

Geology and geodynamics of Iceland

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Iceland is located where the asthenospheric flow under the the NE Atlantic plate boundary interacts and mixes with a deep-seated mantle plume. The buoyancy of the Iceland plume leads to dynamic uplift of the Iceland plateau, and high volcanic productivity over the plume produces a thick crust. (Fig.1). The Greenland-Færöy ridge represents the Iceland plume track through the history of the NE Atlantic. The current plume stem has been imaged seismically down to about 400 km depth (Tryggvason et al. 1983; Wolfe et al. 1997, 2002), throughout the transition zone (Shen et al. 1998; Shen et al. 2002) and more tentatively down to the core-mantle boundary (Helmberger et al. 1998; Bijwaard and Spakman 1999).

During the last 60 Ma Greenland, Eurasia and the NE Atlantic plate boundary have migrated northwestwards at a rate of 1-3 cm/a relative to the surface expression of the Iceland plume (Fig. 1; Lawver and Muller, 1994). Currently, the plume channel reaches the lithosphere under the Vatnajökull glacier, about 200 km southeast of the plate boundary defined by the Reykjanes and Kolbeinsey Ridges (e.g. Wolfe et al. 1997). During the last 20 Ma the Icelandic rift zones have migrated stepwise eastwards to keep their positions near the surface expression of the plume, leading to a complicated and changing pattern of rift zones and transform fault zones.

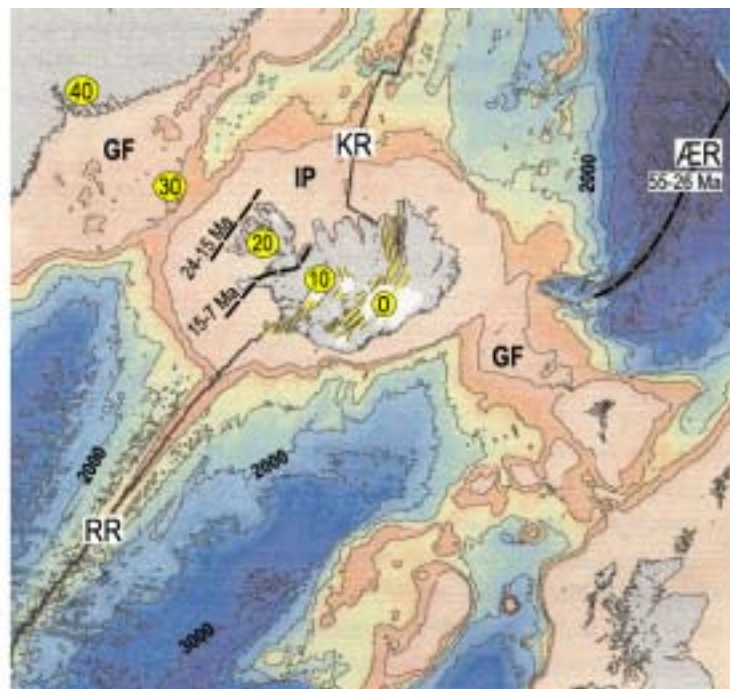


Fig. 1. Bathymetry of the area around Iceland. Depth contours for each 500 m. The position of the Iceland plume relative to Greenland and Iceland at 40, 30, 20, 10, and 0 Ma is indicated by the filled circular yellow fields. Active and inactive spreading axes are shown by continuous and stippled heavy lines. The active rift zones in Iceland are shown as individual fissure swarms in yellow. RR, Reykjanes Ridge; KR, Kolbeinsey R.; ÆR, Agir R.; IP, Iceland Plateau; GF, Greenland-Færöy Ridge. The shallow ridge between KR and ÆR is the southern part of the Jan Mayen Ridge. Map base from J.C. MacLennan (2000).

History of the NE Atlantic

During the opening phase of the NE Atlantic 60-55 Ma ago, central Greenland was positioned above the conduit of the Iceland plume (Lawver and Muller, 1994). Reconstructions of the Iceland plume track relative to the Arctic Ocean, northern Canada and Greenland in combination with the locations of continental flood basalt deposits may suggest that the plume activity goes back to about 130 Ma and caused mid-Cretaceous volcanism along the Arctic Mendeleev and Alpha Ridges and on Ellsmere Island (Forsyth et al. 1986; Lawver and Muller, 1994; Johnston and Thorkelsen, 2000). A large plume head with diameter of more than 2000 km and centered under Greenland accumulated prior to the early Tertiary continental rifting and break-up (Saunders et al. 1997). The accumulation of such a plume head is normally ascribed to the uppermost mantle arrival of a starting plume, but could also result from an existing plume channel by increased plume flux. An extended period of continental lithospheric capping without significant magma tapping or mantle outflow from the plume head region could also contribute to a large plume head accumulation.

The hot plume head material under the continental lithosphere eventually resulted in lithospheric doming and widespread basaltic volcanism (Saunders et al. 1997). Continental breakup, plate separation and incipient ocean crust formation occurred during Chron 24 time (56-53.5 Ma) (Vogt and Avery, 1974). These events were accompanied by the formation of thick seaward-dipping reflector sequences along the E Greenland and NW European margins.

The early seafloor spreading immediately northeast of the Iceland plume track occurred along the now extinct Aegir Ridge. The magnetic anomaly pattern around the Aegir axis is fan-shaped (Talwani and Eldholm, 1977; Skogseid and Eldholm, 1987), and the southern end of the axis bends westwards to link with the Reykjanes Ridge on the southwestern side of the Greenland-Færöy Ridge. The northwestward drift of Greenland reduced the distance between the East Greenland continental margin and the plume stem, leading to another episode of continental rifting along the outermost east Greenland margin at about 36 Ma (Vink 1984). The rifting event detached a microcontinental sliver in the form of the Jan Mayen Ridge from the Greenland margin. The following seafloor spreading along the incipient Kolbeinsey Ridge occurred in parallel with the Aegir axis spreading. Eventually, the Aegir axis became extinct at about 25 Ma. At this stage the Icelandic rift segments became more firmly linked to the Iceland plume stem. The Reykjanes-Kolbeinsey plate boundary passed over the plume stem about 20 Ma ago and has since drifted to a position 150-200 km northwest of the plume axis under Vatnajökull. The Icelandic rift segments of 300-400 km length have jumped repeatedly southeastwards to maintain their position near the plume. These rift jumps at about 24, 15, 7 and 3 Ma, occur by rift propagation within older crust (Fig. 2).

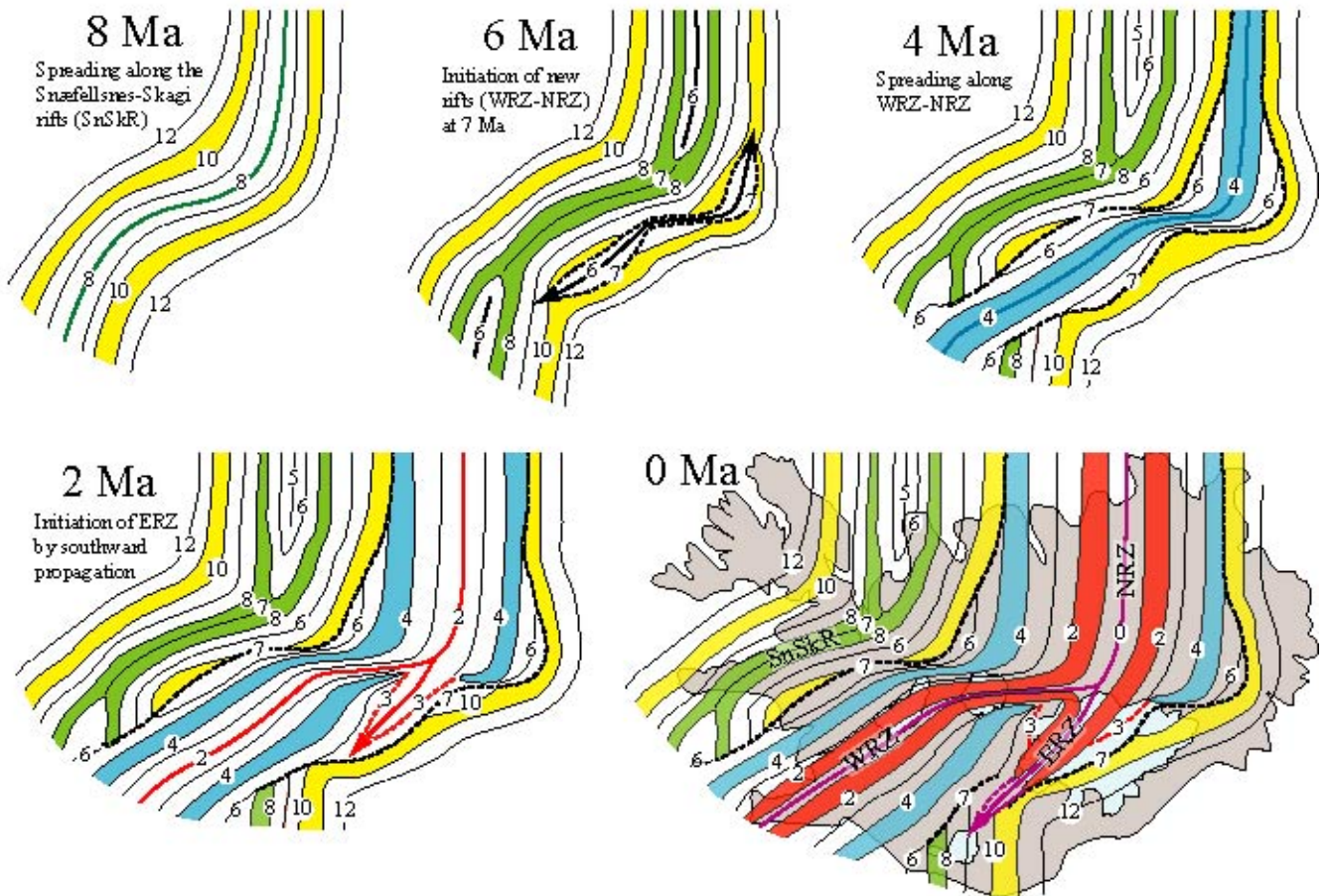


Fig. 2. Crustal accretion, relocation and propagation of the Icelandic rift zones in the last 12 Ma (numbers in Ma). The panels show map views for 8, 6, 4, 2 and 0 Ma. The 8 Ma panel shows the spreading along the Snæfellsnes and Skagi rift zones. The 6 and 4 Ma panels demonstrate the incipient propagation and mature development of the Western and Northern Rift Zones after the new rift initiation at about 7 Ma. The 2 and 0 Ma panels show the southward propagation of the Eastern Rift Zone, initiated at about 3 Ma. Based on data from Sæmundsson (1979) and Jóhannesson (1980) and a synthesis by Ivarsson (1992).

The Icelandic volcanic systems, rift zones and off-rift volcanic zones

The currently active volcanic systems in Iceland are shown in Fig. 3 (e.g. Sæmundsson, 1979; Einarsson, 1991; Jóhannesson and Sæmundsson, 1998). The 40-50 km wide rift zones (Reykjanes, Western, Eastern and Northern Rift Zones) comprise an echelon arrays of volcanic fissure swarms, with 3-4 semi-parallel swarms across the rift zone width. The swarms are 5-15 km wide and up to 200 km in length. With time, they develop a volcanic centre with maximum volcanic production somewhere along their length. The volcanic centres will often develop into central volcanoes with high-temperature geothermal systems, sometimes also with caldera structures produced by large ash-flow eruptions of silicic magma. Each fissure swarm, with or without a central volcano, constitutes a volcanic system. In the non-rifting volcanic flank zones (Snæfellsnes, Eastern and Southern Flank Zones) most of the volcanic centres lack well-developed fissure swarms. The geothermal activity is also generally lower in the off-rift volcanic systems.

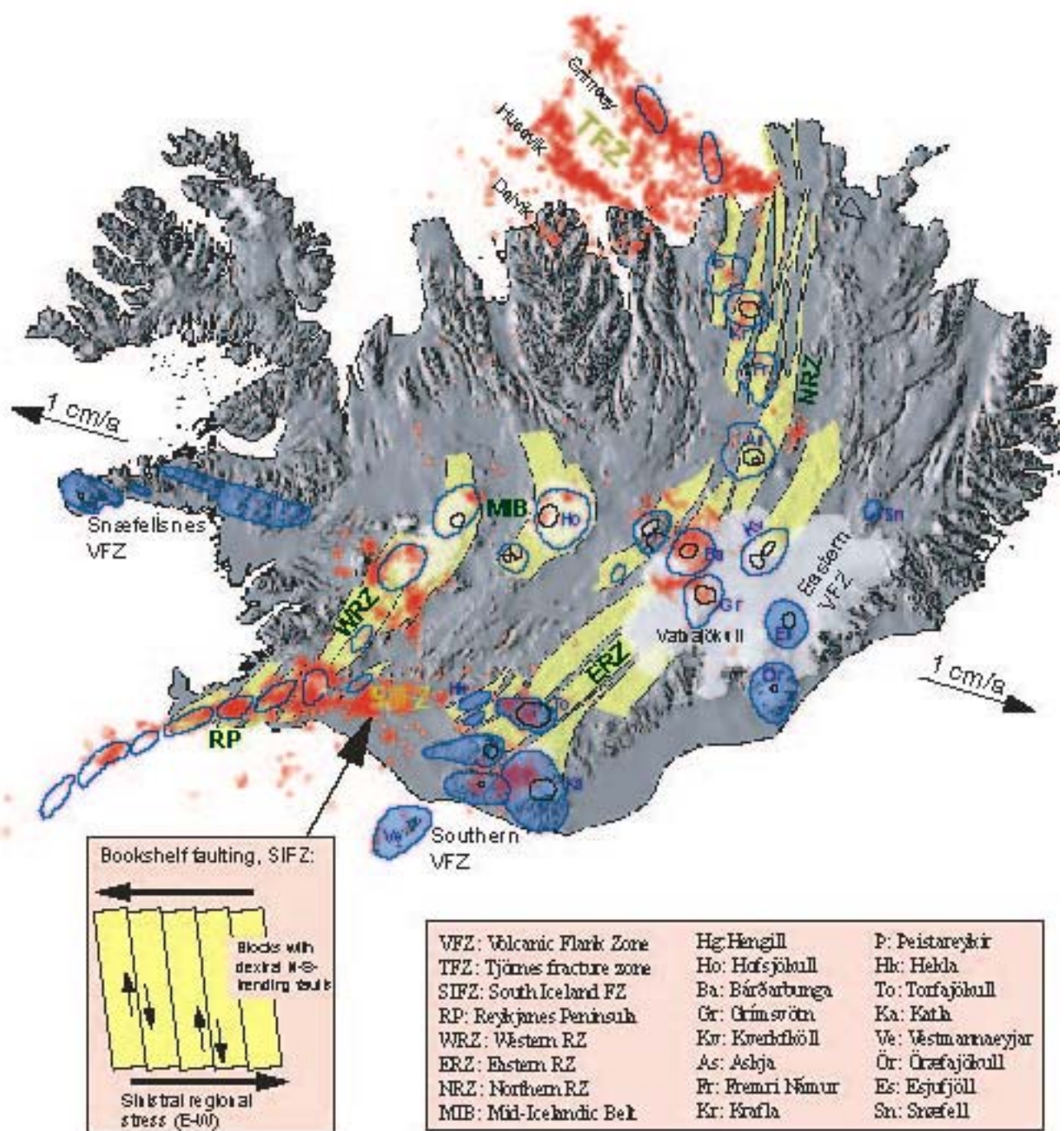


Fig. 3. Volcano-tectonic map of Iceland. Small red crosses (dots) show the epicentres of the 25 000 biggest and best located earthquakes in the period 1994-2000 (from G. Guðmundsson, Icelandic Meteorological Office). Fissure swarms are in yellow. Volcanic centres and calderas are outlined with blue and black lines, respectively. The central volcanic areas of the volcanic flank zones (volcanic off-rift zones) are have blue fill. The bookshelf faulting mechanism is shown in the lower left hand panel.

The Reykjanes Peninsula (RP), with a large component of regional sinistral shear movement, include 4 volcanic systems where the easternmost Hengill system lies at a triple junction between the RP, Western Rift Zone (WRZ) and the South Iceland Fracture Zone (SIFZ). The WRZ continues northeastward from the Hengill central volcano and connects to the Northern RZ via the Mid-Icelandic Belt, which may be considered a “leaky” transform zone. The RP is also in part a leaky transform zone, with dominantly tensional tectonics during periods of high volcanic activity. In volcanically quiet periods, e.g. during the last 760 years, the tectonic situation is mainly that of sinistral shear movement (e.g. Hreinsdóttir et al. 2001).

The Icelandic rift zone system continues northwards from Vatnajökull as the Northern Volcanic Zone (NRZ). The Vatnajökull area is currently the locus of the Iceland plume axis (e.g. Wolfe et al. 1997), and the active rifting along the NRZ is propagating southwards from the plume centre as the Eastern Rift Zone (ERZ). Whereas the Kverfjöll volcanic centre sends its fissure swarms mostly northwards into the NRZ, the fissure swarms of the nearby Grímsvötn and Barðarbunga centres extend mostly southwards along the ERZ. The active rifting parallel to the NE Atlantic plate boundary subsides near the Torfajökull volcanic centre, marking the transition between the ERZ and the Southern Volcanic Flank Zone. The eastward rift jump that initiated the currently active WRZ and NRZ occurred at about 7 Ma, whereas the southward propagation of the ERZ started at 3 Ma. In the period of parallel rifting of the WRZ and ERZ, it appears that the activity is partitioned episodically between the two rift zones. During the last millennium the rifting and volcanism has been almost totally confined to the ERZ. The last eruptive episode in the Hengill volcanic system of the WRZ was in 1000 AD (Svínahruan, Christianity lava), but the western part of the Reykjanes Peninsula had considerable volcanic activity in the thirteenth century.

Whereas the rift zone volcanic systems produce tholeiitic basalts, the major products of the off-rift volcanic zones are mildly alkaline and transitional (tholeiitic to alkaline) basalts (e.g. Sæmundsson, 1979). The tholeiitic rift zone volcanism and mildly alkaline flank zone volcanism in Iceland is equivalent to the main shield-building tholeiitic stage and the pre- and post-shield-building alkaline stages of the Hawaiian volcanoes. The current activity along the Snæfellsnes Peninsula represents dying volcanism unconformably overlying tholeiitic lavas of Tertiary and Pleistocene age, whereas the Southern flank zone has incipient volcanism related to the southward propagation of the ERZ into Tertiary rift zone crust. The mildly alkaline and transitional tholeiitic volcanism along the Eastern Flank Zone may possibly also represent incipient activity associated with an imminent propagation of a new rift zone east of the NRZ-ERZ.

Transform zones and seismic activity

The current offset between the spreading axes of the NRZ and WRZ and the overall plate boundary defined by the Reykjanes and Kolbeinsey Ridges is 100-150 km (Fig. 1 and 3). The Icelandic rift system is connected to the oceanic spreading ridges via the dextral Tjörnes Fracture zone linking the NRZ with the Kolbeinsey Ridge and the sinistral South Iceland Fracture Zone and Reykjanes Peninsula, linking the ERZ to the Reykjanes Ridge. This is the current situation with no rifting in the WRZ.

During periods with the south Iceland spreading activity mainly confined to the WRZ, the Mid-Icelandic Belt must operate as a sinistral transform zone linking the NRZ to the WRZ.

The right-lateral Tjörnes Fracture Zone (TFZ) is a broad area of seismicity where the activity is confined to three parallel seismic zones (Einarsson, 1991, Fig. 3). The northernmost Grímsey zone is the scene of frequent earthquake swarms, lasting for days or weeks. The maximum magnitude of such earthquakes exceeds 5, and the fault-plane solutions show right-lateral strike-slip. The middle Húsavík zone enters land at the town of Húsavík and continues eastwards as the dextral Húsavík fault linking up with the normal faults of the Þeistareykir fissure swarm. A major earthquake sequence occurred along this fault in 1872, causing extensive damage and surface ruptures in town of Húsavík. The third and southernmost Dalvík zone is much less active (Fig.). The main activity appears to be confined to the western part of the zone, but a damaging earthquake occurred in the town of Dalvík in 1934. A 1963 magnitude 7 earthquake with dextral fault-plane solution in the mouth of Skagafjörður represents the westernmost part of the Dalvík zone.

The regional stress field of the South Iceland Fracture Zone (SIFZ) and the Reykjanes peninsula (RP) is EW-directed and sinistral (Fig.3). The earthquakes, however, occur along N-S-oriented, dextral faults. The mechanism for this rotational movement and bookshelf-faulting is illustrated in Fig. 3. The SIFZ extends from the Hekla central volcano in the east to the Hengill central volcano and triple junction. Major seismic events including several (often 2-4) earthquakes of magnitude 6.5 to 7, over periods of a few days to a few years occur with average intervals of 80-100 years. Each of these sequences typically begins with a large earthquake in the eastern part of the SIFZ, followed by slightly smaller earthquakes further west. The last few and well-recorded seismic sequences occurred in 1732-34, 1784, 1896-1912 and 2000. The big earthquakes in 2000 included two magnitude 6.6 events on June 17 and June 21. The second of these occurred about 20 km west of the first one. Both of the earthquakes resulted from up to 2.4 m dextral displacement on near-vertical faults. The fault planes are 15-16 km long and 9-10 km deep (Pedersen et al. 2001). The seismic sequence also included numerous earthquakes along the entire length of the SIFZ and RP. Fig. 4 illustrates the seismic activity in Iceland during selected intervals in the period from February 2000 to May 2002. As shown in the May 2002 diagram, aftershocks defining the two main epicentral faults continue to the present time.

The June 17 main shock triggered three large magnitude deformation events about 65, 80, and 90 km further west on the Reykjanes Peninsula. Accurate seismological estimates of the magnitude of these events are hampered by the near saturation of the seismometers at this time, and the seismic signatures are hidden by the waveform of the main shock. Modelling of high-quality InSAR-data (satellite radar interferometry), however, shows that the Kleifarvatn event at the distance of 80 km from the main shock epicentre corresponds to a magnitude 5.8 earthquake (Pagli et al., *Geophys. Res. Lett.* submitted). The dextral fault movement of this event produced severe ground motion and surface rupturing. Despite its size, the Kleifarvatn event does not appear in the worldwide seismic catalogues, and the aftershock activity along the epicentral fault is anomalously low. Pagli et al. have therefore concluded that the fault slip was largely aseismic (quiet/slow earthquake), in spite of large surface destruction and fault displacement. The events at 65 and 80 km distance from the main shock were both triggered by the arrival of the Love waves and occurred 27 and 31 seconds, respectively, after the main shock. The westernmost event occurred 4 min 16 second after the Kleifarvatn event, which may indicate that the

Kleifarvatn deformation was somewhat protracted. The lake level in Kleifarvatn, which is electronically monitored, dropped dramatically after the earthquake, and the lake is still draining. During the initial 14 months, the lake level fell 4 m, corresponding to about 10% of the lake volume.

Earthquakes during selected weeks in the period February 2000 - May 2002

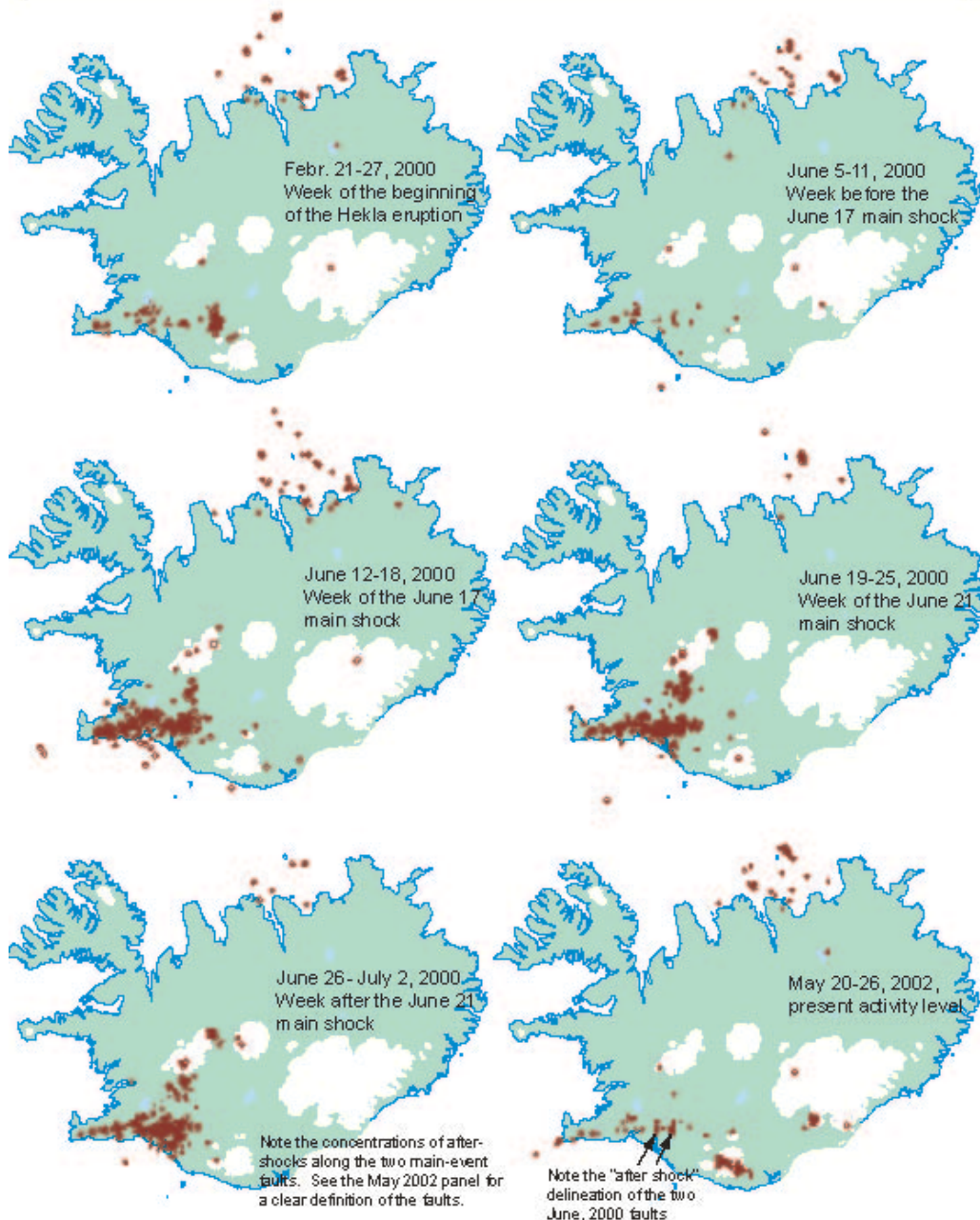


Fig. 4. Earthquake activity in selected weeks in the period 2000-2002. The earthquake magnitudes (ranging from 0 to 6.6) are indicated by different ring sizes. Compiled from the www-page of the Icelandic Meteorological Office (Veðurstofa Islands, <http://www.vedur.is>).

Subglacial and subaerial volcanism

Iceland has been variably covered by large ice sheets and smaller ice caps during the last 3 Ma (e.g. Bourgeois et al. 1998). The heat transfer to the ice during subglacial volcanism is so efficient that the magma enters a subaqueous environment in the form of a water-filled ice cavity or ice-dammed lake (Allen, 1980). The character of the eruption products depends on the hydrostatic pressure at the vent and the internal volatile pressure in the magma. Decreasing external pressure will lead to a transition from pillow lava via pillow breccia to hyaloclastite tuff. Most Icelandic subglacial volcanic mountains comprise cores of pillow lava, overlain by pillow breccia and hyaloclastite tuff, reflecting decreasing hydrostatic pressure as the mountain grows higher during the eruption. If the vent area becomes subaerial, the volcanism may change to lava eruptions. Icelandic hyaloclastite mountains, capped by subaerial lava flows have generally steep sides and flat tops and are referred to as table mountains.

The landforms developed by subaerial and subglacial volcanism are widely different. During subaerial conditions the predominant basaltic eruption products are lava flows from fissure eruptions or gently sloping shield volcanoes. The fissure eruption lavas, in particular, tend to smooth the topography of the rift zone floor. Some of the postglacial lava flows in Iceland have travelled for distances of 50-100 km, and some of these flows have travelled outside the rift zones where they originated. The Eldgja (934-940 AD) and Laki (1783-1784 AD) lava flows with volumes of 19.6 and 14.7 km³, respectively, are examples of such flows (Thordarson and Self, 1993; Thordarson et al. 2001). In contrast, subglacial fissure eruptions and subglacial “shield volcanism”, respectively, produce high and narrow hyaloclastite ridges and steep-sided table mountains, respectively. Subglacial volcanism will therefore tend to build high topography. This mechanism is accentuated by recurrent glaciations followed by interglacial periods of partial ice cover. The high areas under the major Icelandic glaciers, and especially under Vatnajökull, therefore grow more rapidly in elevation compared to the surrounding volcanic zones (Helgason and Duncan, 2001).

The highest area in Iceland, under Vatnajökull, is also the site of the Iceland plume axis. The most recent part of the plume track across Iceland coincides with the two other major glaciers, Hofsjökull and Langjökull (Fig 1, 2).

Crustal structure and upper mantle viscosity

The Iceland Plateau and the Greenland-Færöy Ridge are conspicuous bathymetric features in the NE Atlantic (Fig. 1). Fig. 5 demonstrates that these shallow areas have anomalously thick “oceanic” crust resulting from the high magma production in the plume (Kaban et al. 2002). The Moho beneath Iceland is seismically diffuse because of low densities and seismic velocities in the uppermost mantle and relatively high densities and velocities in the lower crust. Several recent models, however, agree that the crustal thickness varies from about 40 km under Vatnajökull to less than 20 km under the northern part of the NRZ and the Reykjanes Peninsula (Fig. 5; Menke et al. 1998; Darbyshire et al. 2000; Allen, 2001; Kaban et al. 2002). The crustal thickness presumably reflects a volcanic productivity maximum above the plume axis.

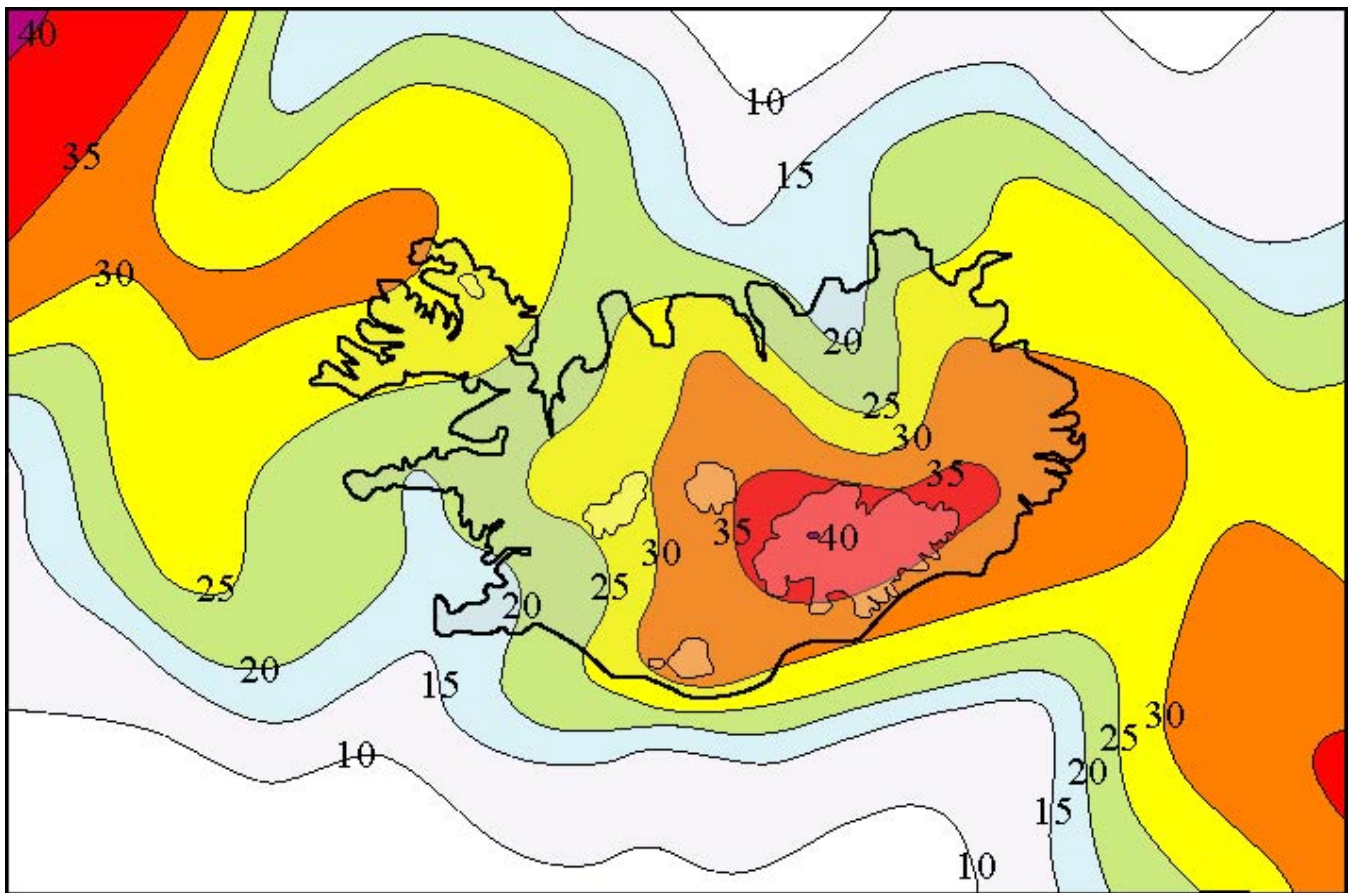


Fig. 5. Crustal thickness (km) variations across Iceland and the adjacent parts of the Greenland-Færøe Ridge (simplified after Kaban et al. 2002).

The maximum elevation of subaerial vents on subglacial table mountains in the Icelandic rift zones constrain the approximate ice thickness during peak glaciations (e.g. Walker, 1965). This analysis gives minimum ice thickness values of more than 1000 m in central Iceland. Offshore terminal morains indicate that the ice margin was 50-100 km outside the present coastline during maximum glaciation (Bourgeois et al. 1998). During the period from 14500 to 11800 BP, however, the margin was inside the present coastline. Modelling of the rapid post-glacial rebound constrains the viscosity of the asthenospheric mantle to be less than 10^{19} Pa s (Sigmundsson, 1991). Similar modelling of the ongoing isostatic rebound around Vatnajökull as a result of 20th century glacial melting gives viscosity values as low as $7 \cdot 10^{16}$ to $3 \cdot 10^{18}$ Pa s at shallow levels corresponding to the lower crust and uppermost mantle (Thoma and Wolf, 2001). Such low viscosity values, compared to an average upper mantle viscosity of $3 \cdot 10^{20}$ Pa s (Lambeck and Johnston, 1998), reflect the thermal anomaly of the Iceland plume.

Mantle melting and crustal reprocessing

The ascending mantle column beneath Iceland melts in response to the pressure release (e.g. MacLennan et al. 2001a, b). Because the adiabatic gradient followed by the ascending mantle material has a steeper dp/dT -slope than the solidus, the melting will proceed continuously after intersection of the solidus. Systematic correlations between major and trace element and radiogenic isotope ratios (Sr-Nd-Hf-Pb-Os-isotopes) in Icelandic lavas demonstrate that the mantle source is heterogeneous. The geochemistry of basalts from the nearby ridge segments, Vesteris seamount, the Jan Mayen area and the Early Tertiary successions of Greenland and the British Isles indicate that the upper mantle in most of the NE Atlantic has the same chemical characteristics as the current Iceland plume source (e.g. Thirlwall et al. 1994; Saunders et al. 1997; Trønnes et al. 1999). It is likely that most of the upper mantle in this area was supplied from the lower mantle by the accumulating early Tertiary Iceland plume head. The first order geochemical heterogeneity of the plume source comprises an enriched and a depleted component. Both of these were derived by Paleozoic subduction of oceanic lithosphere that was hydrothermally altered at a mid-ocean ridge and geochemically reprocessed by dehydration and fluid loss during subduction (Thirlwall 1995, 1997; Trønnes et al. 1999, Chauvel and Hemond, 2000; Skovgaard et al. 2001; Breddam, 2002). Whereas the enriched component derives from the hydrothermally altered basaltic (and sedimentary) upper part of the lithosphere, the depleted component is largely the lower oceanic lithosphere, including cumulate rocks and strongly melt-depleted harzburgite. Preliminary and unpublished Os-isotopic data, however, indicate that the Icelandic mantle source may also contain additional Proterozoic and/or Archean recycled components.

The Icelandic rift zones and the nearby Reykjanes and Kolbeinsey Ridges are characterized by He-isotopic ratios that, in spite of considerable scatter, decreases in both directions away from $^3\text{He}/^4\text{He}$ -values (normalized to the atmospheric composition) of more than 20 near the plume axis (Kurz et al. 1985; Poreda et al. 1986; Breddam et al. 2000). High $^3\text{He}/^4\text{He}$ -ratios are also found in minerals in some of the oldest Tertiary basalts in Iceland located near the plume track, as well as in early Tertiary basalts in east Greenland (Hilton et al. 1999; Graham et al. 1998). However, the volcanic flank zones in Iceland and the alkaline rocks of Jan Mayen have mostly low $^3\text{He}/^4\text{He}$ -ratios corresponding to the degassed MORB-source (Kurz et al. 1982; Poreda et al. 1986; Sigmarsson et al. 1992). The He-isotopic variation therefore appears to be linked both temporally and spatially to the Iceland plume stem. The production of elevated $^3\text{He}/^4\text{He}$ -ratios in strongly melt-depleted and degassed mantle (e.g. recycled oceanic harzburgitic lithosphere) requires only a small amount of primordial He-addition, either from the core or from the lower mantle (e.g. Coltice and Richard, 1999; Porcelli and Halliday, 2001).

The decompressional melting of the ascending heterogeneous mantle under Iceland starts at the intersection of a solidus for a mantle with 600-900 ppm H_2O (e.g. Nichols et al. 2002; Jamtveit et al. 2001). The deepest level of the wet solidus intersection will be near the centre of the Iceland plume stem (Fig. 6). H and He are very incompatible elements and will be efficiently partitioned into the earliest melt fractions, resulting in dehydration of the residual mantle minerals. The dehydration increases the viscosity of the plume material (e.g. Hirth and Kohlstedt, 1996; Ito et al. 1999), presumably leading to partial stagnation and lateral deflection of the plume. Some of the deflected plume material, which has experienced only minor melt extraction will ascend along relatively flat trajectories (Fig. 6).

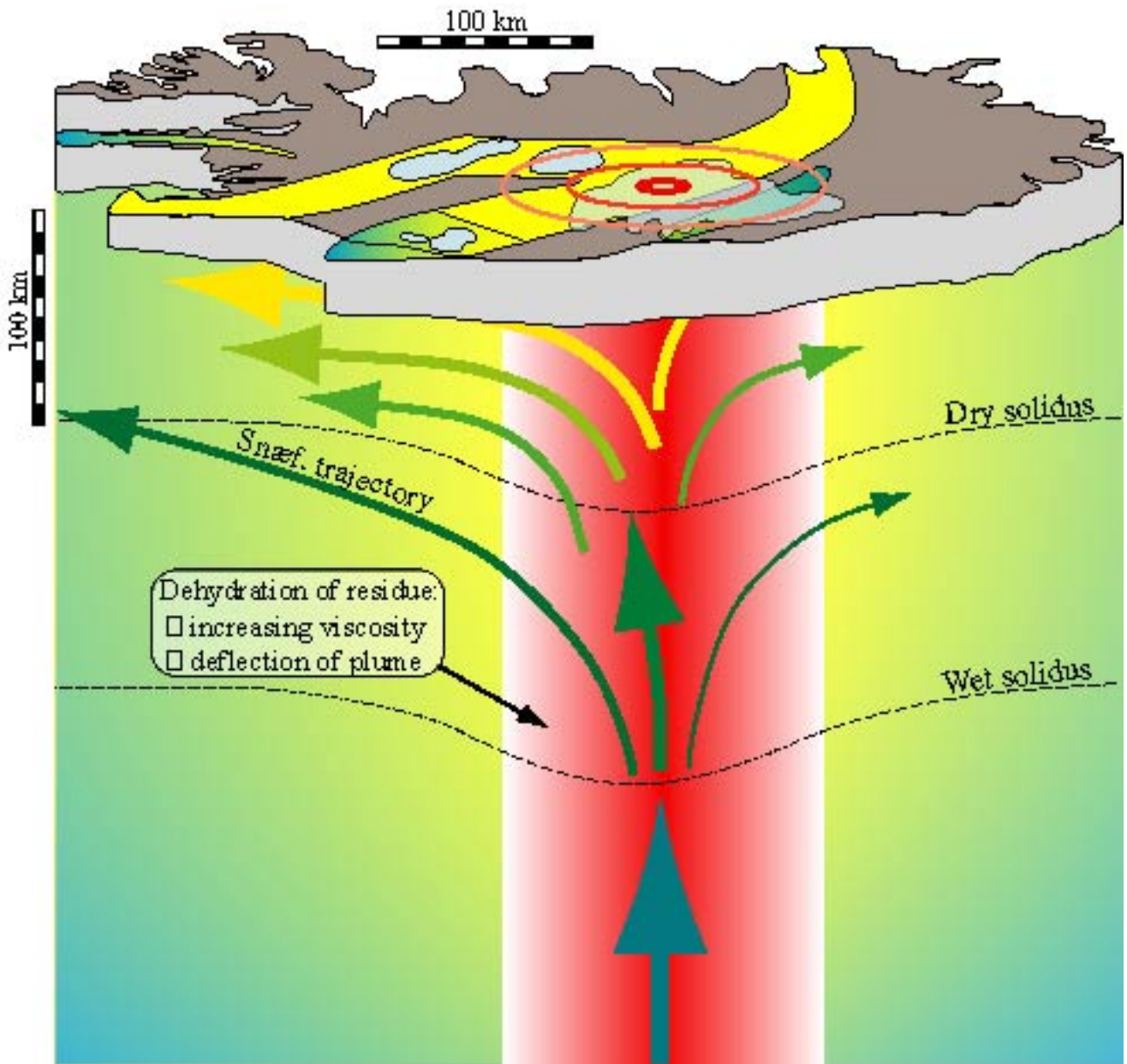


Fig. 6. Schematic illustration of flow trajectories and melting regime of the Iceland plume in relation to the volcanic rift zones (yellow) and off-rift zones (gradients from yellow via green towards blue). The degree of melt depletion of the mantle surrounding the axial plume stem is indicated schematically by colour shades from blue (least melt-depleted) to yellow (most melt-depleted). The same colour coding is used for the flow trajectories. The illustration is a combination of a three dimensional perspective and an east-west vertical cross section, with approximate plume location and dimensions based on Wolfe et al (1997), Shen et al. (2002) and Ito (2002). The crustal thickness is in accordance with Kaban et al. (2002). The initial lateral deflection of the plume flow near the wet solidus is caused by viscosity increase related to the initial dehydration melting (e.g. Ito et al. 1999).

This material may later encounter solidus temperatures at more plume-peripheral positions, e.g. under the Snæfellsnes volcanic flank zone. The bulk of the plume material, however, will rise along steeper trajectories and be deflected at shallower levels, where feeding of plume material into the melting zones under the rift zones also become an important process.

During decompression melting, small melt fractions are continuously extracted from the ascending solid residue as melting proceeds. The initial melt fractions, derived at the greatest depths, are formed from the enriched and fertile component of the source. As melting and ascent continues, the initial enriched melt fractions become progressively diluted with depleted melt fractions from the refractory component of the source.

The melting of the mantle source is terminated when the source is prevented from further ascent by a mechanical boundary layer (lithosphere). Under Iceland, there is probably no mantle lithosphere beneath the rather thick crust. It is even possible that the source ascent under some of the volcanic systems in the Vatnajökull area proceeds to shallower level than the geophysically modelled Moho at 35-40 km depth.

The mantle-derived magmas are fed into crustal magma reservoirs where they can differentiate by a combination of fractional crystallization and anatectic assimilation of wall rock (e.g. Nicholson et al. 1991). The crustal anatectic contributions are especially significant in the most evolved central volcanoes in the rift zones. These volcanoes tend to develop extensive and deep geothermal systems, where basalts and hyaloclastites are heavily altered and hydrated. The volcanic loading of the surface produces a downgoing mass-flow of altered basalts that undergo prograde metamorphism from zeolite- to greenschist- to amphibolite-facies and anatexis under the high geothermal gradient conditions (Oskarsson et al. 1982). The high proportion of rhyolitic extrusives in Iceland is unique in the global "oceanic" context. The Torfajökull central volcano is the largest rhyolite center in the present-day terrestrial oceanic environment. The eruptions of rhyolites and other silicic extrusives are confined to the most evolved central volcanoes. Many of these have also erupted large-volume ash-flow deposits, associated with significant caldera collapses.

The most common type of basaltic volcanism along the Icelandic rift zones are eruptions from fissure swarms connected to a central basaltic magma reservoir under a volcanic centre. The largest fissure eruptions of 10-20 km³ are generally quite evolved and homogenous, demonstrating extensive fractional crystallization and assimilation of hydrothermally altered crust. The most common lava type exposed in the Tertiary volcanic successions of eastern and western Iceland are similarly evolved tholeiitic lavas (Hardarson and Fitton, 1997)

Another important basaltic volcano type is the large shield volcanoes that are scattered along the rift zones and appears unrelated to the volcanic systems and their fissure swarms. These monogenetic shield volcanoes erupted primitive olivine tholeiitic magmas fed by continuous overflow from summit lava lakes. The eruptions appear to have been nearly continuous, with the entire lifetime of the volcanoes completed within about 100 years (Rossi, 1996). The volume of some of the early post-glacial shield volcanoes range up to 20 km³. Similar interglacial shield volcanoes of Eemian age are common along the rift zones. Many of the table mountains are subglacial analogues of the shield volcanoes.

Glacio-isostatic modulation of volcanism

The low viscosity of the high-temperature upper mantle and lower crust makes the pressure release melting very sensitive to rapid glacial unloading and rebound. The link between deglaciation and increased volcanic productivity has been noted in many different parts of Iceland: the Reykjanes Peninsula (Jakobsson et al. 1978), the Veiðivötn fissure swarm in the ERZ (Vilmundardóttir and Larsen, 1986), the Askja (Sigvaldason et al, 1992) and Þeistareykir (Slater et al. 1998) volcanic systems of the NRZ and the Snæfellsjökull system (Hardarson and Fitton, 1991). The unloading effect leads to nearly 100 times higher volcanic productivity than that of recent times (< 5 ka BP) and during glacial periods (Maclennan et al. J. Geophys Res., submitted).

The magmas erupted during the periods of maximum productivity in the glacial rebound stages are the most primitive magmas recorded in Iceland. Almost all of the accessible picritic eruption units and most of the large monogenetic shield volcanoes of primitive olivine tholeiitic composition are of early post-glacial age. These relationships demonstrate that the crustal magma plumbing system has inadequate capacity for melt processing during periods of large magma supply from the mantle.

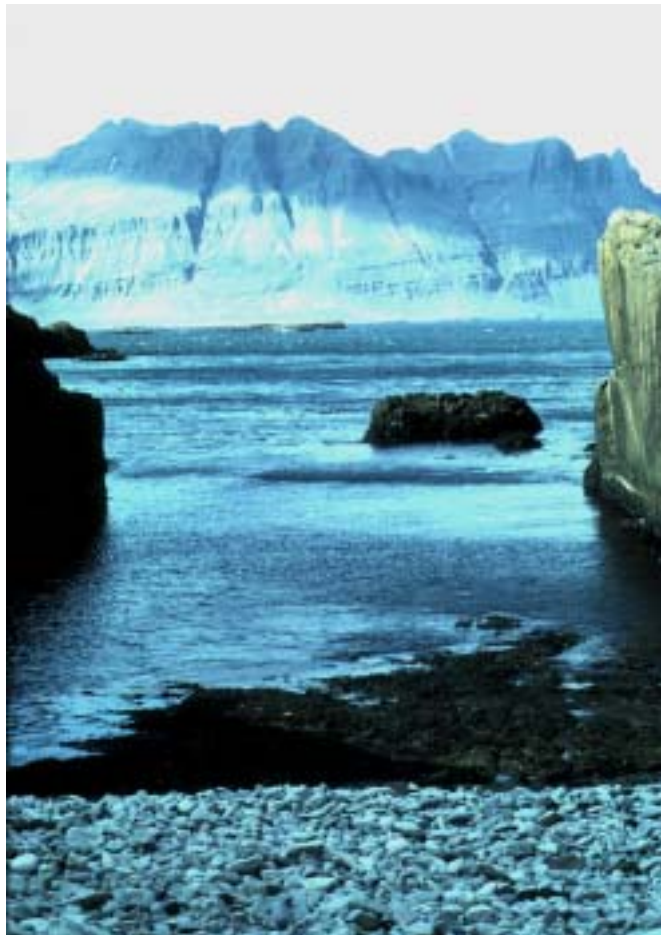


Fig. 7. View towards north across Breiðdalsvík, eastern Iceland. The lava units slope in a westerly direction towards the Northern Rift Zone. The slope decreases slightly upwards from sea level to the highest exposed elevation of 600-700 m.a.s.l. Photo: R. Trönnnes.

Dynamics of crustal accretion

The approximately 50 km wide rift zones with 2-4 parallel volcanic systems in en echelon patterns are continuously covered by new lava flows and hyaloclastite mountains. Nearly the entire rift zone area is covered by eruption units from Eem, Weichsel and Holocene. Older units are subsiding under the new surface load. The volcanic productivity of the Icelandic rift zones is anomalously high relative to the low spreading rate of about 10 km/Ma in each direction. The thin and weak lithosphere cannot support much weight, resulting in the rapid subsidence of the partially altered and hydrated volcanic pile.

Observations from the Tertiary lava areas east and west of the currently active rift zones confirm the continuous subsidence of the volcanic pile. Vertical sections through the Tertiary lava pile in glacially eroded valleys and fjords expose the uppermost 1500 m of extrusive rocks. The lavas dip gently towards the current or extinct rift zones (e.g. Bödvarsson and Walker, 1964; Fig. 7). In coastal sections the dip typically increases gradually from near zero at the highest exposed levels of about 1000 m elevation to about 5-10° at sea level. The increasing dip is matched by thickening of individual lava units (lava groups) in the dip direction towards the corresponding rift zone. The regional flexuring and tilting is a result of the continuous loading and subsidence of the rift zone crust. Whereas the loading is most pronounced under the volcanic centres, the average, time integrated (3-7 Ma) subsidence will be highest along the rift zone axis and decrease towards the rift zone margins. Based on these observations and other geophysical constraints, Palmason (1973) developed a dynamic model for the crustal accretion in Iceland. A simplified sketch of this model is shown in Fig. 8. The subsiding pile of lava and hyaloclastite units, including heavily altered and hydrated rocks, subsides through high geothermal gradients and undergoes prograde metamorphism to zeolite, greenschist and amphibolite facies and partial melting producing rhyolitic magmas (Oskarsson et al, 1982). When the partial melting occurs along the walls of basaltic magma chambers, the rhyolitic melt fractions will mix with the basaltic liquid and promote the magma evolution (e.g. Oskarsson et al, 1982; Trønnnes, 1990; Nicholson et al. 1991). In other areas the rhyolitic melt fractions will segregate and give rise to silicic intrusions and extrusions.

Such crustal reprocessing, which are important in the Icelandic crustal accretion, occur only to a very limited extent along the mid-oceanic spreading ridges. Although the volcanic productivity of the present oceanic ridges is generally too low for such recycling, the current Icelandic scenario may be a good analogue for the formation of Archean greenstone belts.

An interesting corollary of the Palmason (1973) model is that all of the eruption products deposited near the rift zone axis will subside to deep levels and not be exposed by later erosion of the volcanic pile in eastern and western Iceland. Only the lavas that run for a considerable distance toward the margin of or even outside the rift zone have a chance of being exposed some million years later. The studies of Hardarson and Fitton (1997) have demonstrated that the great majority of Tertiary lavas in NW Iceland are uniformly evolved. The chemical composition of the exposed Tertiary lavas corresponds closely to the chemistry of the largest lava flows of late Pleistocene and Holocene age. These lavas, in turn, tend to flow rather long distances towards the rift zone margins or even outside the rift zones.

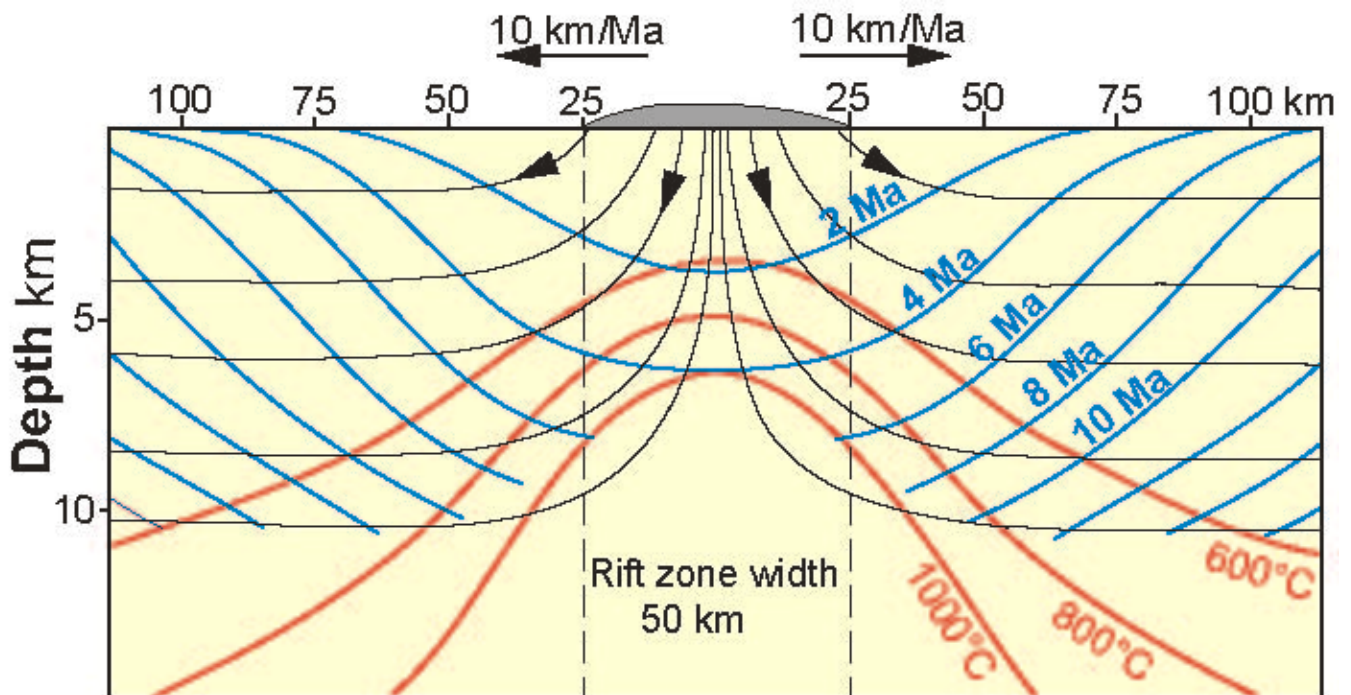


Fig. 8. Simplified model of Icelandic rift zone dynamics (from Palmason 1973, and later studies). The black, blue and red lines are mass trajectories, age contours (Ma) and temperature contours ($^{\circ}\text{C}$). Partial melting of hydrated mafic lithologies will start at about 5 km depth under the central part of the rift zone.

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