33 IGC EXCURSION 5
2008

Geology of Iceland

Pre-Congress Excursion
July 28 – August 4

Post-Congress Excursion
August 15 – August 22

Preface

This guide has been compiled for the Pre- and Post Congress, 33 IGC Excursion no. 5, the Geology of Iceland. A proper reference list is not included and so the guide should not be referred to or cited formally except after consulting with the compilators.

This work is predominantly based on a number of earlier geological guides written by the late Prof. Sigurður Þórarinsson but draws also on some more recent modifications and alterations by e.g. the late Prof. Þorleifur Einarsson. We have also heavily used number of earlier guides by e.g. Dr. Jón Eiriksson, Dr. Leifur A. Símonarson, Dr. Kristján Sæmundsson, Dr. Haukur Jóhannesson, Dr. Helgi Torfason, Dr. Karl Grönvold, Dr. Haraldur Sigurðsson, Dr. Kristinn J. Albertsson, Dr. Hreggvidur Norðdahl and the late Prof. Gylfi Már Guðbergsson.

Apart from the geological guides and roadlogs mentioned above, information on special topics was extracted from the writings of numerous geologists some of which are referred to at the end of this guide.

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Kerið, one of the craters in the Grímsnes volcanic area, was formed at the end or very shorrtly after the eruption when the botom of the crater fell into its own lava conduit.

33 IGC EXCURSION 5

A short guide to the Geology of Iceland

1. Introduction and purpose

The aim of the excursion is to give participants an insight into the fascinating geology of Iceland – a high-latitude volcanic island situated across a rifting plate boundary. The excursion route will be in Southwestern and Southern Iceland, and will focus on rift tectonics and volcanism and glaciers and glaciations through time. Volcanic and glacial/subglacial landscapes will be studied, as well as geothermal fields, geysers and high-energy fluvioglacial and coastal environments. The excursion coverage will be general in nature, suited for earth scientists of varying backgrounds, but there will be a strong emphasis on what makes Icelandic geology so unique – the interaction of ice and fire through time.

Leaders: Hreggviður Norðdahl (*hreggi@raunvis.hi.is*), Ólafur Ingólfsson (*oi@hi.is*) and Ármann Höskuldsson (*armh@hi.is*) at Earth Science Institute, University of Iceland and Gretar Ívarsson (*Gretar.Ivarsson@or.is*); Reykjavík Energy.

2. Geological outline of Iceland

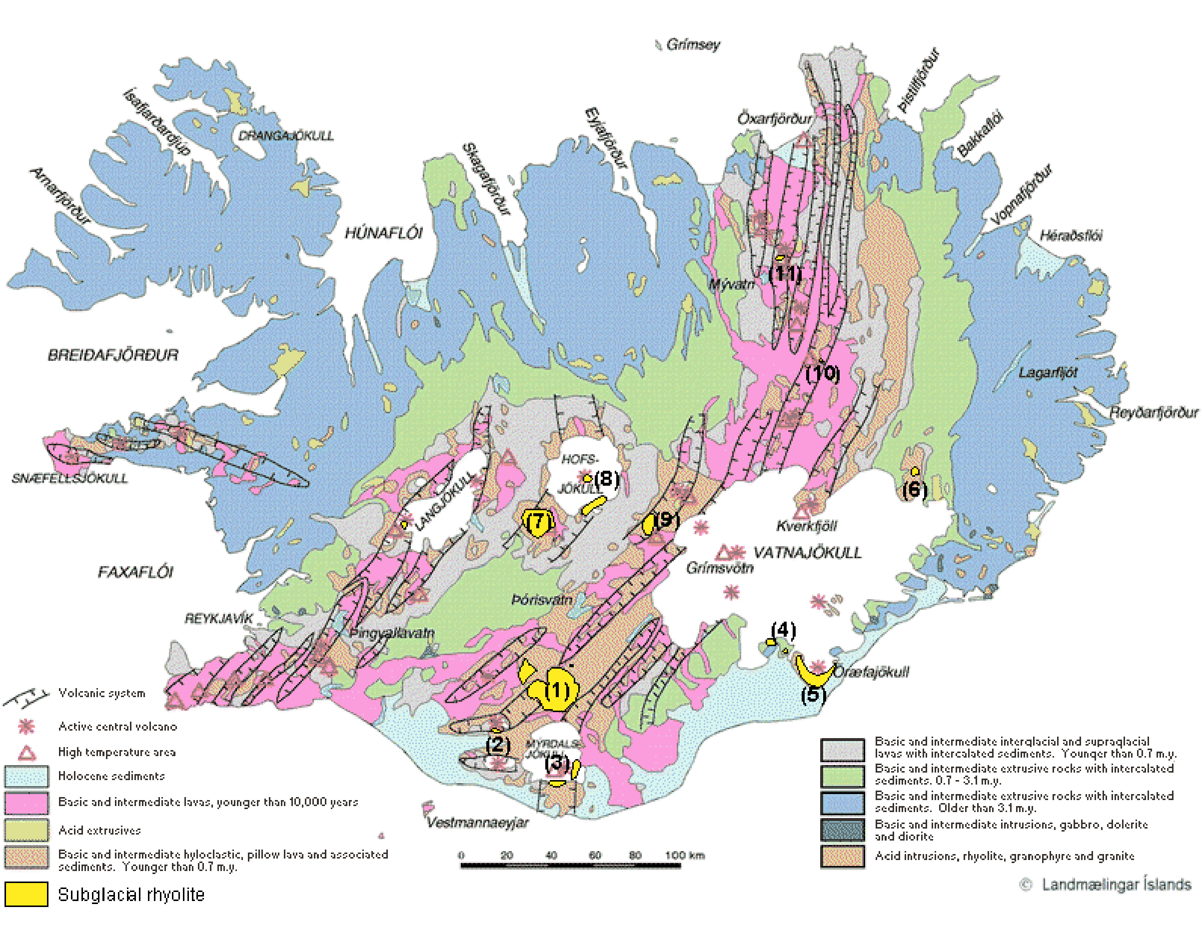
The following Geological outline of Iceland is reproduced from Geological guides compiled and edited by Jón Eiríksson et al. (1994) at the Earth Science Institute, University of Iceland and from a recent paper by Hreggviður Norðdahl and Halldór G. Pétursson (2005).

***Opening of the North Atlantic**:* The geological structure of Iceland is controlled by lithospheric plate margins which divide the island between the North American and Eurasian plates. The plate tectonic theory accounts for the formation, movement, and destruction of lithospheric plates, which are about 100 km thick segments of the earth's upper mantle and crust. The configuration of plate margins in Iceland is complex and is thought to be related to a mantle plume forming a hot spot under Iceland, approximately beneath Vatnajökull, the biggest ice cap in Europe. The Kolbeinsey ridge to the north of Iceland, and the Reykjanes ridge to the south are fairly straightforward spreading ridge segments of the Mid-Atlantic Ridge. Both ridges become obscured by anomalous volcanism and fracture zones as they approach Iceland.

The spreading history of the North Atlantic has been pieced together from dated magnetic anomaly lineations. Greenland and Scandinavia were separated when the Ægir ridge spreading axis was activated some 60 to 70 Ma. The ocean floor of the North Atlantic was formed by lithospheric accretion on both sides of the Ægir ridge up to about 25 Ma, when a new spreading axis opened up westward of the Ægir ridge, between Iceland and Jan Mayen. This was the beginning of the Kolbeinsey ridge which is still active. In the process, the Jan Mayen ridge, a 100 km wide segment of the old Greenland shelf, became isolated between the two spreading ridges.

The Kolbeinsey spreading axis is shifted eastwards along the Tjörnes Fracture Zone and continues across Iceland from Öxarfjörður across Kverkfjöll and Grímsvötn towards Hekla and Vestmannaeyjar. The Tjörnes Fracture Zone is not as well expressed topographically as many of the North Atlantic fracture zones but it is defined by intense seismic activity delineating a 70 km wide WNW-ESE trending segment which is characterized by N-S trending volcanic ridges and numerous faults. Up to 4 km thick sediments have accumulated in local basins in the Tjörnes Fracture Zone, dating at least as far back as the Pliocene.

The Reykjanes ridge is connected to the spreading axis across Iceland through the E-W trending South Iceland Seismic Zone, which extends across the lowlands of South Iceland. Although severe earthquakes occur along the zone it is only expressed in the topography as minor faults and fissures. The zone appears to be a fairly recent phenomenon and may have been gradually shifted southward in the last few million years as a result of a southward propagation of the currently active East Iceland spreading axis (*Eiríksson et al, 1994*).



(1) Torfajökul. (2) Tindfjallajökull. (3) Mýrdalsjökull. (4) Skaftafellsfjöll. (5) Oræfajökull. (6) Snæfell. (7) Kerlingarfjöll. (8) Armafell and Hásteinar. (9) Hágöngur. (10) Eggert. (11) Hlíðarfjall.

Volcanic zones in Iceland: Volcanism and seismic activity in Iceland are closely related to the plate margins across the country. The axial rift zones delineate areas where plate growth or accretion takes place. The axial rift zones are characterized by active volcanoes and extensional volcanic features. The eruptives belong to the tholeiitic rock suite. Two volcanic zones in Iceland are typical axial rift zones the western volcanic zone (Reykjanes-Langjökull Zone) and the eastern volcanic zone (Öxarfjörður-Vatnajökull Zone).

The axial rift zones consist of numerous fissure swarms where spasmodic dilatation of the crust takes place. Many of the fissure swarms are 10-20 km wide and 50-100 km long, and they are frequently associated with a central volcano. Together, a fissure swarm and a central volcano constitute a volcanic system, which has a typical lifetime of 0.3-1.0 Ma. The development of a volcanic system involves intensified dilatation and the formation of a graben system with deep rooted dyke injections into extensional fractures. Volcanic fissures or crater rows are common in these fissure swarms. The central region of many of the fissure swarms is characterized by frequent eruptions within a confined area and the development of a magma chamber in the crust below the system. Processes such as magma differentiation and partial melting of the host rocks within the magma chamber may lead to independent behaviour of the central volcano, which may respond independently of rifting episodes in the fissure swarm. Hekla and Askja are two infamous examples of central volcanoes.

The average dilatation in Iceland amounts to 2 cm/yr, indicating that the velocity of each plate in opposite directions is 1 cm/yr.

Two volcanic zones in Iceland do not have distinct extensional features and have been grouped as flank zones. These are the Snæfellsnes Zone and the South Iceland Zone (the southernmost region of the eastern volcanic zone), These zones are geologically young features and their volcanic products rest unconformably on Early Pleistocene and Pliocene rock units. The composition of the flank zone eruptives in generally transitional or alkalic. South Iceland Zone (the southernmost region of the eastern volcanic zone). These zones are geologically young features and their volcanic products rest unconformably on Early Pleistocene and Pliocene rock units. The composition of the flank zone eruptives in generally transitional or alkalic (*Eiríksson et al, 1994*).

The Tertiary: Tertiary rocks in Iceland consist predominantly of basaltic lava flows that were erupted in volcanic systems on axial rift zones. Rhyolitic and andesitic rocks are common in extinct central volcanoes. The oldest rocks occur in the westernmost and easternmost parts of Iceland, and have been dated to approximately 14-15 Ma with the K/Ar method.

The Tertiary plateau basalts cover well over 50% of the total area of Iceland. Fjords and valleys in western, northern, and eastern Iceland reveal magnificent outcrops of this oldest rock unit in Iceland. Sections in valleys, fjords, and headlands which have been carved into the lava pile by fluvial, glacial, and coastal erosion reveal over 50 buried and extinct central volcanoes. It is generally assumed that the Tertiary plateau basalts were produced by volcanic systems analogous to the currently active ones. Each activated system is believed to have produced an elongate lenticular unit which began to tilt towards the axis of the rift zone as it was removed laterally. The tilting was linked to the growth of the lava pile through subsequent loading and subsidence in the axial rift zones. This model is consistent with several key features of the Tertiary rocks in Iceland, such as the increased dips downwards. Location of fossiliferous deposits which are mentioned in the text the pile, down-dip thickening of lava units, and nearly horizontal hydrothermal alteration zone.

Terrestrial sediments are very common in the Tertiary basalt pile in Iceland. They accumulated as soil on the lava fields, and as volcaniclastic sediments in sedimentary traps and basins at lava margins, in depressions between volcanic systems, elongate graben systems, circular caldera features of central volcanoes, and other valleys. Tephra layers and tuffaceous rocks are frequently interbedded in the sedimentary sequences. Oxidized iron gives many of the Tertiary sediments a reddish tint, particularly where they are relatively thin and interbedded between lava units.

Marine sediments of Tertiary age are exposed on Tjörnes peninsula, but elsewhere they are presumably below sea level off the shores.

Many of the sedimentary horizons contain identifiable plant remains. A Miocene age of the fossil floras was proposed by Oswald Heer, and this has been confirmed by K/Ar dates. The global cooling that has taken place since the Miocene is well documented in the record of fossil floras, owing to the regular spacing apart of plant bearing horizons within the lava pile.

Palaeobotanical studies of Tertiary sediments in Iceland indicate more or less symmetrical horizons on both sides of the Quaternary units. The oldest horizon, older than 14 Ma, is found along the northwest coast and comprises the Selárdalur flora. It consists of a mixed warm-temperate forest of conifers and deciduous trees with *Sequoia, Pinus, Juglans, Alnus, Fagus, Ulmus, Tilia* and *Vitis*. Taxodiaceae dominate over Pinaceae and angiosperms over conifers. The well-known Brjánslækur flora belongs to the next horizon, about 14-13 Ma. The most thermophilic taxa, such as *Vitis,* have disappeared, whereas *Abies, Sequoia, Comptonia, Betula prisca* (Ettingshausen), *Acer,* Alnus, *Magnolia,* and *Sassafras* are prominent. Pinaceae predominate over Taxodiaceae. From 13-10 Ma Pinaceae increased and conifers came to dominate over angiosperms. The oldest horizon in East Iceland (Gerpir) probably belongs to this level. In the next horizon, about 10-9 Ma, comprising the Húsavíkurkleif and Tröllatunga floras in West Iceland and the Hólmatindur flora in East Iceland, Polypodiaceae, *Osmunda, Salix, Juglans, Betula, Acer, Magnolia* and *Carya* predominate. From 9-8 Ma *Alnus, Betula, Acer, Pterocarya, Fagus,* and *Corylus* are the most prominent taxa known from Mókollsdalur in Northwest Iceland.

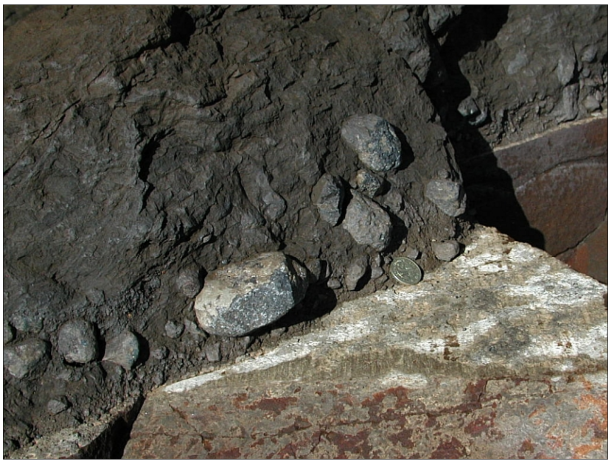
Icelandic Tertiary floras older than 8 Ma are warm-temperate and show close affinity with the recent flora in the Eastern Deciduous Forests of North America. The oldest horizons indicate annual mean temperature higher than 10 °C and frosts were probably rare. Apparently the precipitation was more or less constant throughout the year. A slight cooling may be responsible for the disappearance of *Vitis* at 14 Ma and Magnolia at 9 Ma.

The temperate Hreðavatn flora in West Iceland is about 7 Ma and *Fagus* and other warm-temperate indicators are rare or absent, whereas *Betula, Salix* and conifers are prominent. Apparently the climate grew cooler during the Late Miocene.

The cooling trend continued in the Pliocene and from 6-3 Ma *Betula* and *Salix* shrubs and grasses became more and more common when the forest declined, as indicated by the Sleggjulækur flora in West Iceland and the Pliocene Tjörnes flora in North Iceland. The Early Pliocene climate was probably similar to that in coastal Europe today where the mean temperature of the coldest month is close to 0 °C.

The water temperature (annual mean temperature) in the Tjörnes area during the Early Pliocene, when the marine Tapes and Mactra Zones were deposited, was at least 10 °C, or about 5 °C higher than the present one, as indicated by the presence of *Glycimeris glycimeris* (Linne), *Abra alba* (Wood) and other warmth-loving species.

The first sign of glaciation appears in the stratigraphy of East Iceland within the Fljótsdalur area at approximately 3.8 Ma, and again at 3.4 Ma. Despite detailed stratigraphic work in order to trace these deposits it has not been possible to map any regional distribution. The oldest glacial deposit that can be correlated between the two rock sequences in East Iceland (Fljótsdalur and Jökuldalur) appears within a reversed magnetic event, presumably the Kaena event of the Gauss epoch around 2.8 Ma. Two glacial deposits are identified between 2.8 and 2.48 Ma and altogether six glacial deposits are found in both eastern Iceland sections within the Matuyama epoch from 2.48 Ma to 1.8 Ma. The rock sequences in Fljótsdalur and Jökuldalur are probably the best preserved land sections in Iceland. They represent the most detailed record for the transition from preglaciation through the onset of local glaciation and on to full glaciation of the area. From 4.0 Ma up to 2.8 Ma a period of local glaciation on mountainous volcanoes is inferred, grading into a phase of more regional glaciation which correlates with the uppermost part of the Gauss geomagnetic time after 2.8-2.7 Ma. From then on a certain cyclicity becomes apparent; a cycle from an ice free period is inferred with lava formation, through a glaciation with the formation of basal tillite. The stagnation and retreat of the glacier is marked by marginal glacial deposits, lake deposits and glaciofluvial deposits. The subsequent interglacial period is then indicated by the formation of lava flows on top of the sediments, closing the advance-retreat cycle. The lithofacies studies indicate at least 7 such cycles within the East Iceland sections from approximately 2.8 Ma up to approximately 1.0 Ma.



The transition from local glaciation towards a more regional one in West Iceland is not quite as distinct as in the East Icelandic sections. In the inland section of West Iceland in Borgarfjörður, two glacial deposits are identified within the uppermost part of the Gauss epoch (2.6 - 2.48 Ma) and at least two, possibly three glacial deposits are further identified during the early part of the Matuyama epoch up to the Olduvai event (2.48 - 1.8 Ma). In the coastal section in West Iceland in Hvalfjörður, one and possibly two glacial deposits have been identified within the late Gauss, and 5 glacial - interglacial cycles are identified with certainty within the Matuyama. Lithofacies studies placed in context with the lithostratigraphy of the two areas suggest major glaciation and possible correlation between glacial deposits in the two sections around the Gauss-Matuyama transition.

The oldest glacial deposits in the Flatey and Tjörnes sections are synchronous. This points to a regional glaciation reaching beyond the coastline in northern Iceland at that time.

Preliminary results from a current study in South Iceland indicate that the earliest glacial deposit identified may be correlated with the Reunion geomagnetic time at approximately 2.2 Ma. In a section containing strata from 2.5 Ma to younger than 1.8 Ma only 4 glacial units are identified, two thereof may be from the same glacial cycle suggesting 3 major glacial cycles within that time interval. The Tjörnes and the Flatey sections in North Iceland extend further than the Olduvai event and until the Brunhes epoch, containing ten and two glacial deposits respectively during that time interval. Over 20 glacial deposits/horizons have been identified in the stratigraphic column in Iceland above the 3.8 Ma episode. Full scale glacial-interglacial cyclicity with regional ice-cover in Iceland is correlated with the uppermost Gauss somewhere around 2.6 Ma. A further amplification of the glaciations is identified around 2.4

(Gauss-Matuyama transition), 2.2 Ma (Reunion) and again around 1.8 -1.6 Ma (above Olduvai).

Studies of vertical and lateral facies associations at various sites in Iceland has thus revealed a consistent pattern of environmental changes that are directly related to the growth of a recurring ice sheet in Iceland through the Pliocene and the Pleistocene. An enlarging albeit intermittent ice-sheet is implied, initially expanding from the present Vatnajökull region in SE-Iceland, covering restricted areas during early glacials but eventually capping the whole island (*Eiríksson et al, 1994*).

The Quaternary – The Pleistocene: The lava pile in Iceland has continued to accumulate in volcanic systems during the Pleistocene, much in the same way as before. The main apparent difference is manifested in the radically changed environmental parameters during repeated loading and unloading of ice masses, and concurrent regressions and transgressions of the sea. These factors have dramatic effects on the physical character of the volcanic eruptions.

Rocks of Pleistocene age are mainly exposed within and bordering the volcanic spreading zones but are found also on Snæfellsnes (West Iceland) and Skagi (North Iceland). They differ in many ways from the Tertiary rocks, mainly by their greater variety of rock facies. During the interglacial stages the volcanic activity was mainly effusive as it had been in the Tertiary and was to be in the Holocene. A doleritic texture is a lithological characteristic for the major pan of the basalt (grey basalts). During times of glaciation when the country was covered with ice sheets volcanic products were piled up over the eruption centres as tuff ridges (pillow lavas and breccias) which were capped by subaerial lava flows to form table mountains if the volcanoes were built up through the ice sheets. The older glacial volcanoes are now mostly buried in the pile but those formed during the last two glacials are still impressive morphological features and can be used to define the surface attitudes of the ice sheets of their formation time.

In some regions sediments comprise up to one half of the Pleistocene series mainly fluvial, lacustrine and marine ones representing the interglacials and tillites the glacials. Traditionally, by means of biostratigraphy, palaeomagnetic and K/Ar radiometric work, the Late Pliocene and Early Pleistocene rocks in Iceland have been divided chronostratigraphically into two formations, i.e. the Middle and Late Pleistocene Palagonite Formation (PGF) of Brunhes age, i.e. younger than 0.7 Ma, and the Late Pliocene and Early Pleistocene Pleistocene Grey Basalt Formation (GBF) of Matuyama and probably late Gauss geomagnetic age.

The oldest ice caps or ice sheets formed during "glacials" in the volcanic zones 2-4 Ma ago. The sedimentary sequences are often cyclic, mainly though within the reach of marine transgressions due to glacio-eustatic changes, till, marine silt with arctic molluscs and then with boreal molluscs which in turn are then often covered by fluviatile deposits and with plantbearing sediments (lacustrine sediments or soils or lignite) which then often are covered by lava flows. Fossiliferous cyclic sediments are very useful in determining the full interglacial or interstadial character of the Pleistocene sequences. The Pleistocene sediments are also much more lithified than the Tertiary ones due to an increased admixture of volcanic material produced by subglacial eruptions. Red clayish soil beds, characteristic for the Tertiary rock series, are absent from the Pleistocene series. This and the appearance of tillites and subglacial palagonite i.e. hyaloclastites is very characteristic for the Pleistocene rock series.

Both the flora and the marine fauna were affected by the Late Pliocene and Early Pleistocene climatic deterioration in Iceland. The temperate conifers and deciduous trees became extinct in th Early Pleistocene and did not reappear during interglacials. The most prominent constituents of the Early Pleistocene floras were *Alnus*, *Betula*, *Salix*, and grasses, as indicated by the interglacial floras in Bakkabrúnir, North Iceland (age between 2 and 1 Ma), Stöð, West Iceland (slightly older than 1 Ma), and Svínafell, Southeast Iceland, probably slightly younger than 0.78 Ma.

The Pleistocene floras became gradually more and more similar to the present one, which is clearly of European affinity. *Pinus* became extinct after the deposition of interglacial sediments in Breiðavík on Tjörnes (about 1 Ma), and the sediments in Stöð, West Iceland. *Alnus* has not been found in Middle or Late Pleistocene sediments in Iceland, Apparently the earliest interglacials were milder than the later ones.

The climate during the later interglacial stages was probably similar to the present one, but during glaciations the annual mean temperature was at least 5-10 °C lower than to day and the snowline was about 1000 m lower. During the last glaciation Iceland was nearly completely covered with ice, but the main accumulation, at least towards the end of the Weichselian, was south of the present watershed, which indicates that the precipitation was mainly brought by southerly winds, as is the case today.

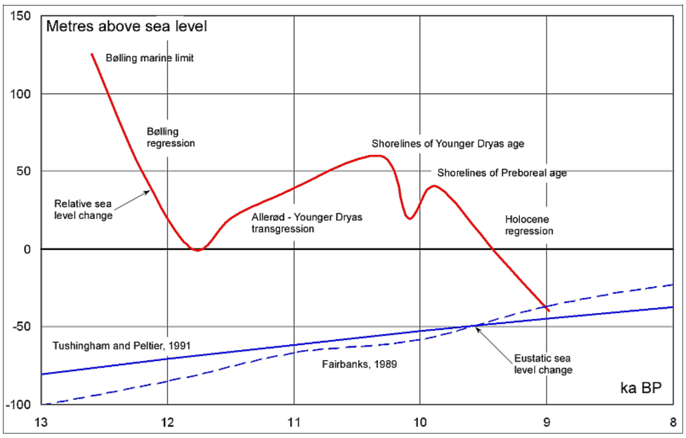
The Late Pliocene and Early Pleistocene marine faunas show a similar trend. The Late Pliocene marine fauna in Skammidalur, South Iceland, comprises warmth loving species such as *Chamelea* (=*Venus*) *gallina* (Linne), *Scrobicularia plana* (da Costa), and *Abra alba* (Wood). These are not found in Pleistocene deposits in Iceland and do not live in Icelandic waters today. This indicates a Late Pliocene water temperature 2-4 °C higher than the present one. In the Tjörnes area the thermophilic species disappeared before the deposition of the Late Pliocene *Serripes goenlandicus* Zone. However, the change in the marine Tjörnes fauna is somewhat obscured by the arrival of Pacific molluscs. The Breiðavík Group on Tjörnes conains several glacial-interglacial cycles of Late Pliocene and Early Pleistocene age. Studies of mollusc and foraminiferal faunas in the Late Pliocene and Early Pleistocene lithological cycles show faunal assemblages ranging from high arctic to boreal in harmony with environmental changes from glacial to ice free conditions. The early interglacial water temperature was extremely low, close to 0 °C but then ameliorated to temperatures similar to the present (*Eiríksson et al, 1994*).

Late Weichselian – Early Holocene: Stratigraphical data from the Quaternary lava-pile of Iceland have revealed repeated stadial–interstadial conditions in the early part of the Weichselian Stage. In Middle Weichselian times coastal parts of Iceland were ice-free and at least partly inundated by the sea during an interstadial that has been dated between 34.7 and 20.3 k ^{14}C yr BP, a period that is probably comparable with the Ålesund Interstadial in Norway. Therefore, the Last Glacial Maximum (LGM) extent of the Icelandic ice sheet most likely occurred at or shortly after, 20.3 k ^{14}C yr BP when ice streams from a central ice sheet reached out onto the shelves around Iceland. The actual extent of the ice sheet has been arrived at on the basis of basal core dates which either post- or predate the LGM extent, leaving it only indirectly and inexactly known.

Considering the present oceanographic circulation around Iceland, with warm watermasses arrived at Southeast Iceland, the sea off West Iceland must have warmed up before the warm sea-current rounded Northwest Iceland and reached the area off North Iceland. Basal dates of about 12.7, 15.4 and 13.6 k ^{14}C yr BP from the shelf off West, Northwest, and North Iceland are, therefore, considered minimum dates for the deglaciation of the Iceland shelf. By comparing these dates to the age of the oldest dated marine shells in Iceland, about 12.6 and

12.7 k ^{14}C yr BP in West and Northeast Iceland, the Iceland shelf seems to have been completely deglaciated during a period of 1000 to 3000 ^{14}C years. The innermost parts of the shelf such as the Faxaflói bay may have been deglaciated in about 400 ^{14}C years, showing an extremely high rate of glacier retreat.

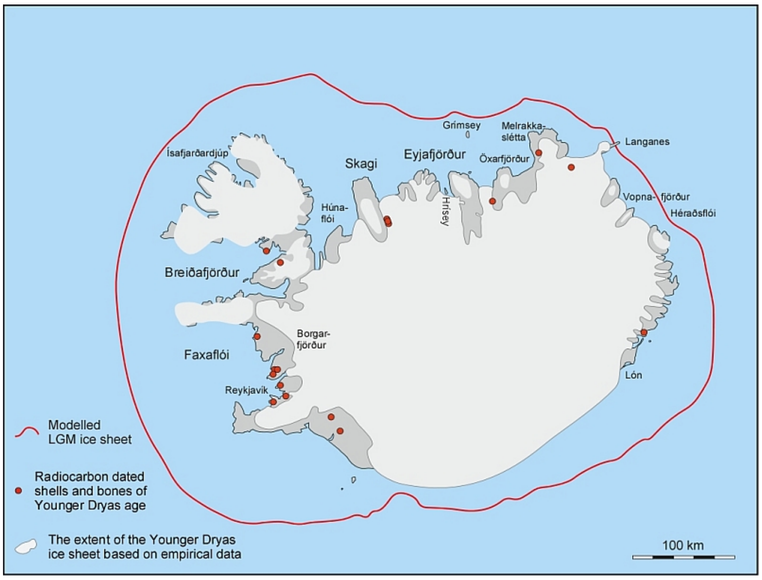
The dating of raised shorelines in West Iceland enables us to determine the amount of glacio-isostatic depression of the coastal areas some 12.6 k ^{14}C yr BP. Adding the height of the ML shoreline (150 m a.s.l.) and the position of eustatic sea-level (ESL) at that time gives an isostatic depression of about 225 to 250 m. The high elevation of the ML shoreline and great mobility of the Icelandic crust must indicate an extremely rapid deglaciation of the shelf off West Iceland. A slow ice retreat from the shelf area along with gradual melting and thinning of the glacier would quickly have been compensated for by continuous glacio-isostatic rebound, resulting in ML shoreline at much lower altitude than 150 m a.s.l. Consequently, greatly elevated ML shorelines of early Bölling age in West Iceland could, therefore, not be formed unless he deglaciation of the shelf and coastal areas in West Iceland progressed very rapidly (*Norðdahl and Pétursson, 2005*).



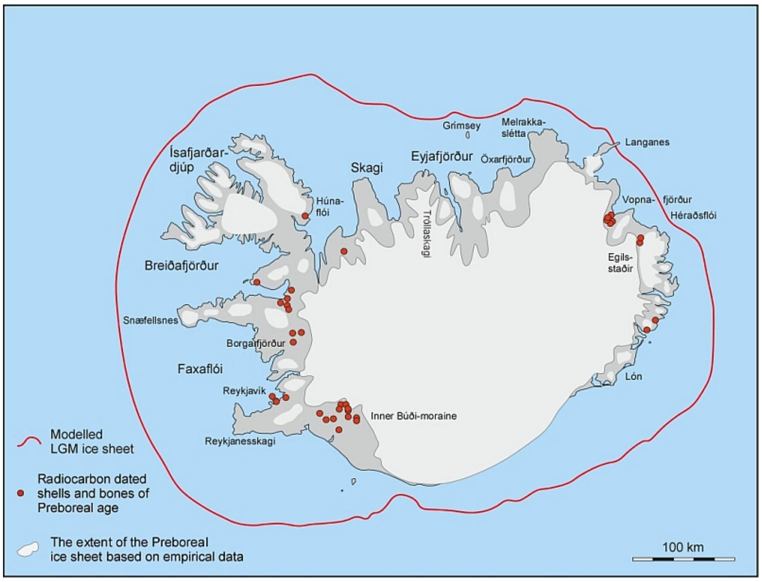
Lateglacial relative sea level change: A general conclusion concerning the initial deglaciation of Iceland, based on dates from the Iceland shelf, West and Northeast Iceland, is that ice covering the shelf around Iceland during LGM was very rapidly withdrawn from the shelf and onto present-day dry land. The deglaciated areas were consequently submerged by the sea and greatly elevated early Bölling ML shorelines were formed in coastal parts of Iceland. Subsequent to the formation of the ML shorelines the rate of glacioisostatic recovery exceeded the rate of Eustatic sea level (ESL) rise and relative sea level (RSL) regressed as evidenced by late Bölling shorelines at lower altitudes.

Data revealing the development of RSL changes and extent of the Icelandic inland ice sheet in Younger Dryas and Preboreal times have been collected in regions covering the majority of coastal Iceland. Data from the western part of North Iceland, Northwest Iceland, and Southeast Iceland are not available at the moment. The most observable fact in these regions is the occurrence of two sets of prominent raised shorelines, the Younger Dryas marine limit shorelines and the subsequent lower set of Preboreal shorelines, except in South Iceland and in Vopnafjörður, where the marine limit shorelines are of Preboreal age (*Norðdahl and Pétursson, 2005*).

Younger Dryas and Preboreal glacier extent: Deterioration of the Bölling – Alleröd marine environment at the end of the Alleröd Chronozone, witnessed e.g. by the appearance of high-arctic mollusc species as *Portlandia arctica* and *Buccinum groenlandicum* in West Iceland, also initiated a positive massbalance change of the Icelandic glaciers. Rising RSL at the end of the Alleröd and beginning of the Younger Dryas Chronozones, was both caused by ongoing global rise of ESL and increased glacio-isostatic load in the coastal regions of Iceland following an expansion of the inland ice sheet and advance of the glaciers. Thus the Bölling – Alleröd marine environment was rapidly changed from a moderate low-arctic/high-boreal to an arctic/high-arctic environment characterized by deposition of glacio-marine sediments frequently barren of mollusc shells and foraminifera tests. Furthermore, a significant decline of inferred mean summer temperature in the North Atlantic seaboard regions, reduced inflow of Atlantic water to the Norwegian sea and to the waters around Iceland, and a southward displacement of the North Atlantic Polar Front, all occurred around the Alleröd – Younger Dryas transition, leading to the onset of the Younger Dryas cold spell (*Norðdahl and Pétursson, 2005*).



Younger Dryas ice sheet: Based on these data we attempt to outline the extent of a Younger Dryas ice sheet in Iceland when only small pieces of present-day dry land may have protruded beyond the margin of the ice sheet. The extent of Younger Dryas glaciers on the Reykjanes and Snæfellsnes peninsulas is still not known in any detail. A first approximation towards the extent of a Younger Dryas ice sheet in Northwest Iceland indicates that the glacier margin extended beyond the present coastline in central parts of the Breiðafjörður north coast and into the main fjords on the west coast of Northwest Iceland when RSL was situated at 60-70 m a.s.l. This attempt to outline the Younger Dryas ice sheet in Iceland differs from earlier attempts mainly in having the margin drawn in greater detail (*Norðdahl and Pétursson, 2005*).



Preboreal ice sheet: Based on data presented above, we have produced the first reconstruction of the spatial extent of an Icelandic Preboreal (ca. 9,800 ^{14}C yr BP) ice sheet. Differences in the configuration and extent between the Younger Dryas and the Preboreal ice sheets are not very marked with the exception of the margin of the Preboreal ice sheet which had retreated a bit further inland in all parts of the country. The difference, though, seems to have been greatest in North Iceland where the major outlet glaciers in Skjálfandi, Eyjafjörður and Skagafjörður had retreated some 30-50 km. On the whole, the Icelandic ice sheet seems to have been reduced by about 20% in a period of about 500 ^{14}C years. Furthermore, indisputable evidence of a transgression of RSL temporarily reaching a Preboreal maximum position in the Reykjavík, Dalir, Skagi, and Langanes area, reflects increased loading on the Iceland crust due to glacier expansion. Evidently, the different size of the two ice sheets is a minimum difference and it is only reflecting a net mass-balance change between the formation of these two ice sheets. The extent of local corrie and valleys glaciers in elevated parts of coastal Iceland and on the Reykjanes and Snæfellsnes peninsulas in Preboreal times is still little known.

An ice cap situated approximately in the middle of the southern part of Northwest Iceland reached down into the inner parts of fjord and valleys in southern Ísafjarðardjúp and northern Breiðafjörður, respectively. Mountainous peninsulas and promontories along the west coast of Northwest Iceland carried at that time numerous corrie and valley glaciers reaching into the sea when RSL was situated at 40-50 m a.s.l. (*Norðdahl and Pétursson, 2005*).

The Holocene: The Lateglacial and early Holocene marine fauna was similar to the present one. However, arctic species like *Portlandia arctica* lived at glacier snouts in fjords. Very few plant bearing deposits (peat or lacustrine sediments) have been found in Iceland from late glacial time. The oldest organic sediments so far found are of Alleröd age and contain a flora similar to the present one with herbs and Graminids. Deposits from NE-Iceland also show low values of *Betula*. About 8500 BP *Betula* spread over the whole country. Several hypotheses have been presented on the origin of the Icelandic flora which contains about 450 species of vascular plants of which over 98% are of European affinity. One of the hypotheses assumes that one half of the flora survived the last glacial in ice-free areas mainly in Northwest, North and East Iceland. About 25% were introduced subsequent to the settlement by cultivation and occasionally like weeds by man and the remaining part through oceanic currents, winds and birds. Similar hypotheses have also been put forward on the origin of the other biotas. Other hypotheses assume that the flora and other biotas have been introduced by drift ice, ocean currents and winds. Recently confirming evidence has been presented on the existence of ice-free areas from last glaciation on a mountainous peninsula between two main fjords in NW-Iceland. Similar conditions may well have existed in numerous places in NW-, N- and E-Iceland and on the outer shelf of the north coast. From late Boreal up to the arrival of man in the late 9th century AD there were no significant changes in the flora except for ecologic changes induced by climatic changes. The only forest forming tree was *Betula* which appeared in W- and S-Iceland 9000 years ago. According to the only macrofossil analysis performed in South Iceland, the appearance of *Betula* occurred in Lómatjörn, Biskupstungur, at 8000 B. P. *Betula* seems to have existed at least during the late Preboreal in Northeast Iceland (Flateyjardalur). Biostratigraphically the Holocene can be divided into the *Betula* free period up to 9000 BP in S- and W-Iceland, the older birch period (6000-8500 BP), the older bog period (4000-6000 BP), the younger birch period (2500-4000 BP) and the younger bog period after 2500 BP. A very sudden climatic deterioration took place 2500 years BP; glaciers descended from high mountains and the birch forest declined rapidly . Probably many of the present ice caps began to form at this time too. Very soon after the beginning of the settlement of Iceland a remarkable floristic change took place. The birch forest was suddenly cleared whereupon a very rapid soil erosion began and cultivated plants and weeds turned up. The late glacial and Holocene volcanism continued mainly in the Neovolcanic zone and on Snæfellsnes. On the average there has been one eruption every five years. The volcanicity has been variable and nearly all kinds of volcanoes have been active, although fissure eruptions have been most common. Postglacial lava flows cover 10,000 km^2 (10%) of the country and the overall volume amounts to 400 km^3. Petrologically, the Holocene volcanic products are mostly basaltic - 90%, but 10% are andesitic and rhyolitic.

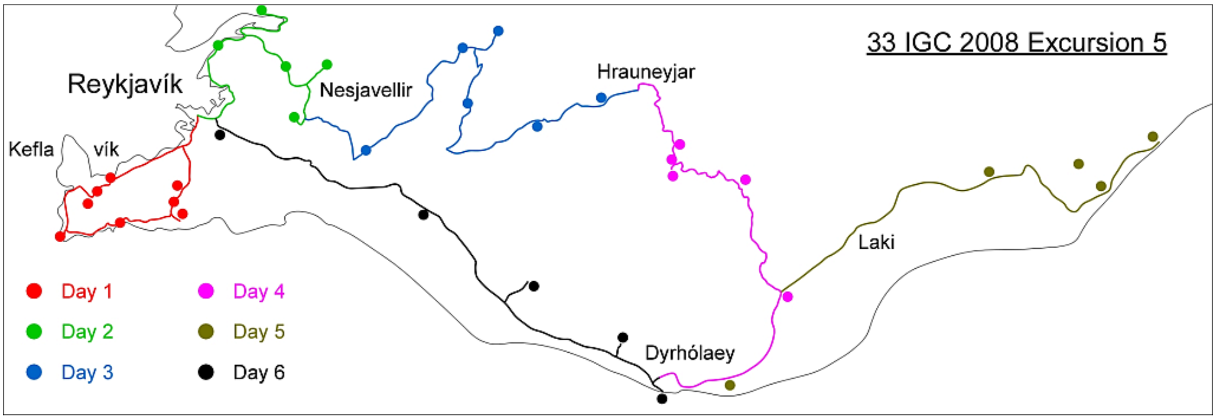
Icelandic soils are bogs or derived from volcanic material either bedrock or loose volcanic material, tephra. Rhyolitic tephra and rock fragments account for some 30% of the soil grains of the tephra loess, which is the main Icelandic dry soil.

The tephra loess is mainly composed of dark brown and altered brown volcanic glass derived from volcanic eruptions and erosion of palagonitic bedrock. In many cases the material has been transported by wind and water into areas far from its origin during the soil formation time. Accumulation of soil has been relatively slow from the beginning of the deglaciation up to the time of the Hekla-3 eruption in 2800 BP. In the latter part of that period the climate was considerably better than at present and vegetation vovered larger areas than at any other time during the Holocene.

The soil accumulation rate doubled from that time up to the Landnám (870 AD) due to deteriorating climate, from 0.04 mm/year to 0.07 mm/year. Since the beginning of the settlement, the accumulation rate increased greatly and is in many areas 0.5 mm/year after that. Since the landnám at least 50 % of the vegetated land has become barren and at least half of that area has lost most of its soil completely (*Eiríksson et al, 1994*).

3. Day by day descriptions

In the on forehand programme and even in this Excursion Guide we have listed a number of geologically interesting sites to which we intend to lead you. Anyway, it will depend on weather conditions, your enthusiasm and the length of discussions at site, whether or not we will manage all the listed sites.



The first day of the excursion will take us clockwise around the Reykjanes peninsula visiting between 6 and 8 sites of geological interest. The day starts and ends in Reykjavík. On the second day we will leave Reykjavík for Hvalfjörður north of the capital, around Mt. Esja to Þingvellir and Nesjavellir. The third day will be spent in the upper regions of the South Icelandic Lowlands visiting Geysir and Gullfoss on our way towards the interior of Iceland and Hrauneyjar. During the fourth day of our excursion we will cross the interior behind the Mýrdalsjökull glacier contemplating different kinds of volcanism and finishing the day on the south side of the glacier. The fifth day will be spent on the south side of the Vatnajökull ice cap with the main emphasis on glaciers and and glacial geomorphology. On the final day we will travel along the south Icelandic coast again with emphasis on glacial geology and Holocene volcanism ending the day in Reykjavík and departure the next day.

3.1 Day 1: Around the Reykjanes peninsula

During this day we will travel from Reykjavík to Krýsuvík and along the southern coast of the Reykanes peninsula to Grindavík. From the tip of the peninsula we will travel to its the northern coast and on our way towards Reykjavík.

En Route: Volcanic activity on the Reykjanes peninsula has been intense in Postglacial times. The number of eruptions is not known but it totals to a few hundreds. The volcanic activity is more or less restricted to the active fissure zones and it seems to be periodic. The time laps between periods is about 1000 years but each period lasts for about 300-400 years. During each period all the fissure swarms are active and it seems that the activity starts at one end of the peninsula and moves to the other. The last period started in the 10th century and lasted to about 1340 AD. The first eruptions took place at the eastern end (in Hellisheiði and Bláafjöll) but spread to the west. Eruptions within each volcanic system behave in similar manner as in the Krafla Fires 1975-1984, i.e. rifting episode which lasts for some years or a few decades, accompanied by a few or numerous eruptions. It seems at each time only one fissure swarm is active. The eruptions which we have been able to establish from historic accounts and by geological mapping and radiocarbon datings are as follows: 950-1000 AD in the Brennisteinsfjöll and Bláfjöll swarm, 1151 and 1188 AD in the Krýsuvík and Trölladyngja swarm, 1210-1240 AD in the Reykjanes swarm and in 1340 AD in the Brennisteinsfjöll swarm. Numerous volcanic eruptions have occured off the Reykjanes peninsula throughout the centuries and the last confirmed one in the year 1926 (*Sigurðsson et al, 1993*).

On the outskirts of Hafnarfjörður are the only historical lava flows within the urban limits of greater Reykjavík. The one closer to Hafnarfjörður is called Hellnahraun and was formed by an eruption at the Tvíbollar cone row in about 950. The vents are out of sight, at a locality called Grindaskörð, and are within the Brennisteinsfjöll volcanic system. Thus, the lava travelled 17 km before coming to halt here at Hvaleyrarholt, flowing across the Krýsuvík volcanic system in the process. It is a pahoehoe lava and it features exceptionally well preserved surface structures, such as tumuli and lava-rise pits, indicative of lava inflation. The tumuli are easily recognizable, as mounds bounded by raised crustal slabs and dissected by a central cleft. The other one is Kapelluhraun, also known as Nýjahraun, and is derived from a cone row located at the foothill of Undirhlíðar, about 7.5 km directly east of the main road. Kapelluhraun is an aa lava flow and was formed by an eruption in the twelfth century during a major volcanotectonic episode on the Trölladyngja volcanic system (*Þórðarson and Höskuldsson, 2002*).

3.1.1 ***Sveifluháls*** (*Hyaloclastite ridge and subglacial eruption*) – When magma reaches the surface underneath a glacier, it melts a cavity in the ice above the volcano. As long as the water pressure is sufficient pillow lavas accumulate in the hollow and if the eruption ceases at this stage a *pillow lava ridge* is formed. If the eruption on the other hand continues, the water pressure decreases and pillow lava can not form. The eruption then changes to being phreatic or explosive and tephra, either pumice, ash or breccia piles up on top of the pillow lava. If the eruption stops at this stage the result is an *hyaloclastic ridge* (móberg ridge) of which Sveifluháls is a good example (*Einarsson, 1994*).

3.1.2 ***Krýsuvík*** (*High temperature geothermal area*) – High temperature areas occur only in active volcanic zones and probably mainly where there are shallow magma chambers or intrusions cooling underneath inactice volcanic areas. Thus high temperature areas are localised areas of upward ground water flow in active volcanic areas. There are mainly two types of edifice in high temperature areas, *fumeroles* and *solfataras*.

Solfataras are common in high temperature areas. They contain blue-grey boiling clay, the colour being due to sulphur compounds of iron which form when sulphuric acid dissolves the rock, the acid forming when hydrogen sulphide (H_2S) reacts with atmospheric oxygen. In the solfataras the clay splashes as gas bubbles burst and the clay builds up a rim around them. The rocks of high temperature areas are usually greatly altered (*Einarsson, 1994*).

En Route: Continuing onwards towards Kleifarvatn. About 1 km down the road is the Grænavatn, the largest explosion crater of a small maar volcano complex that features eight craters in total. The Grænavatn maar is about 300 m in diameter and 44 m deep. The stratigraphy in the inner wall of the maar shows that the eruption began with powerful explosions ejecting tephra and fragments of the country rock. The event culminated with an effusive eruption, producing the small lava flow that caps the tephra sequence. The Grænavatn maar is well known for olivine-gabbro xenoliths, and their occurrence in the tephra deposits suggests that the activity resulted from deep-rooted explosions, perhaps as deep as 2-3 km (*Þórðarson and Höskuldsson, 2002*).

3.1.3 ***Eldborg at Geitahlíð*** (*A spatter ring crater*) – The crater was formed when lava of low viscosity was erupted from a circular vent or short fissure in rather short lived eruption with little or no lava fountain activity. Steep crater walls of very thin lava crust were formed around the vent when lava splashes from the boiling lava lake, while a very low dome of lava surrounds the volcano. Eldborg at Geitahlíð was most likely formed in late Holocene times but before the Norse settlement of Iceland (*Einarsson, 1994*).

En Route: In passing the southern end of the móberg ridge Vesturháls, one enters the rugged lava flowfield of Ögmundarhraun, renowned for the ruins of the old Krýsuvík homestead laid to waste by the lava in the twelfth century. After the eruption, the homestead was rebuilt farther to the east at Bæjarfell. It is also one of the few lava flows in Iceland that takes it name from a person. The folklore tells the story of Ögmundur "*the Berserk*", who cleared a passage-way across the lava for the squire at the Krýsuvík estate, and upon completion of the "*road construction*” he was to be rewarded by marriage to the squire's daughter. However, the squire had no intention of giving his daughter away to the peasant and had Ögmundur killed just as he was about to complete the job. Since then the lava has been known as Ögmundarhraun.

Ögmundarhraun is one of several lava flows formed in the volcano-tectonic episode, known as the Krýsuvík Fires, that raged within the Trölladyngja volcanic system between 1151 and 1188 AD. The lava flows, including Ögmundarhraun in the south and Kapelluhraun in the north, cover 36 km^2 and were produced by a series of volcanic fissures trending north-east for 28 km across the peninsula (*Þórðarson and Höskuldsson, 2002*).

3.1.4 ***Hrólfsvík*** (*Gabbro xenoliths*) – Sometimes volcanic products contain rock fragments which have broken off the feeder conduit deep in the earth. Such fragments are often quite different from the volcanic rocks in which they occur and are either found lying loose in the cinder or coated in solidified magma, or incorporated in the lava. They are called xenoliths or lithics. In Iceland xenoliths composed of gabbro and granophyre are common and often occur where such rocks are not found on the surface near the volcano (*Einarsson, 1994*).

A basaltic lava east of the village Grindavík is of postglacial age, between two and ten thousand years old. The best exposures in the lava are along the shore where the sea has eroded it into the core of the lava and numerous gabbroic xenoliths can be observed. Inland it is not possible to observe the xenoliths due to little erosion.

3.1.5 ***Reykjanestá and Stampar*** (*Submarine and terrestrial eruptions*) – The Reykjanes ridge comes at shore at the tip of the Reykjanes peninsula. The geology is characterized by recent lava flows and open tensional fissures. At this place we are in the central part of the Reykjanes fissure swarm. At the shore we find remnants of a phreatic crater from the year 1226 which formed the Medieval tephra layer. The pinnacle ”*Karlinn*”, just off the tip of the peninsula is also a remnant of this crater. Part of the crater has been overrun by lava flows from the same eruption. In a small cliff a feeder dike is exposed that can be traced to a spatter cone formed in that year. At the horizon towards the southwest the island of Eldey is seen. It was formed in a submarine eruption in the year 1210 or 1211. At the island is one of the largest gannet colonies in the word (*Sigurðsson et al, 1993*).

The annals also mention eruptions at Reykjanes in 1231, 1238 and 1240 AD. The last eruption made the Sun appear "*red as blood*”, which implies volcanic plumes rich in sulphuric aerosols, and atmospheric perturbations on a regional scale.

Such plumes can be generated by effusive basalt eruptions, provided that the volume of erupted magma is reasonably large. The only historical eruptions that fit the requirements of correct setting, age and size are those that produced the Arnarsetur and Illahraun lavas north of Svartsengi, some 18 km to the northeast of Reykjanes. Tephrochronology shows that these two lava flows were formed several years after the Medieval tephra fell in 1226 AD and thus may represent the final episode of the Reykjanes Fires (*Þórðarson and Höskuldsson, 2002*).

3.1.6 ***Stapafell*** (*Pillow lava*) – In an active rock quarry we can see into the base of a submarine volcano that probably was active during the Weichselian glaciation. Stapafell is composed of strongly olivine phyric pillow lava and associated volcaniclastic deposits. Extrusion of magma formed pillow lava above the eruptive vent or fissure. Decreasing confining hydrostatic pressure permitted steam explosions due to interaction of water and magma, with magma fragmentation during phreatomagmatic eruptions, producing thick tuff layers.

3.1.7 ***Rauðamelur*** (*Weichselian deposits and interstadial lava flows*) – The Rauðamelur sedimentary sequence comprises two stadial phases with expanding glaciers and deposition of tills, and two interstadial phases with raised RSL and accumulation of littoral and sublittoral lee-side spit formations. These sediments rest on a striated bedrock suggesting that the outermost part of the peninsula was overridden by glaciers flowing towards the northwest (W35°N). A dated sample of a whalebone collected from the lower spit formation has yielded a finite age of 34,735±1400 ^{14}C yr BP. These sediments are partly capped by a lava flow – the Rauðamelshraun lava – erupted when RSL had regressed to a position below the sediments. Later, the area was again overridden by glaciers advancing in a northwesterly direction (N40°W) and depositing till on top of the lava flow and the lower spit formation. The till is in turn discordantly overlain by the upper spit formation representing a general ice retreat and transgression of RSL. The sediments of Rauðamelur have subsequently neither been overridden by glaciers nor have they been inundated by the sea. Three dates from marine shells yielded a mean age of 12,325±85 ^{14}C yr BP and a single date of 12,635±130 ^{14}C yr BP have produced a weighted mean age of 12,355±80 ^{14}C yr BP for the formation of the upper spit forma-

tion. At that time RSL was approximately on a level with the spit formation now situated at 20-25 m a.s.l, some 40-45 m below the 70 m marine limit (ML) shoreline just north of the Rauðamelur site. This area has tectonically subsided by some 50 m in Postglacial times, explaining the different height of Rauðamelur and an assumed synchronous Bölling marine limit shoreline at about 70 m a.s.l. at Vogastapi (*Norðdahl and Pétursson, 2005*).

3.1.8 ***Vogastapi*** (*Interglacial lava flows, fissures and plate boundaries*) – Vogastapi are the remnants of a shield volcano probably formed during the Eemian, the last Pleistocene interglacial. The morphology of Vogastapi indicates that the centre of the former shield volcano was situated somewhere in the sea not far north off its most elevated part. Glacial striae on Vogastapi show that during the deglaciation of the area the ice was flowing towards the north. Just prior to the Last Glacial Maximum (LGM) the area just north of the town of Keflavík was overran by a westwardly (W5°S) moving glacier forming a basal till containing sedimentary clasts with fragments of *Hyatella arctica* obviously from the bottom of Faxaflói. Similar sedimentary clasts have been found in wave-washed basal till on top of striated (S35°W) bedrock about 6 km west of Keflavík. A radiocarbon date of these shell fragments has yielded an age of 24,145±200 ^{14}C yr BP. This date and other six dates from Njarðvíkurheiði, yielding a weighted mean age of 22,070±90 ^{14}C yr BP, pre-date the LGM glacier advance across and beyond the present coastline of the outer Reykjanes peninsula. An undated shoreline at 70 m a.s.l. on Vogastapi has been related to the general Bölling marine limit in Iceland (*Norðdahl and Pétursson, 2005*).

The historic (about 1226 AD) Arnarsetursrhaun south of Vogastapi, is derived from a 400 m long crater row trending N40°E, and has flowed over much older but postglacial lava flows that are highly fractured by the rifting process that affects most of the Reykjanes peninsula (*Sigurðsson et al, 1993*).

En Route: Hrútagjá and Þráinskjöldur are abutting half shields with their summits at 200-240 m above sea level and tugged onto northwest slopes of the móberg mountain range. They produced pahoehoe lavas that spread radially northwards and form the 20 km long coastline between Straumsvík and Vogarstapi, some 10-11 km from the source vents. The Hrútagjá shield lavas formed some of the largest tumuli found in Icelandic lavas, whereas Þráinskjöldur is heavily dissected by fissures and faults formed by rifting on the Reykjanes volcanic system. These faults are easily seen from the road in the form of northeast-trending linear scarps on the lower slopes of the shield. Farther up slope is the pyramid-shape móberg cone Keilir, an island amid an ocean of lava.

The third and westernmost lava shield, Sandfellshæð, has a more regular shape and covers a large portion of the southwest corner of the peninsula. It is an exceptionally gently sloping shield (<3°) that rises to 90 m above sea level at the summit and its lavas cover 120 km^2. Sandfellshæð has a well formed summit crater, 450 m wide and 20 m deep, which is located about 4-5 km east of the road.

All three shields were formed at the beginning of the Holocene and their cumulative volume is 15 km^3, accounting for more than 75 per cent of the volume of magma erupted onto the surface in this sector of the peninsula in the past 10,000 years. They are also the principal agents responsible for raising the area out of the sea.

3.2 Day 2: Reykjavík – Hvalfjörður – Nesjavellir

During this day we will travel from Reykjavík to Hvalfjörður and Kiðafell on its southern coast where we will have a discussion about interbasaltic sediments and Lateglacial delta formation. From there we will continue Miðsandur to become acquainted with an old central volcano on the north coast of Hvalfjörður. In Þingvellir plate tectonics and continental spreading will be the main issue and in Nesjavellir we will discuss volcanism, geothermal heat and its utilisation.

En Route: The route to Kiðafellsá from Reykjavík passes below the slopes of Mt. Esja (914 m), which dominates the horizon north of Reykjavík. Mt. Esja is the result of Pliocene-Pleistocene volcanism, and contains basaltic lavas as well as gabbroic intrusions and rhyolite, numerous basaltic dykes, clastic (volcanic and glacial) sediments. The Esja succession covers about 1.3 million years, between 3.1-1.8 million years ago. The mountain got its present shape as a result of repeated episodes of glacial erosion during the Pleistocene. The western part of Mt. Esja is an eroded flank of the Kollafjörður caldera volcano, which was active 3-2 million years ago. At the base of the mountain there are basalts and dolorite dykes, but higher in the strata horizons of palagonite (formed by supglacial eruptions) and tills appear. The eastern part of Mt. Esja and Móskarðshnjúkar (rhyolite) are the remains of a younger volcanic centre, the Stardalur central volcano, actice after 2 million years ago.

Reconstruction of isobases for the ML shorelines in the Reykjavik area defines a plane rising towards the northeast with a gradient of about 1.15 m/km. Two samples of whole shells, plates, and fragments of *Balanus balanus* collected from a sandy diamicton at about 20 m a.s.l. at Helgafellsmelar have been dated to 10,215±90 and 10,415±115 ^{14}C yr BP, respectively with a weighted mean age of 10,315±105 ^{14}C yr BP. These samples date the formation of an ice-contact delta graded to about 60 m a.s.l. and a glacier margin in the mouth of the Mosfellsdalur valley, and the ML shorelines in the Reykjavík area. Prominent raised beaches at altitudes below the ML shorelines have been mapped at about 20 m a.s.l. in Hafnarfjörður, at 25-30 m a.s.l. in Garðabær and Kópavogur, at 20-25 m a.s.l. at Austurströnd on the Reykjavík peninsula, and at about 40 m a.s.l. at Helgafellsmelar and Álfsnes northeast of Reykjavík. A single sample of marine shells collected from about 15 m a.s.l. at Helgafellsmelar and two samples from beach deposits reaching 20-25 m a.s.l. at the Austurströnd locality have yielded ages between 9815±150 and 9925±210^{14}C yr BP and a weighted mean age of 9865±170 ^{14}C yr BP. Fossiliferous littoral and sublittoral sediments reaching some 10-15 m a.s.l. at Reykjavíkurflugvöllur airport are related to raised shorelines at about 20-25 m a.s.l.

and dated to about 9895±125 ^{14}C yr BP, a weighted mean age of 4 samples. The 7 Preboreal dates, averaging to about 9875±90 ^{14}C yr BP, are regarded as an approximation to the age of the lower set of prominent raised shorelines found below the ML shorelines in the Reykjavík area. Since the formation of the lower set of shorelines at 9875±90 ^{14}C yr BP, RSL was apparently lowered to a position at least 2.5 m below present sea-level at about 9400 ^{14}C yr BP. The lowermost position of RSL at about -30 m was apparently reached at about 8500 ^{14}C yr BP (*Norðdahl and Pétursson, 2005*).

3.2.1 ***Kiðafellsá*** (*Late glacial and intrabasaltic sediments*) - For getting a glimpse of the Pliocene strata, we will visit a section at the base of Mt. Esja at a site called Kiðafellssá, where clastic sediments of glacial nature outcrop close to sea level, and have a discussion on the origin of the sediments.

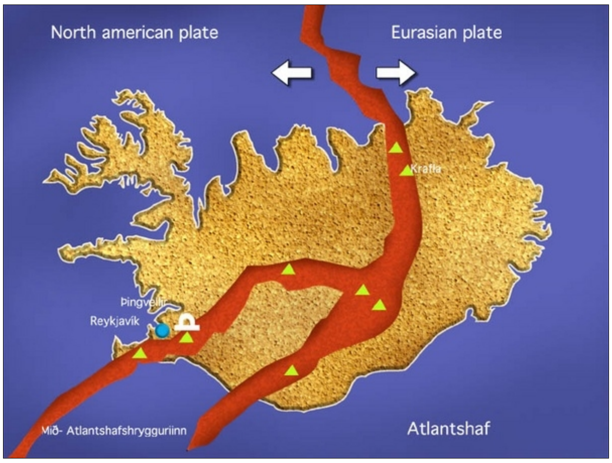
Approximately 120 m of tholeiitic and olivine tholeiitic lavas separate the two earliest diamictite units in Hvalfjörður. The younger unit forms an impressive cliff along the coastline of the fjord Hvalfjörður, and is interpreted as a localized succession of flow units based on its overall stratigraphic framework and composition. The diamictite which in places rests on stratified fine-grained sandstone and conglomerate, displays a facies association consisting of massive, matrix supported diamictite overlain by clast supported diamictite intercalated with finer grained waterlain sediments and sediments of fluvial origin. The overall appearance of this section suggests continuing sedimentation with occasional interruption of single lava flows. Pebble fabric from this exposure is inconsistent and in general shows a unimodal pattern. Post-depositional tectonic and igneous activity is indicated by basaltic dykes and mud dykes that cut through the diamictite (*Geirsdóttir, 1991*).

Two samples of marine shells, yielding Böling ages of 12,335±140 and 12,375±140 ^{14}C yr BP, have been collected at low altitudes in sediment terraces at Ósmelur and Kjalarnes towards the west. The existence of shells in these sediments shows that the southern coast of Hvalfjörður had been deglaciated and submerged by the sea at that time. These terraces have subsequently not been overridden by glaciers (*Norðdahl and Pétursson, 2005*).

3.2.2 ***Miðsandur*** (*An extinct central volcano*) – Between Miðsandur and Ferstikla is a central volcano, the Hvalfjörður Central Volcano that was active in the period between 3.0 and 2.5 Ma. Strata belonging to the central volcano have been traced underneath younger strata towards the west along the coast of Hvalfjörður and to Skarðsheiði and Botnsheiði in the north. In an old rock quarry above Miðsandur, not far from the centre of the volcano, where rhyolite has been mined for silica (SiO_2) and production of cement, we will be able to see an acid intrusion, perhaps the top of a magma chamber fill, as well as basaltic dykes.

En Route: Kjósarskarð (A transect through Tertiary and Quaternary bedrock formations) is on the way from Hvalfjörður to Þingvellir. We will be able to see a transect through Tertiary and Quaternary bedrock formations from the bus as we pass through Kjósarskarð. The strata to the north and west is regular flood basalt sequence with random dykes, which gives way to a more chaotic bedrock pile of lavas, palagonite and tills as we enter the Pleistocene Móberg Formation.

3.2.3 ***Þingvellir*** (*Plate boundary, volcanism and history*) – Þingvellir area is a part of the North Atlantic rift system, almost entirely nested within the Reykjaneshryggur-Langjökull rift system. It can be described as an area of sea-floor spreading, displaying the close association of crustal rifting and volcanism. Þingvellir and the Great Rift Valley of Eastern Africa are the only sites on Earth where the effects of two major plates drifting apart can be observed.



The rift zone constitutes a graben bounded by major faults. The rift valley narrows from about 20-25 km in the NE to about 10 km in the SW. The extension appears as nearly parallel fissures and down-dropped fault blocks running along the length of the valley. On the west (North American) side of the valley, the blocks step down toward the east, while the situation is reversed on the east (European) side. Thus the valley floor is a linear block that has subsided as the valley walls were pulled apart by plate motions. The valley walls are moving apart at a rate of about 7 mm per year, and during the past 9000 years the estimated horizontal extension is of the order of 70 m. The floor is subsiding at about 1 mm per year, with a total subsidence of 40 m for the past 9000 years. Rifting within the graben is episodical, with the last major earthquake activity occurring in 1789. During the 1789 earthquakes the graben floor subsidised 1-2 m.

Almannagjá is 7.7 km long. Its greatest width is 64 m, and its maximum throw is 30-40 m. It marks the eastern boundary of the North American plate. Its equivalent across the graben, marking the western boundary of the Eurasian plate is Hrafnagjá. It is 11 km long, 68 m wide and has a maximum throw of 30 m. The Þingvellir faults are believed to be the surface expressions of deeply rooted normal faults. The numerous fissures encountered on the valley floor are of similar origin.

The bedrock of the Þingvallavatn catchment consists mostly of postglacial lavas that are most extensive in the central part of the graben and Late Pleistocene pillow lavas and hyaloclastites of subglacial origin and sub-aerial lavas. The Þingvellir graben is surrounded by volcanoes, that illustrate the connection between rifting and volcanism. Postglacial lava flows that measure about 30 km^3 have flown into the graben, and fissure eruptions within the graben have left their marks.

The Þingvellir lava (*Þingvallahraun*) floors the northern part of Lake Þingvallavatn and the graben floor north and east of the lake. It originated in a major fissure eruption to the southeast of Hrafnabjörg, around 9100 years ago. The many single flows of this lava are best exposed in the fault scarp of Almannagjá, where numerous sheets of individual lava lobes have been successfully stacked as the eruption progressed.

Mt. Skjaldbreiður (1060 m) is a huge shield volcano that dominates the horizon to the north of the Þingvellir graben. It was formed during a prolonged eruption about 9000 years ago. Apart from constructing the huge shield of Mt. Skjaldbreiður filling the northern part of the graben, the eruption caused a number of lava flows to enter the southern reaches of the graben, partly overflowing the somewhat older Thingvellir lava. Mt. Skjaldbreiður is the prototype of Icelandic shield volcanoes, due to its regularity. The angle of slope is ca. 1° in the distal parts, steepening to 9° near the top. The volume of Skjaldbreiður has been calculated to be about 17 km^3, and the lava covers an area of about 200 km^2. Presumably Skjaldbreiður formed during one eruption, perhaps lasting for 50-100 years.

A third major eruption occurred about 2000 years later, some 7000 years ago, from the Eldborgir fissures. It produced a lava sheet that covers the eastern part of the Þingvellir lava. Together, these three eruptions (Hrafnabjörg-Skjaldbreiður-Eldborgir) filled in the graben and changed the size and shape of Lake Þingvallavatn that occupies the southern deep part of the graben. Þingvallavatn is the largest lake in Iceland, about 82 km^2. With a maximum depth of 114 m, it descends about 10 m below the present sea level. It originated as a glacial lake, during the last deglaciation about 10,000 years ago. The subsequent volcanism stopped any surface runoff from the highlands to the north of the graben towards the lake. The continuing subsidence in the graben has gradually led to the present size and depth of the lake over a period of 9000 years.

The Þingvellir graben is closed towards south by two volcanic systems, the Hengill and Hrómundartindar systems. There have been 4 small eruptions within the fissure swarm north of Hengill in postglacial time: Stangarháls (7500 years old); Hagavíkurhraun (5000 years old) and Nesjahraun/Sandey (2500 years old). The Sandey crater cone formed over a fissure extending into Lake Þingvallavatn, and is the only eruption that has occurred within the lake proper. Hrómundartindar is an active volcanic system east of Hengill, with one known eruption in early postglacial time.

The Hengill Central Volcano: The Hengill volcano is an irregularly shaped central volcano situated some 25 km east of Reykjavík. Most of the volcano itself is built up during the last glacial, and consists of basaltic subglacial volcanics, mostly hyaloclastits and pillow basalts. Only one occurrence of rhyolitic volcanics is known in Hengill. The volcano is extensively transected by northeast-southwest trending faults, in accordance with the regional extensional environment. A volcanic rift zone extends southwest and northeast from the volcano. The rift zones are extensively faulted, forming complex graben structures. The graben structure southwest of the volcano is filled with postglacial lava flows, but towards the northeast the graben is readily observable in the Nesjavellir area, as well as in the Þingvellir area farther north, one of the world's most recognizable graben structures and plate boundary. In postglacial times, three basaltic fissure eruptions have occurred in the Nesjavellir area and with two of then erupting on both sides of the volcano, but not extending through it. They are the Stangarháls lava, Hagavíkurhraun and Nesjahraun. Their age is approximately 7500, 5000 and 2500 years old. The Hengill geothermal field is the second largest in Iceland, following the one at Torfajökull, although only minor surface geothermal manifestations are observable from the Nesjavellir power plant. Two geothermal power plants are in operation in the Hengill area, the Nesjavellir power plant and the Hellisheiði power plant. Two more power plants at Hengill are on the drawing board, at Bitra and at Hverahlíð.

The Nesjavellir geothermal field is a high enthalpy geothermal system within the Hengill central volcano in Southwest Iceland. Geothermal investigations at Nesjavellir commenced in 1946, however, it was not until 1986 that a decision was made to harness the geothermal heat for district heating in Reykjavík. In 1990, following the drilling of a total of 18 wells, the Nesjavellir power plant was commissioned, generating about 100 MWt, by producing 560 l/s of 82 °C hot water for district heating. Initially only four geothermal wells were connected to the plant, but gradually more wells have been drilled and connected, as the capacity was expanded. To date, 25 wells have been drilled and 15 are in full use. In 1995 the capacity was expanded to 150 MWt and in 1998 to 200 MWt and the production of 60 MWe of electricity commenced. Later expansions occurred in 2001 and 2005. Presently the plant produces 120 MWe and 280 MWt.

3.2.4 ***Nesjavellir geothermal plant***: The Nesjavellir co-generation power plant is owned and operated by Reykjavík Energy (Orkuveita Reykjavíkur). An extended stop will be made at the Nesjavellir power plant, where we will be informed about the plant's operation and the geological characteristics of the Hengill volcano. Reykjavík Energy has also invited us to an informal reception at the plant, where we will be exposed to light food and an assortment of beverages.

Plant operation*:* The high enthalpy geothermal field at Nesjavellir is a two phase system with both steam and water coming from wells. The depth of the wells range from 1000 to 2200 metres, and temperatures in the production zone is 320-360 °C. The characteristics of the wells vary, initially the enthalpy of the fluid ranged from 1500-2600 kJ/kg, but the enthalpy has since been influenced by utilisation. The co-generation power plant at Nesjavellir has two func-

tions. The first function is to produce electricity with the geothermal steam. The second function is to heat cold groundwater for district heating. From the wells the geothermal fluid is brought to a separation station where the two phases are separated. The water and steam are piped separately to the power station, but excess steam is released into the atmosphere through a high chimney by a control valve which maintains a constant pressure of 12 bars in the steam supply system. A similar system controls the separate water supply to the power station. The excess water boils to atmospheric pressure after the control valve, and the steam formed is released into the atmosphere. The effluent water is discharged into a nearby stream.

Electricity is generated by four steam turbines, each 30 MW, requiring 240 kg/s of steam in total at a pressure of 12 bar. The steam is condensed in a tubular condenser and cooled to approximately 55 °C with cold groundwater. The condensate waters are disposed of in shallow wells in the nearby lava field (210 l/s) and in reinjection wells (30 l/s). The cooling water is pumped from a shallow fresh water aquifer in the lava field 6 km away from the power plant. Over 2000 l/s is required for the condensers. The cooling water is heated to 55 °C in the condensers, and then piped through heat exchangers for final heating to 87 °C, using 192 °C hot geothermal water from the separators. In the heat exchangers the geothermal water is cooled to 55 °C, and finally discharged into reinjection wells (130 l/s). As the original pipeline to Reykjavík cannot carry all of the heated water produced at Nesjavllir, a cooling tower has been built to cool part of the excess heated waters before discarding them in shallow wells or into the surface stream. The heated water is later boiled under vacuum conditions in deaerators to remove dissolved oxygen to prevent corrosion in the steel pipes. The final treatment before the water is pumped to Reykjavík for district heating is to inject some geothermal steam into it, both to remove the last traces of dissolved oxygen (it reacts with H_2S) and to adjust the water's pH to 8,5.

***Future plans**:* The Nesjavellir geothermal field is considered to be fully exploited and there are no plans on the table for further expansion. Present plans are to continue running the plant for the unforeseen future (30 – 50 years?) and to do that it is estimated that an extra well will have to be drilled every 5 years to compensate for pressure drop in the geothermal system. Other plans include adding to the cooling tower and drilling more reinjection wells. This is part of a plan to minimize chemical/thermal pollution derived either from separate waters, condensate water or excess production water (heated cold water). A general plan on how to reinject uncondensable gases (mostly CO_2 and H_2S) is also being considered.

3.3 Day 3: Nesjavellir – South Icelandic Lowland – Hrauneyjar

On the third day of our excursion we will travel from Nesjavellir and across the upper part of the South Icelandic Lowlands to Hrauneyjar on the south edge of interior Iceland. Our first visit will be lavas emitted from the Seyðishólar craters and from there we will continue to the geothermal Geysir area and the Gullfoss canyon to have a look at bedrock stratigraphy and the products of subglacial volcanism. In Kirkjuskarð volcanic breccia and interbasaltic sediments are used to demonstrate subarial and subglacial environments – cyclic glacial and interglacial environmental conditions. Holocene volcanism and fros crack polygons will then finish of our day shortly before arriving at Hrauneyjar.

En Route: The Þingvellir lava is presumably derived from the shield volcano Skjaldbreiður to the north of Þingvellir. It is probably slightly younger than the Eldborgir lava shield to the north of Hrafnabjörg, which has been dated at the Sog river to 9130±260 years BP.

3.3.1 ***Seyðishólar and Grímsnes lavas*** (*Holocene volcanic area*) – The postglacial Grímsnes volcanic area forms a small volcanic fissure swarm which is made up of 12 eruptions sites with tholeiite lavas covering 54 km^2. The three largest lavas, those of the Seyðishólar, Kerhólar and Tjarnarhólar craters, are roughly contempoaneous. Vigorous explosive activity in Seyðishólar led to the formation of the Seyðishólar tephra (0.09 km^3) which forms a good marker horizon for the age determination of other lavas. The ^{14}C date of the Kerhóll lava yielded 6220±140 BP, 5000-6000 years, based on tephrochronology. The new ^{14}C date shows that the Grímsnes volcanic fissure swarm was formed about 5500-6500 BP (*Albertsson et al, 1997*).

3.3.2 ***Geysir*** (*Geothermal heat, geysers and volcanism*) – Geysir is the most famous of all spouting hot springs in the world and in many languages such springs are called "*geysers* ", the word Geysir means a gusher. But in Iceland Geysir is the name of this particular spring and the general term for a spouting spring is "*goshver* ".

Since 1750 Geysir has been visited by numerous scientists and most of the manifold theories to explain the mechanism of geyser eruption are based on observations of Geysir itself. Geysir lies among many other springs in a field of siliceous sinter deposited by the springs themselves at the eastern end of the hill Laugafjall. Most of the springs are alkaline with clear water. Geysir is by far the biggest of these springs and is surrounded by a regluar siciceous sinter dome in the top of which is a flat saucer shaped basin 8 m in diameter.

The hot springs of geysir are mentioned for the first time in connection with an earthquake in 1224. It is possible that Geysir is one of the springs which the annals mention as having formed as a result of this earthquake. On the NE side of the Geysir dome there is a cut where tephra layer H_3 from Hekla is found underneath the sinter dome. The age of the tephra layer is 2900 BP.

The first widely known theory of geyser eruption is that of MacKenzie (1810), suggesting an outrush of accumulated steam from a subterranean chamber. Bunsen (1846) postulated boiling in an open vertical shaft as the direct cause of the eruption. A model built on this principle works in the laboratory, but the mechanism is seldom realized in nature. The late Icelandic geophysicist Trausti Einarsson regarded superheating of the water in the shaft as the main cause of geysir eruption.

En Route: On our way to Geysir and Gullfoss while acrossing the Torfastaðaheiði we can observe an ancient 6.9 km long turf wall, Þrælagraður, in Biskupstungur, South Iceland, stretches from the Brúará river in the west to the Tungufljót river in the east. The turf wall is more and less submerged in a boggy landscape and in many places not visible at all at the surface. On the other hand sections of the wall were found in newly dug ditches. The age of the wall was determined by its position between two known tephra layers in five soil sections. The wall is immediately above the Landnám (Settlement) layer, which is supposed to have been formed in a mixed eruption and its age has been estimated at 850 to 900 AD. Further on along the road we can see low relief moraines and hummocky moraines most likely formed in early Preboreal times.

In the case of the great Geysir we have a 2-3 m wide cylindrical shaft extending vertically 20 m down from the bottom of the shallow basin. Water at a temperature of about 125 °C enters the shaft near or at the bottom. It does not boil, however, because the boiling point there is over 130 °C, if the water level stands in the basin. In the basin the water has cooled down to some 80 °C and circulation of this cooled water extends deep down into the shaft. At a depth of 10 m (boiling point 120 °C) the temperature comes closest to the boiling point, being often 114-118 °C. In calm, warm weather, when surface cooling is reduced, the whole water body warms up and now boiling may start in the middle section. This boiling is demonstrated by temperture measurements and also by smal vapour bubbles streaming calmly to the surface. It is normal boiling at the respective boiling point and it does not provoke an eruption. But when this has gone on for some time the rise of bubbles becomes intermittent. Superheating is now setting in (superheating by 5-6 °C has been measured) and after a longer pause in the appearance of bubbles a detonation is heard, the earth trembles and the water surface suddenly rises on or two metres in the middle of the basin. This is caused by explosive boiling of the water body (convective tongue) in the middle section of the shaft and initiates the geysir eruption.

Because of the cooling effect of the great surface area of the top basin, geysir is very sensitive to weather. Between 1916 and 1935 Geysir was quiet. When the water level was lowered in 1935, eruptions started again. After an earthquake in 1896 activity in the geysir area increased and also after the eathquakes in 2000.

There are many other hot springs in the vicinity of Geysir. Strokkur, just SW of Geysir erupts regularly and other noteworthy springs are Smiður, Óþerrishola, Sóði and Konungshver. Probably the most beautiful spring in the geysir areea is Blesi, with two basin filled with clear, calm water. One os them is bluish, the other greenish, the colouring being due to presence of colloids in the water. This was changed in the earthquakes in 2000 (*Albertsson et al, 1997*).

3.3.3 ***Gullfoss*** (*tectonics, lava pile with sediment horizons*) – he Gullfoss area lies in the middle run of the river Hvítá, where it descends from the highlands onto the Southern Lowlands. The river has eroded a deep canyon, Gullfossgljúfur, into the edge of the highland where it forms the two stepped 31 m high waterfall Gullfoss. The depth of the canyon is 40-70 m. The avrage discharge of the river Hvítá is 109 m^3/s and the largest recorded flood (March 1930) was 2000 m^3/s. The elevation of the lowlands is 100-120 m but the plateau around Hvítá lies at 200-300 m a.s.l.

The oldest rock series in the Gullfoss area is the Brúarhlöð volcanic breccia, which crops out along the river Hvítá from Brúarhlöð into the mouth of Gullfossgljúfur. The breccia is also exposed in the low mountains to the south and east of the area. The Brúarhlöð breccia consists of fine grained stratified tuff and non-sorted volcanic breccia with intercalated pillow lava and lava flows of basaltic composition. The unit was probably built up as an elongated and subglacial hyaloclastic ridge.

To the north of Brúarhlöð the "volcano" has been partly overlapped or covered by a sequence of basaltic lava flows and sediments, the Gullfoss layers. The oldest unit of this series is the sediment layer. It consists of fluvial sandstone and conglomerate and has a thickness of up to 40 m. It has been covered by 12-20 m thick lava flow, which forms the lower step of the waterfall Gullfoss and is well exposed in the canyon to the south of the waterfall. Later a 4-12 m thick sediment layer, SB, was deposited. It consists of an unsorted silty boulder layer (tillite?), laminated siltstone and fluviatile highest part of a well- to non-cemented conglomerate. Layer SB is covered by a 5-15 m thick basaltic lava flow. This layer forms the higher step of the waterfall Gullfoss. A 2.5-8 m thick foreset bedded slightly cemented conglomerate, rests on top of the upper basalt lava flow. The youngest sequence of the bedrock are the Hólar and Tungufell basaltic lava flows, covering a tillite like deposits seen in the western wall of the canyon above the higher step of the waterfall.

The bedrock in the Gullfoss area is of Pleistocene age. The Gullfoss layers are normally magnetized and probably of early Brunhes age, i.e. little less than 0.7 Ma. The Brúarhlöð breccia is reversely magnetized, i.e. of late Matuyama age. The bedrock has been tilted and faulted. The Gullfoss layers dip 2-3° WNW. The faults and fissures trend NNE-SSW to NE-SW. The most distinctive is the Dimmagróf fissure which cuts through the western part of Tungufell and appears in Gullfossgljúfur below the lower step of the waterfall, where the canyon has been cut along it. The present landscape of the Gullfoss area is glacially sculptured. Glacial striae and roches moutonnées indicate an ice movement towards the southwest. In Lateglacial time, during the Alleröd the glacier retreated from the area but during the Younger Dryas Chronozone the glaciers readvanced when relative sea level was about 100 m higher than at present because of glacio-isostatic loading. When the glaciers finally retreated, a delta was built up at the mouth of Gullfossgljúfur.

Already in early Holocene time, Hvítá occupied its present channel and began to erode the canyon partly along the tectonic pattern and partly controlled by the strike and dip. Gullfossgljúfur south of gullfoss is about 3 km long and has been eroded in Holocene time, i.e. during the last 10,000 years. The average rate of erosion therefore amounts to some 30 cm/year.

It has been suggested that the main part of the canyon was eroded in one or more jökulhlaup during the retreat of the Younger Dryas ice sheet, the water masses would then have originated from ice lakes at Bláfell to the north (*Albertsson et al, 1997*).

3.3.4 ***Kirkjuskarð*** (*Quaternary lava pile with sediment horizons*) – Rocks of Pleistocen age are mainly exposed within and bordering the volcanic spreading zones but are found also in West and North Iceland. They differ in many ways from the Tertiary rocks, mainly by their greater variety of rock facies. During the interglacial stages the volcanic activity was mainly effusive as it had been in the Tertiary and was to be in the Holocene. A doleritic texture is a lithological characteristic for the major part of the basalt. During times of glaciation when the country was covered with ice sheets volcanic products were piled up over the eruption centres as hyaloclast ridges (pillow lavas and breccias) which were capped by subaerial lava flows to form table mountains if the volcanoes were built up through the ice sheet (*Eiríksson et al, 1994*).

In a normal faul escarpment, trending north-south and with the hanging-wall block on its east side, we have three types of beds exposed. Basaltic lava flows, sediments and volcanic breccia. The uppermost basalt lava flow is tightly jointed with an eroded upper surface against a thin bed of light grey silty sediment with angular and dense basalt clasts, obviously deriving from the underlying basalt. Still, some of the clasts are with different petrology, colour and shape and most likely of different origin. The volcanic breccia overlying the light grey sediments is characterized by the great number of semi-angular porous basaltic clast, a pillow breccia in a matrix of brownish, sometimes stratified sand sized tuff.

En Route: The road towards Þjórsárdalur runs parallel to the River Þjórsá, the longest river in Iceland, stretching some 230 km from its source at Hofsjökull to the southern coast, and which has an average discharge of 400 m^3/s. Along this stretch are islets in the middle of river, which are covered by lush vegetation, because their location prevents intervention by man or livestock. The mountains and hills north of the road belong to the Hreppar Formation, a Pliocene – Pleistocene sequence built mainly of subaerial basaltic lava flows and clastic deposits of fluvial and debris-flow origin. Several distinctive tillite horizons and thick glacial lagoon deposits occur within the upper part of the sequence, indicating the onset of full-scale glaciation about 2.4 million years ago. The Hreppar succession is heavily dissected by faults, and the tectonic fabric is characterized by older north-trending strike-slip faults and younger northeast-trending normal faults.

3.3.5 ***Þjórsárdalur and Hjálp*** (*Holocene lava flow and rootless sinter cones*) – In the past 7000 years Þjórsárdalur has been repeatedly devastated by tephra fall from the nearby Hekla volcano, but never more so than by the largest Plinian eruption of Hekla some 2900 years ago.

The whitish specks on the surrounding mountain slopes are not snow but the remnants of the pumice-fall deposit from this Hekla eruption. When the first settlers arrived in Iceland in the ninth and tenth centuries, the valley had completely recovered and it soon became quite densely populated, hosting at least 20 farms. This settlement in Þjórsárdalur was decimated by the tephra fall from the 1104 eruption at Hekla. At the time, the valley was completely blanketed by a 10-30 cm thick pumice-fall deposit, which was enough to make the area uninhabitable.

The valley floor is covered by a 3500 years old lava flow, one of the youngest Tungnaá lavas, which originated from volcanic fissures in the Veiðivötn area. It is a branch of a larger lava flowfield, the Búrfell lava, which covers the high plateau to the east of Skeljafell and Búrfell. In the centre of Þjórsárdalur, the Búrfell lava contains a cluster of small scoria cones, which were formed by rootless eruptions when the lava covered wetlands that once existed in the valley. At Hjálparfoss, river Fossá has eroded a beautiful section through the lava, exposing the roots and conduits of the rootless cones, along with spectacular fanning columns formed by water-enhanced cooling of stagnant lava (*Þórðarson and Höskuldsson, 2002*).

3.3.6 ***Ferjufit*** (*Frost-crack polygons*) – Ferjufit is a flat ground just south of river Tungnaá opposite Búðarháls and 300 m a.s.l. On these flats there is a network of large scale polygons, so-called frost-craks polygons. The polygons, which in Iceland have a diameter of 10 to 25 metres, represent the very discontinuous or sporadic permafrost in Iceland, usually situade above 600 m a.s.l. and thus the 0 °C isotherm.

Frost-crack polygons were first observed in Iceland in 1954. Since then they have been observed in many places, mainly in the interiour. The lower limit of the frost-crack polygons in regolith lies near the 600 m level, but in loessian soils, especially in the thick, tephramixed soil cover near the active volcanoes, the limit is about 300 m lower. Thephrochronological studies indicate that the frost-crack polygons were probably formed mainly during the cold period between about 1550 and 1920 AD. During the cold winters in the late 1960s, cracks again opened up in many of the polygons which had been inactive since the early 1920s.

3.4 Day 4: Hrauneyjar – Fjallabak – Dyrhólaey/Laki

Short stops will be made along the most recent eruptive fissure at the Torfajökull central volcano. The stops will occur at the Ljótipollur explosion crater, the Stútshraun eruptive crater above lake Frostastaðavatn and finally at the edge of the rhyolitic Laugahraun lava at Landmannalaugar. Landmannalaugar is a camping site and has WC facilities. A more detailed description of the stops follws this more generlized section on the Torfajökull central volcano as a whole. Later when leaving Landmannalaugar the road crosses Jökulgilskvísl river and continues across disturbed móberg topography (Kýlingar – Jökuldalir). Shorelines of temporary shorelines in the bed of Tungnaá river can be seen on the slopes of the hills. A stopp will be made in Eldgjá.

Torfajökull Central Volcano – Geology and Petrology: The Torfajökull central volcano is located on the junction between two geologically and petrologically contrasting volcanic regions, the Southern Flank Zone and the Eastern Rift Zone. Immediately to the northeast is the Veiðivötn fissure swarm, which is a normal spreading ridge segment, characterised by such typical rifting features as faults, grabens, linear volcanic features and voluminous tholeiitic volcanics. To the south and west there is a group of large central volcanoes (Hekla, Mýrdalsjökull, Tindfjallajökull, Eyjafjallajökull), as well as individual fissure swarms (Vatnafjöll), that produce relatively large quantities of evolved volcanics (intermediate and silicic magmas), while the basaltic magmas are either transitional in nature or alkaline. No tholeiitic basalts occur south of Torfajökull. The Torfajökull area reflects in many ways the characteristics of both regions. It has through time, produced all the different magmatic types found in this region of Iceland and demonstrates the tectonic characteristics of both a rifting area and a non-rifting one.

The Torfajökull central volcano is a large mountainous area named after a small (15 km^2) glacier (named Torfajökull) in its southeastern part. The Torfajökull massif is elongated in a NW - SE direction, perpendicular to the strike of the Eastern Rift Zone, and has the approximate dimensions of 20 x 30 km. It is transected by narrow (4 - 5 km) NE - SW trending late glacial hyaloclastic ridges and postglacial crater rows (Veiðivötn fissure swarm), that cluster in distinct swarms north and south of it, as well as another cluster of short volcanic fissures in the far west (the Vatnafjöll fissure swarm). A large ring structure partly surrounds the central parts of the Torfajökull volcano and it has been suggested that this ring structure may represent a caldera rim. This author (Gretar Ívarsson) has generally favored a constructural origin for the ring structure, but more recent studies demonstrate that a caldera structure does indeed exist at Torfajökull.

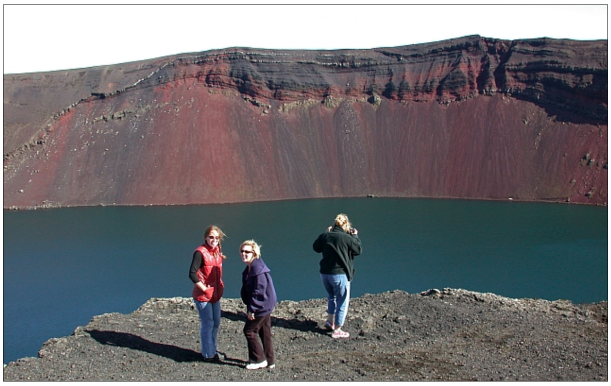
Three tectonic trends are observed in the Torfajökull region. The youngest and perhaps most prominent is the NE-SW trend exhibited by all postglacial eruptive fissures within and outside Torfajökull, and by basaltic hyaloclastic ridges and subglacial silicic mounds northeast and southwest of the area. Probably older and certainly less noticeable is a NW-SE trend observed in subglacial rhyolites and subglacial hyaloclastites northwest and southeast of Torfajökull. Finally there is the arcuate distribution of rhyolites (ring structure) around its marginal parts. The first two tectonic trends are of regional origin, but the last one is of local origin.

The Torfajökull area is the location of the single largest outcrop of rhyolitic volcanics (280 km^2) and contains the largest geothermal field in Iceland (140 km^2). It rises from a 600 meter platform to more than 1200 meters above sea level. Its southern and eastern parts are extensively eroded with broad outwash floored valleys and deep narrow gullies, some with perennial snow, a regular "badland" topography. The northwestern part is more level with conical hills, blocky lava flows and shallow stream beds. Practically no vegetation cover exists, yet outcrops are poor, mainly due to extensive talus slopes, solifluction sheets and outwash deposits, thick volcanic tephra, and snow filled gullies.

***Hot spots and propagating rifts**:* The distribution of the different magma types at Torfajökull in time and space, along with the distribution, types and volumes of magmas in the volcanic zones of Iceland in general, has suggested to this author that a revaluation of the location of the Icelandic hot spot and the nature of the propagating rift in South Iceland is a worth while investment. The established theory is that the hot spot locus is underneath the large ice cap of Vatnajökull and a propagating rift is protruding towards the southwest, its tip being at the location of the islands of Vestmannaeyjar off the coast of South Iceland. One major inconsistency in the traditional theory is that it is unable to explain the presence of 5 large central volcanoes that produce comparable volumes to the most active part of the rift (the proposed hot spot). According to the revised theory the locus of the Icelandic hot spot is located underneath South Iceland, while the Vatnajökull area and the propagating rift connecting the two regions is a thin spot. A thin spot is a hot spot simulator, in the sense that it simulates volcanic activity expected of areas overlying hot spot mantle plumes. The only difference is that the thin spot does not overlie a mantle plume, but receives its supply of partially melting mantle material by lateral flow at the lithosphere - asthenosphere boundary from a hot spot locus located some distance away.

In not so many words this idea assumes that the hot spot is constantly underlying a relatively thick crustal segment. The relatively thick crust underlying South Iceland (10 - 30 km) prevents basaltic partial melts, forming in the hot spot, to rise unmodified to the surface. Instead they cause the crust to partially melt forming silicic magmas which later act as density traps on rising basaltic magmas from below. The hot spot, therefore, constantly melts the overlying crust, forming large volumes of rhyolitic magmas in the progress, which occasionally get dumped on the surface in surface extrusions. Partially melting the crust causes it to become thinner and as time passes it is rafted away from the hot spot locus in accordance with the regional crustal movement and fresh thick crust is rafted over the hot spot to take its place and the cycle continues. Partially melting plume material, which is unable to penetrate a thick crust containing ample low density rhyolitic melts (density traps), will tend to rise towards areas of low pressure, in this case laterally and upwards towards an area of thin crust northeast of the Southern Flank Zone. There, high degree of partial melting will be enhanced, forming tholeiites, due to favourable conditions of high temperatures and low pressure. Thin crustal areas that focus this partially melting mantle material, which originally ascended underneath the Southern Flank Zone, will be areas that have experienced prior crustal thinning and extension, either in a rifting environment or the hot spot trace itself. According to this idea the Southern Flank Zone in not a propagating rift where you have a rifting environment and a melting anomaly extending into an inactive crust, but rather the location of the Icelandic hot spot which is in the process of melting its way through a thick crust, leaving behind it a region of thin crust which favours the formation of tholeiitic magmas.

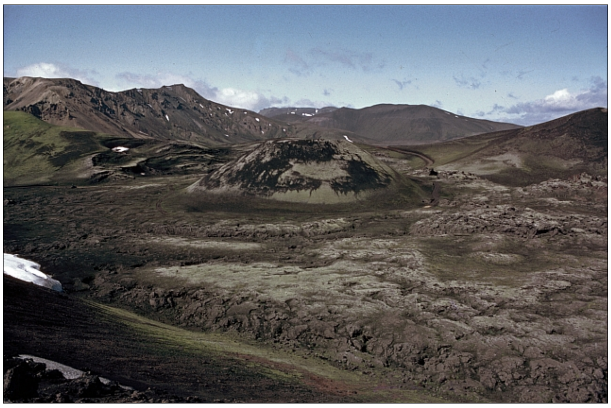
A 40 km long fissure eruption caused the formation of several lava flows and explosion craters in the region of Torfajökull central volcano and a row of spatter/tephra craters in the Veiðivötn fissure swarm. Based on tephrochronological studies this eruption took place in the year 1480 AD, albeit no written accounts survive. The eruption resulted in the formation of the rhyolitic lavas of Laugahraun and Námshraun (Suðurnámshraun), and the basaltic Stútshraun (Norðurnámshraun), Frostastaðahraun and Ljótipollur, along with the eruption of 3.5 km^3 of basaltic tephra (1.1 km^3 calculated as lava) and 0.4 km^3 of lava in the Veiðivötn fissure swarm. The volcanics produced in the Torfajökull area cover an area of 8.4 km^2 and have a volume of 0.084 km^3. Adding the products erupted within the Veiðivötn fissure swarm, a total of nearly 1.6 km^3 (calculated as lava) is reached for this eruptive event.



3.4.1 ***Ljótipollur*** (*Explosion crater*) – A beautiful waterfilled explosion crater that erupted on a 1.2 km long fissure segment through a basaltic hyaloclastic ridge. Northeast of Ljótipollur, where the eruption occurred in the streambed of the river Tungnaá, numerous craters and rootless cones are covering the area due to secondary explosions caused by the association of water and magma. The inside walls of the Ljótipollur crater consists of alternating bands of colorful (often oxidized red) tephra layers and coherent welded spatter layers that apparently flowed down the sides of the crater during the eruption. The eastern side of the crater became unstable during the eruption, sliding down to the Tungnaá river outwash plains. Excellent cross sections, excavated by the river Tungnaá, in the slide reveal coherent partly welded, columnar jointed, spatter units intermingled with variously sorted tephra. Chemical analysis of samples from Ljótipollur suggest that it is a typical tholeiite, similar to the tholeiites erupted within the Veiðivötn fissure swarm farther north. One sample from Ljótipollur is a hybrid tholeiite, suggesting a transitional icelandite component.

En Route: Frostastaðahraun is a basaltic lava flow erupted from 400 meter fissure segment southwest of the Ljótipollur explosion crater. Erupted from the top of a hyaloclastie ridge it forms two contemorary lava flows, the younger one partly covering the older. The lavas cover an area 1.72 km^2 and have an estimated volume of 0.01 km^3. Frostastaðahraun is one of three separate lava flows (the other being Námshraun and the 1800 years old Dómadalshraun) that have contributed in forming the small lake of Frostastaðavatn, by closing the natural outflow. The chemistry of one sample suggests it is a hybrid tholeiite.

3.4.2 ***Stútshraun*** (*Hybrid lava*) – Stútshraun (Norðurnámshraun) erupted through a basaltic hyaloclastic ridge. Initially an explosion crater was formed, followed by the extrusion of a single lava flow from three vents, which spread over the Jökulgilskvísl river outwash plains. Finally a perfectly shaped spatter/tephra cone was accumulated over the main eruption site. The total length of the fissure is about 600 metres, the other two vents visible as small spatter craters and horniot-like features. Previous chemical studies had demonstrated silica content between 50-55%, demonstrating that Stútshraun is a hybrid lava, but two samples from Stútshraun used here are typical tholeiites. Stútshraun covers an area of 2.7 km^2 and has a volume of 0.014 km^3.



En Route: Námshraun (Suðurnáms-hraun) was erupted from a 500 meter long fissure segment on top of a partly subglacial rholitic ridge and a partly basaltic hyaloclastic ridge. The eruption site is exactly on the northern perimeter of the rhyolitic ring structure/caldera rim. It consists of 8 eruptive vents, forming 6 small domes and two larger lava flows. Both flows have cascaded down to the lowlands, one west into lake Frostastaðavatn (consisting of two or three flow units) and the other east down to the Jökulgilskvísl outwash plains. The original explosive opening of the fissure created an elongated crater into the ridge, from which the larger flows were later emitted. Námshraun is metaluminous rhyolitic lava flow containing large volumes of hybrid magma. The hybrid compositions include dacites, Icelandites and basaltic Icelandites. Námshraun covers an area of 0.68 km^2 and has a volume of 0.014 km^3.

3.4.3 ***Laugahraun*** (*Acidic volcanism and geothermal heat*) – Laugahraun consists of two flows, a small flow at the extreme southwest end of the fissure, erupted from a single vent, and a large flow collectively formed by magma emission from at least ten vents. The total length of this fissure segment is about 2 km. The lava covers 1.57 km^2 and has a volume of 0.039 km^3. Neither Laugahraun nor Námshraun appear to have produced any significant amount of tephra, only lava without any explosivity. Laugahraun is a mixed metaluminous rhyolite, containing inclusions of hybrid tholeiite (tholeiite-

rhyolite hybrid) and transitional icelandite, considered here to represent the third end-member participating in the magma mixing in this eruptive event. The geothermal springs at Landmannalaugar, a popular bathing location at the edge of the Laugahraun lava, are sodium-chloride waters representing boiled and variously mixed reservoir water. From Landmannalaugar, beautiful arcuate rhyolitic mountains (peralkaline in composition) can be observed towards the north and east. They represent the ring structure/caldera rim and belong to the Jökulgil series volcanics. Towards the south the dark mountain of Bláhnúkur looms above the Landmannalaugar location. It represents the Bláhnúkur series volcanics and is a relatively recent addition to the volcanic pile at Torfajökull and is made up of rhyolitic pillows or lobes, with intermingled glassy material, frequently altered green. Volcanics (lava flows) from the the volcanoes oldest period (Brandsgil Series) are not readily visible from this location.

3.4.4 ***Eldgjá*** (*Eruption fissure and flow of lava*) – Eldgjá, which was "discovered" in 1883 by Þorvaldur Thoroddsen and described by him and later by the German volcanologist K. Sapper, is one of Iceland's most impressive volcanoes. It is part of the Katla volcanic system, named after the central volcano Katla below the Mýrdalsjökull glacier. The Eldgjá fissure may be as long as 75 km from end to end, but only the middle 30 km form a fairly continuous eruptive fissure.

Eldgjá is most likely a monogenetic feature. The part visited by the excursion is Eldgjá proper. The length of this part of Eldgjá is 8 km, its maximum width about 600 m and maximum depth about 140 m. The inner slopes of the gjá consist of tillite. The topmost part forms vertical walls, 10-15 m high, consisting mainly of spatter, scoria, and tephratic lava, viz. lava formed by the welding together of spatter (ejected lava lumps). The shape of the present gjás, with the spectacular garland contour of its rim, seems to be partly the result of explosive activity, partly the result of slipping of rock and partly the result of the pre-existing topography.

The lava flows from Eldgjá cover an area of about 800 km^2, and according to a recent re-estimate , their voluma may be up to 14 km^3. A thick tephra layer extends mainly towards SE, with thickness up to 100 cm at a distance of 15 km. Tephrochronological studies and studies of electric conductivity of the Greenland ice cores confirm Þorvaldur Thoroddsen's opinion, based on the Icelandic Book of Settlement, that Eldgjá was formed about 934-40 AD.

En Route: Eldhraun (The Lakagígar lava flow of 1783-84) – The Lakagígar fissure eruption in 1783-84 was the biggest lava eruption in Iceland in historical time, i.e. during the last 1100 years and the lava flow the largest extruded on earth during the last 1000 years. The length of the Lakagígar crater row is almost 25 km with the direction N45°E, 500-600 m above sea level. The crater row is divided into two nearly equal parts by Laki, a palagonite mountain formed in a subglacial eruption during the late Pleistocene. The crater row consists of approximately 135 craters and numerous minor vents, the highest being 100-120 m. Most of the craters are built up of scoria and "schweiss-schlacken" (mud pies, cow pats), but two of the bigger ones from the initial phase of the eruption in each part of the crater row are made of fine grained tephra (tephra rings).

Strong earthquakes preceding the eruption were felt in late May and on Whitsunday morning June 8, 1783, the eruption began. During the first 50 days the average production may well have been around 2000 m^3/s. The SW part of the crater row was active from the beginning of the eruption into September of same year, emitting the western lava flow into the course of the river Skaftá through whose canyon it came down to the inhabited lowland plain on June 12. Once there it spread over a large area. The Skaftá branch of the lava came to a halt on July 20 and a few days later, on July 29, the NE part of the Laki fissure started pouring out the eastern lava flow into the course of the river Hverfisfljót. This lava stream reached the lowland plain a few days later on August 7. The lava kept flowing until February 7, 1784 when the crater row ceased erupting.

Recent studies have shown the area of the Lakagígar lava flows to be 580 km^2. The longest lava stream from the SW part is about 60 km and from the NE part about 40 km. The estimated volume of the lava is close to 11.9 km^3. The volume of tephra has been estimated as equivalent to 0.25 km^3 of dense rock. The total volume of volcanic material erupted in the Lakagígar eruption is therefore approximately 12 km^3. The average production of lava in the eruption was therefore approximately 580 m^3/s. During the initial phase of the two main eruption episodes – from June 8 and from July 29 respectively – the rate of flow was much greater, possibly 10 or 20 times greater. The lava is of apalhraun-type, i.e. an aa flow.

Chemical analyses of samples from the Lakagígar lava show a homogeneous chemical composition of major chemical elements. Analytical work on scoria samples shows that on eruption the magma was about 9% crystallized and the amount of crystals fairly uniform. The lava is basaltic of a fairly evolved tholeiite type with re-crystalline phases present as microphenocrysts, i.e. plagioclase, olivine and clinopyroxene.

The evolved nature of the lava, the large volume (12 km^3) and uniform composition pose many questions as to how the evolution from a more primitive mantle derived magma occurred and where. A final evolution in a single magma chamber explains the homogeneous nature of the lava but it is not obvious whether the whole magma volume evolved simultaneously and to the same degree or if there was a major mixing process shortly before eruption? If the strike of the Lakagígar fissure is projected towards northeast it points towards the Grímsvötn area. Along with nearly identical chemistry this once prompted the suggestion that the Lakagígar fissure did drain a magma reservoir beneath the Grímsvötn caldera. The location of the magma chamber is not certain. An obvious place would be below the fissure itself but a parallel with the Krafla fires 1975-1981 suggests that large volumes of basaltic magma may move laterally from central volcanoes situated on the same fault swarm as the crater row. This could either be the Grímsvötn caldera or possibly another central volcano situated between Grímsvötn and Lakagígar with a centre near the rhyolite nunatak, Pálsfjall. That evolution process is likely to have occurred in a relatively shallow holding chamber. There is at present no indication that a shallow holding chamber exists beneath the Lakagígar area and the most likely source for the magma is therefore the Grímsvötn magma area.

Great damage was caused by the lava flow, which covers 580 km^2, but still more by the bluish haze (probably containing SO_2) that lay over nearly the entire country during the summer of 1783. It stunted the grass crop and the result of this and of fluorine poisoning was a disastrous famine, still referred to as the Haze Famine (Móðuharðindin). The effects on life in Iceland were catastrophic, 50% of the island's cattle, 76% of its horses, and 77% of its sheep perished and so did 20% of the entire human population (*Albertsson et. al, 1997*).

3.5 Day 5: Dyrhólaey/Laki – Öræfi - Dyrhólaey/Laki

We will spend our fifth day mainly in the area south of the Vatnajökull ice cap with the main emphasis on glaciers and and glacial geomorphology. On the way we will stop at Jökulsárlón an ice lake in front of the Breiðamerkurjökull, an outlet glacier originating from Vatnajökull and at Kvíárjökull and inspect extremely high lateral moraines. On Skeiðarársandur will visit moraines that were formed at the height of the Little Ice Age and on Mýrdalssandur we will deal with glacier outburst floods from Katla.

En Route: Landbrotshólar – Rootless-cone groups consist of closely packed cones that rest directly on the associated lava flow. Cone dimensions vary from 2 m to 40 m high and 5 m to 450 m wide. They occur in tube-fed pahoehoe flows where the lava advanced over wetlands, such as shallow lakes or swamps. The area of individual cone groups is most commonly between 1 and 10 km^2, but the largest cone group, Landbrotshólar, covers 150 km^2. Rootless cones are formed by hydromagmatic eruptions caused by explosive interaction between molten lava and water-saturated substrate. Their cones represent volcanic vents that have lateral feeders, which are the lava tubes of pahoehoe flows. Thus, they differ from normal vents in that their feeders are not rooted deep within the crust, hence the name "*rootless cones*" (*Þórðarson and Höskuldsson, 2002*).

The Landbrotshólar cone group was formed by rootless eruptions when the Eldgjá lava advanced onto the marshy coastal plains in front of the Síða enscarpment. It is by far the largest cone group in Iceland and at present it covers an area of 60 km^2. Part of the group is now buried beneath the 1783 Laki lava flow and it may have originally covered 150 km^2. In 1793 Sveinn Pálsson examined the Landbrotshólar cone group and described it as pyroclastic cones with a lava cap (i.e. welded spatter). He was the first to suggest that it was formed by secondary eruptions.

En Route: Austur-Síða – Subglacial volcanism; origin and formation of hyaloclastite. The Pliocene – Pleistocene strata of the Síða and Fljótshverfi districts that form the old seacliffs above Kirkjubæjarklaustur consist of a 700 m thick volcanic succession called the Síða Group. It is dominated by 14 large-volume (up to 30 km^3) subaqueous hyaloclastite lava flows emplaced onto a submarine shelf. The Síða Group unconformably overlies an eroded sequence of flat-lying lavas and sedimentary rocks (the Fljótshverfi Group), formed before the Ice Age set in. The distribution of individual flows appears to have been mainly controlled by the existing topography.

The lower part of each hyaloclastite flow consists of columnar jointed lava (5-30 m thick), grading laterally and vertically into pillow lava or cube-jointed lava (5-40 m thick), which in turn is overlain by kubbaberg breccia 10-120 m thick. Each flow sequence is typically capped by a 15 m thick unit of well bedded lapilli tuff. The lava at the base of these flows often exhibits spectacular columnar joints. The cube-jointed lava typically features poorly developed columns with irregular orientation, although locally these fan out to form regular rosette-jointed bodies. The hyaloclastite flows are separated by 5-50 m thick sedimentary units consisting of debris-flow deposits interbedded with marine sandstones and mudstones.

3.5.1 ***Háalda*** (*Glacial and fluvioglacial landforms and jökulhlaups*) – Moraines formed at the height of the Little Ice Age in front of Skeiðarárjökull. Good view along the edge of Skeiðarárjökull, its frontal lakes and drainage pattern with sandur and kettleholes.

Skeiðarárjökull is the largest southern outlet glacier of Vatnajökull, 70 km long. It has an area of some 1600 km^2 (ablation area about 500 km^2, accumulation area 1100 km^2) and a maximum thickness of 1000 m just north of Grímsvötn. Where the glacier is narrowest its width is 8.6 km. There it moves with an average speed of 1.2 m/24 h. The width of the glacier front is 28 km (piedmont type). Since about 1920 the glacier has on the whole receded 2-3 km and the reduction of its area is >35 km^2.

Skeiðarársandur is the largest one of Iceland's sandar. Its area is 1000 km^2, or 1% of the total area of the country. The shortest distance from the glacier to the ocean is 20 km, and the coastline of the sandur is 46 km long. The average slope of the river bed of Sandgígjukvísl between the glacier and the sea is 1:250. Between the present glacier and the end moraines of the 19th century, from which the glacier began to recede in the 1920s, there are now frontal ice lakes.

The three main rivers on the sandur are Núpsvötn near its western margin, Sandgígjukvísl about 6.5 km farther east, and Skeiðará near its eastern border. Skeiðará is the biggest of these. Its average summer discharge is about 200 m/s. Sandgígjukvísl is the smallest one, its average summer discharge amounting to 50-60 m^3/s, but it has increased gradually during the last decades, probably as a result of the gradual heightening of the eastern part of the sandur. At the foot of the slopes west of Skaftafell farms, the surface of the sandur is now about 20 m higher than a century ago.

Jökulhlaups on Skeiðarársandur are of two types: 1) Glacigenic bursts, caused by drainage of laterally dammed ice lakes (Grænalón), and 2) Volcano-limnogenic bursts, caused by drainage of subglacial ice lakes (Grímsvötn).

By far the biggest of the former type are those caused by the draining of Grænalón, which is dammed by the western margin of Skeiðarárjökull. Grænalón is one of the biggest ice lakes in the world. It has a maximum area of 18 km^2, a maximum depth of 180 m, and a maximum volume of some 1.5 km^3. When totally drained, the jökulhlaup reach a maximum discharge of ca 5000 m^3/s. During the "Little Ice Age" the damming glacier was so thick that the ice lake had an outlet over a col on ice free land and was drained into the river Núpsá. The first emptying of the lake in recent times occurred in the autumn of 1898, the next one in 1935. From then on it was emptied two times with four years interval, and then with gradually shorter intervals because of the thinning of the damming glacier. Since 1951 the lake has not been emptied. The water level, though, has been lowered some 18-20 m annually by jökulhlaup. The total volume of these hlaup is 0.2-0.3 km^3, and their maximum discharge some 2000 m^3/s. These hlaup reach a maximum within 15 hours of their beginning and last about 5 days. The Grænalón hlaup emerge from the glacier front into the river Súla. When Skeiðarárjökull was at its maximum in the 18th century an ice lake was dammed by the SW corner of the glacier snout, colliding with the slope of Lómagnúpur. Small jökulhlaup occur now and then in Skeiðará when ice lakes along the eastern margin are emptied.

En Route: Skeiðarárhlaup – The large and sometimes tremendous jökulhlaup that usually start in Skeiðará and at their maximum emerge from numerous outlets along the entire front of Skeiðarárjökull, are caused by draining of a caldera lake, Grímsvötn, in central Vatnajökull. This caldera has an area of about 35 km^2 and is covered by ice, at least 200 m thick. Between the jökulhlaup the water volume in the caldera increases by 0.6-0.7 km^3/year, firstly, because of continuous subglacial melting from solfatara areas within and to the north of the caldera, secondly, by ablation water from those 280 km^2 of Vatnajökull that are drained into Grímsvötn, and, thirdly, to some extent now and then by volcanic eruptions within the caldera.

Until 1934 Grímsvötn were usually drained about every 10th year and the lake level was raised about 150 m between the hlaup. The total volume of these hlaup was probably 6-7 km^3 and their maximum discharge at 40-50,000 m^3/s. These jökulhlaup flooded the main part of Skeiðarársandur. They broke up the glacier front around the outlet and carried icebergs, sometimes of enormous size, out onto the sandur, which in places is dotted with kettleholes resulting from melting of icebergs.

Since 1938 the hlaup have occurred on the average every 5th year or so. The total discharge of those hlaup is 3.0-3.5 km^3 and their maximum discharge about 10,000 m^3/s. The level of the Grímsvötn lake is lowered 80-100 m by these hlaup. The Skeiðarárhlaup increase slowly for about 10 days or so to an euphemeral maximum, after which the discharge drops rapidly to normal. A Skeiðarárhlaup commenced on the 4th of September 1976. The total volume of water discharged was 2.4 km^3, culminating in maximum discharge of some 7600 m^3/s. A still smaller jökulhlaup occurred in January-February 1982. Some of the Skeiðarárhlaup are accompanied by volcanic eruptions. This was the case in 1934. There was a small volcanic eruption in Grímsvötn in May 1983, but it did not cause a jökulhlaup. Subsequently to an eruption in Vatnajökull in early October 1996 a gigantic outburst (jökulhlaup), measuring about 55.000 m^3/s, entered the Skeiðarársandur outwash area in early November.

3.5.2 ***Jökulsárlón*** (*Little Ice Age moraines and glacigenic landforms*) – The recession of Breiðamerkurjökull since about 1930 has resulted in the formation of frontal lakes. The largest one, Jökulsárlón, reaches over 160 m below sea level where seracs are calving off the glacier snout, but the main lake basin is just over 100 m deep. The lake began to appear around 1930.

Sediment loads from Breiðamerkurjökull, previously carried out to the sea by the jökullsá river, are now deposited straight into the proglacial lake. Thus, farther building up of the sandur delta has ceased. Sea erosion is rampant and the shoreline now recedes on average by 8 m/year, creating a threat to the road and bridge over jökullsá on Breiðamerkursandur. Radio echo sounding of Breiðamerkurjökulll show that the glacier bed is below sea level over a distance of 20 km upwards from the lake. This overdeepening which is 2 to 5 km wide and more than 200 m deep was probably created during the cold period from c. 1400-AD to the end of the last century when Breiðamerkurjökulll advanced by some 10 km over soft sediments. The bed is above sea level beneath the less active western Ranches of Breiðamerkurjökull.

Breiðamerkursandur is a collective name of the proglacial area of Hrútárjökull and Fjallsárjökull, both outlet glaciers from Öræfajökull, and Breiðamerkurjökull, the broadest southeastern outlet glacier of Vatnajökull proper. Breiðamerkursandur shows a wealth of proglacial sedimentological and morphological forms which since 1964 have been mapped and studied in detail in numerous exeditions.

En Route: The firn limit of Breiðamerkurjökull is at about 1100 m. The accumulation area is 800 km^2 and the ablation area 470 km^2. The glacier reached its maximum postglacial extent during the 19th century and has on the whole been receding since the middle 1890s. In 1894

the shortest distance between Breiðamerkurjökull's front and the beach, 5 km east of Jökulsá, was only 256 m. At Jökulsá it was about 1 km. In the period 1894-1968 the glacier front retreated up to 2.3 km, exposing 52 km^2 of deglaciated area. During the same period, Breiðamerkurjökull lost about 50 km^3 in volume. The retreat was slow up to 1930, but between 1933 and 1960 the eastern part retreated 1350 m and the western part 1160 m.

Breiðamerkurjökull contains two medial moraines which run from the nunataks Mávabyggð and Esjufjöll. The most active branch of Breiðamerkurjökull lies east of the medial moraine from Esjufjöll. This branch drains ice from the interior of Vatnajökull (Breiðabunga) whereas the western branch of Breiðamerkurjökull only drains ice south of the two nunataks.

Breiðamerkurjökull and Fjallsárjökull coalesced on the lowlands until 1945. During the first centuries of settlement in Iceland the front of Breiðamerkurjökull probably lay not less than 10 km behind the 1894 moraine. During the Commonwealth Period there were two farms on the western part of Breiðamerkursandur. One, Fjall, at the southeast foot of Breiðamerkurfjall, the other one, Breiðá (with a church), somewhat farther east. Both were overrun by the advancing glacier between 1695 and 1720.

3.5.3 ***Kvíárjökull*** (*Lateral and terminal moraine*) – Kvíárjökull is the largest southern outlet glacier of Öræfajökull and it stretches down into the lowland southeast of the Öræfajökull massif. Its lowland part is surrounded by a magnificent moraine amphitheatre, rising up to nearly 100 m above the outside plain. The outermost part of these moraines is prehistoric, but about 1870 the glacier snout rose high above these moraines so that ice blocks now and then rolled down the outer slopes, thus reaching about the same size as its maximum postglacial one.

En Route: Settlement and farming in the rural community between Skeiðarársandur and Breiðamerkursandur has suffered greatly in the past because of the proximity of Vatnajökull, Öræfajökull, their outlet glaciers and glacial rivers as mentioned above. The rural settlement which lies in a semi-circle at the roots of Öræfajökull is different from most other rural areas in Iceland. There are very few isolated or individual farmsteads as elsewhere but rather clusters, where the original farms have been subdivided and the farm buildings are grouped together - Hnappavellir, 4 households, Fagurhólsmýri 4, Hof 6, Svínafell 4 and Skaftafell 2 households - Skaftafell, the site of the National Park (est. 1968) are probably the best known farms in the area. The chieftain Flosi Þórðarson, who was the leader of those who burnt down the farm Bergþórshvoll, killing Njáll and most of his family (cf. Njálssaga), lived at Svínafell around the year 1000. Ingólfshöfði (76 m) is a well known cape on the coast 9 km south of Fagurhólsmýri, named after Iceland's first settler Ingólfur Arnarson who spent his first winter in Iceland there.

Öræfi was very isolated until 1946 when regular flights from Reykjavík started and a further change was brought about when bridges were built across Jökulsá at Breiðamerkursandur in 1967 and across Skeiðará and other rivers on Skeiðarársandur in 1973-1974.

3.5.4 ***Öræfajökull*** (*A stratovolcano*) – Öræfajökull is a huge stratovolcano, second in volume only to Etna among volcanoes in Europe. From the sandur plains near sea level in the south and west it rises with an average angle of slopes about 15° to a crater rim with an average height of 1850 m. This crater rim surrounds an elliptical 400 m deep summit caldera with an area of 14 km^2, which is, along with the upper reaches of the volcano, covered with eternal ice from which outlet glaciers stretch down to the sandur plains.

Reversed magnetic rocks are found in the basement of the Öræfajökull massif. The bulk of the massif is built up partly subaerially by basalt and andesite lavas and partly subglaciaily by pillow lavas and hyaloclastites (móberg). Rhyolite is abundant. Hvannadalshnjúkur, the highest point in Iceland (2119 m), is a rhyolitic peak rising above the northwestern rim of the caldera. The lower part of the massif is deeply dissected by glacial erosion.

The activity of the summit area in postglacial time seems to have been almost exclusively explosive, and tephrochronological studies in the neighbourhood of the volcano show that this activity was rather limited and did not add much to the height nor volume of the volcano. Some radial fissures reaching below the present ice cover have been active in postglacial time, producing at least one lava flow reaching the lowland plain at Kvíárjökull on the southeast side of the volcanic massif.

En Route: Öræfajökull eruption (1362, early June). The most detailed, almost contemporary description of this eruption is to be found in an annalistic fragment from the See of Skálholt, believed to have been written in the monastery at Möðruvellir in North Iceland. It runs as follows:

"*Volcanic eruption in three places in the South and kept burning from Flitting days (= in early June) until the autumn with such monstrous fury as to lay waste the whole of Litlahérað as well as a great deal of Hornafjörður and Lónshverfi districts, causing desolation for a distance of some 100 miles. At the same time there was a glacier burst from Knappafellsjökull into the sea carrying such quantities of rocks, gravel and sand as to form a sandur plain where there had previously been thirty fathoms of water. Two parishes, those of Hof and Rauðilækur, were entirely wiped out. On even ground one sank in the sand up to the middle of the legs and wind swept it into such drifts that buildings were almost obliterated. Ash was carried over the northern country to such a degree that foot-prints became visible on it. As an accompaniment to this, pumice was seen floating off the Northwest Iceland (Vestfirðir) in such masses that ships could hardly make their way through it*".

Other more or less contemporary annals confirm this description. One of these annals states that the flood swept away all buildings of the Rauðilækur rectory except the church. A 16th century annal, based on a lost, much older one, states that "*no living creature survived except for one old woman and a mare*". Tephrochronologicai studies have supplemented the scant information supplied by the old annals.

The 1362 eruption was a typical initial eruption, purely explosive and producing an enormous amount of rhyolitic tephra. The eruption was preceded by a repose lasting at least 500 years. Presumably the eruption occurred mainly, or wholly within the caldera. The regular thickness-distribution of the tephra on land indicates that the main tephra fail was of short duration. It lasted probably only a few days. The passus "kept burning until autumn" in the above-quoted annal fragment need not refer to this eruption. The tephra was carried mainly towards ESE and consequently the main part of it fell into the sea. The tephra has been traced in peat bogs in Scandinavia. The volume of tephra on land as freshly fallen has been at least 2 km^3 and the total volume on land and sea has probably amounted to about 10 km^3, corresponding to ca 2 km^3 of solid rhyolitic rock.

The eruption was accompanied by catastrophic jökulhlaup, which emerged from underneath the outlet glaciers ***Falljökull*** and ***Rótarfjallsjökull*** on the west side of the massif. Some farms were undoubtedly destroyed by these floods, but the tephra fall probably had the main share in the devastation caused by this eruption. The prosperous rural settlements along the foot of the volcanic massif, altogether at least 30 farms, were laid waste so thoroughly that they remained abandoned for decades. When a revival finally came, this district, which before the eruption was called Litlahérað, the name hérað in Icelandic being reserved for extensive and important rural settlements, had received a new name, and a significant one: Öræfi, meaning wasteland.

The tephra fall furthermore damaged rural settlements up to a distance of 70 km east of the volcano so that they were abandoned for a year up to some years. It is fairly certain that the jökulhlaup killed many people living on the west side of the volcano and the tephra fall is likely to have taken its toll in human lives too.

En Route: Öræfajökull eruption (August 3 1727 - April or May 1728). On Sunday the 3rd of August while people were attending divine service in the church of Sandfell in the Öræfi district earthquake shocks were felt. The seismic activity gradually increased. Early the following morning, the 4th of August, the shocks were so strong that everything standing upright in the houses was thrown down. At 9:00 three particularly loud reports were heard and they were almost instantaneously followed by tremendous jökulhlaup from the glaciers ***Falljökull*** and ***Rótarfjallsjökull***. The jökulhlaup destroyed two cottages near the Sandfell parsonage and most of its pastures. They also destroyed two chalets and drowned three people in one of them. Later that day the farms on the western side of Öræfajökull were in complete darkness because of heavy tephra fall which lasted for three days. On the fourth day it cleared up and apparently only a small amount of tephra fell after that. The tephra of the first days was carried towards WNW and the farms of Svínafell and Skaftafell became uninhabitable for a while. However, the tephra production was small compared with that of the 1362 eruption and probably did not exceed 0.2 km^3. The floods and the tephra fall killed about 600 sheep and 150 horses, some of which were found completely mangled by the bomb fall.

The great amount of water discharged by the jökulhlaup points to the eruption having started within the caldera, but already on the 4th of August a fissure with six or seven separate fires opened up on the outer west flank of the caldera, reaching down to about 1100 m above sea level, and in this fissure fire and smoke was seen until April, or even untii late May the following year, but as far as known it did hardly emit any lava.

3.5.5 ***Mýrdalssandur and Kötluklettur*** (*Jökulhlaup from Katla*) - Mýrdalssandur. Hidden beneath the ice cover of the accumulation area of Kötlujökull (Höfðabrekkujökull) is Katla, one of Iceland's most active and destructive volcanoes. The eruptions of this volcano have of-

ten caused great damage both because of heavy tephra falls and the tremendous jökulhlaup (Kötluhlaup) which accompany the eruptions. Written contemporary records exist of all Katla eruptions from 1580 AD onwards, but before 1580 the records are less reliable. Tephrochronological studies have proved that a big eruption occurred in Katla about 1000 AD, another around 1357, and the third around 1490. These eruptions are not mentioned as Katla eruptions in written sources. Since 1580 Katla has erupted seven times, or twice a century. Curiously enough, short and long intervals of quiescence have alternated, so that eruptions have occurred either near the end of the second or sixth decade of each century, with a maximum deviation of only 5 years from this rule. Katla erupted in 1625, 1721, 1823, and 1918, and also in 1660, 1755, and in 1860, so it is no wonder that people have asked "*is not the next Katla eruption overdue?*".

The greatest damage by tephra fall was caused by the eruptions in 1625, 1721, and 1755. Because of the 1755 tephra fall the rural districts Skaftártunga and western Síða were seriously damaged and temporarily abandoned. The districts had not fully recovered when they were hit by the still more disastrous Lakagígar eruption in 1783.

The accompanying jökulhlaup are tremendous, their maximum discharge exceeding that of the Amazon river, but they usually last less than a day. They transport enormous amounts of pumice, debris and ice. The 1721 hlaup deposited over 200 m thick ridges of ice between Hjörleifshöfði and Höfðabrekka. The shoreline gradually progrades. Before 1300 AD Hjörleifshöfði was a promontory washed by the sea and west of it there was a shallow bay. The distance from Hjörleifshöfði to the shore is now 3 km. The Kötluhlaup in October 1918 formed a delta that stretched 3 km into the sea. Gradually the westward currents along the shore eroded this delta away, moving the sand westwards and thereby widening the sandy plain between the village Vík and the sea by more than 200 m.

The farm Hjörleifshöfði was also located on the western side of the "*höfði*" until the farmhouses were swept away by the 1721 Kötluhlaup. The farm was rebuilt where the house ruin now stands, but was abandoned some decades ago. On the highest summit of Hjörleifshöfði there is a burial place which Markús Loftsson, one of the last farmers on Hjörleifshöfði and author of a book on the Katla eruptions, had made for himself. He died in 1906 and rests there (*Albertsson et al, 1997*).

The 1918 eruption began at about 1 pm on October 12, when sharp earthquake was felt. The eruption cloud was first observed at 3 pm with a large plume soon reaching a height of 14 km above sea level, which spread ash chiefly to the east. At about the same time a tremendous flood was seen to emerge from the glacier, over the sandur plain and into the sea, reaching a climax at 530 pm. The flood carried so much ice that "*it looked like snow-covered hills were rushing forward*". Thus the ice was mainly carried on top of the flow. Width of the flow was about 12 to 15 km, but a second flow issued from NW side of the glacier and to the east. Combined area covered by the flow was about 370 to 400 km^2. Eight days after the eruption, the glacier was inspected. It was evident that the first flood had in fact passed over the glacier, whereas the peak flood had also emerged form underneath the ice, fragmenting the glacier and carrying huge ice blocks away. A 2 km long, 500 m wide and 200 m deep canyon had been carved through the tounge of the Katla glacier (Kötlujökull). The eruption continued for 23 days, sometimes with widespread tephra fall. Total tephra fall from the eruption is estimated about 0.7 km^3, but the volume of water-transported tephra is not known. Thickness of the postglacial volcanic tephra and alluvial sands that makes up the Mýrdalssandur plain between Hafursey to the north and Hjörleifshofdi to the south ranges from 100 in the north to 30 m in the south.

Peak flow in the Katla 1918 jökulhlaup has previously been estimated 2-4x10^5 m^3/s but Jón Jónsson presented a higher estimate of 1.5 x 10^6 m^3/s. During the 20 hours of flow, the coast subject to flooding was in parts extended out some 4 km, and to a depth of 40 m. Similarly enormous deposit was laid on top of the sandur plain, with a thickness of about 8 m in some places. About 95% of the flood deposit is volcanic sands and "*pumice*", dominantly 0.5 to 4 mm in diameter, whereas silty material (<0.074 mm) is less than 4 to 6%. The deposit is strongly reversely graded, with sandy and pumice-rich lower part, grading upwards into poorly sorted sandy, gravelly layers rich in large (1 m) boulders. A boulder of about 950 to 1000 metric tons (***Kötluklettur***) was transported by the flow over 15 km. The deposit also contains large clay clasts from hydrothermal deposits near the volcano. Older floods carried a block of soil with the dimensions 90 by 2 m to the coastal area. Jón Jónsson has estimated density of the 1918 flow as high as 2.5 g/cm^3. South of Iceland, submarine sedimentary ridges occur, known as the Katla-ridges. Deep sea cores taken south of Iceland contain volcanic sand layers which most likely originated from turbidity currents from Katla-type events on Iceland.

3.6 Day 6: Dyrhólaey/Laki – South Iceland – Reykjavík

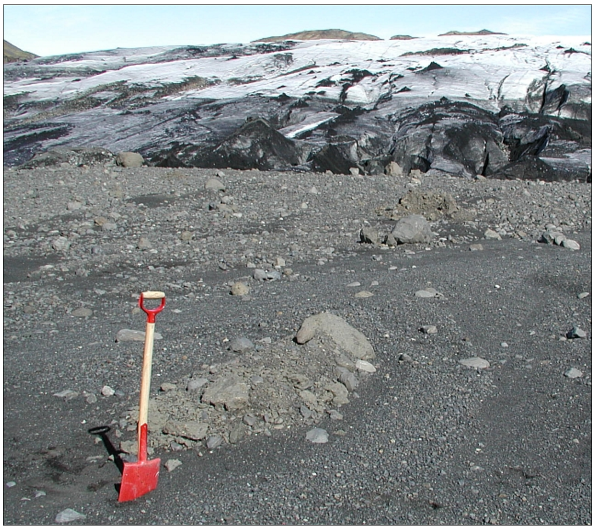
On the final day of our excursion we will travel along the south Icelandic coast again with emphasis on glacial geology and Holocene volcanism ending the day in Reykjavík. We will start with a visit to Dyrhólaey, an old Surtsey with rich birdlife. From there we will travel west along Eyjafjöll and Eyjafjallajökull to Rangárvellir to have a look at an exceptionally large Youger Dryas end-moraine. Closing in on Reykjavík we will try visiting Rauðhólar, a heavily excavated group of rootless cinder cones.

3.6.1 ***Dyrhólaey*** (*An old Surtsey*) - Some of the unusual features of the region around Mýrdalur are the mountains Pétursey, Dyrhólaey and Hjörleifshöfði, which stand as isolated crags in a sea of sand and generally a good distance inland from the current coastline. Now on dry land, these structures are remnants of emergent submarine volcanoes formed when these low-lying coastal plains were fully submerged. Since then, the rivers have transported enough debris to raise the area out of the sea and link the former islands with the main landmass of Iceland.

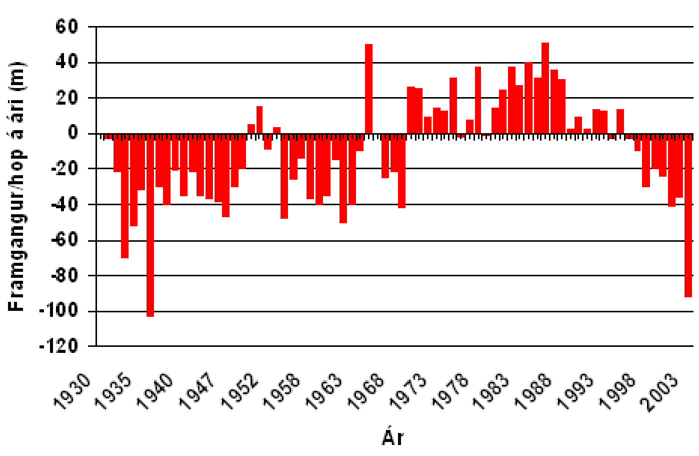
Dyrhólaey (Portland) is the southernmost point of Iceland and is a heavily eroded submarine volcano of the Surtseyan type. The main part of Dyrhólaey is built of well bedded tuff formed by hydro-magmatic explosions and it represents the remains of a larger tuff cone. On the eastern flank of Dyrhólaey, the tuff sequence is capped by compound pahoehoe lava, which in places exhibits cube jointing, indicating water-enhanced cooling of the lava. The lava flows represent the subaerial phase of the eruption when the tuff cone had grown large enough to prevent sea water from accessing the vent. Flowing away from the vent(s), the lava re-entered the sea and cooled rapidly to form the cube-jointed lobes. The Dyrhólaey sequence is typical for Surtseyan eruptions and, if visibility allows, the type-volcano Surtsey can be seen from this location as the southernmost island of the Vestmannaeyjar islands.

3.6.2 ***Sólheimajökull*** (*Moraines, drumlins and glacier response to climatic changes*) – Sólheimajökull is an outlet glacier from the south-western part of the Mýrdalsjökull ice cap. It occupies a valley trough, and is about 8 km long and >1 km broad at its broadest. Its total surface area is ca. 42 km^2. The average thickness of Sólheimajökull is ca. 270 m, but it reaches a thickness of >600 m at most. The subglacial through lies at places >50 below sea level.

Sólheimajökull has generally been retreating since the end of the 19th century. It retreated about 900 m between the years 1930-1964, but had a period of significant advance between ca. 1970 and 1995, when it advanced more than 400 m. Presently it is retreating at the rate of ca. 50 m/year. Continued retreat will lead to the formation of a glacial lake in front of the glacier.



The glacial river that drains Sólheimajökull is called Jökulsá á Sólheimasandi. Since the river partly drains meltwater generated by geothermal fields below the Mýrdalsjökull ice cap, it often contains high concentrations of sulphur. The sulphurous gas, hydrogen sulfide (H_2S), smells like rotten eggs, and gives the river its characteristic smell (the river is sometimes called "Fúlilækur" in Icelandic, which means "Rotten River"). Hydrogen sulfide can be very dangerous in high concentrations, leading to blindness and death. People should avoid visiting the Sólheimajökull ice front/Jökulsá on calm days when the sulphurous smell is strong!



When the ice front was more advanced than today, during the Little Ice Age until early 20th century, the glacier dammed small rivers in ravines and shallow tributary valleys upglacier. These were periodically drained, causing very sudden jökulhlaups in Jökulsá á Sólheimasandi. Before the river was bridged in 1921, the river was considered very dangerous to cross, and at least 20 people are known to have drowned when crossing. The last jökulhlaup occurred during the summer of 1999, in connection with volcanic/seismic activity within the Mýrdalsjökull caldera.

Glacial retreat since 1995 has exposed a number of subglacial landforms, including drumlins and flutes, and dead-ice forms (kettle holes, pitted surfaces) are conspicuous along the ice front. We will also see evidence of the 1999 jökulhlaup, the end moraine from 1995 and a number of small retreat moraines.

En Route: Skógar – A farmstead from the Settlement Age (the name indicates that the slopes were at that time covered with birch woods), it is now the site of the district museum. Skógar is located at the NW border of Skógasandur, the western part of Sólheimasandur, which is the sandur of the glacial river Jökulsá á Sólheimasandi. Skógafoss is one of the most beautiful of the numerous waterfalls of Iceland, and is 60 m high.

Hvammsmúli – Highly porphyritic basic rocks are found exposed at the base of Hvammsmúli. The composition is ankaramitic. It has been suggested that they are intrusive in origin, mainly by virtue of the distinct and different mineral composition as compared to overlying basaltic succession. Veins, ramifying the hyaloclastite above are of somewhat similar nature, but differ significantly in the relative abundance of phenocrysts fer significantly in the relative abundance of phenocrysts they contain.

Holtsós – The former delta area of the glacial river Markarfljót. The active delta is now about 15 km farther to the west.

Seljalandsfoss – A 60 m high, plunges from the head of an overhanging abrasion cliff.

Stóra Dímon – Beautifully jointed intrusive basalt. Dimon is a Celtic place name and means a twin mountain. Dimons are always in pairs, a big one and a smaller one nearby. There are many Dimons in Iceland, as well as a lot of other Celtic place names, although the Nordic ones are predominant.

Markarfljótsaurar – Until a few decades ago the river Þverá, mainly a branch of Markarfljót, flowed along the northern border of the sandur and cut away every year a strip of the thick loessial soil cover of the fertile Fljótshlíð lowland. In order to prevent this, a dam was built directing all the water of the sandur to the main bed of Markarfljót. Buried some 20 m below the sandur surface is an end-moraine which has been tentatively correlated to the Younger Dryas moraines in Rangárvellir, the next site to be visited.

3.6.3 ***Stokkalækur*** (*A Younger Dryas moraine*) - The Búði moraine can be correlated with other moraine ridges to form *"a belt of terminal moraine ridges"* that have been traced to Vatnsdalsfjall, some 25 km towards the southeast across the South Lowlands, and northwestwards for about 16 km to Grámelur at river Hvítá. Later the Búði moraine was extended to Efstadalsfjall, about 28 km northwest of the Búði moraine at river Þjórsá. It is to be emphasized that the Búði moraine is a complex of up to seven more or less parallel ridges with considerable age difference between the outermost and innermost ridges. A recent study shows that the Búði terminal zone is made of two sets of end-moraines some 5 km apart. The inner (younger) moraine crosses river Þjórsá at the Búði waterfall and can be traced for about 15 km towards the southeast where it disappears underneath a postglacial lava flow. It is the inner moraine that has been followed from the Búði waterfall to Efstadalsfjall. The outer (older) moraine has been followed from Mt. Vatnsdalsfjall and for about 25 km towards the northwest where it disappears in the Holt area (*Norðdahl and Péturson, 2005*).

It has been concluded, on the basis of structural and sedimentological data collected at river Þjórsá (Búðaberg and Þrándarholt) and at river Stóra-Laxá (Hrepphólar), that a glacier had advanced across glacio-marine sediments to form moraine-like ridges consisting of disturbed beds of glacio-marine diamicton. Four samples of marine shells collected from sediments in close relation to the Búði moraine at Þjórsá and Stóra-Laxá have yielded a weighted mean age of 9855±90 ^{14}C yr BP for the formation of the moraine and, hence, the advance of the glacier. Fragments of barnacles, collected from a diamicton with ice-rafted clasts in the upper parts of the Búðaberg section, have yielded dates of 9925±140, 9755±150, and 9765±170 ^{14}C yr BP, confirming that the formation of the Búði moraine at river Þjórsá inner moraine in the Rangárvellir area were apparently completed in early Preboreal time. The outer set of moraines was, therefore, formed during a glacier event preceding the formation of the Búði moraine proper. The altitude of ML shorelines in the Holt area is close to 110 m a.s.l, and about 100 m a.s.l. below Skarðsfjall, just north of the Búði moraine proper (*Norðdahl and Pétursson, 2005*).

At Stokkalækur we have a moraine reaching more than 100 m a.s.l. and in a section "*behind*" the moraine we can see glacially deformed sandy sediments between the moraine itself and a Holocene lava flow that stopped at the moraine.

3.6.4 ***Þjórsárbrú*** (*The Þjórsá lava, origin, extent and its age*) - The Great Þjórsá lava is the oldest flow of the Tungnaá lavas, a series of eight lava flows that emanated from fissures within the Veiðivötn volcanic system to the north of Landmannalaugar during the Holocene (*Þórðarson and Höskuldsson, 2002*).

The Þjórsá lava is the most voluminous lava in Iceland and the largest Holocene lava flow on Earth, both with respect to volume and area. It was erupted from a crater row in the Veiðivötn volcanic area in Central Iceland. The craters and the uppermost parts of the lava seem to be mostly buried in younger erup-tives, lavas and tephra. Some 75 km southwest of the volcanic site the lava crops out and covers an extensive area between the glacial rivers Þjórá and Hvítá. These two, the largest rivers of Iceland, have found their way along the eastern and western edges of the lava all the way to the sea. The lava forms the reef-bound coast between the river mouths and stretches several hundred meters out from the shore below sea level. The total length of the lava is at least 140 km, making it the longest flow in Iceland. Its thickness in the farming areas of South Iceland is fairly well known, being 15-20 m thick in its distal parts in the Flói district. In the Skeið district further north it is 20-25 m thick and in the Land district the thickness is about 30 m. The average thickness calculated from 40 boreholes along most of its length is 22 m. The lava is estimated to be about 950 km^2 in area. The total volume is therefore 21 km^3 (*Árni Hjartarson, 1988*).

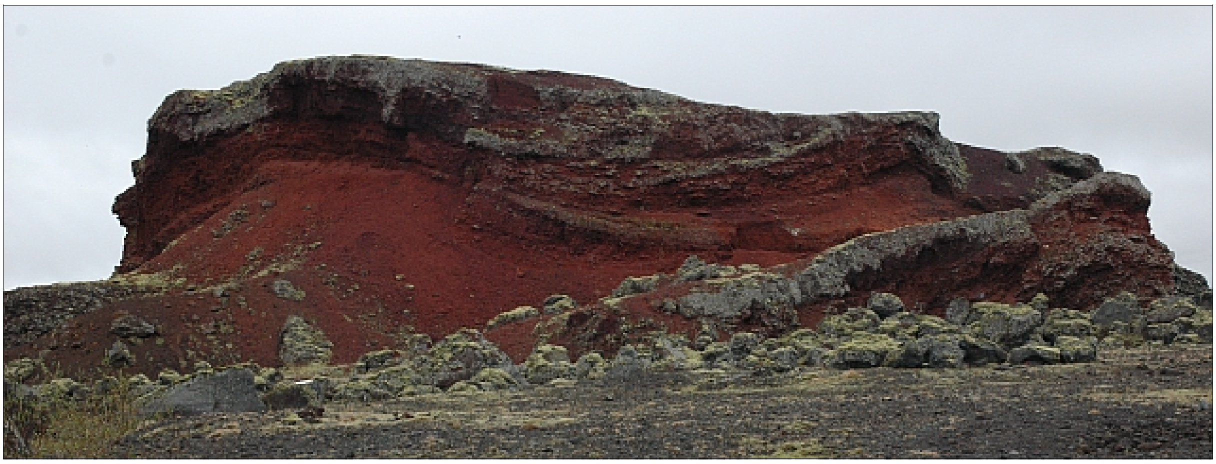
The ^{14}C age of the Þjórsárhraun (Þjórsá lava flow), was first determined in peat underlying the lava on the west bank of the river, just north of the mainroad bridge, to be 8065±400 BP when relative sea level was at least 15 m lower than at present (*Albertsson et al, 1997*).

No ash layer has been related to this eruption, but in ice cores from Greenland there are indications of a violent eruption somewhere in the northern hemisphere around the date 6675±150 BC. A sample of charcoal from below the lava yielded an age of 7800±60 BP which fits fairly well with the other one when converted to calebrated years, about 6630±70 BC (*Árni Hjartarson, 1988*).

3.6.5 ***Rauðhólar*** (*Inside rootless cinder-cones*) – Rauðhólar occur as a well confined field (1.2 km^2) of small scoria and spatter cones within a 2 km wide lava branch derived from the 4700 year old Leitin lava shield of the Brennisteinsfjöll volcanic system. This branch of the lava extends some 27 km from its source at Bláfjöll to the sea at Elliðavogur in Reykjavík. Several smaller cone groups occur in the Leitin lava, including the minute but well known Tröllabörn cone group about 5 km east of Rauðhólar. The Rauðhólar cone group is located where the lava flowed into and covered a small lake just north of present-day Lake Elliðavatn. The cone group is confined to a raised (5 m high) scoria platform that rests on 7 m of solid lava. In turn, the lava rests directly on l-2 m thick mudstone that covered the floor of the lake before the lava arrived.

The excavated part provides excellent exposure to examine the internal structure of the cone group and individual cones. Each cone has a crudely funnel-shape conduit (not visible) extending from the base of the flow and up through the coherent lava, terminating in a bowl-shape crater. Another noteworthy feature is that later cones lap onto and sometimes partly conceal earlier cones. Most importantly, individual cones exhibit distinctive internal layering, showing that each cone was built from multiple explosions during a period of sustained activity.

In section, the cone ramparts typically feature a well bedded lower sequence of ash and scoria and a crudely bedded upper sequence of coarser-grain agglutinates. The lower sequence consists of 0.2-0.6 m thick layers composed of highly fragmented lapilli scoria alternating with thinner (<0.2 m) crudely laminated beds of red-baked lacustrine silt and black ash. The scoria layers also contain, in variable abundance, lumps of red-baked mud and clasts that have cores of either vesicular lava fragments or cooked mud armoured with a skin of quenched lava. The upper sequence features multiple 0.5-1.5 m thick layers of brick-red agglutinates, mainly composed of spatter bombs with fluidal surfaces and variable twisted or distorted shapes. The distinctive brick-red colour of the deposit comes from a micrometre-thick coating of red-baked mud that was welded to the surface of the clasts while hot.



When considering the eruption mechanism that produced Rauðhólar and other rootless-cone groups, the following facts need to be kept in mind. Firstly, although the cone groups rest directly on top of the host lava, their cones show no evidence of having been deformed or modified by moving lava. Secondly, the occurrence of lacustrine muds as distinct beds within the lower sequence is a clear indication of their involvement in generating steam explosions by providing the water (i.e. the coolant). Obviously the lava was the fuel. Thirdly, the internal stratification of individual cones shows that they were formed by multiple explosions of decreasing intensity, whereas overlapping arrangements of cones within the group imply a certain time sequence and duration for the explosive activity. Fourthly, we have to consider the apparently random distribution of the cones and the fact that the lava continued its advance despite the rootless eruptions. In other words, we have to get the lava across the lake bed and at the same time initiate rootless hydromagmatic eruptions through contact between hot lava and water, forming cones on top of the lava that are in no way modified by its movements (*Þórðarson and Höskuldsson, 2002*).

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