **between aa and pahoehoe**. Close to Road 425 it is a pahoehoe lava, whereas further inland parts of it are aa lava. The morphological characteristics of lava flows depend on many factors, such as volumetric flow rate (the amount of lava issued by the volcanic fissure over a given time period, say a second), viscosity (how easily the lava flows), temperature, and the slope of the land on which the lava is flowing. The lava composition has also strong effects. The same volcanic fissure can issue aa and pahoehoe, and commonly a pahoehoe lava may change into aa when the lava cools. Earlier we saw that the Nesjahraun lava also varies in character between aa and pahoehoe (Chap. 12).

The Eldvörp is a typical volcanic fissure or **crater row** (Fig. 13.25). It is composed of many segments or parts, whose nearby ends are separated by distances as long as 500 m. In addition, the segments are not in the same straight line, are **not collinear**, but rather out of line, that is, **offset**, either to the east or west of an (imagined) straight line by as much as 200 m. Earlier we saw normal faults and tension fractures composed of many offset segments (Chap. 5). There are many

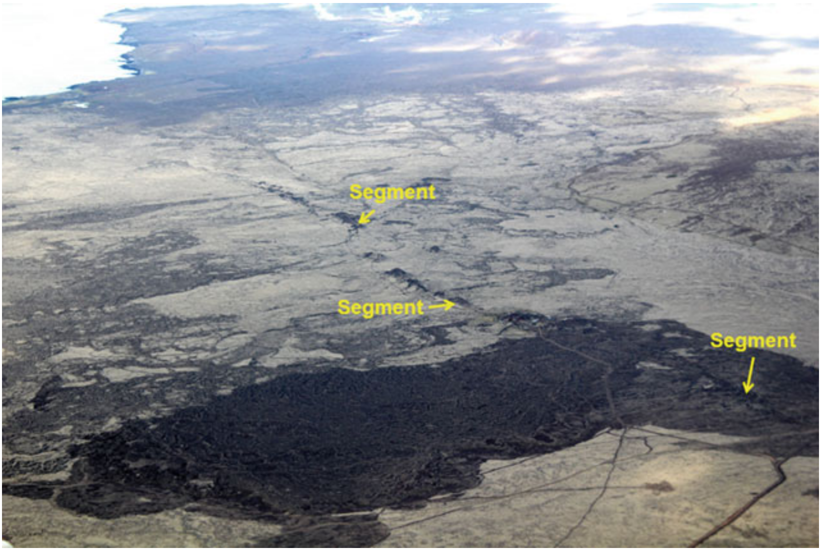

Fig. 13.25 Aerial view of part of the Eldvörp Volcanic Fissure, formed around 800 years ago. Like most volcanic fissures, it is composed of many segments, some of which are indicated here. The dark hillocks are crater cones. View southwest, Reykjanes and its geothermal fields are seen in the distance. The total length of the fissure is close to 11 km

small hills, or **crater cones**, along the fissure segments. Yet, only a fraction of the total length of the volcanic fissure is covered with crater cones.

There are several reasons for these geometrical features of volcanic fissures. First, they are segmented and offset because that is the way the **feeder-dikes** themselves (and, in fact, all dikes) are. Within the zone of high stress (concentrated or raised stress) in the crustal layers above an upward-moving or propagating dike, the layers rupture where they are weakest, that is, have the **least tensile strength**. Because of variations in the tensile strength within each layer, rupture normally occurs not along a straight line above the top of the dike at any time but rather at different points which may be far apart—resulting in the dike front being composed of many separated segments or '**fingers**' (Figs. 13.26 and 13.27). The eruption then begins when the first finger reaches the surface. Subsequent fingers are initially far apart, forming a discontinuous fissure composed of offset segments. As the eruption intensity increases, and more fingers reach the surface, some of the segments may link up to form longer segments. But it is very rare to see the entire volcanic fissure erupting at the same time—normally only parts of the fissure are active at any given time.

So why is the eruption confined to parts of the fissure? The reason is that from the beginning the opening of the fissure, its **aperture**, differs somewhat from one measurement point to another along the fissure. This we know from simply observing tension fractures or open or gaping normal faults, such as at Thingvellir (Chaps. 5 and 6). All such fractures show significant variation in their opening along their lengths. Similarly, dikes followed laterally along their lengths show significant variations in thickness and thus in original opening or aperture of the dike-fracture. The volume of magma or lava issued through a part of the fissure/feeder-dike depends very strongly on the opening or aperture of that part. In fact, the volume of magma issued depends on the opening of the fissure in the 3rd power. This means that if one segment of the fissure has an opening of 1 m and the adjacent equally-long segment an opening of 3 m (everything else being the same),

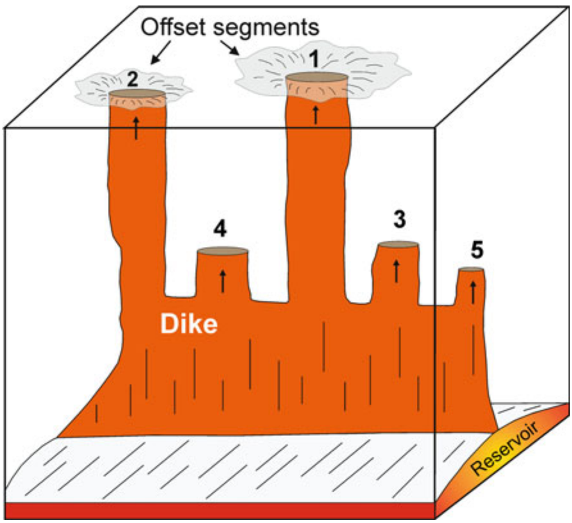

Fig. 13.26 Dike propagates as offset, separate segments, whose vertical propagation fronts are referred to as 'fingers'. Eventually, the fingers may or may not become laterally connected. In this schematic illustration, fingers or segments 1 and 2 have already reached the surface, whereas segments 3–5 are propagating towards the surface. Segments 3–5 may all reach the surface; alternatively, some or all of them may become arrested, stop their vertical propagation, at some crustal depth. This is a regional dike whose source is a large magma reservoir, as is appropriate for the Eldvörp feeder-dike (Fig. 13.25)

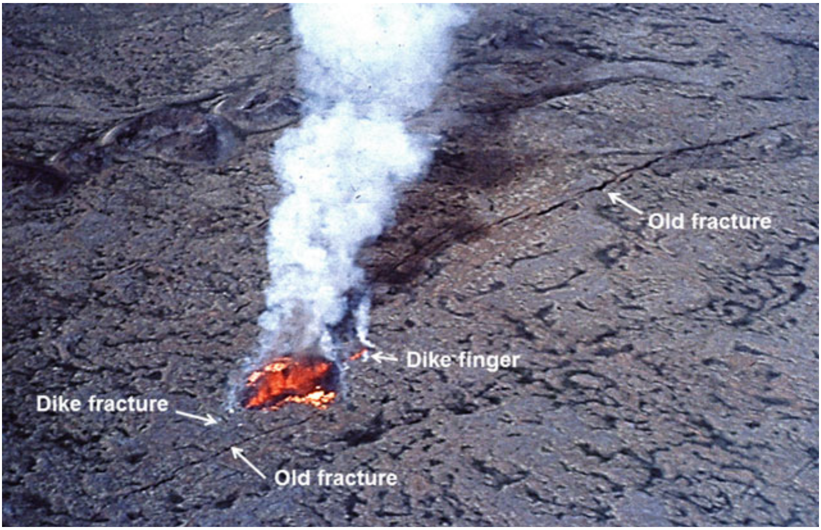

Fig. 13.27 Example of the first dike fingers to hit the surface, as seen here in the July 1980 eruption in the Krafla Volcanic System in the North Volcanic Zone (the zone is located in Fig. 2.2). The first dike finger or segment to reach the surface initiates the resulting fissure eruption. The first segments at the surface are commonly short, tens of metres, but they commonly propagate laterally at the surface and may eventually link up into larger fissure segments (Fig. 13.25). The indicated 'dike fracture' is a tension fracture forming ahead of a dike segment that has still not reached the surface as a finger

so that they differ by a factor of 3, then the second segment would issue $3^3 = 3 \times 3 \times 3 = 27$ times as much lava/magma as the first segment. This is a **universal law** that controls fluid transport in fractures, whether the fluid is groundwater, geothermal water, gas and oil, or magma. The law is named the **cubic law**, because a cube has the dimensions of length in the third power. The cubic law is the main reason why fluid transport, including magma transport, in fractures tends to be focused on, or channelled to, the large-aperture parts of the fracture. This process of channelling the flow to the large-aperture fractures of the fracture is referred to as **flow channelling**. Like the cubic law, flow channelling is a universal principle that applies to the flow of any type of fluids in crustal fractures.

How does the cubic law relate to the crater cones (Figs. 13.20d and 13.25; see also Fig. 12.21)? Answer: The crater cones tend to form along the segments, or those parts of the segments, where the opening is largest. Because of the cubic law, many times more magma is from the beginning issued through the **large-opening**

parts. It follows that these parts produce most of the lava and other eruptive materials. These are also the parts which the magma continues to flow through during the waning states of the eruption, when the magma pressure has decreased and the overall volumetric flow rate has fallen. Since the volume of hot magma flowing through the large-opening parts is much greater than through the adjacent narrower parts, the large-opening parts may somewhat increase their apertures through partial melting and removal of parts of the rock in the walls of the fissure. This melting and removal of wall rock is referred to as **thermal erosion** and may result in gradual increase in apertures during the course of the eruption of those parts of the fissure that had large openings from the beginning. All these factors contribute to much of the magma being transported through certain parts of the fissure; parts that develop into specific conduits that pile up lava, spatter, and scoria so as to become **crater cones** (Figs. 12.21, 13.20d and 13.25; Sect. 13.12).

13.12 Reykjanes—Lava Shields and Lava Fields

We will see close-by examples of crater cones, the result of the cubic law and flow channelling, at the next stop. So we move on and drive to the west towards the toe of the Reykjanes Peninsula, which is our **ninth stop (9)**. There are many things to look at here, but after parking the car perhaps the best approach is to walk first to the south edge of parking area to take a look at the rocks that constitute the hill of **Valahnjukur** (**Valahnjúkur**, the name is also given on maps as the plural, Valahnjúkar) which we will walk up. While the small hill adjacent to the parking place is somewhat separated from the main hill to the east, they constitute the same hyaloclastite mountain. As seen in the sea cliffs, the mountain is primarily of **pillow lavas**, **pillow breccia**, **and tuff** (Fig. 13.28). The pillow lava is very clear, with layers of hyaloclatite tuff (basically ash layers) in between but becomes more broken into smaller rock fragments higher in the cliffs.

From here we follow the path to the top of the hill of Valahnjukur. I do not recommend that you go to the very edge of Valahnjukur since the rocks there are loose and the wind can be strong. You get as good overview even if you stand at some distance from the very edge. On the way you may notice that there are several points of land projecting into the sea, including the one where we observed the pillow lava at the parking place. All these points trend roughly

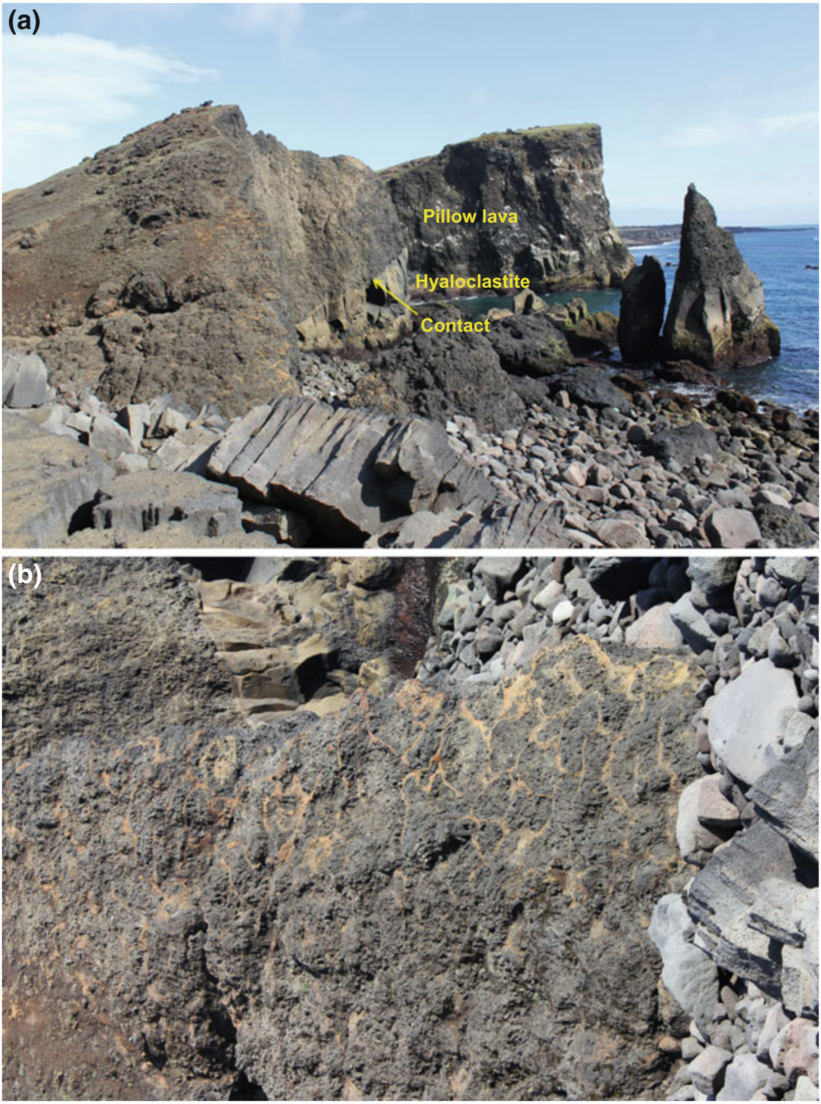

Fig. 13.28 Pillow lavas and hyaloclastite (tuff) seen at Valahnjukar (Valahnjúkar). **a** View east, the contact between the tuff and the pillow lava is very sharp. **b** Close-up of part of the pillow lava in (**a**). Compare Figs. 6.14–6.18

northeast-southwest, that is, parallel with the main faults on Reykjanes and are related to **normal faults**. The clearest ones are on either side of the 60–70 m wide cove that you pass on the path to Valahnjukur itself. Valahnjukur is a hyaloclastite mountain, primarily made of pillow lava, pillow breccia, and tuff (Fig. 13.28). The mountain reaches only 43 m above sea level and is formed in an **eruption in the sea**, a submarine eruption, most likely at the end, or just after the end, of the last ice period—perhaps some 13 thousand years ago when the sea level was for a while many tens of metres higher than it is today (see the discussion about sea-level changes at the sixth stop above). Although smaller, Valahnjukur is in many ways similar to Festarfjall. About half of each volcano has been eroded away through the action of the sea, so that both now constitute high and close to vertical sea cliffs.

Looking to the southeast and east from Valahnjukur, we see many geological features of interest. The 'toe' of Reykjanes, **Reykjanesta (Reykjanestá)**, is seen in Fig. 13.29, as well as the eastern boundary fault of the main graben that comes on land on Reykjanes. The main graben, which may be regarded as an on-land continuation of the mid-ocean Reykjanes Ridge, has a width of 5–6 km. The boundary fault dissects a pahoehoe lava flow which constitutes a part of a small **lava shield**, named **Skalafell (Skálafell)**. Small lava shields of this type (Fig. 13.30) are common on the Reykjanes Peninsula. They all have the same 'shield geometry' as Skjaldbreidur (Fig. 6.7; Chap. 6) but are much smaller. This one, Skalafell, issued the pahoehoe (basaltic) lava flow which forms the 'toe' of Reykjanes and, as said, is seen dissected by the boundary fault in Fig. 13.29. Its volume, however, is small; perhaps about 0.2 km^3 or about one per cent of the volume of Skjaldbreidur. Even if Skalafell is small, it is not formed in a single eruption; there is evidence of later eruption from its top crater. The main eruption that formed the lava shield, however, presumably occurred some 9–10 thousand years ago.

While most lava shields in Iceland appear to have formed in single eruptions, there is, in principle, nothing extraordinary with their being the result of more than one eruption. It is well known that many dikes, including feeder-dikes to eruptions, are '**multiple**', that is, formed by **many magma injections** into the same fissure. Many such dikes have been observed in deeply eroded sections, such as in the Hvalfjördur area (Chap. 11), and some dikes supplying magma to active volcanoes are known to be multiple. Perhaps the best known is the feeder-dike to the summit eruptions of the Hekla Volcano, as we discuss in Chap. 14.

If we now take the northeast view, we see the hill on which the lighthouse stands, the lava fields, and the nearby geothermal field (Fig. 13.31). The hill is

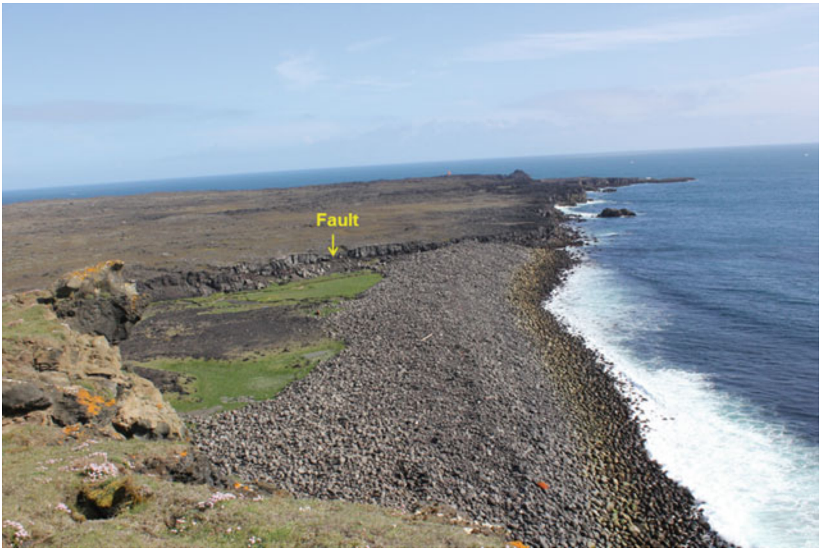

Fig. 13.29 View southeast to the toe of Reykjanes (Reykjanestá). The normal fault seen here is the southeastern boundary fault of the main 5–6 km wide graben that comes on land here (extending from the mid-ocean Reykjanes Ridge offshore). The other boundary fault is seen in Fig. 13.39

named **Baejarfell (Bæjarfell)**. Baejarfell reaches 50 m above sea level and, like Valahnjukar on which we stand now, is largely composed of pillow lava, and was most likely formed in **submarine eruption** some 13 thousand years ago. The lighthouse used to be located at Valahnukar, but because of the action of the sea which gradually erodes the hill, it was moved inland, to Baejarfell, in the early 20th century.

To the right (east) of Baejarfell, there is a geothermal field, referred to as **Gunnuhver**. This is a small geothermal field with many vents of steam and boiling mudpots, including one of the largest mudpots in Iceland. Nearby is the Reykjanes Geothermal Power Plant, producing some 100 MW of electricity, using steam from 12 wells whose bottom temperatures, at the depth of around 2700 m, are about 300 °C. While the geothermal field is certainly worth visiting, I do not include it as a part of this excursion since we have already explored the larger geothermal fields in Krisuvik today and, earlier, the fields at Geysir (Chap. 7).

Fig. 13.30 Continuation of the normal fault in Fig. 13.29. View east, the lava shield Skalafell is 9–10 thousand years old

Fig. 13.31 View northeast, the hyaloclastite mountain Baejarfell (Bæjarfell) reaches 50 m above sea level and was most likely formed in a submarine eruption some 13 thousand years ago. It is largely made of pillow lava. To the east is the geothermal field Gunnuhver

13.13 An 800-Year-Old Fissure Eruption

Close to you in Fig. 13.32 are black lava fields. The pahoehoe lava flow seen here is from the youngest eruption in this part of the Reykjanes Peninsula, namely the **Yngri Stampar (Younger Stampar)** eruption. This eruption is a part of the Reykjanes Fires that also generated the lava flows of Eldvörp (eight stop), Illahraun (seventh stop), and Arnarsetursharaun (seventh stop). These fires took place between the years 1210 and 1240, that is, about, 800 years ago. The Yngri Stampar lava flow issued from a volcanic fissure, a crater row with many crater cones, which has also an exposed feeder-dike, all of which we will see in a moment. The lava closest to you in Fig. 13.31 is a part of a narrow stream, between Baejarfell (with the lighthouse) and Valahnukur (where you stand). The lava flowed down to the south until it met with the normal fault, the eastern boundary fault of the main graben in this area (Fig. 13.29). The lava flow may have reached the sea, but it cannot be determined because at the coast it is covered with gravel.

Turning now to the northwest, we can see the much of the Yngri Stampar lava flow, as well as most of its volcanic fissure with numerous large and small crater cones (Fig. 13.32). The fissure and its feeder-dike extend into the sea, so that the total length is unknown, but the on-land the length is about **4.8 km**. One of the largest crater cones is seen in Fig. 13.32b, and I show you a close-up of that one in a moment. Two dikes related to this eruption are well-exposed in the sea-cliffs seen in Fig. 13.32c, one of which is the feeder-dike. Since feeder-dikes are rarely exposed, and since this one is only about 800 years old and clearly connected to its volcanic fissure, it is worthwhile to take a walk to the cliffs to observe the dikes. There is a path that can be followed most of the way. During that walk the artic tern (a small bird) may be quite aggressive defending its territory if you happen to be there during the nesting time. Otherwise the path is easy, but the walk down to the coast should be done with care, and there are commonly large boulders on the coast that may make walking there somewhat difficult. The coast is continuously changing due to the sea waves, so that no hard-and-fast rules can be given as to how easy it is to get to the dikes. Walking on the coast itself always requires great care.

On the way to the cliff section with the dikes, we come closer to one the largest **crater cones** of the Yngri Stampar crater row (Fig. 13.33). This is a typical **spatter-and-scoria cone**, such as are commonly associated with most volcanic fissures or crater rows in Iceland (Fig. 13.25). You may recall from Chap. 9 that **scoria** constitutes broken, stiff lava fragments, mostly from millimetres to centimetres in size (diameter), derived from the spray of the lava fountains. Scoria

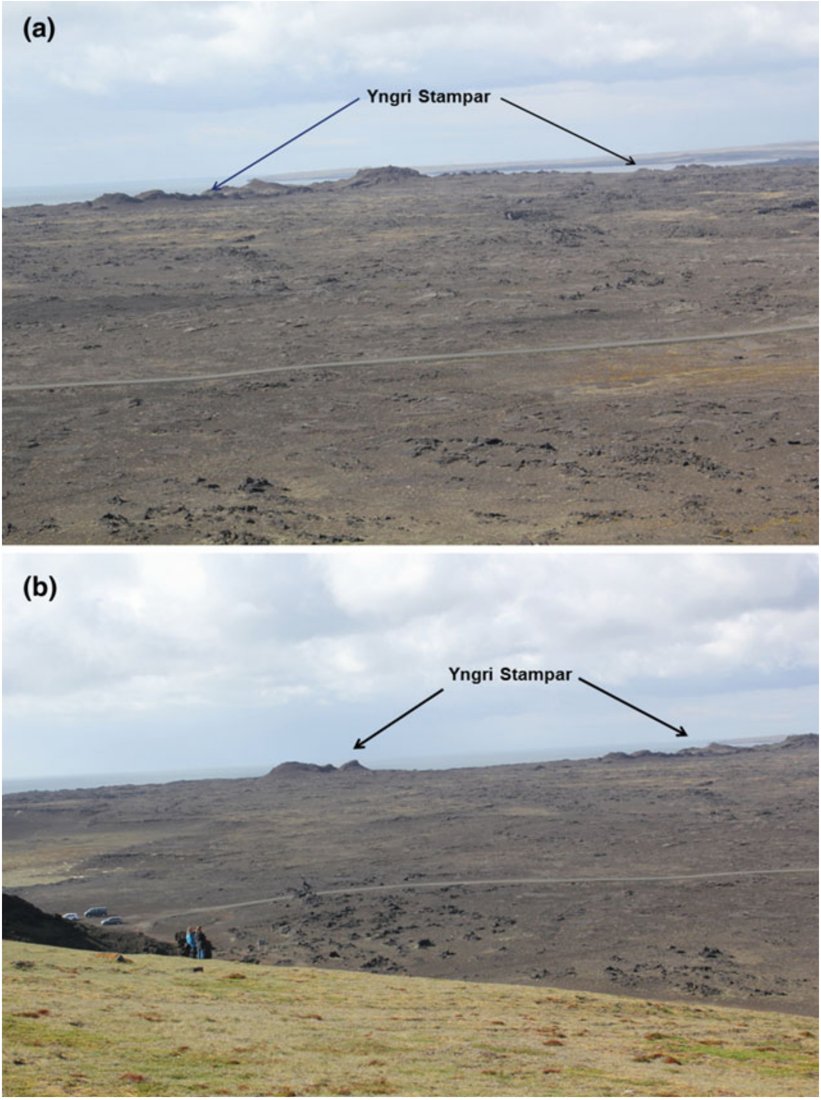

Fig. 13.32 View northwest showing the volcanic fissure of the Yngri Stampar eruption which happened some 800 years ago. Panorama showing **a** the middle and northern crater cones, **b** the southern crater cones, and **c** the southern end exposed in sea cliffs with both the feeder-dike to the eruption as well as a non-feeder, that is, an arrested dike (indicated)

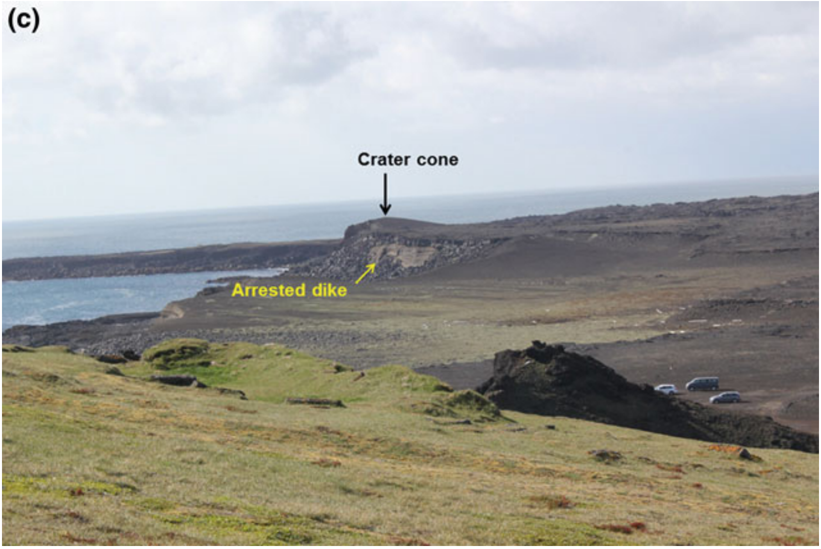

Fig. 13.32 (continued)

fragments are basically solidified and stiff when they hit the ground. By contrast, when the **spatter** clots hit the ground, they are still hot and plastic. We see here both black and red scoria. The black colour is the normal colour while red is due to oxidation, that is, the process of oxygen in the atmosphere combining with iron in the rock. Close to you in Fig. 13.33 is the **Yngri Stampar lava flow**. As discussed at the eighth stop, the reason that crater cones like those on the Yngri Stampar volcanic fissure form, including the one in front of you, relate to **flow channelling** which, in turn, depends on the cubic law that controls fluid transport in fractures (here a feeder-dike). The process of channelling the flow of magma to the large-aperture parts of the volcanic fissure results in the formation of crater cones.

While standing here close to the lava edge, and before descending to the coast and the dikes, it may be worthwhile to face the southwest and take a look at the rock pillar or sea stack of **Karl** (meaning a man). The rock pillar Karl (Fig. 13.34) is a part of **crater cone**, now mostly eroded away by the sea, that formed during the same Reykjanes Fires as the Yngri Stampar volcanic fissure, that is, some 800 years ago. The pillar, about 50 m tall, is just what remains of the island, the

Fig. 13.33 View northwest on the path to the dikes, one of the larger crater cones of the Yngri Stampar eruption. A typical spatter-and-scoria cone. Also seen, next to the photographer, is a vertical section through part of the Yngri Stampar lava flow

crater cone, that formed in a submarine eruption. In fact, the beginning of the Reykjanes Fires is supposed to have been in the sea, including the formation of crater cone of Karl. It is possible that the lower part of the lava flow with which the dikes discussed in a moment meet was erupted from the crater cone of Karl.

13.14 Details of 800-Year-Old Feeder and Non-feeder Dikes

Descending to the coast and walking along the sea cliff to the northwest, we first meet a **non-feeder** dike, that is, a dike segment that never reached the surface to supply magma to a fissure eruption. This follows because the dike ends vertically, its top or tip becomes **arrested**, on meeting a lava flow (Fig. 13.35a, b). This lava flow, at least its lower part (it is divided in two parts), may have been erupted from the crater cone of Karl early in the Reykjanes Fires. But the entire lava field is referred to as **Yngri Stampar**, and that is the name we use here for the lava flow as seen in the sea cliffs.

The general direction or trend or strike of the dike is 25° east of north, that is, north-northeast, and thus somewhat different from the general trend of the volcanic fissure of Yngri Stampar which has a more easterly trend (37° east of north). The trends of dike segments and fissure segments, however, vary. For example, the uppermost part of the dike trends 46° east of north. The dike is of basalt and with numerous crystals or minerals of plagioclase—white spots—as well as vesicles, that is, holes from escaping gas (Fig. 13.35c), as you saw in the lava flow in the walls of Almannagja (Chap. 5) and some dikes in Hvalfjördur (Chap. 11). The lowermost visible exposure of the dike depends on when you are here—the action of the sea is continuously changing the coast and how much of the dike can be seen. Normally, however, the lowermost part of the dike that is seen is about 35 cm thick, and from there it gradually thins to a few centimetres where it ends

Fig. 13.34 View southwest, the rock pillar or sea stack of Karl, the remnants of an eroded crater cone formed at the same time as the Yngri Stampar eruption

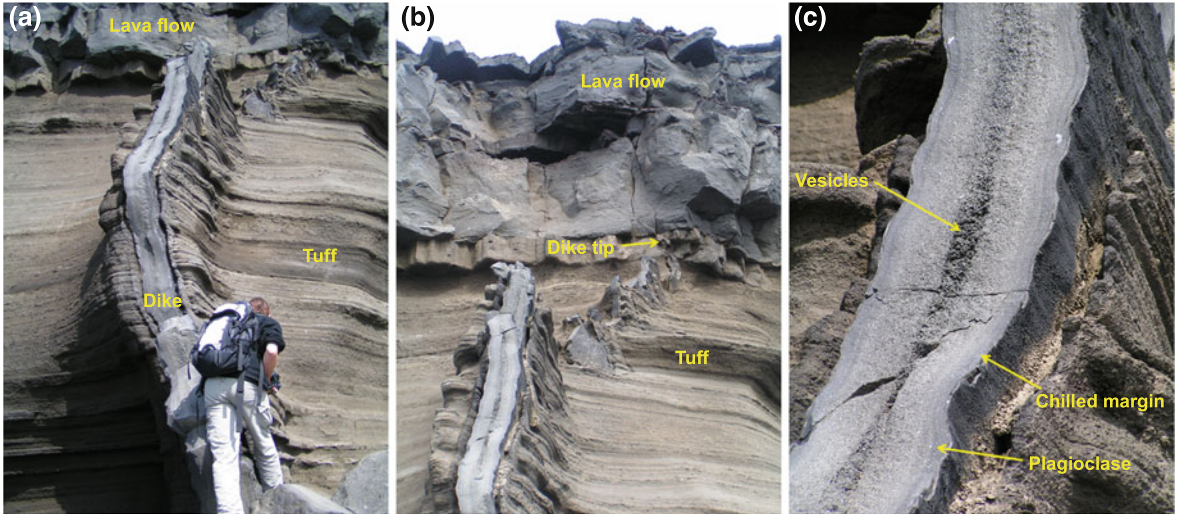

Fig. 13.35 View northeast, the non-feeder arrested at 5 m below the surface of the Yngri Stampar lava flow. **a** The dike becomes arrested, stops its vertical propagation, on meeting the contact between the soft tuff and the stiff basaltic lava flow. **b** Close-up of the tip of the dike where it becomes arrested. Also seen is the entire thickness of the lava flow, showing that the dike became arrested, most likely some 800 years ago, just below the surface of the volcanic zone. No normal faults or grabens occur ahead of the arrested tip. **c** Close-up of the dike rock. The rock is basalt with numerous vesicles, whose frequency is highest in the centre of the dike, as well as plagioclase phenocrysts. Also indicated is the chilled margin or selvage, that is, the darker very fine-grained to glassy margin formed during very rapid cooling as the hot magma (1100–1200 °C) came into contact with the surrounding host rock, the layered tuff

vertically, becomes **arrested**, on meeting the **Yngri Stampar lava flow**. The dike cuts through the **layered tuff**—a so-called lapilli tuff (Fig. 13.35c), but does not enter or penetrate the lava flow. Since the dike ends vertically on meeting the flow, it is clear that this dike, or dike segment, is not a feeder, but rather a **non-feeder** that became arrested just about 5 m below the surface some 800 years ago.

The second dike, about 25 m northwest of this one, is a clear **feeder-dike**. In contrast to the non-feeder, this dike reaches to about the **centre** of the Yngri Stampar lava flow, which is here about 8 m thick (Fig. 13.36a). This dike is also about twice as thick as the non-feeder, with a maximum thickness of some 70 cm. Its thickness changes not much until it enters the lower part of the lava flow, where the thickness abruptly decreases from about 50 cm to about 30 cm. This is exactly where the dike front leaves the **soft** or compliant tuff and enters the much **stiffer** lower part of the lava flow. The dike is then clearly connected with the lava itself in the central part of the layered lava flow. It is possible that the lower part of the lava flow was erupted from the crater cone of Karl, and then the upper part from this feeder-dike, which is almost certainly the feeder to the main part of the Yngri-Stampar volcanic fissure (Fig. 13.32).

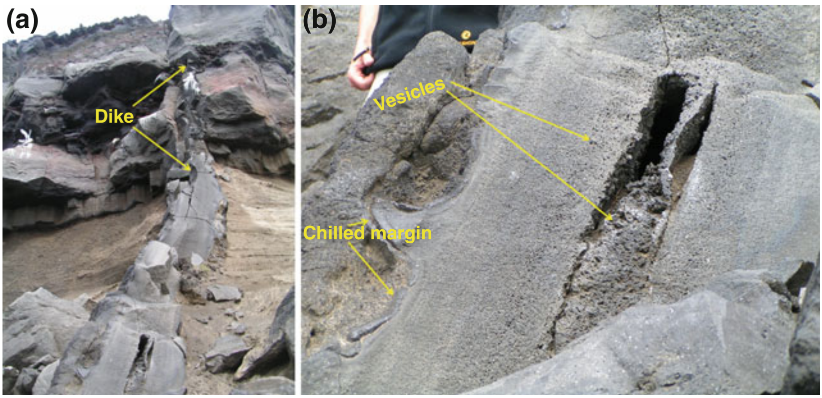

Fig. 13.36 View northeast, the feeder-dike to the Yngri Stampar volcanic fissure. **a** The feeder-dike enters the lower part of the lava flow and clearly feed at least the upper part of the flow. **b** Close-up of part of the feeder-dike showing wavy chilled margin and zones of vesicles, with particularly frequent and large vesicles in the central zone

While this dike is similar to the non-feeder in its internal structure, or texture, there are also certain clear differences. Similarities include the general rock type, glassy (chilled) margins—indicating that the host rock (the tuff) was comparatively cool when the dike-magma entered—numerous plagioclase crystals (phenocrysts) and vesicles (holes from escaping gas). In both dikes the vesicles occur in **bands**, parallel with the dike margins, and are most common in the dike centre. Differences between the dikes include, first, that the **vesicles** are more common and, in particular, much larger in the feeder-dike (Fig. 13.36b) than in the non-feeder dike (Fig. 13.35c). Second, the feeder-dike is much **thicker** than the non-feeder. Third, the non-feeder **thins** rapidly towards the bottom of the lava flow, by a factor of 3–4. By contrast, the feeder-dike reduces its thickness only slightly towards the bottom of the lava flow, that is, by about one-third or a factor of 1/3. We should keep these differences in mind when we discuss the dike formation below.

We have often discussed the relationship between earthquake fractures, particularly normal faults and grabens, and dikes. In view of these two dikes here, it is of interest that there are **neither grabens** nor **normal faults** associated with the dikes. The non-feeder clearly had essentially no magmatic pressure where it became arrested, and the feeder-dike was not able to generate or reactivate any faults. In fact, there is no graben seen to be associated with the Yngri-Stampar volcanic fissure. Part of the Yngri-Stampar lava flow seen in the cliff section (Figs. 13.35 and 13.36) had clearly formed when the non-feeder met it, or else the lava flow could not arrest the dike. Since the Reykjanes Fires, of which the Yngri-Stampar lava flow is a part, may have lasted many years—during the period from 1210 to 1240—there was plenty of time for the lava flow to solidify and become stiff before the non-feeder met it.

We have now spent considerable time and effort on describing these dikes, so you might ask in which way they are so important. There are several reasons for their importance. First, it is very rare to see a clear **feeder-dike connected to its lava flow** (Fig. 13.36). Second, it is even rarer to see side-by-side a feeder-dike and a non-feeder. Third, neither dike induces a graben or normal faults at the surface, an observation of great importance when trying to understand dike movement or propagation, with implication for **eruption forecasting**. And, fourth, there are few if any other dikes in the world known to be arrested only 5 m below the surface (Fig. 13.35). This observation has great implication for understanding and forecasting eruptions. It means that even if a dike reaches very close to the surface, there is no certainty that it will break through to the actual surface and erupt. The properties of the rocks, particularly the contrast in mechanical properties such as stiffness, and the local forces or stresses in the crust, even at very shallow depths, can result in dike arrest.

Which brings us to the question: Why did one of the dikes reach the surface and erupt, whereas the other one did not? The answer to this question, when generalised, is of fundamental importance in reliable **forecasting of volcanic eruptions**, so I shall here explain what seem to me the main reasons for the success of one and failure of the other dike in reaching the surface. The reasons may be listed as follows:

- The magma in the feeder-dike has many more and, in particular, larger vesicles. This means that it was more gas-rich, and thus presumably with a lower density (lighter) than the magma in the non-feeder. Light magma is **buoyant**, which normally means greater **driving pressure** or overpressure (pressure to rupture the rock and advance the dike front) than denser or heavier magma.
- The feeder-dike is **twice as thick** as the non-feeder at their deepest exposure, and 10–20 times as thick on meeting the bottom of the Yngri-Stampar lava flow.
- The plagioclase crystals (phenocrysts) are less common in the feeder-dike than in the non-feeder. When the proportion of crystals in magma increases, the magma **viscosity** normally increases so that ease of flow decreases. Based on the available data, the viscosity of the magma in the feeder-dike was presumably somewhat lower than that of the magma in the non-feeder.
- The feeder-dike is presumably older and cuts through the lower half of the lava flow (Fig. 13.36). The driving pressure or overpressure of the feeder thus generated horizontal **compressive stress** in this lower part which may have contributed to the arrest of the (assumed later-formed) non-feeder.
- All these factors indicate that the driving pressure was greater, and the viscosity somewhat less, in the feeder-dike than in the non-feeder, which may, partly at least, explain why the feeder was able to reach the surface and erupt, whereas the non-feeder became arrested.

Before we leave these cliffs and our discussion of feeder-dikes and non-feeders, we might ask the question: are these really two separate dikes, or just an offset part of the **same dike**? In standard methods of dike and fracture studies, these would be regarded as **two dikes**. That is simply because they are separated by some 25 m. It is possible, however, that they join at some depth in the crust, perhaps at a depth of several hundred metres, even kilometres, and then represent two segments or ‘fingers’ of the same dike (Fig. 13.26). Then the dikes seen in the sea cliffs could be **overlapping segments**. That idea, while possible, is perhaps made less plausible by the fact that the lava flow already existed as solidified cool rock body when the non-feeder met and became arrested at the lava-tuff contact. In any case it is clear that one dike/dike segment is a non-feeder that ends vertically at 5 m below the surface after propagating vertically from depths of many kilometres—most likely from a depth of some ten kilometres. That a dike/dike segment became arrested at such a shallow depth after propagating for a long vertical distance is, indeed, a geologically **very rare**—perhaps unique—event.

13.15 Bridge Between Two Continents

We now walk back to the car park, drive to Road 425, and follow that road to the north. On the way we cross the Yngri-Stampar volcanic fissure and see many crater cones, some primarily spatter cones, two of which are known as **the Stampar** (from which the fissure derives its name). We follow the road through several young lava flows until we come to the final main stop in this excursion, namely the **tenth stop (10)**, a parking place to the east of the road from which you can walk to the so-called **Bridge Between Two Continents**. The bridge is across a tectonic fracture, more specifically a pure **tension fracture** (Fig. 13.37). The fracture is formed by the same plate-tectonic tensile forces or stresses as the tension fractures at Thingvellir (Figs. 5.11–5.13; Chap. 5).

However, the present fracture is not filled with ground water as are the fractures at Thingvellir, but rather by sand. The direction or trend of this tension fracture is northeast and the maximum opening or aperture as great as **30 m** (the opening is around 15 m where the bridge itself is). The opening of 30 m is among the largest on any tension fracture in Iceland. Such large fractures are in some ways like **narrow grabens**, as we discussed in connection with Almannagja, the western boundary fault of the Thingvellir Graben (Chap. 5). Like most other large tension fractures in Iceland, this one is located in a pahoehoe lava flow which, in this case, derives from the lava shield **Langholl (Langhóll)**. The structure of the lava flows, as well as the flow units and **tumuli** in cross-section, are seen in the walls of the tension fracture (Fig. 13.38). You may recall seeing lava tumuli at the surface of some of the pahoehoe flows on your way from Keflavik to Reykjavik (Fig. 2.8; Chap. 2).

Fig. 13.37 Tension fracture seen from the 'Bridge between two continents'. View southwest the maximum opening or aperture of the tension fracture is about 30 m, but about 15 m where the bridge itself crosses the fracture. This is a pure tension fracture, as indicated by the orange arrows and as can be seen in that the fracture walls on either side of the opening are at the same elevation. The person on the left (east) fracture wall provides a scale

Does the bridged tension fracture really mark the separation between two continents? Hardly, because Iceland is not a continent or a part of a continent. The tension fracture is, however, without doubt a part of the on-land continuation of the Reykjanes Ridge which is a **boundary** between **two tectonic plates**: namely, the Eurasian Plate and the North-American Plate. If you want to mark that plate boundary by one fracture, then this one is a good as any other. In reality, however, the bridged tension fracture is part of a much larger set of tension fractures and normal faults which together constitute the northwestern edge of the main **5–6 km wide graben** that marks the on-land continuation of the mid-ocean Reykjanes Ridge. You saw the southeastern boundary fault of that graben from Valahnukur (Fig. 13.29), and the fracture set here may be regarded as the northwestern boundary of that graben. Part of the main fault, the normal fault that actually marks the boundary of the graben, is seen in a close-up in Fig. 13.39, which also shows a large **tumulus** and flow units of the lava flow from the Langholl lava shield.

How deep into the crust does the fracture in Fig. 13.37 extend? Using the theory explained in Chap. 5, the maximum depth of this tension fracture is **several hundred metres**. If the fracture tried to extend to greater depths it would change into a normal fault such as the one in Fig. 13.39. The stresses forming the tension fracture in Fig. 13.37 are not large, for the simple reason that rocks are very weak in tension. The tensile strength of pahoehoe lava flows such as the one in Figs. 13.38 and 13.39 is a few mega-pascal, and that is the tensile stress responsible for their formation. You may recall that one mega-pascal corresponds to the pressure or stress your body would feel at the depth of about 100 m in a lake or the sea (Chap. 5).

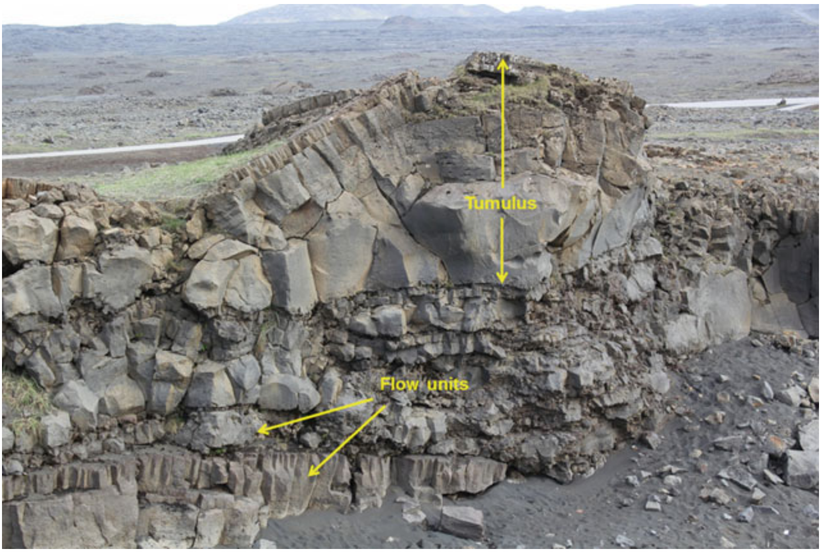

Fig. 13.38 View south of a tumulus and pahoehoe flow units seen in a vertical cross-section in the southeastern wall of the tension fracture in Fig. 13.37. Compare the flow units in the walls of Almannagja (Figs. 5.5 and 5.6) and the tumulus, seen at the surface, in Fig. 2.8 and in a cross-section in Fig. 13.39

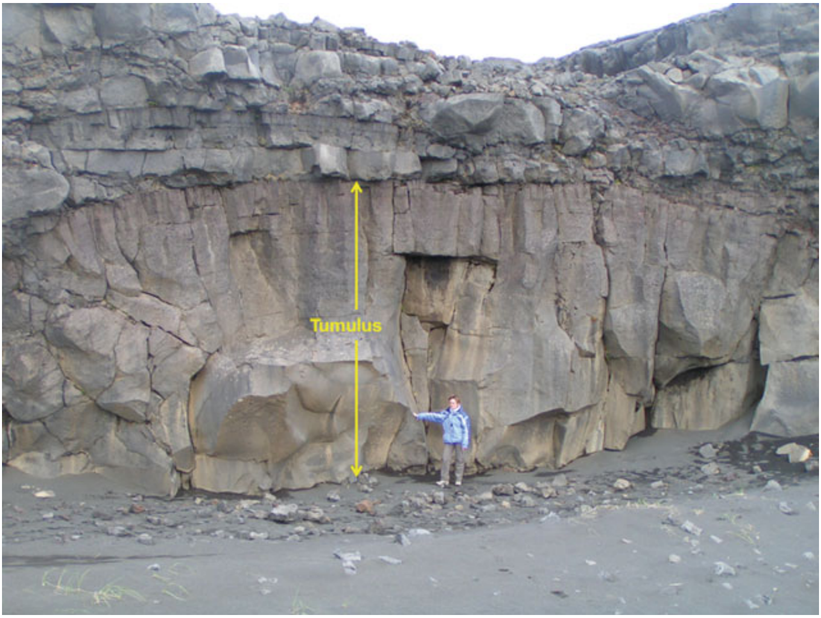

Fig. 13.39 Main northwestern boundary fault of the 5–6 km wide graben on Reykjanes, an on-land continuation of the mid-ocean Reykjanes Ridge (the southeastern boundary fault of the graben is seen in Fig. 13.29). Also seen is a vertical cross-section through flow units and a large tumulus (compare Figs. 2.8 and 13.38). The person provides a scale

This is our last formal geological stop in this excursion. We now return to Road 425 and drive first north along the edge of the Reykjanes Peninsula, to the village of **Hafnir** and from there along Road 44 to the northeast until it meets Road 41 close to the town of **Njardvik (Njarðvík)**. We are then on the same road as on the first day, namely the road from the Keflavik Airport to Reykjavik. All the main geological features along that road are already described and discussed in Chap. 2.

14 Reykjavik-Eyjafjallajökull-Reynisfjara

This last excursion takes us through some of the recently active areas of Iceland, both in terms of volcanic eruptions and earthquakes. In particular, the South Iceland Seismic Zone, already mentioned in Chaps. 9 and 13, and the famous volcanoes Hekla (erupted in 1991 and 2000) and the majestic Eyjafjallajökull (erupted in 2010), will be discussed. In addition, there are the beautiful waterfalls of Seljalandsfoss and Skogarfoss, as well as the coast of Reynisfjara, famous for its impressive rock columns and offering a view of the nearby point of Dyrholaey, also a popular touristic site (Fig. 14.1).

We follow Road 1 to South Iceland. We first drive along areas already described and discussed in Chap. 9. Thus, in case you would like to take a better look at some of the lava flows and mountains on the way, they are all described in Chap. 9. There are also several towns on the way. These include Hveragerdi (**Hveragerði**), with its greenhouses, and **Selfoss** at the river of **Ölfusa (Ölfusá)** which is a continuation of the glacier river **Hvita (Hvítá**, White River), in which the waterfall Gullfoss is located (Chap. 8) and the spring-fed river Sogid (**Sogið**), which originates in Lake Thingvallavatn (Chap. 5). Since we have already discussed all the main geological interests on the way to the cross-road between Road 35 and Road 1 (Fig. 4.1, Chap. 9), we may regard the real excursion here as starting at that cross-road, as we do in Fig. 14.1.

14.1 Largest Lava Flow Erupted on Earth in the Past 10 Thousand Years

Our **first stop (1)** is on north bank of the river **Ölfusa** (Fig. 14.2). There are several reasons for this stop. The first one is that, in terms of discharge, Ölfusa is the **largest river** in Iceland. Its average discharge on entering the sea is about

A. Gudmundsson, *The Glorious Geology of Iceland's Golden Circle*, GeoGuide, DOI 10.1007/978-3-319-55152-4_14

Fig. 14.1 Map showing the excursion to Eyjafjallajökull and Reynisfjara. The suggested main stops are indicated by encircled numbers, 1–10. The names of some of the landmarks and sites discussed in this excursion are indicated here, but many more in Fig. 14.12b. Figure 14.18 shows the southernmost part in a comparatively large scale and without names

Fig. 14.2 View south across the river Ölfusa, the largest river in Iceland in terms of discharge. Its south bank seen here is composed of the Thorsarhraun (Þjórsárhraun) lava flow, the largest lava flow erupted on Earth in the past 10–12 thousand years

420 $m^3 s^{-1}$ (cubic metres per second). To put this discharge into context, it is about 6-times the discharge of Thames in England and around 1/6 of the discharge of the river Rhine in Western Europe. Ölfusa itself is only about 25 km long,

because inland it soon splits into the other two main rivers, Hvita and Sogid, the latter being the largest spring-fed (groundwater-fed) river in Iceland.

The second reason for the stop is that the cliffs on the south and east banks of Ölfusa constitute part of the largest lava flow known to have erupted on Earth during the past 10–12 thousand years (Fig. 14.2). This is the **Thorsarhraun (Þjórsárhraun)** lava which was issued by a volcanic fissure somewhere in the East Volcanic Zone 8600 years ago. The exact location of its origin, the feeding volcanic fissure, is unknown because the fissure is covered by younger lava flows in that zone. The Thorsarhraun lava flow covers close to 1000 km^2, that is, about **1%** of the total area of Iceland, is more than 20 m thick, and has an estimated volume of some **25 km^3** (cubic kilometres). For comparison, typical eruptive volumes in the volcanoes Hekla and Eyjafjallajökull, both of which we discuss later today, are about one percent or 1/100th of the Thjorsarhraun lava.

The largest rivers in Iceland, in terms of discharge, mark the boundaries of the Thorsarhraun lava. In the west the river Ölfusa, as mentioned above, and in the east it is **Thjorsa (Þjórsá)**, the **longest river** in Iceland, 230 km, and second in terms of discharge (about 350 $m^3\ s^{-1}$, cubic metres per second). It is from Thjorsa that the lava flow derives its name, the river following the eastern and southern margin of the lava for tens of kilometres.

Another interesting aspect of this lava flow is that when it flowed all the way down to the coast (the foundation of the villages Eyrarbakki og Stokkseyri, south of Selfoss, is also Thorsarhraun), **sea level** was as much as **15 m lower** than today. As a consequence, the Thjorsarhraun lava extends as an ordinary lava (not as a pillow lava) to about 1000 m south of the present coast. So why was the sea level lower 8600 years ago than it is today? The answer is primarily the slow melting of the world's main ice sheets at the end of the last glacial period. Melting of the ice sheet in Iceland was completed some 9000 years ago, but some of the large ice sheets in North America and Scandinavia did not completely melt away until 6000 years ago. Thus at the time of the eruption of the Thorsarhraun lava much water was still frozen in the continental ice sheets and therefore not in the sea. Consequently, the sea level stayed lower than it is today until these ice sheets had melted and their water entered the sea. It follows that 8600 years ago, at the time of the eruption of the Thorsarhraun lava, the sea level was still much lower than today.

We now drive along Road 1 on from Ölfusa and through the town of Selfoss and to the east. As we do so, we are part of the way on the Thorsarhraun lava flow. You notice it on either side of the road in the surface being rough and contrasting with the more smooth grass fields further to the east where we approach the river Thjorsa. We are now entering the most active part of the South Iceland Seismic Zone. Unfortunately, the earthquake fractures are very poorly seen in the

Thjorsarhraun lava and other lava flows here, but it is worth stopping at the parking place where Road 1 and Road 30 meet and consider some of the photographs of the earthquakes in the year 2000.

14.2 Earthquake Fractures from the Year 2000

The **second stop (2)** is thus at the parking place (north of the road) at the cross-road between Road 1 and Road 30. This is exactly where one of the two large earthquake fractures ruptured the ground in June 2000, each fracture generating an earthquake of magnitude 6.6. In the north you see close to you two mountains, both located in Fig. 4.1. One is **Hestfjall**, which was dissected by the western earthquake fault in 2000. The other is **Vördufell (Vörðufell)** which has many faults from earlier earthquakes in the area (Figs. 9.2 and 9.3), but was not ruptured in the June 2000 earthquakes.

The main earthquake faults associated with the June 2000 earthquake in this area were 20–25 km long and reached depths in the crust of 8–10 km. The faults were close to vertical and of the type referred to as **strike-slip faults**. Then the main movement of the rock walls on either side of the fault is **horizontal**, in contrast with the main movement of the walls on either side of the normal faults, such as at Thingvellir (Chap. 5) and Reykjanes (Chap. 13) which is vertical (Figs. 4.5, 4.14 and 5.9). Strike-slip faults are very common—perhaps the most famous one is the San Andreas Fault in California in the United States—and their fracture pattern at the surface tends to be somewhat more complex than that of normal faults.

So what do strike-slip faults look like at the surface? I first show you what the faults formed in the June 2000 earthquakes looked like just after the earthquakes; you cannot see them as such today because they have long been filled with soil and covered. At the time there was a summerhouse southwest of the second stop (south of the road) which moved about 1 m horizontally in the earthquake. There were also numerous fractures in the asphalt of the car park where you stand now.

Figure 14.3 is an aerial view of a part of the earthquake fracture that hit the summerhouse. Figure 14.4 shows part of the same fracture on the ground. In both the earthquakes, the one on 21 June seen above and the earlier one, on 17 June (Iceland‘s National Day), which occurred some 18 km to the east of the present one, buildings were damaged (Fig. 14.5). While the earthquakes thus caused considerable damage, such as of buildings in the town of Hella through which we pass later today, fortunately no people died in these two earthquakes. Earthquakes with magnitude 6.6, or more specifically those with magnitude between 6.1 and

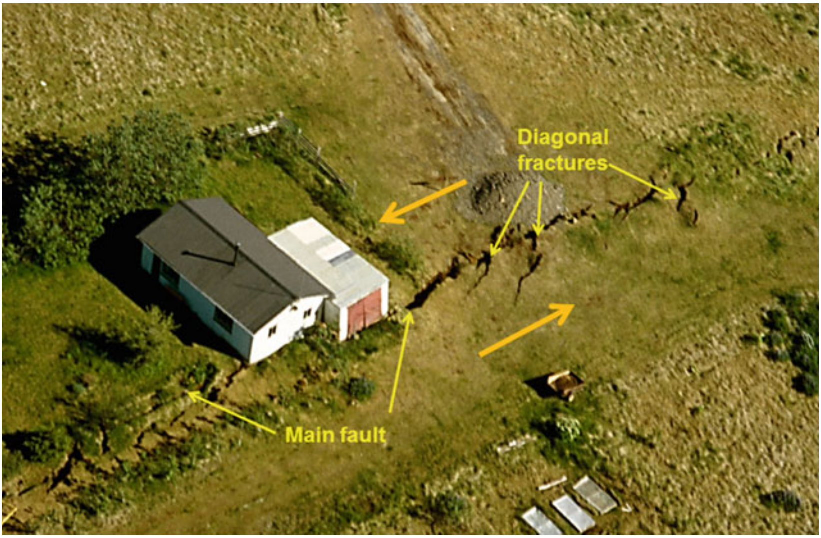

Fig. 14.3 Aerial view of one of the earthquake fractures formed in the magnitude 6.6 June 2000 earthquakes in the South Iceland Seismic zone. This is a strike-slip fault of type A. The same type of fault is seen in Figs. 8.3, 9.2, 13.7, 14.6 and 14.10

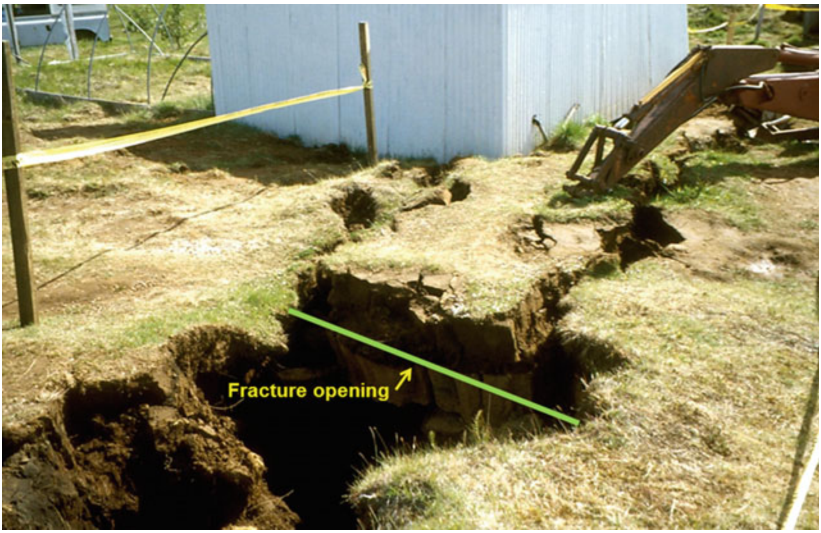

Fig. 14.4 Ground view of an open fracture generated during the June 2000 earthquakes

Fig. 14.5 Earthquake fracture formed in the June 2000 earthquakes running through a building at a farm in South Iceland. The upper part of the building has moved to right relative to the lower part. This fracture is a strike-slip fault of type C. The same type of fault is seen in Figs. 8.3, 9.2, 13.7, 14.6 and 14.10

6.9, are regarded as **strong** and may cause much damage in urban areas. They are rather common, with about 100 strong earthquakes every year worldwide.

Larger earthquakes do, however, occur in the South Iceland Seismic Zone, about once every century. These earthquakes reach magnitude over 7, possibly 7.2. Here I should mention that the energy radiated in an earthquake, and partly available for destruction of buildings and other human-made constructions, increases by a factor of about 32 for each added magnitude. In other words, moving up or down the earthquake-magnitude scale by one magnitude changes the radiated energy by a **factor of about 32** (strictly 31.6). Similarly, moving up the scale by two units, say from an earthquake of magnitude 6 to an earthquake of magnitude 8, the radiated energy increases 1000 times (31.6 times 31.6). Fortunately, earthquakes of magnitude 8 are rare—about one every year worldwide—and do not occur in Iceland. However, as said, earthquakes of magnitude in excess of 7 do occur in South Iceland, and I now show you photographs from one of these.

The associated earthquake fractures occur at the farm/hotel **Leirubakki**, which is located far inland close to Road 26, and thus outside the stops of this excursion. However, it is worthwhile to look at the earthquake fractures, because their surface features and sizes indicate what can be expected from time to time in South Iceland. The exposed or visible length of the main earthquake fracture is about 8 km, and it cuts through smooth basaltic lava flows formed sometime in the past 10 thousand years. The earthquake fracture must be younger than the lava flow that it dissects, and is thus younger than 10 thousand years—but the exact age is unknown. The direction of the main C-fault is 10° east of north, which is very common in the South Iceland Seismic Zone (Chap. 9).

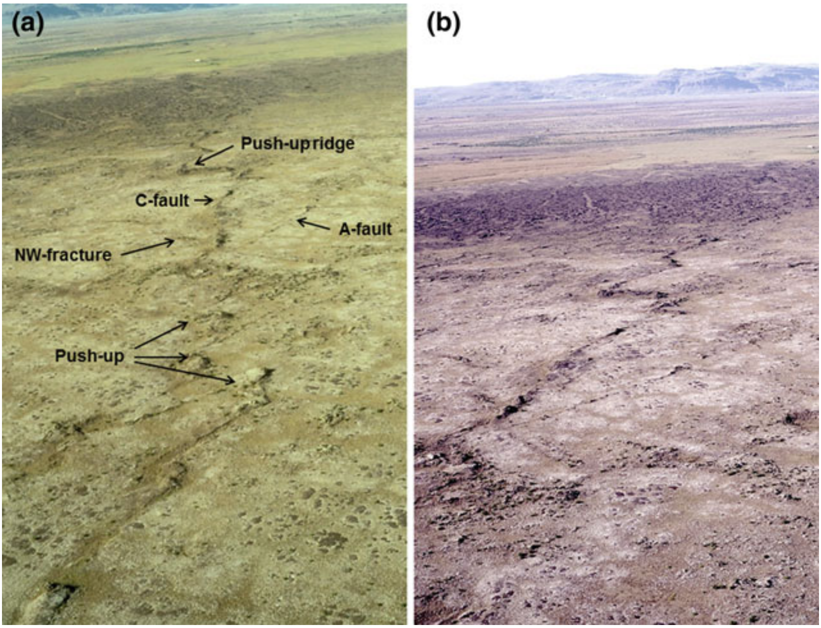

Fig. 14.6 Aerial view north of the main earthquake fractures at the farm/hotel Leirubakki in the South Iceland Seismic Zone. **a** Strike-slip faults are characterised, like this one, by hillocks and ridges referred to as push-ups. The main fault here is a north-trending C-fault, whereas northeast-trending A-faults as well as a NW-trending fractures occur also. **b** Close-up of the northern part of the fault seen in **a**. All the main structures are seen here: the C-fault segments, several A-faults, the NW-trending fracture, as well as the push-ups, including the push-up ridge seen in Fig. 14.7. All the fracture types seen here are also observed at Kleifarvatn (Fig. 13.7) which is located within an extension of the South Iceland Seismic Zone to the Reykjanes Peninsula

Fig. 14.7 View north, close-up of the push-up ridge seen in Fig. 14.6. The height of the push-up is 4.3 m. The push-up is formed by the compressive stressed generated during the fault slip. From this and other push-ups and structures along the fault (Fig. 14.6) it is possible to estimate both the associated fault displacement, about 3 m, and the earthquake magnitude, about 7.1

The surface features of the fault are typical of **strike-slip faults** (Fig. 14.6); you see similar features at the surface of the San Andreas Fault in the United States, the Anatolian Fault in Turkey, and other similar famous strike-slip faults. The main characteristics are **hillocks** and small **ridges**, together with various **fractures**. The hillocks and ridges are formed through compression and crushing of the rock along the fault when its walls move abruptly during the earthquake. One of the ridges/hillocks is indicated in Fig. 14.6, as seen from the air, and on the ground in Fig. 14.7. Observing these bent and broken-up layers of hard basaltic lava, we understand better the **great forces** and stresses that operate during earthquake rupture.

The bent layers can also be used to estimate the displacement, the relative surface movement of the walls, across the fault. The results indicate that movement was close to **3 m**, or much larger than that of the June 2000 earthquakes. From the displacement we can infer the earthquake magnitude as 7.1 and the total length of the earthquake fracture as about **50 km**. As regards energy, this earthquake released around 6-times as much energy as each of the earthquakes in June 2000.

South Iceland is one of the best **monitored** earthquake areas in the world. We know the rate of movements of the crustal plates in the area, we record all the earthquakes down to so small ones that humans cannot feel them—in fact, down to negative magnitudes. (Earthquake magnitudes are exponents, so that negative magnitudes simply mean very small earthquakes in a similar way as the number 10 to the power of −1 (here corresponding to the magnitude) is $10^{-1} = 0.1$.) There are numerous recording instruments and objects on the ground (seismometers, strainmeters, geodetic points), as well as measurements by satellites (GPS, InSAR) from above. Yet, in South Iceland as elsewhere, we **cannot forecast** earthquakes (their time, location, and magnitude) with any accuracy. Modern buildings in Iceland, however, are normally very strong; most are made of reinforced concrete. Thus, even in strong earthquakes, with magnitudes between 6 and 7, where the earthquake fracture passes through the building and causes considerable damage to it (Figs. 14.3, 14.4 and 14.5) the building rarely collapses. Fortunately, fatalities are thus rare in earthquakes in Iceland.

14.3 Hekla Volcano

We now continue to drive on along Road 1 to the east. The next stop depends on visibility and can be at several locations close to the town of Hella. Assuming that the visibility is good enough to see the most famous volcano in Iceland, we can stop either just after or before reaching the town of Hella, our **third stop (3)**. This stop is to see the volcano **Hekla** from a distance of some 40 km (Fig. 14.8). Hekla is in many ways a unique volcano. First, all the proper Hekla eruptions occur along the **same fissure** (Fig. 14.9) which has a length of about 6 km and is named the **Hekla Fissure**. This is in contrast with many central volcanoes where different eruptions occur on different volcanic fissures or feeder-dikes (Figs. 11.2, 11.3 and 11.4) or through an elliptical or circular conduit (Fig. 11.13). Second, and partly as a consequence of all the eruptions in Hekla itself being along the same fissure, the volcano has a **ridge shape** (Fig. 14.9), is a volcanic ridge, rather than the typical shape of a stratovolcano—which tends to be circular or somewhat elliptical in plan view. Third, despite intensive research, using geodetic and seismic measurements and monitoring over many decades, no magma chamber has been detected beneath Hekla. We know that there must be a magma chamber—for the simple reason that the time from the first earthquakes, indicating magma movement up along the Hekla Fissure or feeder-dike, until the eruption begins is very short; sometimes only about **half an hour** or so. Thus, the source for the magma, some sort of a

Fig. 14.8 View northeast, the Hekla Volcano as seen from Road 1 and located in Figs. 14.1 and 14.12b. Hekla is a central volcano of a rare shape, namely a ridge with the unusual trend of about 65° east of north. Part of the Hekla Fissure can be inferred from the white line of unmelted snow. It is the second most active volcano in Iceland in terms of eruption frequency

chamber, is likely to be at a shallow depth (of the order of kilometres or less), but it has not been detected so far.

Why does the feeder-dike always follow the same path to the surface, to the Hekla Fissure? The most likely explanation is that the Hekla Fissure is reopened again and again by external tectonic forces. More specifically, the Hekla Fissure has a very unusual direction, about **65**° east of north, whereas nearly all surrounding fractures in the East Volcanic Zone, including volcanic fissures and hyaloclastite ridges (old volcanic fissures), are directed 45° east of north (Fig. 14.1; see also Fig. 14.12b). There is thus a striking 20° difference in the direction of the Hekla Fissure and nearby fissures. Furthermore, the direction of the Hekla Fissure is similar to that of one of the main trends of earthquake fractures in South Iceland—and west to the Reykjanes Peninsula, namely that of the so-called **A-faults** (Figs. 8.2, 9.2, 13.7, 14.6, 14.10 and 14.11). It is thus likely that the Hekla Fissure is an **old earthquake fracture**, a strike-slip fault, that is reactivated again and again by plate-tectonic forces. Since this old earthquake fracture extends into a volcanic zone, where there is magma at depth, the magma uses it as a path to the surface. It follows that many dikes use the same fracture to reach the surface, thereby generating the unusual shape and eruptive activity of Hekla.

Hekla has erupted many ash (tephra, pyroclastic) layers as well as numerous lava flows, all of which are **aa flows** (Figs. 13.3, 13.4, 14.9 and 14.11). The volcano rises to about 1490 m above sea level and is thus easily seen from a distance of 40 km, so long as the visibility is good (Fig. 14.8). Roads 26 (already passed) and Road 271 take you to Hekla. To reach the volcano, however, is a considerable drive that is outside the present excursion. In fact, Hekla is in some ways best appreciated from the air; the close-up photographs of Hekla in Fig. 14.9 are taken from the air and show the snow-covered Hekla during its eruption in 1991. Since that time, there has been one more eruption, in 2000, and new

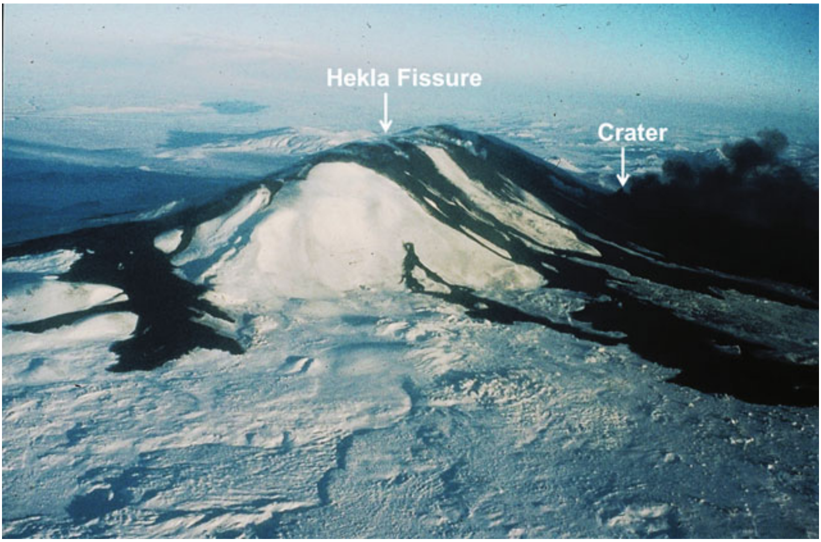

Fig. 14.9 Aerial view of snow-covered Hekla Volcano during the 1991 eruption. The ridge shape is clear on this photograph. Black lava issued during the eruption is in sharp contrast with the older, snow-covered lava flows. The Hekla Fissure is indicated and so is the crater cone that was active during the entire 53-day duration of the eruption. The erupted lava covered 23 km^2 and had an estimated volume of 0.15 km^3. During the peak of the eruption the volume of lava issued was about 800 m^3 s^{-1} (cubic metres per second), or about double the discharge of Ölfusa (Fig. 14.2)

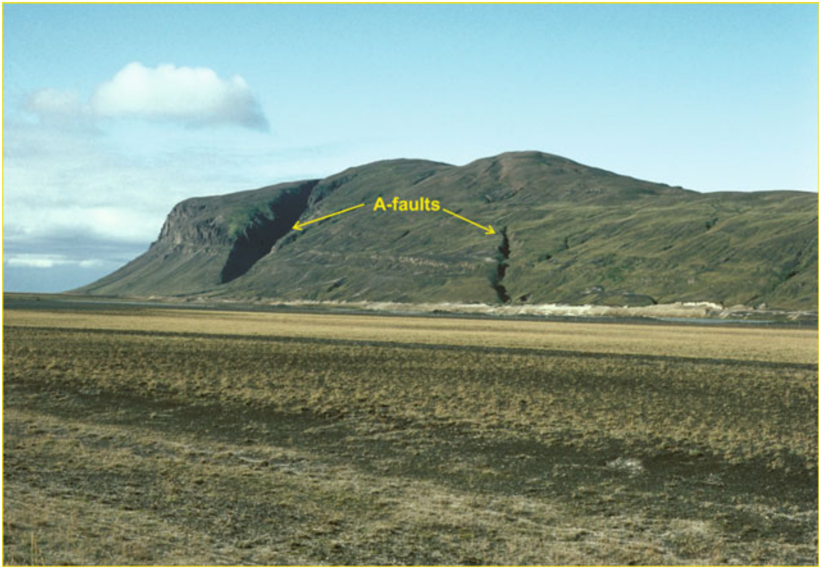

Fig. 14.10 View southwest of two east-northeast trending A-fault cutting through the hyaloclastite mountain Burfell (Búrfell), located close to the Hekla Volcano (both mountains are located in Fig. 14.12b). The Hekla Fissure has the same trend as these faults, 65° east of north, and is almost certainly a reactivated C-fault

eruptions can happen any time. Hekla is well-monitored with various instruments, but the time from the start of the unrest (usually indicated by earthquakes) to the beginning of the eruption is unusually short—tens of minutes—so that Hekla, the Queen of Icelandic volcanoes, should be approached with utmost respect and care.

From roughly this stop, good overview photographs can often be obtained of the famous **Eyjafjallajökull Volcano** (Fig. 14.12), to which we are heading. Sometimes, depending on visibility, it is better to take the overview photographs when you are closer to the volcano, that is, further to the east along Road 1. The photograph I show you was taken soon after the end of the eruption in May 2010. The gently inclined column is, like an eruption column, primarily the effect of heating of the atmosphere. Here the heat comes from the cooling eruptive materials in the main crater of the volcano, the one that resulted in the airspace in parts of Western Europe being closed for many days because of ash (the fine ash particles may damage the jet engines of the aeroplanes, hence the closure).

Fig. 14.11 Typical aa lava flow, here seen flowing in the 1991 Hekla eruption. All the proper Hekla lava flows are of the aa type. The 1991 lava flow is mostly with a thickness of 6–7 m. The person provides a scale

14.4 Seljalandsfoss Waterfall

We describe the structure of Eyjafjallajökull and its 2010 eruption in a moment, but first we make our **fourth stop (4)** at one of the better-known waterfalls in Iceland, namely **Seljalandsfoss** (Figs. 14.13 and 14.14). To reach this waterfall, we first cross the bridge across the glacial river Markarfljot (**Markarfljót)**. The glacial water is mainly derived from the ice caps of Eyjafjallajökull and Myrdalsjökull (**Mýrdalsjökull)**, both of which we discuss further today (Fig. 14.12b). Markarfljot is normally small, such as many glacier rivers are, but can become large during floods. In particular, eruptions beneath these two glaciers can result in discharges of many thousand cubic metres per second.

A second feature to mention just as you approach the river Markarfljot is the archipelago of **Vestmannaeyjar** (Figs. 14.1 and 14.12b). The main island, **Heimaey**, can usually be seen when you are that close to the coast. You may also see part of the town of Vestmannaeyjar. If the visibility is very good you may, here or earlier on Road 1, see some of the other islands—including the southernmost one,

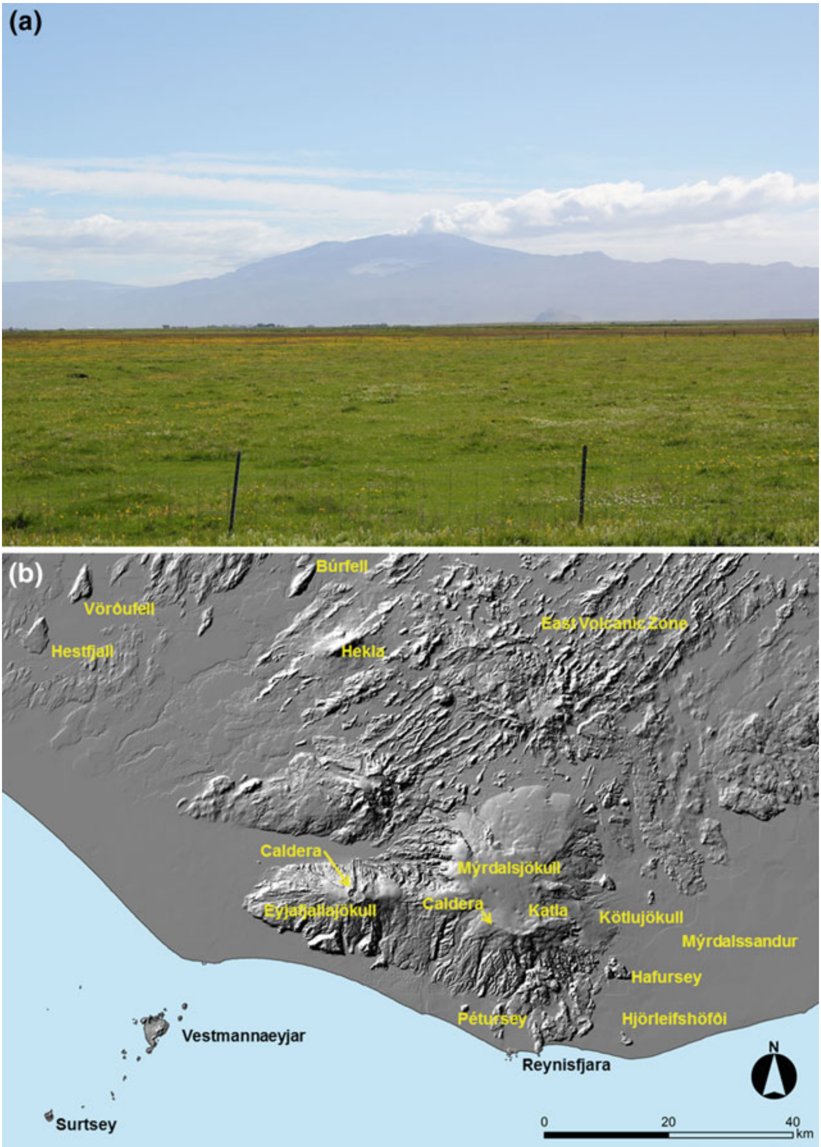

Fig. 14.12 **a** View east, the Eyjafjallajökull Volcano seen from a distance. This photograph is taken soon after the 2010 eruption. The inclined while column is due to heating from the eruptive materials in the caldera. The mist or haze is due to the ash in the air, carried by the wind when the photograph was taken. **b** Location of Eyjafjallajökull, showing its caldera, as well as the names and locations of other main mountains and landscape features discussed in this chapter

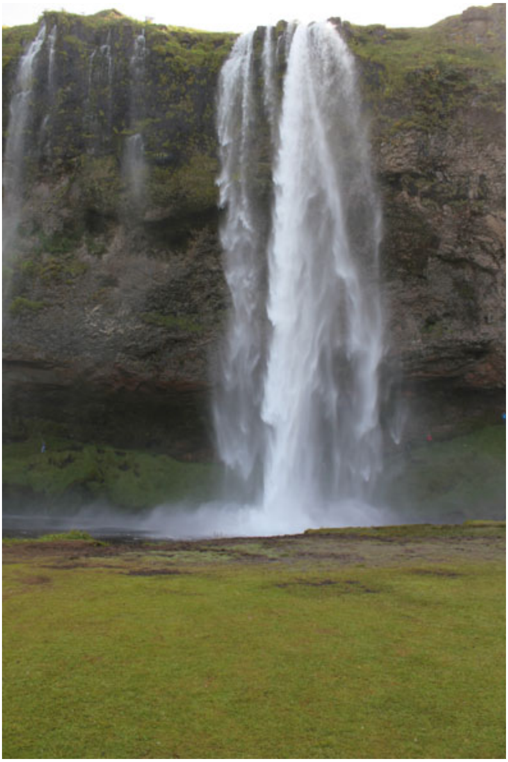

Fig. 14.13 View north, the waterfall Seljalandsfoss. The 60 m high cliff from which the water falls is an old sea cliff, formed by wave erosion when the sea level was much higher some 13 thousand years ago

namely the island of **Surtsey** (Fig. 14.12b), a table mountain formed in the sea (Fig. 6.6, Chaps. 6 and 13) in an eruption that lasted from 1963 to 1967.

Driving now to the fourth stop, the cliff that makes the Seljalandsfoss waterfall is an old **sea cliff**—here with a height of about 60 m. As indicated earlier (Chap. 13), at the end of the last ice period, some 13,000 years ago, the sea level in Iceland rose to some 100–150 m above the present sea level. The vertical 60-m-high cliffs, from which the water falls, are thus formed primarily by wave erosion when the sea level was, for a while, much higher than it is today. The same applies to most of the vertical cliffs that constitute the lower part of Eyjafjallajökull, and which we will see in a moment.

Fig. 14.14 Seljalandsfoss is in the river Seljalandsa (Seljalandsá), here seen on the ground in front of the waterfall

The rocks that constitute the cliffs, here as in much of the exposed parts of Eyjafjallajökull, are of hyaloclastite, that is, basaltic breccia, formed during eruptions under ice sheets—during the ice periods—and, possibly, also (the lowermost parts) in eruptions in the sea (see Chap. 13 for a discussion on eruptions in the sea). Seljalandsfoss is well-known for being a waterfall that can be viewed not only from the front, but also from the back, from a footpath **behind the waterfall**. Some people walking along that path are seen in Fig. 14.13. The water in the cascade is beautifully clear and clean and constitutes the small river Seljalandsa (**Seljalandsá**) (Fig. 14.14).

14.5 Eyjafjallajökull Volcano—Internal Structure and the 2010 Eruptions

From here we drive on along Road 1 to the east along the high cliffs of **Eyjafjallajökull** (Fig. 14.12b). The name means literally ´the ice cap of the island mountains', the islands being Vestmannaeyjar (Figs. 14.1 and 14.12b). If the visibility is good, the part of the mountain that is seen can be extraordinarily impressive and beautiful, in addition to being geologically of great interest. To explore and understand the interior of the volcano, and how it relates to its 2010 eruption, there are many places that you could stop at. The one I have chosen, the **fifth stop (5)**, is selected just because there the sea cliffs are very clean and clear and the structure can be easily observed and understood. I show photographs from this stop, and one additional one a short distance further east, but the photographs are all basically from the same large cliff exposure in the vicinity of the fifth stop.

What we see are major cliffs which range in height from about 200 m (Fig. 14.15) to a maximum of about **350 m** (Figs. 14.16 and 14.17). These cliffs, as indicated above, are primarily sea cliffs, even if the sea never reached to heights of more than about 100–150 m above the present sea level. As the action of the sea erodes the lower parts of the cliffs, the upper parts collapse, thereby keeping the entire cliff section close to vertical. These cliffs form the lowermost exposed part of the volcano which rises to a maximum elevation of 1666 m above sea level. The top part is the site of an ice cap, Eyjafjallajökull, and also of a small caldera with a diameter of about 3 km (Figs. 14.12c and 14.18).

Fig. 14.15 Vertical cliff, partly an old sea cliff formed where the sea level was much higher some 13 thousand years ago, in the southwestern side of Eyjafjallajökull (the mountain is located in Figs. 14.1 and 14.12b; see also Fig. 14.18). The vertical cliff is about 200 m high and primarily composed of hyaloclastite

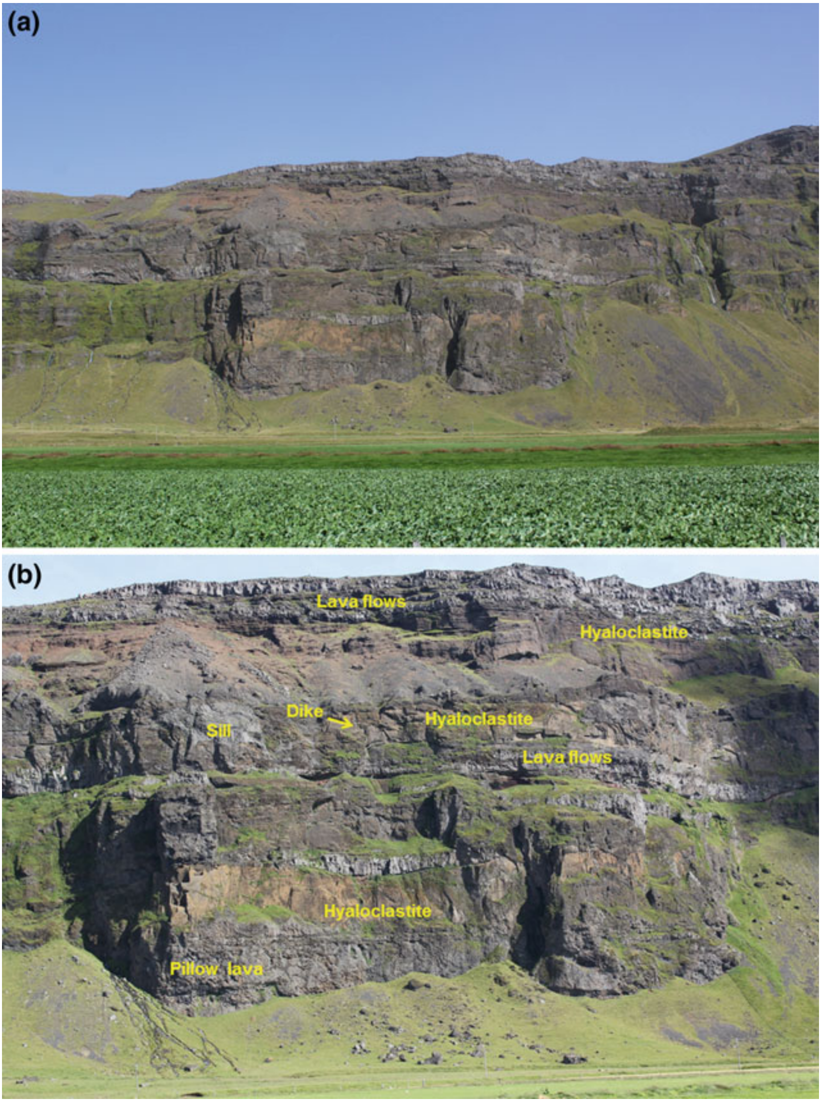

Fig. 14.16 Vertical cliffs showing the internal structure of part of Eyjafjallajökull (located in Figs. 14.1 and 14.12b) as seen from Road 1. **a** he cliff section seen here reaches height of about 350 m and is composed of a variety of rock layers and units. **b** Close-up of a part of the cliff-section in **a**. Eyjafjallajökull is primarily composed of hyaloclastite and lava flows, including pillow lava in the lowermost section. There are also many intrusions, primarily sills and dikes

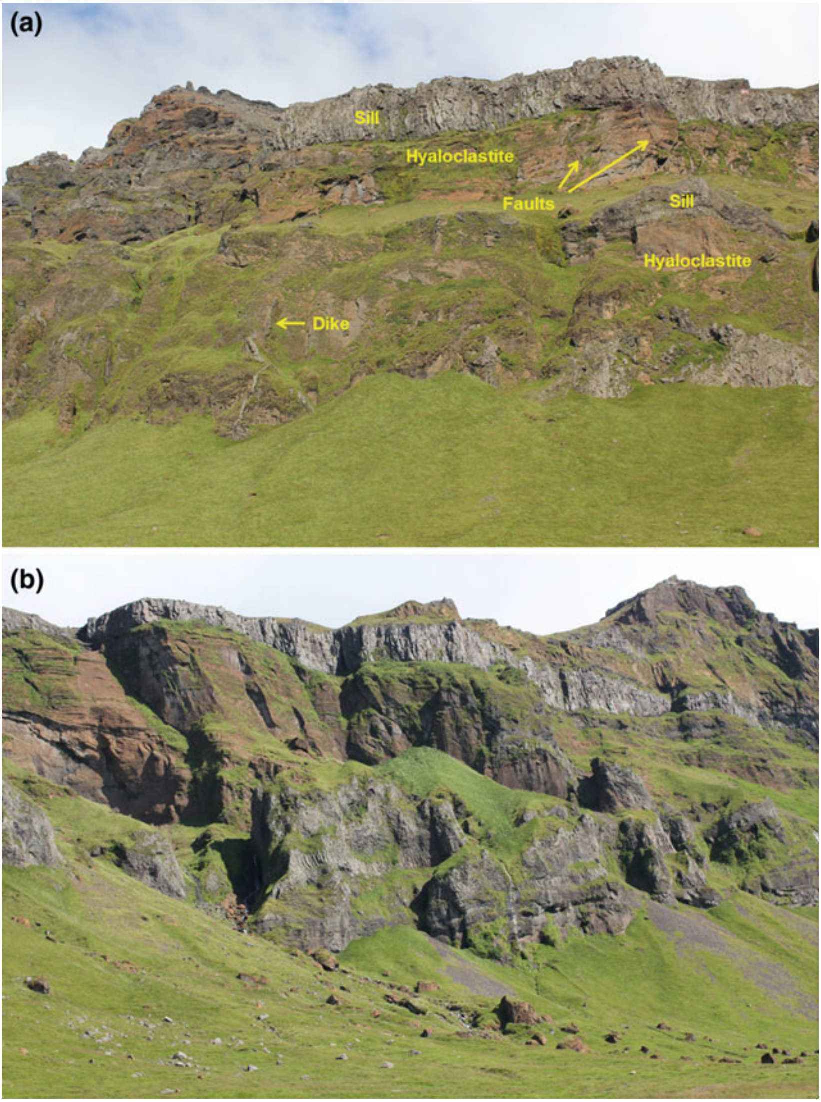

Fig. 14.17 More details on the internal structure of Eyjafjallajökull (located in Figs. 14.1 and 14.12b) as seen from Road 1. **a** The internal structure of Eyjafjallajökull, as seen here, is characterised by thick layers of hyaloclastite, together with lava flows (seen in Fig. 14.16). In addition, the volcano is intruded by many, mostly thin, dikes and, normally much thicker, sills. **b** Direct continuation (panorama), to the east, of the cliff section seen in **a**, showing the same rock layers and intrusions

Figures 14.15, 14.16 and 14.17 show us that the volcano is made of a variety of rock layers. These include hyaloclastite layers, lava flows, cube-jointed lavas (Icelandic: kubbaberg), pillow lavas, sedimentary layers (glacial sediments, tillites), sills, and dikes. Eyjafjallajökull is a volcano composed of strata (layers) of widely different properties, and is thus a **stratovolcano/composite volcano**. The volcano has been active for about **800 thousand years**, and is therefore comparatively old. Most central volcanoes (stratovolcanoes and calderas) in Iceland are active for about five hundred thousand to million years—only a few are active for longer than a million years.

Eyjafjallajökull erupts infrequently when compared with, for example, Hekla and Katla (Figs. 14.1 and 14.12b). In historical time in Iceland, that is, the past 1100 years, Hekla has erupted 23 times, Katla 21 times, but Eyjafjallajökull only 4 times. The most frequent eruptions, during historical time, however, have neither been in Hekla or Katla but rather in the caldera **Grimsvötn (Grímsvötn)**. Grimsvötn is located in the Vatnajökull ice sheet (located in Fig. 2.2), and thus

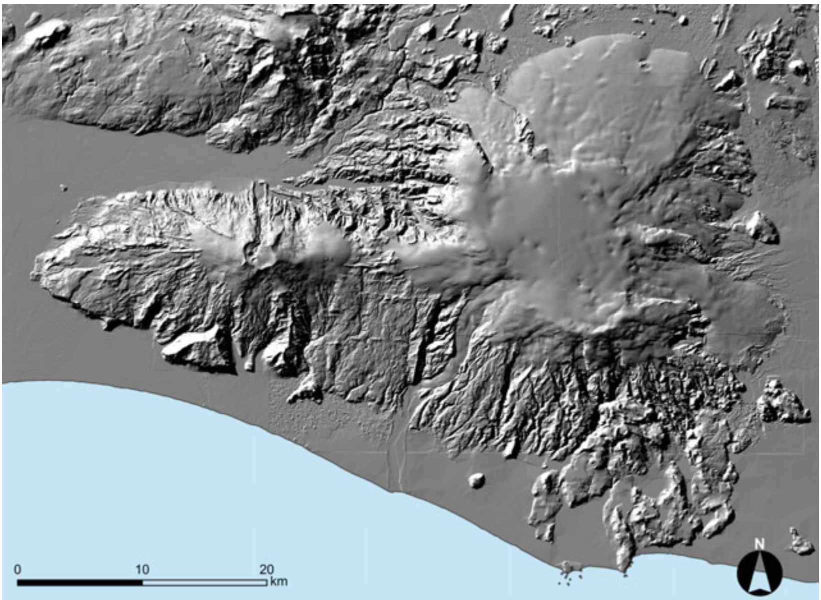

Fig. 14.18 Digital elevation map of Eyjafjallajökull and Myrdalsjökull (Mýrdalsjökull) showing the shapes of the volcanoes and their calderas. The names and structures are identified in Figs. 14.1 and, in particular, in Fig. 14.12b

outside this excursion, but is the most active volcano in Iceland in terms of frequency of eruptions. During the past 1100 years Grimsvötn has eruptive about 70 times, or once every 16 years. The four historical eruptions in Eyjafjallajökull include the eruptions in 2010, which we shall now describe briefly.

A typical basaltic **effusive fissure eruption** began 20 March 2010 in the pass between the glaciers Eyjafjallajökull and Myrdalsjökull (Figs. 14.1, 14.12b and 14.18). This fissure eruption, producing low-viscosity primitive basalt, came to an end 12 April. The associated volcanic fissure has a north-northeast direction and is about 500 m long, with an estimated maximum opening (aperture, feeder-dike thickness) of about 1 m. The length/opening ratio of this fissure/feeder-dike is in good agreement with measured ratios elsewhere in Iceland (Chap. 11) and shows that the overpressure or driving pressure of the magma when the feeder-dike reached the surface was low (several mega-pascals). Consequently, the eruption was a very small (total volume 0.02 km^2) 'gentle' outpouring of aa lava (Fig. 14.11)—the kind of eruption that some refer nowadays to as a 'touristic eruption'.

The first eruption faded away and finally came to an end on 12 April. Then a **second eruption** began 14 April 2010, but this time beneath the ice cap in the summit **caldera** (Figs. 14.12b and 14.18). This eruption melted its way through the 250 m thick ice cap, was highly explosive, and produced ash of an intermediate (andesitic) composition, which means that the magma erupted was more gas-rich and viscous than the one in the earlier (March) eruption. The magma supplied during the second eruption came from a shallow **magma chamber**, at about 4–5 km depth which is associated with the summit collapse caldera. The eruption column reached a maximum altitude of about 11 km, and the resulting ash clouds had major disrupting effects on **air traffic** in Europe for many days. The second eruption came to an end on 22 May 2010.

It is convenient to regard the first and second eruptions as two phases of a single eruption. The second, explosive, phase produced about 0.18 km^3 of ash and lava. The total volume of ash (calculated as solid rock) and lava in both phases of the eruption was thus about 0.2 km^3. This is thus a typical **small eruption** from a central volcano in Iceland in terms of volume of eruptive materials calculated as solid rock (or as magma volume) The volume of ash or tephra produced, however, was large in comparison with the size of the eruption. The volume of fresh ash (not calculated as solid rock or magma) was about 0.3 km^3. To give you an indication of what that means we can consider the following hypothetical case (which could never happen in the real world). Imagine all this ash falling on Reykjavik (Figs. 2.1 and 3.1) with a total area of about 275 km^2. Then that total area would have been covered with ash of an average thickness just over 1 m. In reality, much of the ash fell on Eyjafjallajökull itself and its close surroundings, whereas part went to Europe. The ash clouds carried to Europe had major disrupting effects on air traffic, particularly in West and North Europe, for many days. More specifically, many airports had to be closed for days on because of the ash clouds. Part of the ash was particularly **fine-grained** (composed of very small particles), which was one reason for its potentially damaging effects on the jet engines of the aeroplanes, hence the closure of the airspaces.

There were three fissures associated with the 2010 eruption, all of them several hundred metres long (maximum length 6–700 m). Two of the fissures are directed north-northeast; one is inside the caldera, the other is outside the caldera. The third fissure, which produced most of the eruptive material, is directed east-west. These three fissures thus reflect the main tectonic trends in Eyjafjallajökull, namely the east-west trend, which is also the direction of the long axis of the volcano itself (Figs. 14.12b and 14.18), and the northeast trend which is the general trend of the East Volcanic Zone (Fig. 14.12b).

The 2010 eruption was the culmination of an unrest process lasting **decades**. More specifically, there were several earthquake and deformation episodes in Eyjafjallajökull prior to the 2010 eruption. There were at least four earthquake swarms in the 1990s, two of which were clearly related to doming or uplift (inflation) of the volcano, namely in 1993–1994 and in 1999. There were few earthquakes and little deformation in the volcano from the year 2000 until 2009, when earthquakes and deformation began again. The rate of deformation and earthquakes increased much from early March 2010 until the eruption began on 20 March.

All these main earthquake and deformation episodes in Eyjafjallajökull have been interpreted as being related to **dike injections** from great depths (some 20 km) that became deflected into **sills** (Figs. 14.16, 14.17 and 14.19) at depths between about 3 and 6 km below the top of the volcano. Thus all the injected dikes during these 17 years became **arrested** in the sense that they changed into sills, or alternatively just stopped their vertical propagation at contacts, and thus did not erupt until the dike that fed the March eruption of 2010. It has been suggested that some of the sills may reach lateral dimensions (diameters) of as much as 17 km. One remarkable thing is that it appears that the dike that fed the March 2010 was deflected into sill at a shallow depth, but then deflected again into a dike to reach the surface. Such changes in dike paths are, in fact, common, but this one apparently had an unusually long (lateral) sill part before it changed again into a dike and erupted.

Is there anything unusual about these dike-sill intrusive events in the past 17 years? Certainly not. The cliffs sections show **exactly the same structures**,

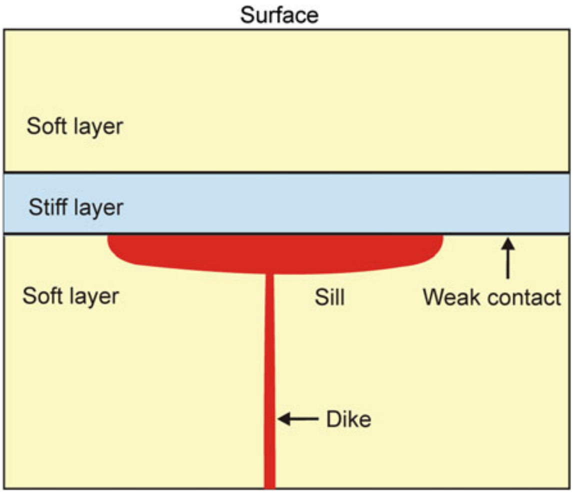

Fig. 14.19 Dikes are commonly deflected into sills at contacts between mechanically dissimilar rock layers. Here the dike was transformed into a sill at the contact between a stiff layer (for example, a basaltic lava flow or an earlier sill) above and a softer (more compliant) layer below (for example, hyaloclastite or tuff). Many of the sills in Eyjafjallajökull are formed at such contacts. Alternatively, if the driving magmatic pressure (overpressure, see Chap. 11) in the dike is low when the dike meets the contact, the magma is not able to open the contact to form a sill and the dike simply ends vertically (Fig. 13.35)

even at shallow depths, namely dikes and sills. In particular, Figs. 14.16 and 14.17 show many sills and dikes. As is common, the dikes tend to be much thinner than the sills (Fig. 14.19). Most sills are supplied with magma through dikes, and very few sills reach the surface. This means that most of the sills you see in the cliffs of Eyjafjallajökull were fed by dikes that never reached the surface to erupt. It follows that the processes that occurred at depths of 3–6 km beneath the top of Eyjafjallajökull in the period from 1993 to 2010 are exactly the same as seen at depths of 1–1.6 km below the top of the volcano in the cliff sections in Figs. 14.16 and 14.17; namely injected dikes failing to reach the surface—becoming arrested—and many changing into sills (Fig. 14.19).

Why is dike arrest and dike deflection into sills so common in Eyjafjallajökull (and stratovolcanoes in general). Answer: because Eyjfjallajökull and other similar volcanoes are composed of layers with widely **different mechanical properties**. Some of the layers are stiff and brittle, others are soft (compliant) and more ductile. For example, the hyaloclastite layers tend to be rather soft, particularly when young, whereas the lava flows (and existing sills) tend to be stiff. Such materials in general are referred to as **composite materials**, of which plywood is perhaps the best-known example. Composite materials are used in a variety of things, including aeroplanes, to make them strong. They are strong in the sense that the contacts between layers tend to arrest, stop or deflect, fractures. Stratovolcanoes are also called **composite volcanoes** (Fig. 11.3) and behave mechanically in a very similar way to artificial

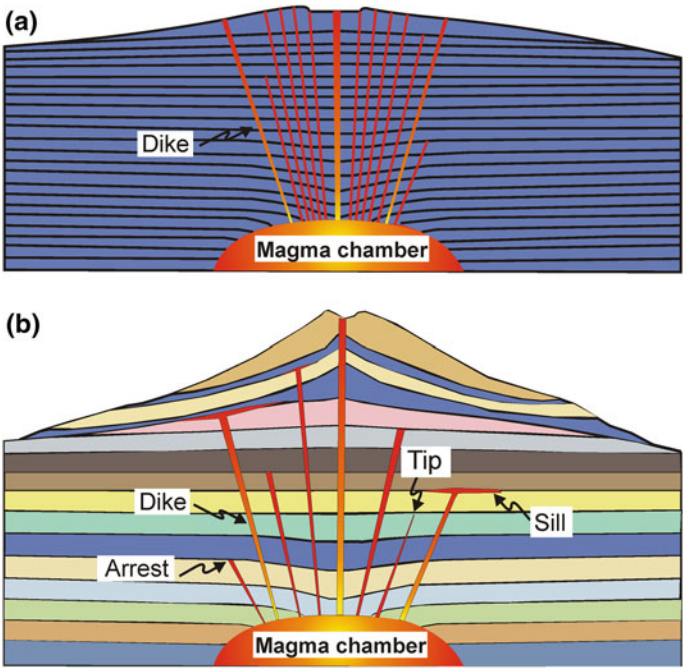

Fig. 14.20 Basaltic edifices are composed of rock layers all of which have very similar mechanical properties (basaltic lava flows) whereas stratovolcanoes (central volcanoes, composite volcanoes) are composed of rock layers with widely different mechanical properties (including lava flows of various compositions, layers of pyroclastics (e.g. hyaloclastites) and sediments, and intrusions). It is thus easier for any type of fracture to propagate through a basaltic edifice **a** than a stratovolcano/composite volcano **b** for the simple reason that many more fractures, such as dikes, become arrested or deflected into sills, and therefore do not reach the surface, in stratovolcanoes than in a basaltic edifice. The shape or profile of the basaltic edifice in **a** is based on the shield volcano of Mauna Loa, Hawaii, whereas that of the stratovolcano in **b** is based on that of Mt Fuji in Japan

composite materials. Thus, when a dike is propagating it induces stresses ahead of its tip or top. The stresses may be high in the stiff layers, and encourage dike propagation, but suppressed or very low in the soft layers, encouraging dike arrest. Commonly, when the dike meets a contact between stiff and soft layers it thus either becomes arrested altogether—its tip stops propagating—or the dike become deflected into a sill along the contact (Figs. 14.19 and 14.20). This latter is what we see as common in the present cliff sections (Figs. 14.16 and 14.17) and was inferred from geodetic and seismic measurements prior to the 2010 eruptions in Eyjafjallajökull.

Because composite volcanoes, composed of widely different layers such as Eyjafjallajökull, more easily stop or deflect fractures—not just dikes but all rock fractures—than volcanoes whose layers have all basically the same mechanical properties, such as lava shields and basaltic edifices (for example, the shield volcanoes of Hawaii), composite volcanoes/stratovolcanoes are mechanically **stronger** than **basaltic edifices** (Fig. 14.20). It is therefore easier to fracture, say, a lava shield than a stratovolcano. This is one reason why normal faults and tension fractures are much more common in lava shields of Iceland (Chaps. 2, 5, 6 and 13) than in areas composed of widely different rocks types, such as hyaloclastite mountains and stratovolcanoes.

14.6 Skogarfoss (Skógarfoss) Waterfall

We now drive on along Road 1 to the east along the cliffs of Eyjafjallajökull until we come to the **sixth stop (6)**, namely the river **Skoga (Skógá)**, which is roughly at the boundary between Eyjafjallajökull and the next volcano to the east, namely **Katla**. The river Skoga (Fig. 14.21a), however, not only marks the boundary between these two large and active volcanoes, but also hosts the famous and beautiful waterfall **Skogarfoss (Skógarfoss)**, seen in Fig. 14.21b. It is worthwhile to drive the road to Skógarfoss to have a closer look at this exceptionally well-formed waterfall. Like Seljalandsfoss (Fig. 14.13) Skogarfoss falls off a 60-m-high cliff, an old sea cliff, formed when the sea level was much higher some 13,000 years ago. The width of the waterfall is about 25 m. There is commonly a rainbow produced by the interaction of sunrays with the drizzle from the waterfall (Fig. 14.21b). The rock that constitutes the cliff is mainly of hyaloclastite, similar to that found along the cliffs of Eyjafjallajökull and at Seljalandsfoss. There is a path to the top of the waterfall.

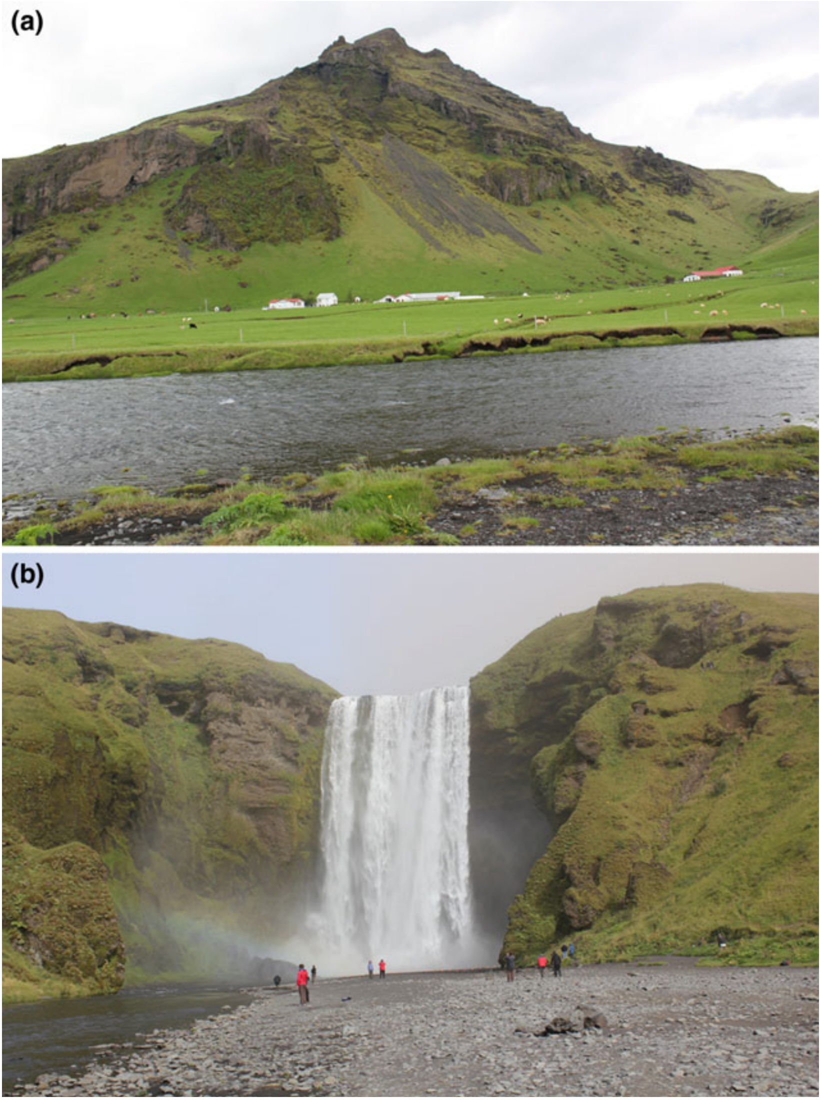

Fig. 14.21 River Skoga (Skógá) and its waterfall Skogarfoss (Skógarfoss). **a** View west, Skoga is roughly at the boundary between the volcanoes Eyjafjallajökull in the west and Myrdalsjökull/Katla in the east. **b** The waterfall Skogarfoss, 60 m high, is one of the best-known landmarks in Iceland

14.7 Petursey (Pétursey)—An Island on Dry Land

From the Skogarfoss waterfall we drive back to Road 1 and then continue to the east. You may want to drive Road 221 to the glacier **Solheimajökull (Sólheimajökull)**—the associated river which you cross on Road 1 is well known for its strong sulphur smell—but that drive is not a part of the present excursion. Instead we drive on until we reach the small mountain of **Petursey (Pétursey)**, our **seventh stop (7)**. The mountain, located in Fig. 14.12b and seen in Fig. 14.18 also, can be viewed either from the west or from the east; your choice will presumably partly depend on the time of the day (and the direction of the sunlight) when you reach the mountain. Here we view Petursey from the east (Fig. 14.22). South of the mountain (not seen on Fig. 14.22) is a small intrusion, a solidified conduit, a plug, similar to the one in Fig. 11.15 but smaller.

In which way is Petursey remarkable? Strictly, it is a typical small hyaloclastite mountain, most likely formed in an eruption in the sea, a submarine eruption, sometime in the past one or two hundred thousand years. The mountain is not very tall, reaches a maximum height of about 274 m above sea level, and is in most ways

Fig. 14.22 View west, the hyaloclastite mountain Petursey (Pétursey). The mountain is located in Fig. 14.12b and also seen on Fig. 14.18

similar to several hyaloclastite mountains we have seen earlier (e.g. Chaps. 4–6 and 13) except for one thing: it may have been an island, and is now on dry land. The name Petursey means **Peter's Island**. The southernmost cliffs of Petursey are now about 3 km from the coast. Like most landform names in Iceland, the name of Petursey is about 1100 years old. It was given to the mountain during the settlement (primarily by Skandinavian, British, and Irish people) some 1100 years ago. Thus 1100 years ago Petursey may have been an island (although this is not certain), whereas today it is 3–4 km inland (the island is just over 1 km in diameter). Petursey therefore could be one indication that the south coast of Iceland at this location has migrated to the south—has extended into the sea—at a rate of about 3 m per year during the past 1100 years. We see later today that there are other indicators of the southward growth of the south coast, and Petursey may possibly be regarded as one such.

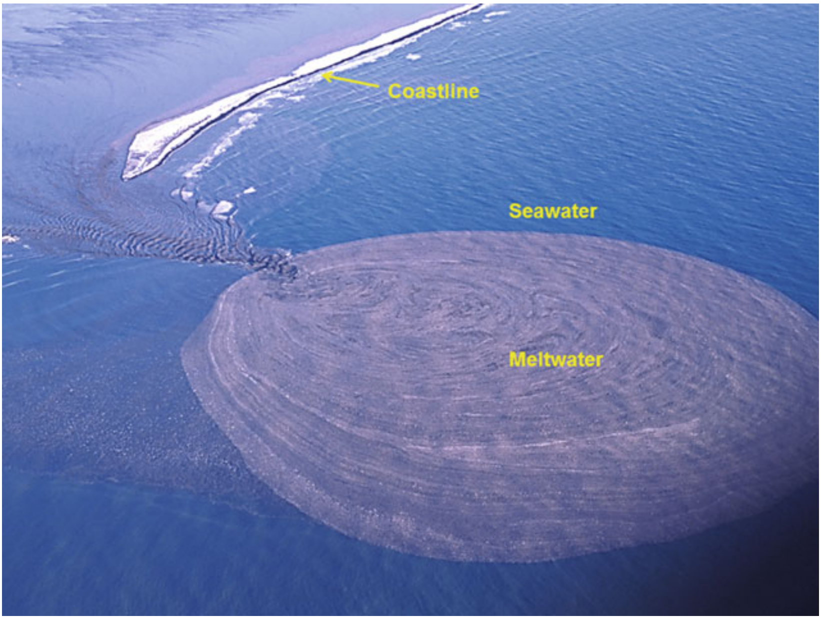

Fig. 14.23 Aerial view of a muddy flood entering the sea during the 1996 Gjalp (Gjálp) eruption in the Vatnajökull ice sheet (Vatnajökull is located in Fig. 2.2). The floods during Katla eruptions are much larger than this one

So why is the coast advancing to the south? The answer is partly the sediments carried by the glacial rivers but mainly the sediments carried by the floods that occur during eruptions in the nearby glaciers, particularly in the **Katla Volcano**. Katla is located within the ice cap of Myrdalsjökull (Figs. 14.1, 14.12b and 14.18). We will talk a bit more about Katla at the second last stop today. However, during eruptions from that volcano, there are normally enormous melt-water floods (the eruption melts the ice) that carry sediments (sand) to the coast, thereby adding to and advancing it. The sediments are partly carried by sea currents to other nearby coasts, such as the one close to Petursey, helping those costs to advance. In addition, there are, from time to time, melt-water floods on the outwash plains adjacent to Petursey, primarily **Solheimasandur (Sólheimasandur)**, just west of Petursey and related to the glacier Solheimajökull. An example of such a muddy flood, much further to the east along the south coast, occurred in the 1996 **Gjalp (Gjálp) Eruption** in the Vatnajökull ice sheet (Fig. 14.23). The floods during Katla eruptions are much larger than the one generated by Gjalp, as discussed later today. As a consequence of these muddy floods, the coast advances to the south.

14.8 A Village Threatened by the Eruptions of the Katla Volcano

We now drive on to the east. Depending on visibility, the ice cap Myrdalsjökull may be seen anywhere on the way to the next stop. Thus, I do not give any specific suggestions as to where it is best to take photographs of this great ice cap, but provide some taken from the Sandur plains later today. On our way to the next stop we pass through the valley of **Myrdalur (Mýrdalur)** and Road 215 on the eastern side of that valley. Road 215, which leads to the coast of Reynisfjara, will be our last stop in this excursion, and is very suitable as a final stop. Thus, we drive on to our **eighth stop (8)**, the village of **Vik i Myrdal (Vík í Mýrdal)**.

There are two main geological features to look at in Vik i Myrdal: one is the beautiful **layering** in the hyaloclastites behind the summerhouses (Fig. 14.24). The other is the coast and the rock pillars or sea stacks (Fig. 14.25). We will see the same rock pillars from a different perspective at our tenth stop later today, at Reynisfjara. The layering is typical for hyaloclastite mountains, but is exceptionally clear here (compare Figs. 6.9, 6.10 and 6.11). Each ash (tephra or pyroclastic) layer normally corresponds to one explosion in the crater while the mountain is forming. Some layers, however, are folded later during sliding of the ash layers down the slopes of the forming mountain (compare Fig. 13.6). These layers, like those in Petursey (seventh stop), in Reynisfjara (tenth stop) and, generally, in the

Myrdalsjökull Volcano (Mýrdalsjökull Volcano)—of which Katla is a part—are mostly younger than 200 thousand years. Thus, all the cliffs and mountains in this area are primarily of hyaloclastite, but, as said, the layering is exceptionally clear in the small cliffs seen behind the summerhouses and may be worth a short walk to take a closer look.

The rock pillars or stacks, named **Reynisdrangar**, reach a maximum height of about 66 m (Fig. 14.25). All the pillars are made of basalt and, like the mountain of which they are a continuation, namely Reynisfjall, they are of a mixture of extrusive and intrusive rocks. In Iceland most pillars or stacks of this type are former conduits, either cylindrical ones (Fig. 11.13), in which case the rock is commonly a mixture of breccia (fragmented rock glued into hard rock) and thin intrusions (mostly dikes), or elongated, and then most commonly rather thick, regional dikes (Figs. 11.7–11.10). Here the main pillars are three. One is very narrow, like a finger, and somewhat to the west of the other two, which are in line. One of those is cone-shaped (close to the coast), the other is with three peaks. We discuss the pillars again, and show more photographs of them, at the tenth stop.

Fig. 14.24 View northeast, layered hyaloclastite close to the village of Vik i Myrdal (Vík í Mýrdal). The summer houses provide a scale

Fig. 14.25 View southwest, the Reynisdrangar rock pillars or stacks, made of basalt, reach a maximum height of 66 m. Figures 14.35 and 14.36 provide close-ups of the pillars

14.9 Myrdalssandur (Mýrdalssandur)

From Vik i Myrdal, we drive further to the east along Road 1 until we cross the river **Mulakvisl (Múlakvísl)** and enter the plains of **Myrdalssandur (Mýrdalssandur)**, which will be our **ninth stop (9)**. The first feature to notice is the hyaloclastite mountain **Hjörleifshöfdi (Hjörleifshöfði)** in the southeast (Fig. 14.26). Its maximum height above sea level is about 220 m and it was most likely formed in an eruption in the sea, just like Petursey, and is also of similar age to Petursey. The mountain is named after one of the first two settlers in Iceland, Hjörleifur Hróðmarson, who is supposed to have arrived in Iceland together with his foster brother (as well as his first cousin) Ingólfur Arnarson and their families in about 874, that is, over 1140 years ago. Ingólfur settled in Reykjavik, but Hjörleifur in Hjörleifshöfdi (Hjörleifur was subsequently killed by some of his slaves and his killing was revenged by Ingólfur, but that story is outside the scope of this excursion). At that time, that is, in the year 874 there was a fjord west of Hjörleifshöfdi, so that Hjörleifshöfdi extended as a true cape into the sea (the word höfði means cape). The south part of Hjörleifshöfdi still reached into the sea as late as the 14th century. Then eruptions in Katla became larger and, presumably, the associated outburst floods also became larger, carrying with them sediments (sand and silt of volcanic origin) that extended the coast to the south. For example, measurements show that during the 1918 eruption of Katla, which included a large flood (see below), the coast closest to Hjöleifshöfdi extended south by a maximum of about 1.5 km, and some 14 km^2 of new land were added. Today the south edge of Hjörleifshöfdi is about 3 km from the coast, but that distance will no doubt increase considerably during the next eruption of Katla (Fig. 14.27).

The view to the north gives the second remarkable feature to notice, namely the hyaloclastite mountain **Hafursey**. This mountain reaches a maximum height of just over 580 m above sea level, and is most likely formed in eruptions beneath the ice sheets of the glacial periods during the past one or two hundred thousand years. The name means 'Oats Island', so the question is: was it once an island? Today Hafursey is about 14 km from the south coast. The elevation of the Sandur plains

Fig. 14.26 View east, the hyaloclastite mountain Hjörleifshöfdi (Hjörleifshöfði) has a maximum height of about 220 m and it was most likely formed in a submarine eruption

Fig. 14.27 View north, the hyaloclastite mountain Hafursey reaches a maximum height of just over 580 m above sea level and is most likely formed in one or more subglacial eruptions

around the mountain is close to 100 m, but those elevations are presumably mainly the result of the floods from Katla in the past centuries. There is little doubt that Hafursey really was an island at the end of the last glacial period, say 13 thousand years ago, when the sea level was much higher than it is today. But it is unlikely to have been an island in a normal sense during and soon after the settlement, some 1100 years ago, because at that time and for some centuries after there were farms on the lowlands south of Hafursey. These farms were later destroyed by the outburst floods from Katla eruption. But in the first centuries after the settlement, the Sandur plains we see today were, to a degree, covered with vegetation. The Sandur plains, at that time, were presumably mainly adjacent to the main rivers, such as Mulakvisl, which we just crossed. There are indications that until the 15th century the floods were both smaller and tended to be more confined to the eastern part of the present Myrdalssandur plains and had therefore more limited effects in the area between Hafursey and Hjörleifshöfdi, thereby making it possible to farm in these areas. Beginning in the late 15th century, however, the floods became larger and more destructive and tended to affect the western part of the plains as well as the eastern part. The outburst floods are called **jökulhlaup** in Icelandic. The word means glacier run or running glacier, and is used increasingly in the technical literature.

14.10 How Large Are the Outburst Floods?

So **how large** are the outburst floods or jökulhlaups during Katla eruptions? There are written descriptions of some that happened centuries ago, but these descriptions are normally brief. By far the best description, including photographs, is of the flood associated with the 1918 eruption of Katla. Estimates vary, but the maximum discharge during the jökulhlaup is generally regarded as between 200 thousand and 300 thousand cubic metres per second and covering much of the Myrdalssandur plains. For comparison, the maximum discharge during the 1996 Gjalp eruption (Fig. 14.23) was about 50 thousand cubic metres per second. The largest eruptions in Myrdalsjökull might generate even larger floods—possibly with maximum discharge rates as high as **million** cubic metres per second.

Let us now put these discharge values into context. As we discussed earlier today, the average discharge of the largest (in terms of discharge) river in Iceland, namely Ölfusa (Fig. 14.2) is about 420 $m^3\ s^{-1}$ (420 cubic metres per second). Thus, the discharge of the flood during the 1918 eruption of Katla was around **500-times larger** than that of the largest river in Iceland. So we have to look elsewhere for a reasonable comparison, always referring to the discharge where the river enters the sea. Volga is the largest river in Europe, with an average discharge of about 8100 $m^3\ s^{-1}$. Saint Lawrence and Mississippi are the largest rivers in North America, both with an average discharge of about 16,800 $m^3\ s^{-1}$. Ganges is the largest river in Asia, with an average discharge of about 38,100 $m^3\ s^{-1}$, and Congo is the largest one in Africa, with an average discharge of about 41,200 $m^3\ s^{-1}$. Thus, all these large rivers have average discharges less than **one-fifth** of the discharge during the jökulhlaups associated with major Katla eruptions. In fact, the only river in the world that has an average discharge similar to that of the peak discharges of the Katla jökulhlaups is the greatest river in South America (and in the world in terms of discharge), namely Amazon, whose average discharge is about 209,000 $m^3\ s^{-1}$ or close to the 200 to 300 thousand cubic metres per second peaks of the Katla outburst floods. Thus, the peak discharge of jökulhlaups during Katla eruptions is **similar to** the average discharge of the largest river in the world, **Amazon**.

Knowing the magnitude of the floods, the next question is: what are their velocities—how quickly does the water flood the Sandur plains and reach the sea? Written descriptions as well as model calculations suggest that the velocity may be several metres per second (around **15 km h^{-1}**, kilometres per hour). For comparison, a person may walk at the pace of about 1 $m\ s^{-1}$ (one metre per second), and average jogger runs at the speed of 12 km h^{-1}. If the next jökulhlaup followed similar paths as in the 1918 eruption, the flood would reach the place where you are currently standing, after escaping from the edge of the glacier **Kötlujökull** (Fig. 14.12b) in somewhat less than an hour. This is one reason why Road 1 across the river Mulakvisl (Múlakvísl) and Myrdalssandur in general is closed as soon as there is any unusual earthquake activity in the crustal segment beneath the glacier of Myrdalsjökull.

14.11 Katla Volcano

To see the location of the culprit, namely that of the Katla Volcano, somewhat better, we can drive a bit further to the east along the sandur plains. Then we have a reasonable view of Myrdalsjökull which hosts the **Katla Volcano** (Fig. 14.28). The ice cap is also seen in detail in Figs. 14.12b and 14.18. So what do we know about this destructive and rather notorious volcano? Here are some of the main points. First, Katla is located in the southeast corner of the much larger central volcano

Fig. 14.28 View west the ice cap and central volcano Myrdalsjökull (Mýrdalsjökull), part of which is the Katla Volcano. Located in Figs. 14.1 and 14.12b, and seen clearly in Fig. 14.18

named **Myrdalsjökull** (Fig. 14.12b). Myrdalsjökull, with a maximum elevation of 1450 m above sea level, is the second largest central volcano in Iceland—second only to Hofsjökull (in central Iceland). (Central volcanoes, however, are parts of volcanic systems, Fig. 2.2, and Hofsjökull is not the largest system, which is that of Bardarbunga (Bárðarbunga). An eruption occurred in Bardarbunga in 2014–2015. Second, Katla is a collapse caldera (Fig. 14.12b, compare Figs. 4.10 and 11.14), the second largest in Iceland (following the Torfajökull Caldera in the East Volcanic Zone). The Katla caldera has a maximum diameter of about 14 km and an area of about 110 km^2. This caldera is thus many times larger than the one in Eyjafjallajökull, as is seen in Figs. 14.12b and 14.18. Third, there is a magma chamber with a roof at the depth of about 2 km below the caldera bottom (surface of the crust)—that is, at about 2.5 km below the surface of the ice sheet that fills and covers the Katla Caldera. This is clearly a shallow magma chamber whose thickness reaches at least 1 km and the lateral diameter at least 5 km.

Katla is the third most active volcano in historical time (the past 1100 years) in Iceland. Only Hekla (23 historical eruptions) and Grimsvötn (in Vatnajökull, 70 eruptions) have higher eruption frequencies than Katla (21 eruptions). In the past 11–12 thousand years, that is, after the last glacial period, more than 400 eruptions have occurred in Katla and its surroundings. Some of the historical eruptions have produced much ash or tephra, largely because of the interaction of the meltwater from the ice cap with the magma—similar to what happened during the second 2010 Eyjafjallajökull eruption. For example, in the 1918 Katla eruption, some 700 million cubic metres of ash or tephra were produced, which is about twice the amount produced in the 2010 Eyjafjallajökull eruption. Calculated as magma, however, 700 million cubic metres is only about 300 million cubic metres (0.3 km^3), which means that the eruption was a typical small central-volcano eruption.

Most of the historical eruptions have been similar to the 1918 eruption, or smaller, and primarily produced ash or tephra, except one. This one is the **Eldgja (Eldgjá)** eruption in the year 934. This eruption took place on a volcanic fissure which is located partly beneath the Myrdalsjökull ice cap, but mostly north of it (so you cannot see it from here). The volcanic fissure is about 75 km long, but composed of segments (thus not continuous for 75 km—see Fig. 13.25 and the discussion about fissure segments in Chap. 13). The Eldgja eruption produced around 4.5 km^3 of fresh (newly fallen) ash, or 15-times as much as the 2010 Eyjafjallajökull eruption (4.5 km^3 of fresh ash corresponds to roughly 1.5 km^3 of magma, because the density of the ash is much less than that of the magma from which the ash is generated). The main production of the Eldgja eruption, however, was a basaltic lava flow covering some **800 km^2** and with a volume of about

18 km^3. So as regards production of eruptive materials, the Eldgja eruption was **100-times larger** than the 2010 Eyjafjallajökull eruption.

The Eldgja lava flow is partly covered by the outwash sediments, the sandur, from the more recent jökulhlaups from Katla. It is clear, however, that the lava flow was very extensive—as indicated, its estimated area is 800 km^2—and thus devastated and destroyed much agricultural land and associated farms in this part of Iceland. The eruption may have lasted many years. Eruptions of this magnitude, fortunately, are rare—perhaps one every thousand years in Iceland.

14.12 The Beach of Reynisfjara

We now drive back to the west along Road 1 through the village of Vik i Myrdal and continue until we come to Road 215. Then we drive that road south to the beach of **Reynisfjara**, which is our **tenth stop (10)**. As we walk from the parking place down to the beach we see the famous landmark **Dyrholaey (Dyrhólaey)** in the west (Fig. 14.29). Some of you might like to visit that locality as well. This is easy along

Fig. 14.29 View west, Dyrholaey (Dyrhólaey) with its famous semi-circular hole. Dyrholaey is the remnant of a hyaloclastite mountain and reaches a maximum height of 120 m above sea level

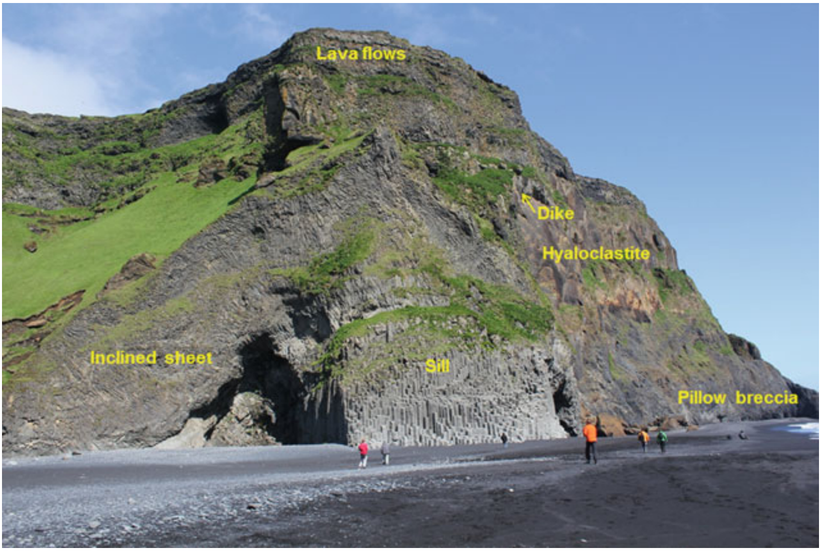

Fig. 14.30 View east, the southernmost part of the hyaloclastite mountain Reynisfjall and the Reynisfjara beach. The mountain is 5 km long and reaches a maximum height of 340 m above sea level. However, at the coast here the height is about 150 m. On the top is a cap of lava flows supplied with magma by a feeder-dike (both indicated). In the lower part of the mountain are an inclined sheet and a sill. Close-ups of the sill are in Figs. 14.31 and 14.32 and of the feeder-dike in Fig. 14.34. The people also provide a scale

Road 218. Dyrholaey is currently the southernmost point of Iceland (for a long time following the 1918 Katla eruption, Kötlutangi, the point of land south of Hjörleifshöfdi (ninth stop) was the southernmost part of Iceland). Dyrholaey is the remnant of a hyaloclastite mountain formed in an eruption in the sea, very similar to Petursey, Hjörleifshöfdi, and other mountains located close to the coast and of similar age. The maximum height of Dyrholaey is 120 m above sea level. It was formerly an island (hence the ending 'ey', meaning island, in its name) but is now a point. It is primarily of hyaloclastite, which is partly covered with a cap of pahoehoe basaltic lava flows that extend into the sea as pillars or sea stacks. The famous **semi-circular hole** through the southernmost part of Dyrholaey is formed by sea erosion.

As for the coast of Reynisfjara itself, the first thing to mention is that there can be sudden large and very dangerous **waves**. So please be aware of the waves all the time and take care. With this warning, we first take a look at the mountain itself, **Reynisfjall** (Fig. 14.30). We saw its eastern slopes from a distance at the eighth stop

(Fig. 4.25), and here we take a much closer look at its western slopes. Reynisfjall is a typical hyaloclastite mountain, just like Petursey, Hjörleifshöfdi, and Dyrholaey. The length of the mountain is about 5 km. It is elongated, with a maximum width of about 0.7 km (700 m). Its top is remarkably flat; the maximum height is about 340 m above sea level and occurs in its northern part. Close to the coast (Fig. 14.30) the height is about 150 m above sea level. Reynisfjall is of similar age to the mountains mentioned above, and most likely formed in several eruptions, partly in the sea and perhaps partly beneath ice caps, during the past 200 thousand years.

Reynisfjall is composed of a variety of rock units and layers. These include units of hyaloclastites (basaltic breccias and tuff), basaltic lava flows (in the top part), pillow breccias, and intrusions (dikes and sills/sheets). Perhaps the most striking features that catch the eye, however, are the rocks that form the beautiful **columns** at the base of the cliffs (Fig. 14.30). Columns of this type form as the magma/rock body (here a basaltic intrusion) contracts or shrinks during the cooling of the magma—from the original temperature of 1200–1300 °C. The shrinkage is because solids (including rocks) normally occupy less volume than fluids of the same material (the well-known exception being frozen water, ice, which occupies larger volume than liquid water).

The fractures are called **columnar** or **cooling** joints, and we have seen them many times before in sills (Fig. 4.9), lava flows (Figs. 5.5 and 5.6), and dikes (Figs. 11.6 and 11.10), but not as beautiful as here. The joints that form the beautiful rock columns begin to form when the magma has cooled down to about 800 °C and continue to develop as the rock cools further. Heat transport or 'flow' is always primarily along the steepest temperature gradient, in a similar way as a fluid flows downhill along the steepest slope of the hill. (This is, however, only an analogy; in heat transport there is strictly nothing that 'flows'. Heat is disorderly **transfer of energy** through uncoordinated motion of particles (atoms and molecules), that is, heat is energy in transit.) The steepest temperature gradient is in the direction of the greatest temperature difference between the hot magma and the surrounding cold rock. The columnar joints thus initiate at the contact between the magma and the surrounding rock, the host rock, and the columns that form are oriented perpendicular, at right angles, to the contacts. This is the reason why the columnar joints, and thus the rock columns, are horizontal in vertical dikes (Figs. 11.6 and 11.10) and vertical in horizontal sills (Figs. 4.9 and 14.17) and lava flows (Figs. 5.5 and 5.6).

In Fig. 14.30 we see that in the upper part the columns are steeply inclined, meaning that the cooling surface, the contact with the host rock, at the time of their formation was also inclined—in fact the contact was sloping similar to the present grass field in the photograph and belongs to an inclined sheet. In the lower part,

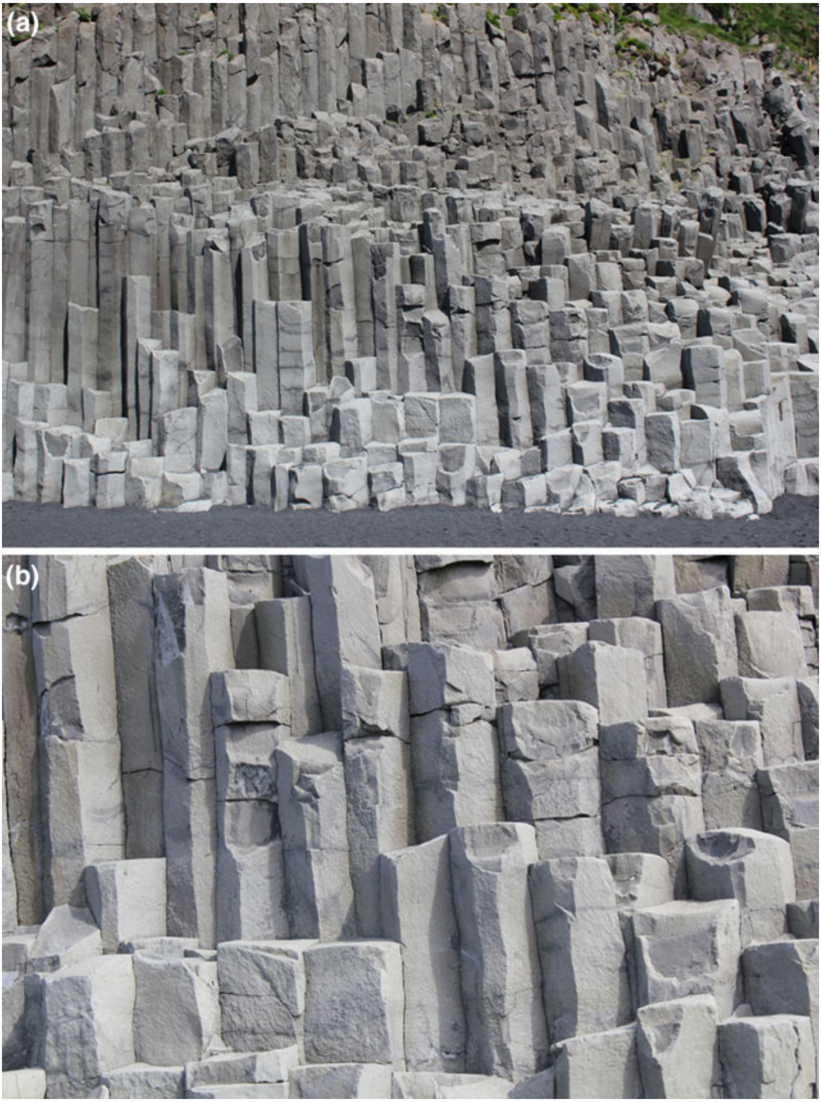

Fig. 14.31 Close-ups of the sill in Fig. 14.30. **a** Vertical rock columns indicate that the cooling surfaces were horizontal, as is typical for a sill intrusion. **b** Close-up of some of the columns seen in **a**. All the rock columns are generated through the development of cooling joints, columnar joints, as the magma cools down following emplacement as a sill

down at the beach, the columns are close to vertical. So this part had a horizontal contact with the host rock when the columns formed—and is a sill.

If we now take a closer look at this lower part (Fig. 14.31) we see that the columns are exceptionally well formed. So well-developed columns are very rare in lava flows and are normally found only in intrusions. The main reason is that in order to develop so regular and well-shaped columns, the rate of solidification ('freezing') of the magma and subsequent cooling of the rock to the same temperature as that of the host rock, must be slow—the cooling must take long time. If the magma is emplaced at the surface, the cooling is normally rapid (the magma being in contact with air or water or both), whereas if the magma is emplaced at depth inside older rocks, then the poor conduction of heat through rocks ensures that the cooling of the magma, that is, of the intrusion, will be slow. Given that the

Fig. 14.32 View southeast, some of the rock columns in the sill seen in Figs. 14.30 and 14.31 are very tall. The height of these columns is about 10 m

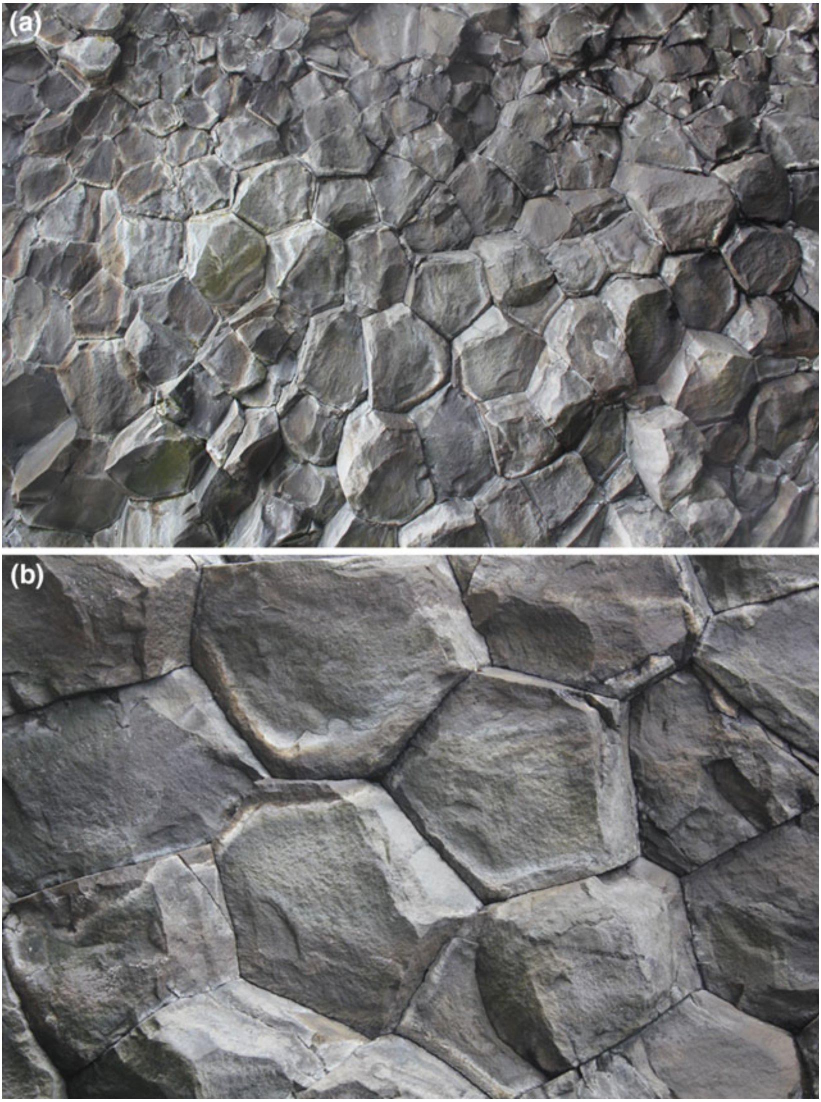

Fig. 14.33 Cross-sectional shapes of the rock columns as seen in the ceiling of the cave of Halsanefshellir (Hálsanefshelli). While the columns have a variety of shapes most are 5-sided (pentagons) or 6-sided (hexagons). **a** General view of the ceiling. **b** Close-up of some of the columns

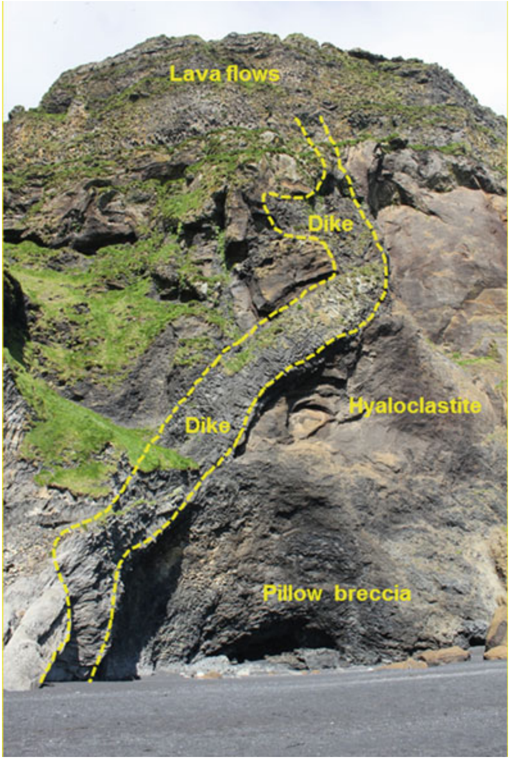

Fig. 14.34 View north of the basaltic feeder-dike to the lava flows on the top of Reynisfjall. The lower part of the dike is steeply inclined, mostly 1–2 m thick, but its uppermost part is much thinner and close to vertical

rock columns in this part are very well shaped, the cooling must have been slow, so that this is an intrusion. And from the vertical orientation of the columns we infer that the intrusion itself was horizontal, and thus a **sill**. Some of the columns at Reynisfjara are as tall as 10 m (Fig. 14.32). Similar columns are found in many sills in Iceland (Figs. 4.9 and 14.17), although rarely as well-developed and finely shaped as here

To get a better three-dimensional view of the columns, we can enter the cave **Halsanefshellir (Hálsanefshellir)** which is just around the corner (Fig. 14.33). In the ceiling of the cave we see that in plan view the columns have a variety of shapes. Most are 5-sided (pentagons) or 6-sided (hexagons). When the intrusions are uniform in composition, thickness, and other properties, and the cooling surfaces are straight and of uniform properties, then **hexagons** are most common. But there are normally, as here, some variations in intrusion properties and thickness, as well as in those of the host rocks, in which case pentagons and other geometries may also be common.

A short walk to the east of Halsanefshellir allows us to observe a basaltic **feeder-dike** to the lava flows at the top of Reynisfjall (Fig. 14.34). The lower part of the dike is steeply inclined, mostly 1–2 m thick, but its uppermost part is close to vertical and thinner. This change in geometry of dikes as they approach the surface is very common and follows from the fact that the forces and stresses in the crust demand that a dike meets the Earth's surface at (roughly) a **right angle**. The surrounding rock, the host rock, is mostly hyaloclastite, breccia and tuff.

We end the excursion by viewing the rock pillars or sea-rock pillars **Reynisdrangar**. These basaltic pillars or sea stacks, reaching a maximum height of **66 m**, were already described at the eighth stop (Fig. 14.25). But because we are much closer to them here, and view them from a different angle, they are really worth a closer look. In Fig. 14.35 we view them from a greater distance, and all the three

Fig. 14.35 View southeast, the three sea-rock pillars that constitute Reynisdrangar. All the three pillars can be seen here, although the third one (whose two peaks are seen) is somewhat hidden behind the one closest to us. The pillars reach a maximum height of 66 m

Fig. 14.36 View southeast, close-up of two of the sea-rock pillars that constitute Reynisdrangar. All are made of basaltic rocks

are seen, although the third one (whose two peaks are seen) is somewhat hidden behind the one closest to us. In Fig. 14.36, however, the third one is completely hidden from view, so only two of them are seen. Reynisdrangar are a famous landmark in Iceland, and an appropriate one for the last formal geological stop in this excursion.

From here, we drive back to Road 1 and then to the west to our final stop in this excursion, Reykjavik.

Glossary

Aa lava One of the two main morphological types of lava flows (the other being pahoehoe). Aa lava is characterised by a rough surface composed of angular and irregular blocks; it normally flows more slowly and has a higher viscosity than pahoehoe lava. Vesicles in aa lava flows tend to be more elongated and irregular (angular) than those in pahoehoe lava flows because of the higher viscosity of the aa lava

Active volcano A volcano that has erupted at least once during the past 10 thousand years

A-fault A term used here for a strike-slip fault of the type referred to in geology as **sinistral** or left-lateral (see strike-slip fault)

Amygdale Mineral deposit in a vesicle. A vesicle is the cavity in an igneous rock, such as a lava flow, that forms when gas expands in the magma and later escapes. Amygdales form when geothermal water flows through the rock and some minerals 'fall out' of the water, that is, precipitate, as the water becomes cooler. Amygdales have generally shapes similar to those of the hosting vesicles

Aperture The aperture of a fracture is the separation of the fracture walls or, more specifically, the shortest distance between matching notches and jogs on the fracture walls. It is also called **opening**

Ash Volcanic ash is the finest airborne material produced during an eruption. Ash is composed of particles that are 2 mm or less in size (diameter), similar to that of common dust particles. Ash forms a part of the class of pyroclastic material/rock (defined below) and is mainly composed of crystal fragments and volcanic glass. Volcanic ash is also sometimes used in a more general way as a synonym for pyroclastic material

A. Gudmundsson, *The Glorious Geology of Iceland's Golden Circle*,
GeoGuide, DOI 10.1007/978-3-319-55152-4

Basalt Basalt is dark-coloured (that is, mafic), fine-grained (with small grains/crystals) igneous (solidified from magma) rock. The most common rock on Earth; for example the oceanic crust is composed of basalt. Basalt occurs mostly as lava flows, but also as intrusions such as dikes and sills. All pahoehoe lava flows are of basalt and many aa flows as well. Large intrusions of basaltic magma, such as fossil shallow magma chambers, form a coarse-grained version of basalt named gabbro

Basaltic edifice A large polygenetic (formed in many eruptions) volcano, primarily or entirely made of basalt. A basaltic edifice is also commonly referred to as a **shield volcano**, with particular reference to the shields that constitute Hawaii. Notice that elsewhere the term shield volcano is also used about (primarily monogenetic—single eruption) lava shields. In this book the terms 'shield volcano' and 'basaltic edifice' mean the same and refer to a polygenetic volcano

Boundary faults The outermost and normally (but not always) the largest faults of a graben

Breccia Rock composed of angular and irregularly shaped fragments—broken rocks—in a matrix of finer or smaller particles. Here we discuss mostly volcanic breccia, but there are other types such as fault breccia which occurs on all large faults (generated through friction between the fault walls)

Brittle deformation When a rock (or other solids) response to force or stress by breaking and forming fractures, the deformation is said to be brittle. Brittle means that the material has a tendency to break (rather than bend, fold, or flow) when subject to stress or force

Buoyant A buoyant magma is lighter (less dense) than the surrounding rock (or magma) and tends to rise towards the surface. Acid or felsic (e.g. rhyolite) magma is generally more buoyant than basic or mafic magma (e.g. basalt)

Caldera Most calderas form through collapse or subsidence of part of a volcano as a piston into an associated shallow magma chamber. They are thus also named **collapse calderas**. The subsidence or collapse occurs along a fault that is commonly close to circular or somewhat elliptical in plan view, as is the caldera itself, and referred to as a ring-fault. Many calderas form or reactivate in connection with eruptions, but some collapses are not related to eruptions

Central volcano A polygenetic volcano, normally with frequent eruptions and a shallow magma chamber. The term covers both stratovolcanoes (composite volcanoes) as well as collapse calderas. In Iceland a central volcano is often, but not always, at the centre of a volcanic system and is the place where eruptions are most frequent in the system—hence the formation of a volcanic edifice that may later collapse to form a caldera

C-fault A term used here for a strike-slip fault of the type referred to in geology as **dextral** or right-lateral (see strike-slip fault)

Collinear Here used about fractures whose segments are all along a single straight line

Columnar joint A narrow fracture formed during the solidification and shrinkage (contraction) of the magma/lava body, such as a lava flow or an intrusion (e.g. a dike or a sill). Ideally, columnar joints form a network characterised by hexagonal columns. They are also called **cooling joints**

Complex network A network is a large collection of interconnected sites or points. The sites, also referred to as nodes or vertices, are connected by bonds or links, also referred to as edges. Partly because of the great number of bonds and sites, the network behaviour is complex; it cannot be predicted from the behaviour of individual sites or bonds. Common functions of networks include the transportation of matter, energy, force/stress, and information. For example, computer networks transport information, power networks transport electricity, railway and street networks transport vehicles (and people and goods), and fracture networks transport fluids

Composite materials These are materials made of two or more component parts that are bound together. Many composite materials are human made, but others are natural, such as natural wood and bone and rock. Rocks are composite on a small scale (crystals in a finer matrix) and on a large scale (layers or units constituting a volcano or a crustal segment). **Composite volcanoes/ stratovolcanoes** are excellent examples of structures made of composite materials (rock layers and units) on a large scale

Composite volcano This type of volcano, also known as a **stratovolcano**, is the most common type of polygenetic volcano and includes as members some of the best-known volcanoes in the world (such as Mayon in the Philippines, Merapi in Indonesia, Fujii in Japan, St. Helens in the United States, Teide in the Canary Islands, and Hekla in Iceland). Composite volcanoes are composed of layers or strata, of lava, pyroclastics, sediments, and intrusions such as dikes and sills. The layers have much more widely different compositions and mechanical properties than those that constitute basaltic edifices

Compressive stress Stress generated by a compressive force, that is, a force that seeks to compress the rock and generally reduce its volume

Continental drift An old version of the modern theory of **plate tectonics**

Cooling joint Same as columnar joints, discussed above

Crater cone A general term for the spatter and scoria (or cinder) cones that constitute common monogenetic volcanoes and also characterise volcanic fissures/crater rows. The cone is a steep-sided hill or hillock, normally circular or somewhat elliptical in plan view, formed around a lava fountain. Most of the cones have a central depression, the 'crater', hence the name

Crustal dilation A term used to indicate expansion of the crust as a result of spreading. It is normally given as the ratio of the new dimension divided by the original dimension, or as percentage change. For example, in measured sections or profiles in the Hvalfjördur area (Chap. 11) the crustal dilation due to cumulative thickness of dikes and displacement on normal faults varies from about 1% to about 6%

Cubic law The law states that the volume of fluid transported through a fracture, such as a volcanic fissure/feeder-dike, depends on the opening or aperture of the fracture in the 3rd power, that is, on the cube of the aperture. Thus, a feeder-dike, or a feeder-dike segment, whose opening or aperture is 4-times larger than that of another dike/dike segment (all other factors being the same) transports $4 \times 4 \times 4 = 4^3 = 64$-times more magma to the surface

Deformation band A narrow zone or a band where the rock has become deformed, but not yet developed a fault. Usually the deformation band is characterised by crushed grains making the band more compact and less permeable than the surrounding rock. Deformation bands occur primarily in sedimentary rocks

Dike A tabular body of igneous rock normally close to vertical or steeply inclined. Dike is an intrusion; more specifically, it is formed by a magma-driven fracture. Dikes normally dissect or cut through the layers of the surrounding host rock, such as lava flows and pyroclastic layers, and are thus discordant. Originally the word dike referred to the solidified rock in the fracture, but is now also used for the propagating magma-driven fracture itself, in phrases such as dike propagation and dike emplacement. Dike is the American spelling, and is also used in British spelling; **dyke** is exclusively British spelling

Dike swarm An elongated cluster or zone where dikes are much more common than in the surrounding parts of the crust. A typical dike swarm contains from many hundred to several thousand dikes and is a subsurface expression of a **volcanic system**

Dip The geological term for inclination of layers and fractures and other structures. Dip always refers to the angle in degrees between a horizontal plane and the inclined layer or fracture. We mostly use the more general word **inclination** in the book

Ductile deformation When a rock (or other solids) response to a force or stress by flowing rather than breaking the deformation is said to be ductile. Ductile means that the material has a tendency to flow, bend, or fold when subject to stress or force

Earthquake As the name implies, an earthquake is vibrations in the Earth. The vibrations are due to the passage of seismic waves carrying energy. Earthquakes therefore refer to events that generate such vibrations, that is, to seismic waves. For most scientists, however, an earthquake is the rupture or slip on the associated earthquake fracture, the fault, rather than the resulting ground movement or shaking

Earthquake fracture The fractures associated with or generating most earthquakes are **faults**

Earthquake magnitude There are two main earthquake-magnitude scales. One relates to the surface effects of the earthquake, primarily on human-made constructions, has the range from 1 to 12, and is named the Mercalli intensity scale. We do not use the Mercalli scale in this book. The other scale relates to the energy released or transformed during the earthquake. The most commonly used scale today is the moment-magnitude scale which is based to the size of the rupture area and the fault slip associated with the earthquake. Magnitude is commonly given with the symbol M followed by a number. Earthquakes of M2.5 or less are normally not felt by people. Earthquakes less than M5.5 usually cause little damage and may not rupture the Earth's surface (the fault slips only in the subsurface). Earthquakes of M6 and larger may cause much damage and those of M8 and larger a total destruction in a large area around the fault. Based on their magnitudes, earthquakes are characterised as follows: M3-3.9 minor, M4-4.9 light, M5-5.9 moderate, M6-6.9 strong, M7-7.9 major, M8 and above, great

Eruptive fracture Denotes the same as the term **volcanic fissure** and is the surface expression of a feeder-dike. Many eruptive fractures/volcanic fissures generate many crater cones and are referred to as **crater rows**

Explosion crater A depression formed by explosion when hot magma comes into contact with groundwater. When subsequently partially filled with groundwater the depression is referred to as a **maar**

Fault A rock fracture where the movement of the rock walls on either side of the fracture plane is parallel with the plane. When one wall moves up and the other down the fault plane, the fault is named **dip-slip** (the slip or displacement is parallel with the inclination or dip of the fault plane). **Normal fault** is the common dip-slip fault in areas of extension (reverse and thrust faults are common in areas of compression). When one fault wall moves horizontally to the left and the other to the right along the fault plane, the fault is named **strike-slip** (the slip or displacement is parallel to the direction or strike of the fault plane). Suppose you stand on one of the strike-slip fault walls. If the opposite wall has moved to the left during the slip, the fault is named left-lateral or **sinistral** (here an **A-fault**), whereas if the wall has moved to the right during the slip the fault is named right-lateral or **dextral** (here a **C-fault**)

Feeder-dike A dike that reaches the surface and supplies magma to an eruption. Also called **feeder**

Flow channelling A process whereby much of a fluid flow tends to be focused on, or channelled to, the large-aperture fractures or segments or parts of fractures. The **cubic law** is the main reason for flow channelling

Flow unit A layer that constitutes a part of a pahoehoe lava flow

Fold Bent or flexured rock layers; the result of ductile deformation

Forecast In geology the term is mainly used in connection with research on earthquake and volcanic-eruption forecasts. Generally speaking, there are, as yet, no reliable methods for forecasting earthquakes. The same applies broadly speaking to forecasting volcanic eruptions, except that over certain periods of time and for some volcanoes that behave very regularly, eruption forecasting has sometimes been successful

Fossil magma chamber A solidified (frozen) magma chamber and thus no longer active. Fossil magma chambers are normally seen as plutons, that is, large intrusive rock bodies (commonly of gabbro or granite), which are exposed through erosion

Fracture Rock fracture is a mechanical break, a discontinuity, in a rock body. The fracture separates the rock body into two or more parts

Gabbro A dark-coloured or grey, coarse-grained igneous rock, the plutonic equivalent of **basalt**. Coarse-grained means that the crystals that constitute the rock are easily visible to the naked eye. Many fossil shallow magma chambers are partly or entirely made of gabbro. A somewhat finer-grained version of gabbro is **microgabbro**

Gaussian distribution Another name for the well-known and common **normal** (or bell-shaped) probability distribution

Geothermal Related to the heat of the Earth's interior. As applied to geothermal fields, reservoirs, and energy the term geothermal refers primarily to the heat that can be extracted (as water or steam) from hot water circulating in the uppermost part of the Earth's crust

Geyser A hot spring that repeatedly erupts boiling water and steam

Glacier A body of ice, made of recrystallised and compacted snow, that flows in a ductile or plastic manner

Graben A subsided, elongated crustal segment or rock body bounded by two main parallel normal faults

Grey Basalt A name given to certain (mostly) pahoehoe lava flows in Iceland partly because of their grey weathering colour. The grey basalts are almost all formed in eruptions ice-free (interglacial) periods of the **Ice Age**

Groundwater Fresh water that fills cavities and fractures (pore space) below the Earth's surface. The depth to the top of the groundwater zone, named **water table**, varies considerably. It is shallowest in deep valleys with much **precipitation** (snow and/or rain), where the surface may be below the water table, in which case part of the valley becomes a lake, but deeper in mountain areas and deserts

High-temperature field A geothermal or hydrothermal field where the temperature at the depth of 1000 m is above 200 °C

Hyaloclastite A rock layer or unit formed as an aggregate of glassy fragments. The fragmentation occurs when lava/magma comes in contact with water (as in submarine and subglacial eruptions), cools instantly, and shatters. Also called **moberg** (**móberg**) and palagonite

Ice Age Informal name for the most recent major glacial epoch, the Pleistocene, whose duration was from about 2.6 million to 11–12 thousand years ago. There were short interglacial (ice free) and much longer glacial (ice) periods within the Ice Age. The term is also a general one for epochs when the poles and large parts of the continents have been covered with ice sheets

Ice sheet A thick body of glacial ice that covers the terrain (normally for a long period of time) and whose area exceeds 50 thousand square kilometres. Presently, the only ice sheets are on Greenland and Antarctica. A localised ice sheet smaller than 50 thousand square kilometres is referred to as an **ice cap**

Igneous A rock (or a mineral) generated as a result of solidification (freezing) of magma

Inclined sheet One of the three main types of sheet-like or tabular intrusions (magma-driven fractures) in volcanoes. An inclined sheet is neither vertical (like a dike) nor horizontal (like a sill) but inclined, commonly at 30–60° to the horizontal. Also called **cone sheet**

Intrusion A body of magma emplaced (intruded) into existing rock. The word refers both to the original magma body as well as to the solidified (frozen) body. The existing rock is referred to as the surrounding or host rock

Jökulhlaup An outburst flood generated by melting of glacial ice through volcanic and/or geothermal activity. All large jökulhlaups are the result of subglacial eruptions

Landslide Downslope movement, under gravity, of masses of rock and soil. The movement is normally comparatively rapid. The term is general and covers various specific mass-movement processes such as debris flows and **rockslides**

Lava The name given to magma once it reaches the Earth's surface and flows on it as a viscous liquid, eventually forming stream of lava, a lava flow

Lava fountain On reaching the surface, rapidly expanding and escaping gas may spray lava into the air to form a fountain above the conduit. Lava fountains commonly reach the height of tens of metres, but occasionally several hundred metres. On falling to the surface again the lava droplets become either **spatter** or **scoria**

Lava lake Fluid lava filling part of a crater, a caldera, or other depressions in a volcano. A small lava lake is commonly referred to as a **lava pond**

Lava shield Shield-shaped basaltic volcano. A lava shield has a similar shape as a **shield volcano**, but the lava shield is much smaller and normally monogenetic (and has no shallow magma chamber). By contrast, a shield volcano is polygenetic and has a shallow magma chamber

Lava tube A sort of tunnel that transports lava once its surface has solidified into a tunnel roof. Very common in pahoehoe lava flows; rare and short in aa lava flows. Also called **lava cave** and **lava tunnel**

Low-temperature field A geothermal or hydrothermal field where the temperature at the depth of 1000 m in the crust is less than 150 °C

Maar Explosion crater. Most maars are partly or totally filled with groundwater

Mafic Primarily used about igneous rocks rich in dark minerals (olivine, pyroxene, etc.), therefore dark coloured. Mafic corresponds to the word **basic**

Magma Molten rock, normally containing, in addition to the melt liquid itself, crystals and, particularly close to the surface, gas bubbles. Once magma reaches the surface as a flowing liquid it is called **lava**

Magma chamber A crustal region, basically a cavity, holding magma, part of which is eventually erupted. When unspecified, the word magma chamber means a shallow magma chamber, whose roof depth is normally not greater than 5 km below the Earth's surface. Deeper magma storage regions are commonly referred to as **magma reservoirs**

Mantle plume A large cylindrical upwelling of hot solid material and melt in the Earth's mantle. The surface expression of a mantle plume is a **hot spot**. Iceland is a classic example of a mantle plume, whose depth is at least 400 km and may be 2900 km. In the latter case, the plume would extend down to the Earth's outer core

Marker layer An easily identified rock layer that can be used to measure the displacement or slip along faults

Mechanical properties For solid rocks these refer mainly to the elastic constants, the most important of which is Young's modulus. Young's modulus is a measure of the springiness or stiffness of a rock (the higher the Young's modulus the stiffer is the rock) and is commonly also referred to as **stiffness**

Mid-ocean ridge A swell or ridge forming a part that rises above the surrounding ocean floor and is a site where new oceanic crust is formed through intrusions and volcanic activity. Mid-ocean ridges (or ocean ridges) are divergent plate boundaries where the crustal plates move apart through spreading

Mineral vein A fracture (normally as thin as a few millimetres or centimetres) filled with secondary minerals precipitated from circulating geothermal water

Moberg (móberg) The Icelandic (and now also somewhat international) term for hyaloclastite and palagonite

Neck A volcanic neck is a solidified part of an old cylindrical conduit, normally close to, or at, the centre of a volcano. Also known as a **plug**

Normal distribution A common continuos probability distribution with a bell shape. This means that most sizes or other values concentrate or cluster in the middle part of the distribution and then gradually decrease in frequency—taper away—to either extreme. The mean (average value), the median (the middle value in a sequence of values), and the mode (the value that occurs most often in the sequence) are all the same

Normal fault A dip-slip fault characterising areas of extension such as mid-ocean ridges and divergent plate boundaries in general. Two normal faults inclined towards a common centre constitute a **graben**. The most common fault type in Iceland

Offset Offset fractures or fracture segments are out of line, that is, are not collinear

Opening Opening of a fracture is the separation of the fracture walls or, more specifically, the shortest distance between matching notches and jogs on the fracture walls. It is also called **aperture**

Outburst flood A flood generated by melting of glacial ice through volcanic and/or geothermal activity. All large outburst floods are the result of subglacial eruptions. Also called **jökulhlaup**

Pahoehoe lava One of the two main morphological types of lava flows (the other being **aa**). Characterised by smooth ropy surface and flow units. Pahoehoe lavas normally flow faster, commonly in tubes, and have lower viscosity than aa lavas. Vesicles in pahoehoe lava flows tend to be circular or somewhat elliptical in cross-sections and more regular in shape than those in aa lava flows

Pascal A unit for pressure and stress, given as newtons per square metre. One pascal is very small so that the megapascal (million pascal) is commonly used in geology

Precipitation The total amount of rain, snow, and hail received by any particular part of the Earth's surface. Commonly given in millimetres per month or year.

The yearly precipitation in Iceland ranges mostly between about 400 and 4000 mm, being highest on the ice caps

Permeability The ability of rock to transmit fluids; alternatively, how easily fluids are transported through the rock

Pillow lava A heap or pile of pillow-like structures; the typical morphology of basaltic lava when erupted under water

Plate tectonics The modern version of the theory of continental drift, established in the 1960s. The outermost (mostly solid) layer or shell of the Earth, the lithosphere, is thought to be broken into 7 major and about 12 minor moving plates. Most of Earth's earthquakes, intrusions, and eruptions occur at the boundaries between the plates, at which the plates can converge (collide), diverge (pull-apart), and slide laterally past each other. The Eurasian and North-American plates diverge across the greater part of the volcanic zones of Iceland

Plug A volcanic plug is a solidified part of an old cylindrical conduit, normally close to, or at, the centre of a volcano. Also known as a **neck**

Pluton A large intrusions, that is, a subsurface body of igneous rock. Many plutons are fossil magma chambers

Power law A very common probability distribution. The distribution is characterised by a very high frequency of small objects or features or processes (for example, volumes of lava flows) and a very low frequency of large objects or features or processes of the same type and in the same distribution. Small and large are always relative and refer to the particular distribution being studied. Power laws are 'scale free', which implies that there is no size of objects or events that is typical for the distribution

Pseudocrater A hillock formed during hydromagmatic explosion in lava flows that meet water—a shallow lake or a moor. Their material is entirely derived from the lava flow in which they occur. More specifically, on contact between the lava, at a temperature of 1100–1200 °C, and the cold water, the water is transformed at an instant into rapidly expanding vapour, resulting in hydromagmatic explosions. The resulting hillocks are what we call pseudocraters or **rootless craters**

Pyroclastic rock/material Rock/material composed of fragments generated by explosive volcanic activity. More specifically, a rock composed of consolidated volcanic fragments. Depending on the sizes of the fragments, pyroclastic materials are classified (in order of decreasing fragment sizes) as bombs, lapilli, and ash

Regional dike A comparatively thick and long dike, commonly injected from a deep-seated reservoir rather than a shallow magma chamber. Regional dikes in Iceland form swarms that reach many tens of kilometres in length, whereas in some other countries such swarms reach the lengths of hundreds of kilometres and occasionally a few thousand kilometres. By contrast local swarms close to shallow magma chambers are mostly composed of inclined sheets and thin, rather short radial dikes. The composition of regional dikes is more ´primitive´ or basic (mafic) than that of the local radial dikes and inclined sheets. Regional dikes have, for example, supplied magma to all lava shields in Iceland

Rifting event Sudden extension across a part of a divergent plate boundary; for example, part of the volcanic zones of Iceland. Such an event happens when the tensile stress due to divergent plate movements reaches the tensile strength of the rock. During a rifting event there is normally formation and development of tension fractures and normal faults at the surface, and dike injection in the subsurface. Most dike injections, however, do not result in eruptions since the dikes become arrested (stop on their paths to the surface)

Rockslide A landslide during which a rock mass slides over an inclined surface of weakness, such as a soil, tuff, or hyaloclastite layer between inclined or dipping lava flows

Scoria Broken, stiff (often frothy) lava fragments, mostly from millimetres to centimetres in size (diameter). Scoria derives from the spray of lava fountains

Secondary mineral A mineral formed (normally much) later than the rock in which the mineral occurs (the host rock). In Iceland, many secondary minerals form **amygdales**, whereas others form **mineral veins** of quartz, calcite, and zeolites that derive from the circulation of geothermal water through the rock

Seismogenic layer The (brittle) part of the crust and lithosphere where earthquakes occur

Shear movement Fault displacement or slip

Shield volcano A basaltic edifice, a central volcano, and thus with a shallow magma chamber. The largest volcanoes on Earth (and in the Solar System) are shield volcanoes

Sill A sheet-like or tabular intrusion that is normally emplaced at contacts between horizontal or gently inclined or dipping layers. Most sills are thus close to horizontal

Slip Fault movement during a single earthquake. All the slips added together constitute the total **displacement** on the fault

Spatter Clots or 'droplets', from the spray of lava fountain, that are still hot and plastic when they hit the ground. The spatter clots commonly combine into larger flattened 'pancake-like' masses which, eventually, may flow for a while

Spreading The slow lateral divergent movement of the Earth's tectonic plates. More specifically, the crust/lithosphere in Iceland spreads horizontally at a rate of about 2 cm per year away from the volcanic rift zones where new crust is generated through intrusions and eruptions

Spreading vector Denotes the direction and velocity of spreading. In Iceland the spreading vector is directed northwest-southeast and the rate of spreading, as indicated above, is about 2 cm per year

Strain rate The amount or magnitude of strain, for example crustal dilation in a graben of a given width, over a specific time period (usually a second)

Stratovolcano Another name for a **composite volcano**

Stress Force per unit area; a measure of the intensity of a force on a given surface. Stress is given as the force unit newton per square metre, a unit whose name is **pascal**

Strike-slip fault When one fault wall moves horizontally to the left and the other to the right along the fault plane, the fault is named strike-slip (the slip or displacement is parallel to the direction or strike of the fault plane). Suppose you stand on one of the strike-slip fault walls. If the opposite wall has moved to the left during the slip, the fault is named left-lateral or **sinistral** (here an **A-fault**), whereas if the wall has moved to the right during the slip the fault is named right-lateral or **dextral** (here a **C-fault**)

Subaerial eruption Volcanic eruption 'under air', that is, on dry land

Subaquatic eruption Volcanic eruption 'under water', that is, in a lake (on a lake bottom)

Subglacial eruption Volcanic eruption 'under glacier (ice)', that is, within an ice cap/sheet

Submarine eruption Volcanic eruption 'under sea', that is, in the sea (on a sea bottom)

Table mountain A hyaloclastite mountain formed either in a subglacial or a submarine eruption. The lowermost part of the mountain is mostly pillow lava, followed by pillow breccia and then hyaloclastite at higher elevations. On the top of the hyaloclastite is a lava cap, a small lava shield

Tectonic plate In the theory of plate tectonics, the outermost layer or shell of the Earth, the lithosphere, is divided into (7 major) tectonic plates. Each plate is a massive slab of (mostly) solid rock whose shape is determined by the geometries of the plate boundaries at its edges. Some plates are purely oceanic (the Pacific Plate, for example), but most are composed of both oceanic and continental lithosphere. Also called **lithospheric plate**

Tensile strength The maximum tensile stress a rock can tolerate before it ruptures and forms an extension fracture such as a dike or a tension fracture

Tensile stress Stress generated by a tensile force, that is, a force that seeks to extend the rock and generally increase its volume

Tension fracture A fracture formed by tensile stress. The movement of the fracture walls is simple opening, pulling the fracture walls apart. In contrast to fault slip, there is no fracture-parallel movement during tension-fracture formation, and therefore no friction between the fracture walls

Thrust fault Fault formed under compression, compressive stress, where the fault plane has a gentle (shallow) inclination or dip. Thrust faults characterise convergent (colliding) plate boundaries and generate the largest earthquakes on Earth

Tuff Pyroclastic rock composed primarily of consolidated volcanic ash

Tumulus A hillock that forms when pahoehoe lava ponds in a small surface depression. Soon a solidified or frozen surface crust forms on the pond. If lava continues to flow into the pond, the solidified surface bends or domes so as to make space for the incoming lava, thereby forming a tumulus

Unrest Volcanic unrest refers to any significant increase in various detected geophysical, geochemical, and geological signals indicating changes in the rates or styles of associated physical processes within the volcano. The signals include changes in earthquake activity, ground deformation, emission of volcanic gases, and flow and chemistry of groundwater and associated streams. Volcanic unrest is often related to new magma being injected into the shallow chamber of the volcano resulting in its expansion and volcano **inflation** and may result in magma-chamber rupture and dike injection. Most unrest periods, however, do not result in a volcanic eruption

Vesicle A small cavity or hole in a lava flow or an intrusion at shallow depths generated by the expansion of a gas bubble during solidification or freezing of the magma that forms the rock. Most vesciles are a few centimetres or millimetres in diameter

Viscosity A measure of the resistance to flow in a fluid such as, for example, magma, lava, and water

Volcanic eruption In simple terms, the transport of lava (effusive or lava eruption), pyroclastics (explosive eruption), and volatiles (mainly water and various volcanic gases) from a volcano

Volcanic rift zone A volcanic zone where divergent plate movements—rifting—is the dominant volcanotectonic process. In Iceland, the North and West Volcanic Zones are rift zones, as well as the East Volcanic Zone roughly south to the northern edge of Myrdalsjökull

Volcanic system A swarm of tectonic fractures and volcanic fissures associated with a specific magma reservoir as a source. In the volcanic rift zone of Iceland most volcanic systems are highly elongated, commonly with lengths of 50–150 km and widths of 10–20 km and characterised by tension fractures, normal faults, and volcanic fissures at the surface, and dikes and normal faults at greater depths. Most volcanic systems develop a central volcano which is supplied with magma from a shallow chamber. All volcanic systems in Iceland have been active in the past 10–11 thousand years and constitute the most active parts of the volcanic zones

Volcanic zone A zone where a volcanic eruption may occur. The three main volcanic zones in Iceland are mostly 20–80 km wide and covered with rocks formed in the past 800 thousand years. Within the zones, however, volcanic eruptions and fracture formation (and earthquakes) are mostly confined to the volcanic systems

Water table The top or surface of the groundwater zone or an open aquifer

References and Further Reading

Much of the information provided in the book, including almost all the geological field measurements, derive from my own studies in the excursion areas during the past decades. It is not appropriate to list many tens of scientific papers here, but these are provided in the reference lists of my textbooks and monographs cited below. Here I list some of the books, special issues of journals, scientific papers, and maps that I have consulted and/or which may be helpful for further exploration of the geology of Iceland. All the edited volumes contain many scientific papers by different authors. I include many publications in Icelandic. Many of these have short summaries in English. In addition, I hope that many Icelanders will read the present book. For the publications in Icelandic I provide an English translation of the Icelandic titles which are then given in parentheses.

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