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DYNAMICS OF VOLCANIC SYSTEMS IN ICELAND: Example of Tectonism and Volcanism at Juxtaposed Hot Spot and Mid-Ocean Ridge Systems

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■ **Abstract** Volcanic systems are swarms of tectonic fractures and basalt volcanoes formed as a result of plate-pull (as the plates are pulled apart) associated with the mid-ocean ridges and the magma dynamics of the Iceland Mantle Plume. Most systems are 40–150 km long, 5–20 km wide, and develop a central volcano. They supply magma to all eruptions in Iceland. Data obtained in the last few years have greatly improved our knowledge of their volcanotectonic environment; as a result, the geometry of the plume is better constrained, and the crust, previously considered thin (~10 km), is now modeled as thick (~20–40 km). Depending on the location of the volcanic systems, their activity either decreases or increases faulting in the two main seismic zones. From this, we can infer that emplacement of the feeder-dike to the largest historical eruption in Iceland (that of Laki in 1783) increased shear stress in the South Iceland Seismic Zone and almost certainly triggered the largest (M~7.1 in 1784) historical earthquake in Iceland.

*Addendum

INTRODUCTION

Iceland owes its existence to the mantle plume that supplies magma to its volcanic systems (Figure 1). Most of these systems are giant swarms of cracks that develop within the volcanic zones in response to crustal spreading by plate pull. Their dynamics are controlled partly by the plate dynamics of the North American and Eurasian plates, and partly by the fluid dynamics of the Iceland Mantle Plume. Thus, they offer a fascinating opportunity to study the evolution of volcanotectonic systems that develop at a juxtaposed mantle plume and mid-ocean ridge system.

Understanding of the volcanic systems of Iceland has improved greatly in the past 20 years since the concept was introduced (Jakobsson 1979a, Saemundsson

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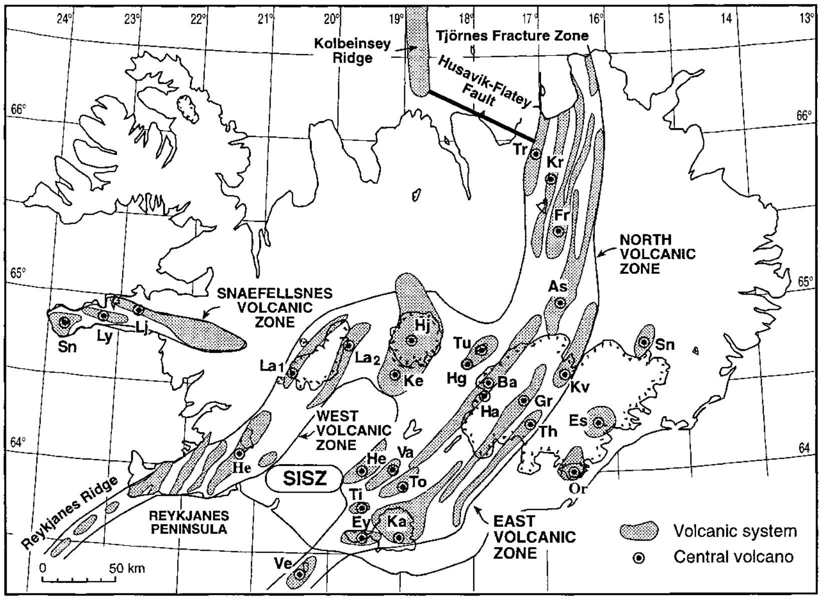

Figure 1 Volcanic zones and systems in Iceland. The systems are: Tr = Theystareykir, Kr = Krafla, Fr = Fremri-Namur, As = Askja, Kv = Kverkfjoll, Th = Thordarhyrna, Gr = Grimsvotn, Ha = Hamarinn, Ba = Bardarbunga, Tu = Tungnafellsjokull, Hg = Hagongur, Ka = Katla, Ey = Eyjafjallajokull, Ve = Vestmannaeyjar, Ti = Tindfjallajokull, Va = Vatnafjoll, He = Hekla, Hj = Hofsjokull, Ke = Kerlingarfjoll, La1&2 = Langjokull, He = Hengill. The other systems on the Reykjanes Peninsula (from east to west) are Brennisteinsfjoll, Trolladyngja, and Reykjanes. Off-coast are Eldey, Geirfuglasker, and Eldeyjarbodi. In the Snaefellsnes Volcanic Zone are the systems Sn = Snaefellsjokull, Ly = Lysuskard, and Lj = Ljosufjoll. Outside the East Volcanic Zone are the systems Or = Oraefajokull, Es = Esjufjoll, and Sn = Snaefell. The rift zone (its axis is indicated in Figure 3) comprises the North Volcanic Zone, the West Volcanic Zone, and the East Volcanic Zone to the south tips of the volcanic systems of Bardarbunga and Grimsvotn. Also indicated are the main ocean-ridge discontinuities, the Husavik-Flatey Fault of the Tjornes Fracture Zone, and the South Iceland Seismic Zone (SISZ), located between the overlapping West and East Volcanic Zones. Data are from Jakobsson (1979a), Saemundsson (1979), AT Gudmundsson & Kjartansson (1996), and Johannesson & Saemundsson (1998).

1979). Not only have we obtained a wealth of new data on their structure, geochemistry, and dynamics, but the volcanotectonic environment in which they develop is now much better understood. During these two decades, extensive studies have been done involving active and inactive volcanic systems, accurate measurements of local and regional plate movements, and geochemical and geophysical exploration of the Iceland Mantle Plume.

Most references in this review define the current state of knowledge of the dynamics of volcanic systems in Iceland and their environment. A few references guide the reader to topics that, although important and connected to the dynamics of the volcanic systems, are not of central concern in this paper. Many citations refer to papers published in the last few years, but I have also included references to several older papers of historic importance. Although this is primarily a review paper, I decided to present a number of suggestions, some of which are new, so as to provide a consistent overall picture of the volcanic systems in Iceland and their interaction with the mantle plume and the ocean-ridge discontinuities. Many of these suggestions will doubtless have to be refined and improved in the future.

I begin with a review of the volcanotectonic environment in Iceland and how the volcanic systems interact with it. This part summarizes relevant knowledge of the plate boundaries and plate movements in Iceland and its mantle plume, as well as models on locking and unlocking of the main seismic zones by activity in the volcanic systems. Next, I summarize the systems' infrastructure, focusing on the rift-zone systems. These summaries form a basis for understanding the dynamic evolution of the volcanic systems, the subject of the third main part of the paper. In that part, I focus on the general evolution of the systems—from birth, through maturity, to death—and the effects that the formation of a shallow magma chamber has on their subsequent development.

PLATE BOUNDARIES AND MOVEMENTS

Iceland is located at the junction between the Kolbeinsey Ridge in the north and the Reykjanes Ridge in the south (Figures 1 and 2). The Kolbeinsey Ridge trends roughly N10°E, subperpendicular to the spreading vector. Its seismic structure, essentially uniform, indicates a crustal thickness exceeding that of a normal mid-ocean crust by 1–1.5 km (Kodaira et al 1998). In contrast to the Kolbeinsey Ridge, the Reykjanes Ridge, trending roughly N36°E, is an oblique-spreading ridge. Its neovolcanic zone consists of axial volcanic ridges, each of which trends subperpendicular to the spreading vector, with an average spacing of 14 km and an overlap of roughly one third of their lengths in an overall en echelon arrangement (Searle et al 1998).

The surface expression of the mid-ocean ridge in Iceland is the zone of active volcanism: the neovolcanic zone. This zone, covered with rocks of the Bruhnes magnetic epoch ($<$0.8 Ma), corresponds to the magnetic plate boundaries of the mid-ocean ridges, not to their neovolcanic zones which, by definition (Macdonald 1982), are of Holocene age. The neovolcanic zone has three main segments: the North Volcanic Zone, the West Volcanic Zone and the East Volcanic Zone (Figure 1). The Snaefellsnes Volcanic Zone, an old Tertiary volcanic zone reactivated at 2 Ma, is propagating to the southeast, whereas the southern part of the East Volcanic Zone is propagating to the southwest.

In addition to rift zones and propagating rifts, the plate boundaries in Iceland include a transform fault and an overlapping spreading center (Figure 1). In North Iceland the Tjornes Fracture Zone, which is partly exposed on land, is a transform

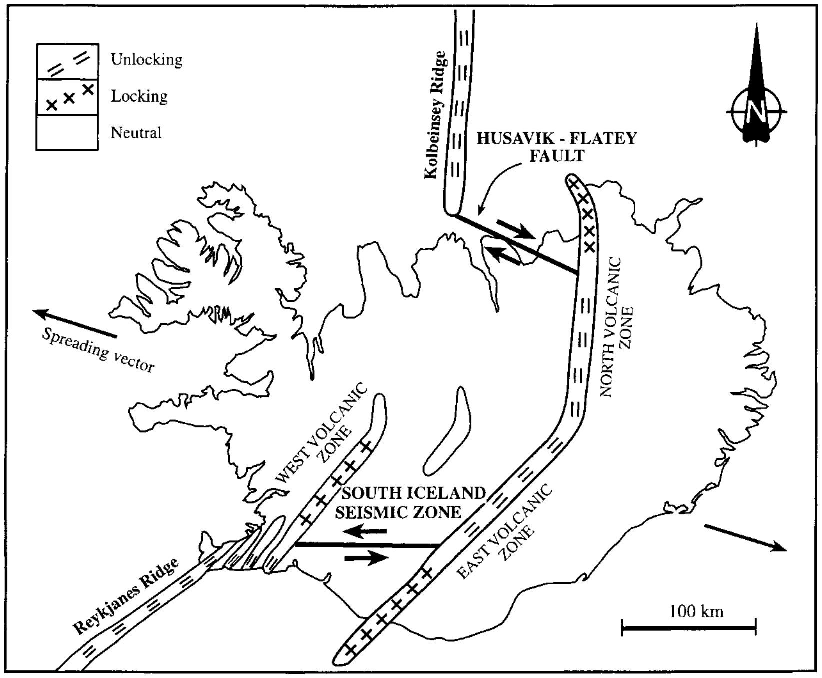

Figure 2 Dike injections in the volcanic systems can temporarily lock (suppress earthquakes) or unlock (trigger earthquakes) in the main seismic zones in Iceland, the Husavik-Flatey Fault, and the South Iceland Seismic Zone. Static magma overpressure in a regional dike can reach 10–50 MPa, resulting in a high horizontal compressive stress (cf Figures 3 and 4). Dike injection in the parts of the volcanic zones marked by crosses would tend to suppress, whereas dike injection in the parts marked by parallel ticks would tend to trigger, earthquakes in the seismic zones

fault connecting the plate boundary of the Kolbeinsey Ridge with that of the North Volcanic Zone. The Fracture Zone's main section is the Husavik-Flatey Fault, a major dextral strike-slip fault that corresponds to the transform-tectonized zone of a typical oceanic transform fault. The South Iceland Seismic Zone (SIZS) is a zone of complex faulting located between the overlapping parts of the West Volcanic Zone and the East Volcanic Zone (Figure 1).

The magnitude and direction of plate movements in the Iceland area are now well known (DeMetz et al 1990, 1994). The size and direction of the spreading vector varies slightly from the north to the southern part of the country: At the Tjornes Fracture Zone the total spreading (opening) rate is 1.79 cm/year in the direction N106°E, but 1.85 cm/year in the direction N104°E at the South Iceland Seismic Zone (DeMets et al 1990, 1994; Sigmundsson et al 1995; Jonsson et al 1997; Vadon & Sigmundsson 1997). For Iceland as a whole, the average spreading rate is 1.8 cm/year in the direction N105°E.

MANTLE PLUME AND CRUSTAL STRUCTURE

A widely accepted conceptual model of mantle plumes is that they are long, subvertical, and narrow columns of hot, buoyant, low-density material that flows upward from deep regions in the mantle or the core-mantle boundary. The surface manifestations of such plumes are regions of great volcanic activity, high heat flow, and broad swells, referred to as hot spots—one of which is Iceland. The existence of the Iceland Mantle Plume, an idea proposed several decades ago (Wilson 1965, Morgan 1971), was supported by the results of Tryggvason et al (1983), and later refined by Ryan (1990). Extensive studies in the past few years by Bjarnason et al (1996), Wolfe et al (1997), Shen et al (1998), and Allen et al (1999) have greatly improved our understanding of the structure of the Iceland Mantle Plume.

In accordance with the general picture of mantle plumes, the Iceland Plume is modeled as a cylindrical zone of hot material with low seismic (P-wave and S-wave) velocities that extends to a depth of at least 400 km. The plume may reach a depth of more than 600–700 km (Shen et al 1996, Shen et al 1998) and may even reach the core-mantle boundary at a depth of 2700 km.

Early attempts to model the general geometry of the Iceland Plume yielded a picture surprisingly similar to that of the plume associated with Hawaii (Ryan 1990). A recent model based on seismic tomography shows the plume as a vertical column, slightly elliptical in plan view, and narrowing toward the surface (Wolfe et al 1997). In this model, the diameter of the plume is ~300 km, although a seismic study by Allen et al (1999) indicates a diameter of ~200 km.

The plume, with a central temperature 150–200°C greater than that of the surrounding mantle (White et al 1995), provides most of the melt that eventually reaches the surface as magma during eruptions in the volcanic systems. On its way to the surface, the melt is modified in various ways and it stays, for longer or shorter periods (Sigmarsson 1996), as magma in deep-seated reservoirs and shallow crustal chambers. The plume is also responsible for Iceland being above sea level; its has generated a broad swell in the North Atlantic with bathymetric anomalies that extend for more than 1000 km from Iceland (Staples et al 1997).

The plume generates the Iceland crust. Interpretation of the structure and thickness of this crust is currently being revised: in the thin-crust model (Palmason 1971, Palmason & Saemundsson 1974) the crust varies in thickness from 8 to 16 km; however, recent seismological studies support a thick-crust model where the thickness is ~10–40 km (Bjarnason et al 1993a, Staples et al 1997, Darbyshire et al 1998, Menke 1999). The ~10-km-thick crust is associated with the volcanic zones of the Reykjanes Peninsula, whereas the ~40-km-thick crust occurs beneath the central highlands and the Tertiary areas.

The thick-crust model recognizes three crustal layers (Staples et al 1997): (*a*) at the surface of the neovolcanic zone is an unconsolidated layer; this corresponds to Palmason's (1971) layer 0 and is thus, on average, ~0.5 km thick; (*b*) below this is the upper crust; this corresponds to Palmason's (1971) layers 1 and 2 and is therefore, on average, ∼3–4 km thick; (*c*) deepest of all is the lower crust, which includes Palmason's (1971) layers 3 and 4. In the thick-crust model, layer 4, interpreted by Palmason (1971) as an anomalous upper mantle, is considered part of a lower crust ∼15 to 30 km thick. The crust is normally thinnest beneath the currently active Holocene volcanic systems. For example, in Northeast Iceland the crustal thickness is as great as 35 km beneath the Tertiary lava pile, but as little as 19 km beneath the Krafla Volcanic System (Staples et al 1997). The structure of the deeper part of the crust is poorly constrained. It has been suggested, however, that the Tertiary basaltic lava pile dipping toward the axis of the current rift zone can be traced as a seismic reflector down to depths of 15–18 km beneath Reydarfjordur in East Iceland (Smallwood et al 1998).

In seismotectonics and volcanotectonics, the thickness of the brittle or seismogenic crust is of great importance. Tryggvason (1998) has estimated the thickness of the brittle crust, defined as the depth above which 90% of earthquakes (of magnitude –0.5 to 4.5) occur, for Southwest and South Iceland. He finds that the brittle crust varies in thickness from 5 km beneath parts of the Reykjanes Peninsula (Figure 1) to a maximum of 12 km at the eastern end of the South Iceland Seismic Zone. At these depths the crustal temperatures, as inferred from geothermal gradients outside the rift zone, are 550–750°C. This agrees with the observation that seismogenic faulting in Iceland is rarely deeper than 12 km (Einarsson 1991, Rognvaldsson 1994).

OCEAN-RIDGE DISCONTINUITIES

The Tjornes Fracture Zone and the South Iceland Seismic Zone (Figures 1 and 2) generate the largest earthquakes in Iceland, some of which exceed magnitude 7. Their seismicity is primarily controlled by shear stresses that become concentrated because of plate-pull (the plates being pulled apart) in the adjacent segments of the volcanic zones (Figures 2, 3, and 4). Dike injections in the associated volcanic systems can either increase (unlock) or decrease (lock) the seismicity in these zones (Figure 2).

Structure of the Tjornes Fracture Zone

This 70-km-wide (N-S) and 120-km-long WNW-trending zone contains the seismic lineaments of Grimsey and Dalvik (Rognvaldsson et al 1998), located a few tens of kilometers north (Grimsey) and south (Dalvik) of, and running subparallel with, the Husavik-Flatey Fault (Figure 3). The lineaments of Grimsey (Rognvaldsson et al 1998) and Dalvik (Langbacka & Gudmundsson 1995) are sets of NNW-trending sinistral faults, whereas the Husavik-Flatey Fault is the principal structure of the Tjornes Fracture Zone, a dextral strike-slip fault, active for ∼9 Ma, with a cumulative transform-parallel displacement of ∼60 km (Saemundsson

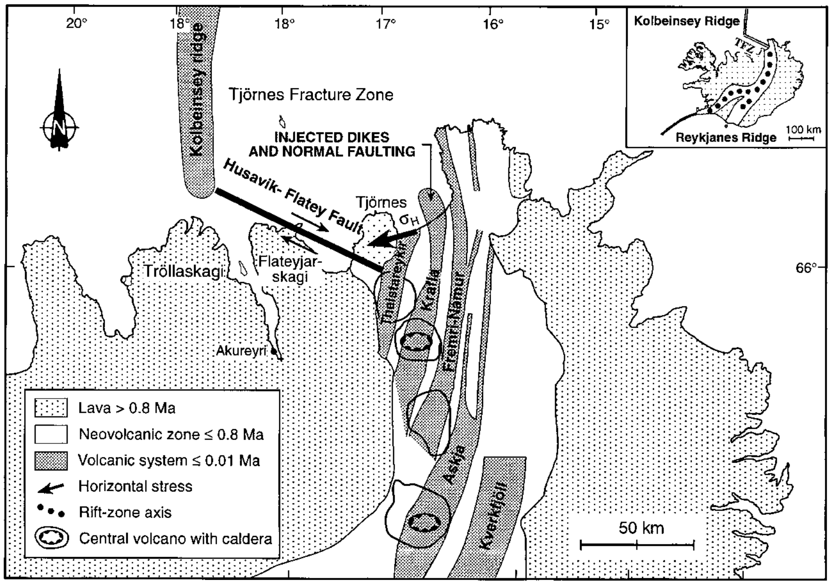

Figure 3 Seismicity in the Husavik-Flatey Fault became suppressed when the fault was essentially locked by dike injection in the northernmost part of the Krafla Volcanic System in early 1976. The dike generated high horizontal compressive stress (σ_H) in a direction opposite to the long-term dextral plate movements across the Husavik-Flatey Fault. Data on the 1976 dike location from Bjornsson et al (1977); data on the geology from Saemundsson (1974, 1978) and Johannesson & Saemundsson (1998).

1974). Offshore, it is marked by a fracture-zone (transform) valley that is 5–10 km wide and 3–4 km deep; onshore, it outcrops as a 3- to 5-km-wide zone of intense crustal deformation with numerous strike-slip and normal faults, located in a lava pile that exceeds the regional tilt by 15–35° (Fjader et al 1994, Rognvaldsson et al 1998). The Husavik-Flatey Fault contains extensive sets of mineral veins, transform-parallel basaltic dikes, and zones of completely crushed rocks, referred to as fault cores. The trend of the dikes suggests that the Husavik-Flatey Fault must occasionally be subject to transform-perpendicular tensile stress, as are many other transform faults (Gudmundsson 1995a,b). However, most normal faults and mineral veins strike perpendicular to the trend of the Husavik-Flatey Fault.

The stress field associated with the junction of the Husavik-Flatey Fault and the Kolbeinsey Ridge generates a curved fabric in the northernmost part of the Trollaskagi Peninsula (Figure 3). The fabric is composed of oblique-striking normal faults, mineral veins, and tilted lava flows (Gudmundsson 1995a). On the Tjornes peninsula (Figure 3), the Husavik-Flatey Fault runs for 25 km as a dextral strike-slip fault characterized by pull-apart structures until it joins Holocene normal faults of the Theistareykir Volcanic System (Figure 1).

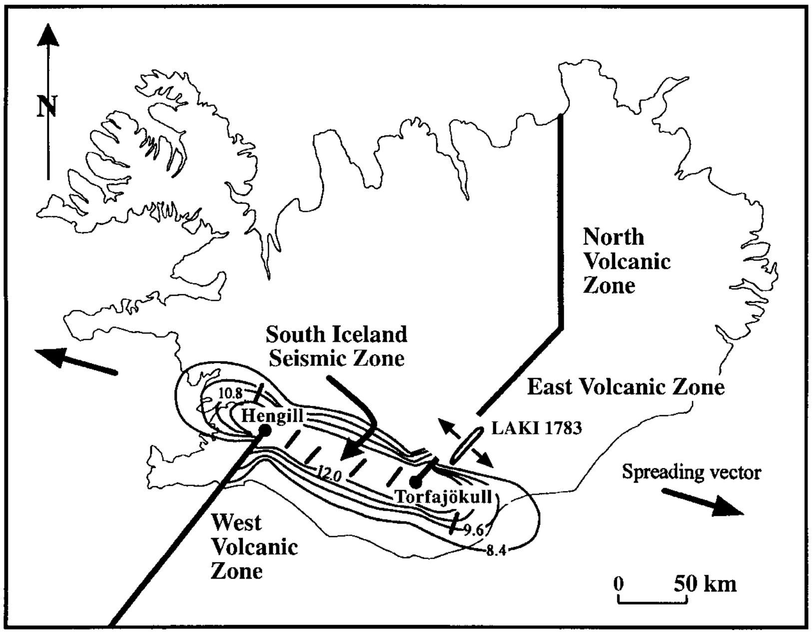

Figure 4 Plate-pull generates shear stress that favors overall sinistral movement across the South Iceland Seismic Zone (cf Figure 2). The Laki 1783 eruption generated a dike that is 27 km long (Figure 6) and a horizontal stress of 10–50 MPa subparallel with the plate movements. This eruption almost certainly triggered the earthquake sequence of 1784, which included a magnitude 7.1 earthquake, the largest historical earthquake in Iceland. The shear-stress contours, in mega-pascals, are the result of a boundary-element study using a plate pull of 6 MPa for the part of the West Volcanic Zone south of the Hengill Central Volcano and for the East Volcanic Zone north of the Torfajokull Central Volcano (modified from Gudmundsson & Homberg 1999). The ticks represent the orientation of the maximum compressive principal stress. The Laki 1783 feeder dike would have added to the shear-stress intensity.

Structure of the South Iceland Seismic Zone

During the last 800 years, destructive earthquakes have occurred at least 33 times in the South Iceland Seismic Zone (Bjornsson 1975, Rognvaldsson 1994). Earthquake sequences where the largest shocks can reach or exceed magnitude 7.1 occur in this zone at intervals of 80–100 years (Einarsson et al 1981, Stefansson et al 1993). For short periods of years and decades at a time, the seismicity is essentially confined to a zone that is ~10–20 km wide (N-S) and ~60–70 km long (E-W) (Rognvaldsson 1994, Sigmundsson et al 1995, Bergerat et al 1998) and centered at 64°N (Figures 1 and 2). It has been proposed that this zone is related to bookshelf faulting (Sigmundsson et al 1995). The zone is partly covered by Holocene lava flows in which the seismogenic faults consist of conjugate arrays of fractures trending NNE and ENE with push-ups between their nearby ends (Bjarnason et al 1993b). The zone is partly covered by Pleistocene rocks in which traces of the seismogenic faults are less easily recognized. However, recent studies show that active seismogenic faults occur tens of kilometers north of 64°N (Gudmundsson 1995b, Luxey et al 1997, Passerini et al 1997, Bergerat et al 1998, Johannesson & Saemundsson 1998, Gudmundsson & Homberg 1999). Strike-slip faults worldwide that give rise to earthquakes of magnitude 7.1 are commonly 40–80 km long (Wells & Coppersmith 1994, Ambraseys & Jackson 1998). Consequently, the NNE-trending dextral fault arrays associated with the largest earthquakes may be expected to reach N-S lengths of tens of kilometers, indicating that during major destructive periods the N-S width of the South Iceland Seismic Zone may be as much as 50–60 km (Gudmundsson 1995b, Gudmundsson & Homberg 1999).

The location of the South Iceland Seismic Zone between two overlapping volcanic zones both subject to plate-pull (Figures 1 and 2) gives rise to shear-stress concentration that largely controls its seismicity (Figure 4). The intensity and geometry of this shear-stress region depend greatly on the geometrical configuration of the adjacent volcanic zones. Because the East Volcanic Zone has been propagating to the southwest during the last 3 Ma, the geometrical configuration, as well as the geometry of the region of shear-stress concentration, has changed during this time (Figure 5). Although the present configuration is as shown in Figure 4, the stress intensity can be temporarily changed by dike injections in the nearby volcanic zones.

Locking and Unlocking of Seismic Zones by Dike Injection

It is proposed that dike injection (and normal faulting) in the volcanic systems can lock or unlock the central parts of the South Iceland Seismic Zone as well as the Husavik-Flatey Fault (Figure 2). Dike injection in the parts of the North Volcanic Zone and the East Volcanic Zone that are between the Husavik-Flatey Fault and the South Iceland Seismic Zone tends to open (unlock) these zones and trigger seismogenic faulting. For example, there are indications of a positive correlation between volcanic activity in parts of the East Volcanic Zone and seismic activity in the South Iceland Seismic Zone (Larsen et al 1998). By contrast, dike injection in the parts of these zones north of the Husavik-Flatey Fault and south of the South Iceland Seismic Zone tends to lock these faults and suppress seismogenic faulting. Similarly, dike injection in the north part of the West Volcanic Zone tends to lock the South Iceland Seismic Zone, but dike injection in its south part (including the Reykjanes Peninsula) tends to unlock the South Iceland Seismic Zone. Locking by dike injection is always temporary, however, because plate pull gradually relaxes the compressive stresses generated by the dikes.

In terms of this model, the largest historical eruption in Iceland, Laki 1783 (Figure 6), may have triggered the largest known earthquake sequence in South Iceland, that of 1784 (Figure 4). The largest shock in this earthquake sequence is

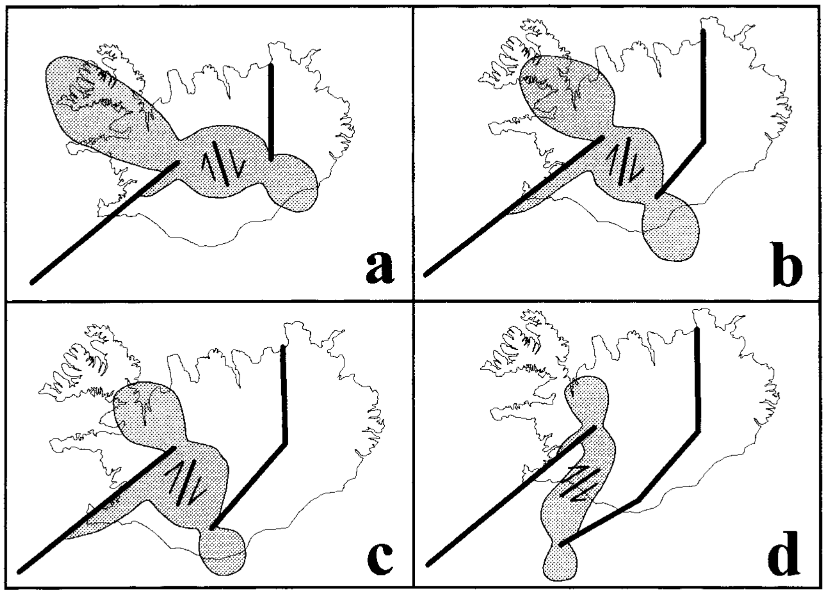

Figure 5 Evolution of the main volcanic zones during the last 3 Ma, and the changes in the regions of shear-stress concentration (shaded) and main trend of dextral faulting. Illustration (a) indicates where the tip of the East Volcanic Zone (and the associated shear-stress geometry) was located at ∼3 Ma; (b) and (c) show where the current tip is located; and (d) shows where the tip will be when it has reached approximately to the island of Surtsey in the geologically near future. Data on the evolution of the volcanic zones from Saemundsson (1979). Modified from Gudmundsson (1995b).

estimated at magnitude 7.1 (Stefansson & Halldorsson 1988). The feeder-dike of the Laki eruption is at least 27 km long (Figure 6) at the surface (Thordarson & Self 1993) and perhaps longer at depth (Gudmundsson 1990). Field observations indicate that the near-surface dike thickness is at least 6–10 m (Thordarson 1990). Such a dike, coming from a depth of 10–20 km (Sigmarsson et al 1991), can develop a static magmatic overpressure of 10–50 MPa (Gudmundsson 1990). Displacements and compressive stresses of this magnitude, generated over a period of only 8 months (the length of the Laki eruption), certainly increased the shear-stress intensity in the South Iceland Seismic Zone. This shear-stress increase may have caused small earthquakes already at the end of the eruption, in early 1784 (Thordarson 1990, 1991); it is also suggested that this stress increase triggered the main earthquake sequence in the summer of 1784.

By contrast, the Husavik-Flatey Fault has recently experienced locking by dike injection. Considerable seismicity was associated with the Husavik-Flatey Fault until early 1976, when dike injection and normal faulting of the 1975–1984 rifting episode of the Krafla Volcanic System (Bjornsson et al 1977) occurred in its

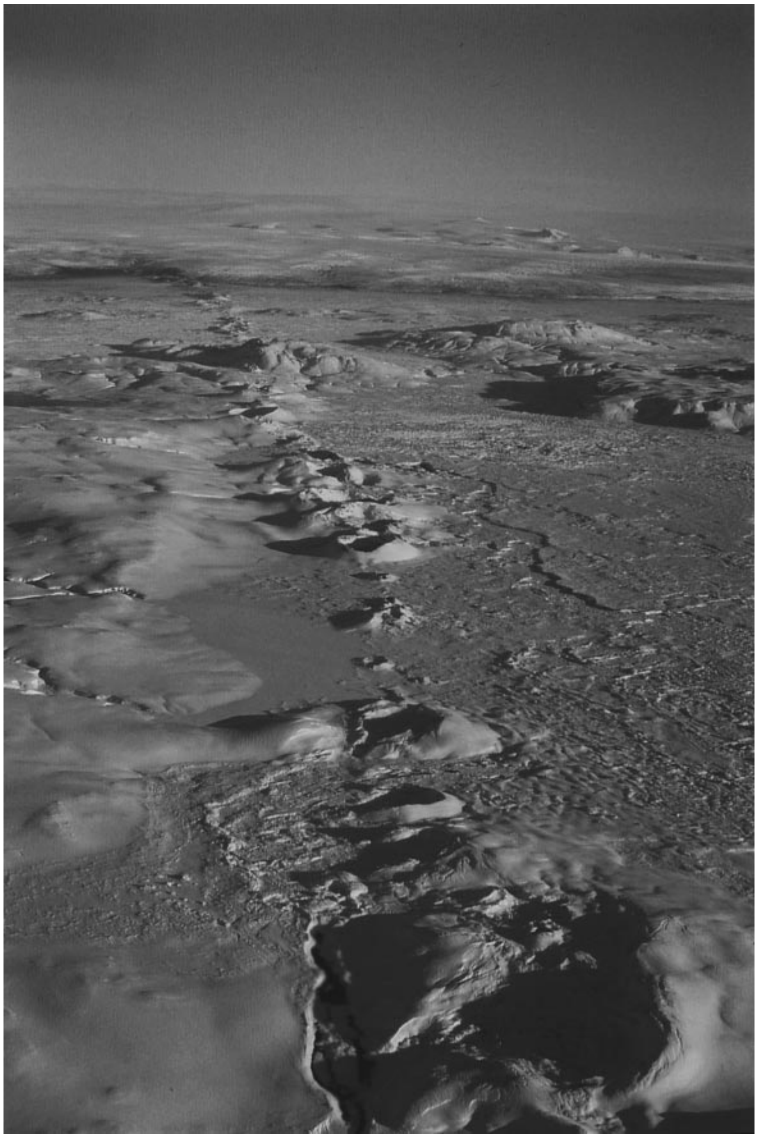

Figure 6 Aerial view (in winter) of the Laki 1783 volcanic crater row. It is 27 km long (the greater part is seen here), and belongs to the Grimsvotn Volcanic System (cf Figures 1 and 4). View NE, the snow-covered crater row extends all the way to the margin of the Vatnajokull ice sheet. This eruption produced the largest historical lava flow (cf Thordarson & Self 1993), which has a very homogeneous composition (Sigmarsson et al 1991). Its feeder dike almost certainly triggered the largest historical earthquake in Iceland (cf Figure 4).

northernmost part (Figure 3). Dike injection in this part generated horizontal compressive stresses that encouraged sinistral movement on the otherwise dextral Husavik-Flatey Fault, thereby locking the fault.

Renewed seismicity on the Husavik-Flatey Fault, following a magnitude 5.5 earthquake in February 1994 at its junction with the Kolbeinsey Ridge (Rognvaldsson et al 1998), indicates that the Husavik-Flatey Fault is currently being unlocked. Seismogenic faulting and unlocking that migrate southwest along the fault are caused by normal plate-pull movements that gradually relax the horizontal compressive stresses generated by the 1976 dike injection (Figure 3). The unlocking began in the westernmost part of the Husavik-Flatey Fault, at the greatest distance from the 1976 dike.

EVOLUTION OF VOLCANIC ZONES

The volcanic zones in Iceland are dynamic systems that evolve through time (Figure 5). The mid-ocean ridge in the Iceland region migrated over the center of the Iceland Plume (or vice versa) at ~24 Ma, during magnetic anomaly 6 (Vink 1984). One of the primary driving forces for shifting and relocation of the volcanic zones is the overall westward movement of the plate boundary relative to the center of the Iceland Plume. A general, consistent model of the evolution of the volcanic zones and ocean-ridge discontinuities in Iceland during the past 16 Ma is still lacking. Recent models on the evolution of the volcanic zones include the following.

Johannesson (1980) proposed a model for the evolution of the volcanic zones in the western part of Iceland, primarily based on geological data, and Hardarson et al (1997) provided geochemical and chronological data and a model on the relocation of the volcanic rift zones in that part. In this latter model, two relocation cycles are identified where the time between rift-zone extinctions is ~8 Ma. Oskarsson et al (1985) proposed a model, partly based on a model by Vink (1984), for the general evolution of the volcanic zones in Iceland, and Helgason (1985) proposed their frequent shifts.

The most recent changes in the plate-boundary configuration in Iceland are well established (Saemundsson 1979). For example, the southern part of the East Volcanic Zone has been propagating to the southwest during the last 3 Ma (Figure 5). It is likely that this propagation will continue until the East Volcanic Zone joins the Reykjanes Ridge (Gudmundsson 1995b). At that stage, one of the volcanic zones in South Iceland becomes extinct, the other takes over as the principal rift zone, and the South Iceland Seismic Zone becomes inactive.

The Snaefellsnes peninsula was a part of a major Tertiary volcanic zone until ~7 Ma (Johannesson 1980). The current Snaefellsnes Volcanic Zone became active at ~2 Ma. Its eastern tip is currently propagating to the southeast. This propagation, which will eventually lead to the Snaefellsnes Volcanic Zone joining the West Volcanic Zone, is expected to gradually increase the seismic activity in the region between these two volcanic zones.

STRUCTURE OF VOLCANIC SYSTEMS

Based on petrological (Jakobsson 1979a,b; Jakobsson et al 1978) and volcano-tectonic (Saemundsson 1978, 1979) data, some 30 volcanic systems have been defined within the volcanic zones (Figure 1). The exact number and boundaries of active systems (cf Figures 1 and 3) are still unclear; this is partly because the Holocene is a very short time compared with the eruption frequencies of these systems, and partly because Holocene systems can only be studied at the surface. However, analogous, extinct systems occur in Iceland's deeply eroded Tertiary and Pleistocene lava pile. Some 15 such systems have been mapped, and between 40 and 55 identified (Walker 1960, 1966; Saemundsson 1979). These extinct systems provide detailed information on the likely three-dimensional structure and evolution of the active systems, which is of great importance when trying to understand their dynamics.

Rift-zone volcanic systems, of which some 20 are related to the current rift zone, are the main focus here (Figure 7). Each rift-zone system contains numerous basalt volcanoes and commonly a central volcano, in addition to a swarm of faults and fissures, and at deeper crustal levels, dikes and inclined sheets.

Basalt Volcanoes

Basaltic lava flows, aa and pahoehoe (Self et al 1998), constitute the bulk of the products of the volcanic systems. In the Tertiary and Pleistocene systems, two main lithological types, tholeiite and olivine tholeiite, are recognized (Walker 1960), but they do not correspond to definite petrological types. The basaltic volcanoes form in either subaerial or subglacial/submarine eruptions. The most common subaerial volcanoes are crater rows (Figure 6) and shield volcanoes; less common are lava rings, tephra rings, and maars. The most common subglacial/submarine volcanoes are hyaloclastite ridges and tablemountains (the subglacial analogs to crater rows and shields, respectively); less common are hyaloclastite cones.

Most of the ~45 Holocene shield volcanoes in Iceland are composed of olivine-tholeiite or other primitive magmas and are older than 3500 years (Jakobsson et al 1978; Rossi 1996, 1997). Beginning on a fissure, a shield eruption gradually becomes concentrated on several vents, generating overlapping shields, and then on a single vent, resulting in a main shield that buries the overlapping, smaller shields. A shield consists of two morphological units: a central cone and a lava apron (Rossi 1996). The cone, which may consititute as little as 1–3% of the total volume of the shield, is commonly its main exposed part. Therefore, although the largest shields (such as Skjaldbreidur in the West Volcanic Zone)

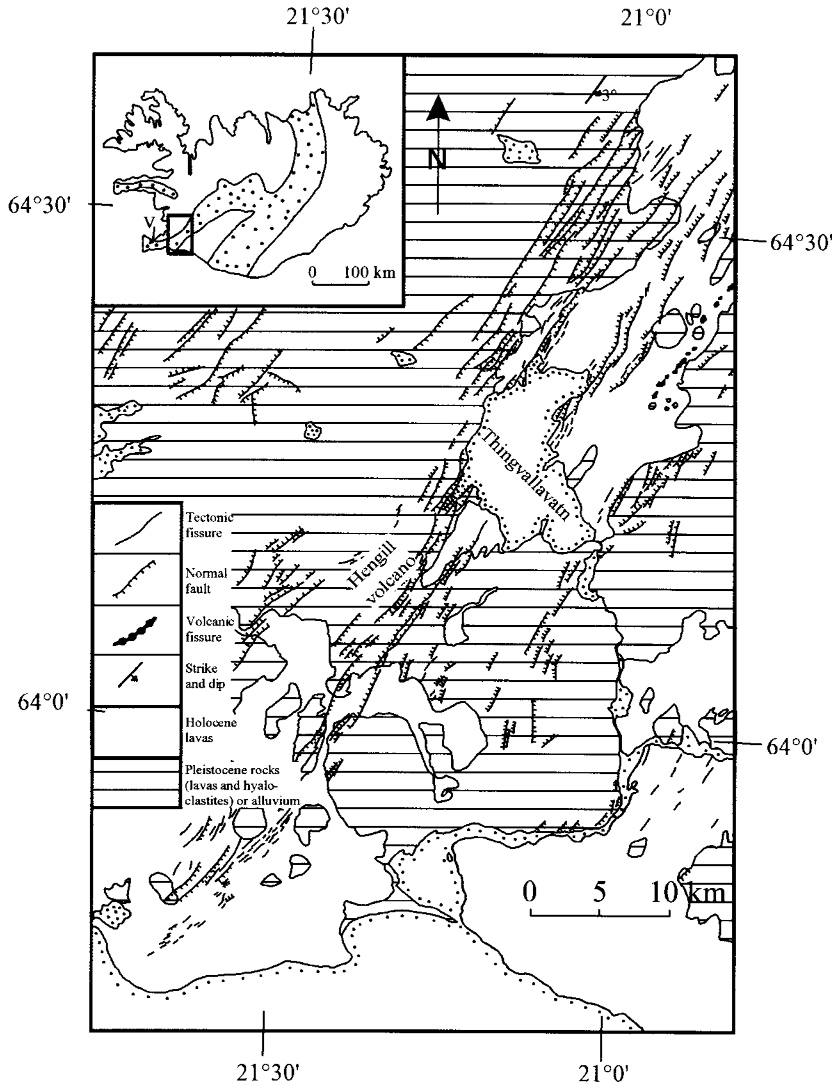

Figure 7 Surface of the Hengill Volcanic System, a rift-zone system in the West Volcanic Zone (cf Figures 1 and 9). The Hengill Central Volcano is indicated (geologic data from Saemundsson & Einarsson 1980, cf Saemundsson 1992). At the surface, the volcanic system consists of numerous tectonic fissures (mode I fractures) and normal faults, with several volcanic fissures (crater cones). The system is ~80 km long and ~10 km wide; the part located in the Holocene lava flow north of Lake Thingvallavatn is referred to as the Thingvellir Fissure Swarm, and its main western boundary fault is Almannagja (cf Figure 9).

have volumes of $\sim$17 km^3 (Saemundsson 1992), the volumes of many shields mostly reflect the cone volumes and may be underestimates. Buried shield volcanoes occur throughout the Tertiary and Pleistocene lava pile. For example, a cross-section through two Pleistocene shields is in the mountain Thyrill in the fjord of Hvalfjordur in Southwest Iceland.

The Holocene crater rows are composed of as many as 100 crater cones of scoria and spatter (Figure 6). All long rows are discontinuous. The longest include the Sveinar-Randarholar row (50–70 km long) in the North Volcanic Zone. In the East Volcanic Zone, they include the AD 871 Vatnaoldur and AD 1480 Veidivotn rows, both of which exceed 65 km in length (Larsen 1984, Larsen & Gudmundsson 1998), and the AD 1783 Laki (Figure 6) and AD 934–940 Eldgja rows with lengths of 27 km and >57 km, respectively (Thordarson & Self 1993, Thordarson et al 1999). Holocene shields occur only in the North Volcanic Zone and the West Volcanic Zone, whereas crater rows occur in all the zones but produce the largest volumes in the East Volcanic Zone. Where they occur together, crater rows are much more numerous than shields, but the total volume of lava produced by the shields exceeds that produced by the crater rows (Gudmundsson 1986).

During the last 3 Ma, Iceland has been subject to many glaciations. Subglacial eruptions partly melt the ice, forming a water-filled cavity, a lake, within which the volcanic products, mainly pillow lava and hyaloclastite, pile up. Many tablemountains and hyaloclastite ridges, earlier thought to be formed in single eruptions, are now considered to be generated in several eruptions. This interpretation is supported by the observation that the hyaloclastite ridge generated in the October 1996 Gjalp eruption in the Vatnajokull ice sheet (Figure 8) is on top of a ridge formed (partly or wholly) in a similar eruption in 1938 (MT Gudmundsson et al 1997). Also, the tablemountain Herdubreid in the North Volcanic Zone appears to have formed in several eruptions (Werner et al 1996). However, the tablemountain of the island of Surtsey in the Vestmannaeyjar Volcanic System of the East Volcanic Zone (Figure 1) was generated in a single eruption lasting from 1963–1967 (Thorarinsson 1967).

A widely accepted model on basalt volcanoes formed during subglacial eruptions is as follows. Normally, the first products are pillow lavas, formed under high hydrostatic pressure due to the water and ice in the cavity. A fissure eruption that ends at this stage produces a pillow ridge (Hoskuldsson & Sparks 1997). When water from the cavity escapes under the surrounding ice sheet (MT Gudmundsson et al 1997), the hydrostatic pressure decreases below that of the gas in the pillows, resulting in the generation of pillow breccia and pyroclastic glass, which is subsequently altered into hyaloclastite tuff. Fissure eruptions ceasing at this stage, such as the 1996 Gjalp eruption (Figure 8), produce hyaloclastite ridges. When the hyaloclastite tuff builds up above the lake level, a subaerial eruption generates a shield on top of the tuff, which results in the formation of a tablemountain.

The subglacial 1996 Gjalp eruption provides some constraints on this model (Figure 8). The eruption lasted for 13 days and produced more than 0.75 km^3 of

Figure 8 Aerial view of a part of the fissure generated in the Vatnajokull ice sheet by the 1996 Gjalp eruption. The photograph shows the fissure after the eruption had come to an end. Located in the vicinity of the Grimsvotn Central Volcano (Figure 1), this is the best-documented subglacial eruption (MT Gudmundsson et al 1997, Sigmarsson et al 1999) and has provided important data on the formation of subglacial basalt volcanoes in the volcanic systems.

volcaniclastic material composed of basaltic andesite, equivalent to $\sim$0.4 km^3 of magma. It formed a ridge 6–7 km long and 200–300 m high beneath an initial ice thickness of 500–750 m (MT Gudmundsson et al 1997). During its 12 days of subaerial eruption, the upper part of the cavity maintained its width of 200–300 m. In contrast, the subaerial tops of tablemountains are commonly several kilometers in diameter. Thus, an eruption much longer than that in Gjalp (or several short eruptions) is needed to form a tablemountain. This is not surprising because Gjalp was a short-lived fissure eruption, whereas eruptions producing tablemountains (and large shield volcanoes) would commonly last many months or years. Alternatively, some large tablemountains may be produced by several small but clustered eruptions, similar to the common clusters of Holocene shield volcanoes on the Reykjanes Peninsula (Jakobsson et al 1978, Gudmundsson 1986).

Subglacial basalt volcanoes, in particular hyaloclastite ridges, have widely different distributions in the volcanic zones. Hyaloclastite ridges are common in the central parts of the West and East Volcanic Zones, whereas nearly all the ridges in the North Volcanic Zone occur at its eastern margin. This marginal area is a continuation of the Kverkfjoll Volcanic System (Figure 3) and is referred to as Fjallgardur (Helgason 1989). Bourgeois et al (1998) suggest that subglacial hyaloclastite ridges are primarily preserved in the parts of the volcanic zones that coincide with the ice divides of the Weichselian ice cap. The ice divide in the North Volcanic Zone, as determined by Einarsson & Albertsson (1988), coincides with the axis of Fjallgardur, where the hyaloclastite ridges are preserved, whereas the other ridges in the North Volcanic Zone were eroded away by the flowing glaciers. By contrast, the ice divides in the West and East Volcanic Zones coincide roughly with their central parts, so here the hyaloclastite ridges are preserved.

Central Volcanoes

Central (composite) volcanoes, either as stratovolcanoes or collapse calderas, are the largest structures of the volcanic systems. They erupt on average once every several hundred years, produce acid and intermediate rocks in addition to basaltic rocks, and have lifetimes of 0.5–1 Ma. Over short periods, the eruption frequency of a central volcano may be much greater than this average. For example, the Grimsvotn Central Volcano (Figure 1) has erupted once every decade during the last 400 years—most recently in December 1998 (MT Gudmundsson et al 1999) while the Gjalp 1996 eruption (Figure 8) was in its vicinity (MT Gudmundsson et al 1997, Sigmarsson et al 1999).

Intermediate and, in particular, acid rocks are almost exclusively produced in central volcanoes. In the Tertiary and Pleistocene lava pile, extinct central volcanoes are easily recognized from the occurrence of large volumes of light-colored acid rocks surrounded by black basaltic lavas. The silicic magmas producing the acid rocks appear to result from processes that operate within, or in the vicinity of, the shallow crustal source chambers of the central volcanoes (Gunnarsson et al 1998).

At least five stratovolcanoes have been active during the Holocene: Snaefellsjokull, Eyjafjallajokull, Tindfjallajokull, Hekla, and Oraefajokull (Figure 1). In addition, the stratovolcano Snaefell, north of Vatnajokull, previously considered extinct or dormant, is now considered active; its last eruption was in late Weichselian or early Holocene time (Hoskuldsson & Imsland 1998). Most stratovolcanoes are roughly circular or slightly elliptic in plan view. The Hekla Volcano is unusually elongate, however, due to its erupting repeatedly on the same ENE-trending fracture. It is proposed that this fracture is largely controlled by the stress field generating the ENE-trending sinistral faults in the South Iceland Seismic Zone (Figures 1 and 4). The stratovolcanoes are commonly 5–20 km in diameter at their base and reach heights of 1–2 km and volumes of many tens of cubic kilometers.

Most of the ~20 collapse calderas identified in active central volcanoes (Saemundsson 1982) are normal-fault calderas (Gudmundsson 1998a) and slightly elliptical in shape; their average major semi-axis is 4 km, and average semi-minor axis is 3 km. The caldera associated with the Torfajokull Central Volcano in the East Volcanic Zone (Figure 1), where more acid rock has been produced than elsewhere in the volcanic zones (Gunnarsson et al 1998), has a major semi-axis of 9 km and a minor semi-axis of 6 km. This is the largest active caldera in Iceland. Vertical displacements on the caldera faults, commonly several hundred meters, reach 500–700 m in some subglacial calderas (Bjornsson 1988). Calderas associated with the extinct Tertiary and Pleistocene central volcanoes are generally of similar dimensions, displacements, and shapes to those in the active volcanoes.

All central volcanoes in Iceland that are associated with collapse calderas and high-temperature areas are presumably supplied with magma from shallow chambers. Geodetic, seismic, and gravimetric studies of active central volcanoes such as Grimsvotn (MT Gudmundsson & Milsom 1997), Krafla (Brandsdottir et al 1997, Arnadottir et al 1998), Katla (O Gudmundsson et al 1994), and Askja (Camitz et al 1995) indicate the presence of shallow magma chambers. Gabbroic bodies and other plutons in the roots of deeply eroded Tertiary and Pleistocene central volcanoes, as well as the occurrence of dense sheet swarms, lend further support to the model of a shallow chamber feeding central volcanoes (Gudmundsson 1998b). The largest plutons are $\sim$10–25 km^2 in area, with exposed volumes (of gabbro, granophyre, or both) of $\sim$10 km^3 (Gudmundsson 1995c).

Active rift-zone central volcanoes are associated with well-defined swarms of tension fractures, normal faults, and volcanic fissures (Figure 7), whereas the off-rift central volcanoes lack such swarms (Gudmundsson 1995c). In the Tertiary and Pleistocene lava pile, the deeply eroded extinct volcanoes are represented by local swarms of inclined (cone) sheets whereas the parts of the volcanic systems that are outside the central volcanoes are represented by swarms of regional dikes and normal faults. The sheet swarms are commonly circular or slightly elliptical in shape, measure several kilometers in radius, and are confined to the extinct volcano. Commonly, the sheet swarms can be traced down into gabbro bodies, exposed at 1.5–2 km depth, that form the uppermost parts of the extinct shallow magma chambers which supplied magma during the active time of the volcano (Gudmundsson 1998b).

Fissure Swarms and Fault Swarms

The Holocene swarms, which consist of many short tension (mode I) fractures and less-common but longer normal faults (Figures 7 and 9), are 5–20 km wide and 40–150 km long. The tension fractures and normal faults develop from columnar (cooling) joints in the lava pile and are vertical at the surface, indicating near-surface development under absolute tensile stresses (Gudmundsson 1992). The largest Holocene faults reach widths (openings) up to 60–70 m, throws of 30–40 m (Figure 9), and lengths of $\sim$10 km (Gudmundsson 1995c). The longest faults are thus much shorter than the volcanic systems within which they develop.

Below the Holocene surface, the fissure swarms consist of normal faults and dikes (Figure 10). Tension fractures are limited mainly to the uppermost few hundred meters of the crust, and dikes dominate in the deeply exposed (to nearly 2 km) parts of the swarms. The fault dip, averaging 75° in a lava pile dipping 2–

Figure 9 Aerial view of the Almannagja normal fault (*center*) and associated tension fractures (*left corner*) in the Thingvellir Fissure Swarm of the Hengill Volcanic System in the West Volcanic Zone (cf Figure 7). View SW, Almannagja is as wide as 60 m (*close to the waterfall*), with a maximum throw (*down to the left*) of 40 m; it forms the western boundary fault of the Thingvellir Graben (Saemundsson 1992), which is 6 to 7 km wide. All the fractures are vertical at the surface and develop from columnar joints in the 9000-year-old pahoehoe lava flow. Cars and houses (*to the left of Almannagja*) provide a scale.

8°, is either toward or away from the rift zone axis. Common throws are ~10 m, and maximum throws are ~150–200 m.

Dike Swarms and Sheet Swarms

Outside central volcanoes, the regional dikes are mostly subvertical, subparallel, thick, and basaltic (Figure 10). They occur in swarms that are analogous in dimensions to the Holocene fissure swarms. Each swarm contains hundreds of dikes at the level of exposure. The dike intensity in the swarms generally increases with depth in the lava pile (Walker 1960). The majority, and presumably most, of the dikes are nonfeeders (Gudmundsson et al 1999).

Dike thickness varies from a few centimeters to ~60 m. The thickness-size distributions for individual swarms are normally power laws, less commonly log-normal laws (Gudmundsson 1995c). The mode thickness in individual swarms is commonly 1–2 m, whereas the arithmetic average thickness is 4–6 m in the Tertiary swarms and 1–2 m in the Pleistocene swarms. Crustal dilation (extension) due to the dikes, in profiles 5–10 km long, ranges from 1–2% to 28% but is most commonly 5–6%.

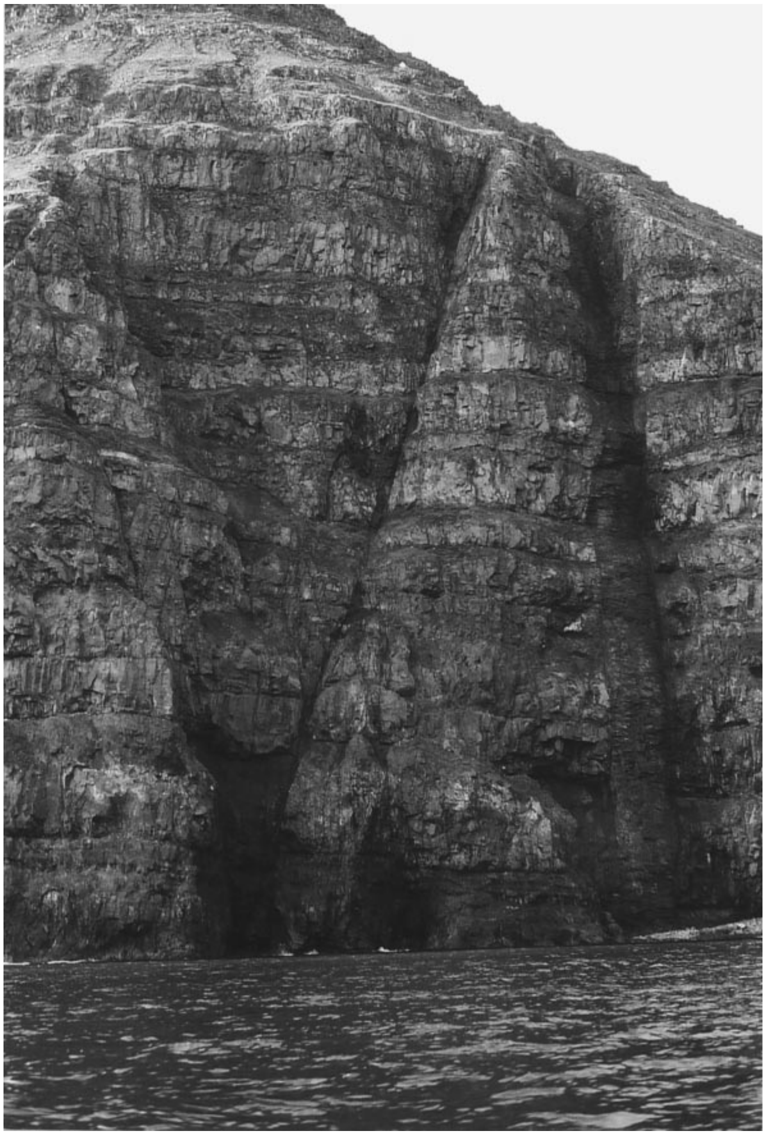

Figure 10 Volcanic systems outside central volcanoes consist primarily of dikes and normal faults (and tension fractures at shallow depths). View SW, two normal faults, forming a graben (*center*), occur to the side of a ~6-m-thick basaltic dike (*right*) in a ~12 Ma basaltic lava pile on the east coast of the Trollaskagi Peninsula in North Iceland (see Figure 3). Further geological data is provided by Langbacka & Gudmundsson (1995) and Gudmundsson (1995d).

The measured lengths of 16 Tertiary dikes in Iceland are 3–22 km; their length/thickness (aspect) ratios are 300–1500, with a typical value close to 1000. Lengths of intact volcanic fissures are equal to the surface lengths of their feeders. The longest (discontinuous) Holocene feeder dikes are 65–70 km (Larsen & Gudmundsson 1998). The length-size distributions of feeder-dikes are power laws (Gudmundsson 1995c,d). It has been suggested that many dikes propagate laterally as single, blade-like structures (Sigurdsson & Sparks 1978, Einarsson & Brandsdottir 1980, cf Rubin 1995). Alternatively, dikes may grow by repeated magma injections as sets of discontinuous structures—each of which propagates radially from its individual point of origin—that link up to form the dikes (Gudmundsson 1995d).

Dense local swarms of many thousand inclined, mostly basaltic sheets in the roots of central volcanoes are normally circular or slightly elliptical in plan view, and are commonly several kilometers in radius. The attitude of the sheets in a particular swarm varies widely; the dip distribution commonly has two peaks: one representing steeply dipping sheets, the other shallow dipping sheets. Sheets dipping 75–90° occur mainly in the central part of the swarm, whereas sheets dipping 20–50° are largely confined to its marginal parts. Most sheets dip toward the central part of the volcano to which they belong.

Sheet thickness ranges from a few centimeters to $\sim$14 m. The most common thickness, however, is $\leq$0.5 m, much less than that of the regional dikes. Adjacent to the exposed gabbro pluton (extinct chamber), the crustal dilation due to sheets is 80–100%, with the whole rock consisting of sheets (Gudmundsson 1998b). Apart from sheet attitude, this part of a sheet swarm is similar to a typical sheeted dike complex of an ophiolite. The number of inclined sheets falls off rapidly with distance from the gabbro source, however, and most of the sheets, like the regional dikes, are nonfeeders.

DYNAMICS OF VOLCANIC SYSTEMS

The Iceland Plume provides the magma that is erupted in the volcanic systems. The magma becomes divided between the systems as a result of fluid-dynamic processes that are partly controlled by the plate movements and associated stresses (Gudmundsson 1987). Magma flow into and within the reservoirs that supply magma to the volcanic systems can be modeled as fluid flow in porous media; flow of magma from the reservoirs to the shallow source chambers of the central volcanoes, and to the surface, can be modeled as fluid flow in fractured media. The shallow chambers in Iceland do not appear to have partly molten, porous roots (O Gudmundsson et al 1994, Brandsdottir et al 1997).

No quantitative, testable models exist on the fluid dynamics of magma flow in the mantle below Iceland. However, the theory of fluid flow in porous media (Bear 1972), rigorously tested on empirical results from hydrogeology (Domenico

& Schwartz 1998) and petroleum geology (Dahlberg 1995), can be used to make some general, qualitative suggestions (A Gudmundsson, unpublished data).

The potentiometric surface, here defined as the level to which magma can rise in an open conduit, may be either below or above the land surface of the volcanic zones. In the former case, no eruptions can occur; in the latter, eruptions are theoretically possible, but magma traps and stress conditions in the crust will determine whether they do occur. For magma or melt to flow, its potentiometric surface must be dipping. Inside the volcanic zones, the potentiometric surface roughly coincides with the tops of the highest mountains. For example, the maximum heights (above sea level) of late Pleistocene hyaloclastite mountains decrease by some 1500 m along the North Volcanic Zone from Kverkfjoll (Figure 1) to the coast (Walker 1965). This elevation difference may reflect a decrease in the Pleistocene ice thickness toward the coast (Walker 1965), but it is here taken as a measure of the slope of the associated potentiometric surface that drives magma along this volcanic zone, toward the Tjornes Fracture Zone. Thus, the potentiometric surface dips away from the junction between the volcanic zones, with the volcanic systems of Bardarbunga, Grimsvotn, and Kverkfjoll (Figure 1), located roughly above the central part of the Iceland Plume. Magma is driven from the plume to the coast. Most of the magma flow (in the mantle or lower crust) follows the volcanic zones, because they have much higher permeability than their surroundings.

Birth

All volcanic systems are born, reach maturity, and then die. The longevity of a system depends partly on its tectonic environment and partly on its fluid-dynamic environment. Many, perhaps most, systems develop double magma chambers—that is, shallow chambers fed by deep-seated reservoirs (Figure 11). The shallow chamber supplies magma to its central volcano, and the deep-seated reservoir supplies magma to the shallow chamber and to many basalt volcanoes outside the central volcano. In particular, shield volcanoes, which are commonly located in the marginal parts of volcanic systems and erupt large volumes of primitive lava, are nearly all supplied with magma from deep-seated reservoirs. By contrast, some crater rows are supplied with magma from shallow chambers, others from deep-seated reservoirs (Sigmarsson et al 1991).

A volcanic system is born when magma accumulates in a reservoir at the boundary between the upper mantle and the crust or in the lower crust. The magma accumulates in the regions of local minimum potential energy (Gudmundsson 1987). Under static conditions, these regions coincide with the regions of local minimum depth to the top of the partially molten upper mantle or lower crust below the surface of the volcanic zone. For reservoirs at the crust-mantle boundary, local minimum depths may be $\sim$10 km (Menke 1999) for the volcanic systems on the Reykjanes Peninsula (Figure 1), but $\sim$19 km for the Krafla Volcanic System of the North Volcanic Zone (Staples et al 1997). A static magma

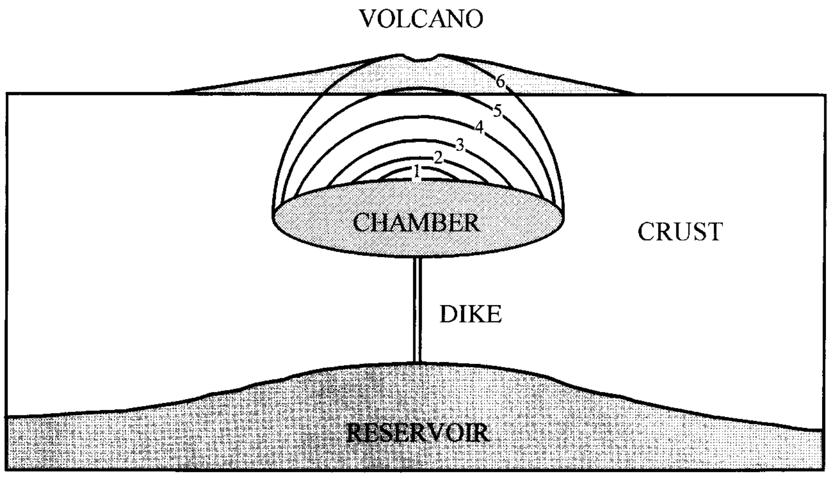

Figure 11 Most volcanic systems are supplied with magma through a double chamber: a deep-seated reservoir, located in the upper mantle or lower crust, supplies primitive magma to a shallow chamber that is the source of a central volcano (cf Figure 12). Most sheet injections during a stress cycle do not reach the surface but become arrested at various crustal depths (levels 1–5). The injection of sheets, however, gradually homogenizes the stress field to the surface (level 6), at which stage one or several eruptions may occur. Each stress cycle is approximately equal in length to the average time between eruptions in the central volcano.

reservoir becomes gradually stratified, with low-density magma and gas at the top underlain by high-density magma, and with all the fluid contacts horizontal. Partly owing to crustal bending and uplift due to magma pressure and plate pull, rift-zone reservoirs are likely to be dome-shaped, with cross-sectional shapes similar to those of their volcanic systems. By contrast, reservoirs supplying magma to off-rift areas (Figure 1) can be of any shape, although many would be roughly circular or slightly elliptic in plan view.

The static conditions described above are an ideal case that may be significantly altered by the magmadynamic conditions at the crust-mantle boundary beneath Iceland. The effects of hydrodynamic environments in trapping oil and gas and tilting their contacts with flowing water are well known in petroleum exploration (Bear 1972, Dahlberg 1995, Selley 1998). For reservoirs in Iceland, the capability of magmadynamic environments for trapping low-density magmas and gas (assumed immiscible) and tilting their contacts with high-density magmas is currently being explored (A Gudmundsson, unpublished data), but includes the following results.

General flow of high-density, primitive magma, from the center of the Iceland Plume along the crust-mantle boundary below the volcanic zones to the Kolbeinsey Ridge and the Reykjanes Ridge, may tilt the contacts between this primitive magma and the more evolved magmas in the reservoirs. The amount of tilt depends on the intensity of flow of the high-density magma, but it is in the direction of the flow. It follows that beneath the West Volcanic Zone and the East Volcanic Zone, the contacts would generally be tilted to the southwest, but in the North Volcanic Zone they would tilt to the north. The greatest magma accumulation is not necessarily in the central part of a reservoir; it can be near either of its ends, upstream or downstream. Its location depends on along-axis variations in the slope of the reservoir roof, the reservoir permeability, and the intensity of the regional magma flow beneath the volcanic zone to which the reservoir belongs. This location has implications for the subsequently formed shallow chambers and central volcanoes.

Formation of a Shallow Chamber

At its initial stage, a volcanic system has a reservoir but no shallow crustal magma chamber. All intrusions and eruptions at this stage are thus fed directly from the deep reservoir. Some volcanic systems in the rift zone, particularly those where the spreading rate is low, may never develop a shallow crustal magma chamber nor produce any acid rocks. Current examples include volcanic systems on the Reykjanes Peninsula (Figure 1).

A shallow crustal chamber is likely to be generated above great accumulations of magma in the source reservoir. The magmadynamic considerations indicate that such accumulations are not always below the center of a rift-zone volcanic system but can be near either of its ends. For example, currently the central volcanoes, and by implication the source chambers and great accumulations in the associated reservoirs, are near the ends of the volcanic systems of Fremri-Namur, Askja, and Kverkfjoll in the North Volcanic Zone, and Grimsvotn in the East Volcanic Zone (Figure 1). Changes in the intensity of the regional flow of primitive magma may shift the main magma accumulation from one locality to another along the axis of the reservoir (Dalhberg 1995). Location of the shallow chamber and associated central volcano may be expected to change accordingly, resulting in more than one central volcano in the same volcanic system. Although rare, this appears to be the case in the volcanic systems of Hofsjokull and Tungafellsjokull (Figure 1) and is also known from some Tertiary volcanic systems (Johannesson & Saemundsson 1998).

Magma accumulation in a reservoir is a necessary condition for the formation of a shallow magma chamber, but it is not a sufficient condition. For a chamber to form, there must be a trap at some crustal depth that collects magma. Discontinuity traps occur at lithological or structural boundaries, for example between layers of hyaloclastite and basaltic lava flows (Gudmundsson 1987, 1990). The contacts remain as discontinuities, rather than welding together, because of the contrast in the mechanical properties of the different rocks. When discontinuity traps stop upward propagating dikes, the dikes may change into sills or laccoliths. However, such magma bodies evolve into larger magma chambers only if they receive frequent input of magmas from accumulations in the underlying reservoir (Gudmundsson 1990). The low density of hyaloclastite and sediments also helps to trap basaltic magma. Low-angle faults are discontinuities that may trap upward moving dikes and lead to the formation of magma chambers. Such faults are rare, however, in the exposed part of the lava pile in Iceland.

Probably the most common traps for shallow magma chambers in Iceland (and elsewhere at divergent plate boundaries) are stress barriers (Figure 12), crustal layers where the horizontal compressive principal stresses are greater than the vertical stress (Gudmundsson 1990). In a rift zone, the vertical stress is normally the greatest principal stress, but horizontal compressive stresses can temporarily exceed it as a result of dike injection. Dikes from reservoirs at a depth of 10–20 km can develop static magmatic overpressures of 10–50 MPa (Gudmundsson 1990); when added to the dike-perpendicular horizontal compressive stress, initially the minimum stress, the dike-perpendicular stress easily becomes greater than the vertical stress. When the dike solidifies, this stress field is largely maintained in the rock until relaxed by plate movements. Subsequent dikes, trapped by the stress barrier to form sills, may develop a large-scale crustal magma chamber that generates its space by host-rock anatexis, stoping, and elastic-plastic expansion.

Deflection of the crust due to glacial loading can also give rise to stress barriers that trap magma. The number of active central volcanoes (and thus shallow chambers) per unit time appears to be greater in the Quaternary than in the Tertiary (Gudmundsson 1986). Downward deflection of the crust by an ice sheet forms a crustal stress barrier under the greater part of Iceland, with compressive stresses above and tensile stresses below the neutral surface (cf Gudmundsson 1986). Dike injection is therefore favored below the neutral surface, and sill injection at and above the neutral surface. Much of the magmatism during the glacial periods of

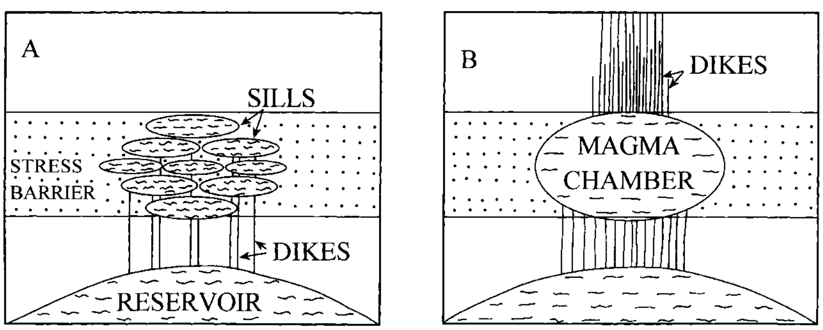

Figure 12 Many magma traps are stress barriers where the horizontal compressive principal stresses are greater than the vertical stress. These can trap upward-propagating dikes and deflect them into sills. (A) Several sills that combine into a larger one (or an initial single large sill) that receives magma frequently (through dikes) may grow into a shallow magma chamber, (B). Modified from Gudmundsson (1990).

the Pleistocene may have thus resulted in magma-chamber development rather than volcanic eruptions, whereas eruptions were more common during interglacial periods (Sejrup et al 1989, Sjoholm et al 1991, cf Lacasse et al 1998) as well as during the Tertiary.

A shallow chamber acts as a sink for magma injections from the deeper reservoir, and at the same time as a source for magma injections and eruptions in the subsequently formed central volcano (Gudmundsson 1998b). The sink behavior of the chamber is due to three related factors. First, when tensile stress concentrates around the chamber, its potential energy becomes lower and it draws in magma. Second, the stress trajectories are commonly inclined so as to capture upward propagating dikes and converge them toward the chamber. Third, while partially or completely molten, the chamber acts as a weak interface—a stopper—in the path of upward-propagating dikes. By contrast, the chamber acts as a source when it becomes ruptured by magma overpressure and injects sheets. The sheets channel magma to a limited area on the surface where a central volcano builds up.

Eruption and Intrusion Frequencies

The eruption and intrusion frequencies of a central volcano are very different from those of the other parts of its volcanic system (Gudmundsson 1998b). Eruptions and sheet intrusions occur frequently in a central volcano; commonly, the volcano erupts once every few hundred years. Most eruptions are of small volume (normally less than 0.1 km^3) and fed mainly by the thin ($<$0.5 m) inclined sheets. Outside its central volcano, a volcanic system erupts normally at intervals of several thousand years. The eruptions, however, are commonly of large volume ($>$1 km^3) and fed mainly by thick (2–5 m) regional dikes. This inverse relationship between volume and frequency of eruptions and intrusions is partly a result of the shallow chamber rupturing much more frequently (because of stress concentration) than other parts of the volcanic system, and partly a result of the chamber's small size. Thus, a single magma flow from the source reservoir may overpressure and rupture the chamber many times, resulting in many small sheet intrusions and eruptions.

Most sheets and dikes are nonfeeders (Gudmundsson et al 1999). Stress fields around magma chambers commonly favor sheet injection next to the chamber but sheet arrest at greater distances from it (Figure 11). The sheets relax the absolute and relative tensile stresses in the vicinity of the chamber and make the stress field homogeneous. When tensile stresses build up repeatedly at the chamber, for example by plate pull, the sheets homogenize the stress field in gradually larger regions around the chamber. Thus, the crustal volume around the chamber into which sheets can propagate, for a given size and excess pressure in the chamber, gradually increases with time (Figure 11). When the conditions for sheet propagation reach any part of the surface, whereby the stress field has been homogenized to the surface, one or several sheet-fed eruptions will occur. The feeders, however, bring the state of stress back to what it was at the beginning of the stress cycle, which thereby starts again. Generally, each cycle is roughly equal in length to the average time between eruptions, commonly a few hundred years. This model implies that during most volcanic unrest periods with sheet injections, there will be no volcanic eruptions.

Extinction

Volcanic systems become extinct when they no longer receive magma from a deep reservoir or a shallow chamber. Commonly, the chamber is the first to become inactive. Subsequent to its initiation, a magma chamber grows in volume. The maximum volume erupted from a chamber during a single rupture depends on its magma volume (Gudmundsson 1987). It follows that as long as a chamber's magma volume grows, there is a gradual increase in the combined maximum volume of extrusions and intrusions that the chamber can produce in single events.

When the chamber has reached a size where the influx of magma is in equilibrium with the magma that it injects or otherwise solidifies, the chamber has reached maturity. Normally, a chamber spends most of its lifetime at this stage, with eruption and intrusion frequencies and volumes essentially constant. The final stage in the lifetime of the chamber occurs when it no longer receives magma from the source reservoir at a rate sufficient to maintain equilibrium with its surroundings. A chamber depends on a reservoir as a source and becomes extinct when that source no longer exists. Nevertheless, a chamber can be active for some tens of thousands of years, depending on its size, shape, and depth (Spera 1980), after the magma supply from the reservoir is partially or completely cut off. The chamber reacts to the decreasing magma supply by changing its shape, whatever its mature shape, to that of a sphere. This is one reason so many extinct chambers are roughly spherical, whereas active ones have different shapes.

Eventually, however, tectonics processes, fluid-dynamic processes, or both bring about the extinction of the chamber. Plate movements in Iceland may transport a chamber with a lifetime of 0.5–1 Ma laterally by as much as 10–20 km. If the source reservoir remains stationary, progressively fewer dikes will reach the chamber, resulting in its gradually becoming extinct.

Many reservoirs are likely to be nonstationary, however, and may fluctuate as a consequence of the dynamical changes in the flow of melt from the mantle plume and toward the axis of the reservoirs. The accumulation of melt, necessary for maintaining the chamber as well as the volcanic system as a whole, may thus be shifted along its axis, or laterally, out of the current reservoir, which thereby becomes inactive.

FUTURE RESEARCH

The following is a list of research topics of current interest and importance for understanding the dynamics of volcanic systems in Iceland and their volcanotectonic environment. This list is not meant to be exhaustive, however.

1. As regards the structure of the Iceland Plume, it is particularly important to put better constraints on the margins of the plume. Although these are perhaps not likely to be well defined, they may significantly affect the development of the volcanic zones and systems and the two main seismic zones.
2. The crustal structure of Iceland is currently the subject of intensive research, an important part of which is mapping of the ‘‘brittle’’ or seismogenic crust. Location of magma reservoirs in the lower crust or upper mantle (so far not very successful) and the magma residence times (Sigmarsson 1996) are clearly key issues in improving the understanding of the volcanic systems.
3. The proposed general flow of melt along the volcanic zones and its effects on magma accumulation in reservoirs needs to be explored in detail. An obvious starting point is mapping of the potentiometric surface, with its volcanotectonic implications quantified and explored. A more careful definition of the boundaries of the individual volcanic systems—which is clearly needed (cf Figures 1 & 3 and Johannesson & Saemundsson 1998)—can be provided once their dynamic behavior is better understood.
4. The effects of glaciation and deglaciation on the volcanotectonics of Iceland as a whole (Jull & McKenzie 1996, Bourgeois et al 1998), its individual volcanic zones (Hardarson & Fitton 1991, Gee et al 1998), and individual volcanic systems (Gudmundsson 1986), along with their relation to mantle dynamics and viscosity (Sigmundsson 1991), are topics for potentially fruitful research projects.
5. Any general model of the dynamics of volcanic systems must relate the development of the systems to plate movements and seismicity. The close connection between volcanic activity and seismic activity, proposed in this paper and supported by data (Thordarson 1990, Larsen et al 1998), needs to be explored in more detail. Such a connection, when quantified, is likely to help us forecast earthquakes in the seismic zones and improve our understanding of the development and dynamics of the volcanic systems.

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