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Infrastructure and mechanics of volcanic systems in Iceland

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Abstract

The divergent plate boundary in Iceland consists of more than two dozen systems where most of the volcano-tectonic activity takes place. At the surface the volcanic systems are characterised by 5–20-km-wide and 40–100 km-long swarms of tension fractures (~ 10^2 m long), normal faults (~ 10^3 m long) and volcanic fissures. The Holocene fissure swarms are confined to less than 10,000-year-old basaltic lava flows, mostly pahoehoe, occurring near the centre of the active rift zone. In any particular swarm, the number of tension fractures exceeds that of normal faults. All tension fractures and normal faults are vertical at the surface, indicating that the surface parts were generated by an absolute tension. In addition to a fissure swarm, many volcanic systems have a central volcano some of which have developed collapse calderas. In the late Tertiary and Pleistocene lava pile of Iceland, extinct volcanic systems are represented by local sheet swarms and regional dyke swarms. The sheet swarms are normally circular or slightly elliptical, several kilometres in radius, and are confined to the extinct central volcanoes. Many swarms are associated with large plutons, exposed at 1–2 km depth beneath the initial top of the rift zone and presumably the uppermost parts of extinct crustal magma chambers, and in short traverses up to 90% of the rock may consist of sheets. The sheets have a very variable strike, dip on average 45–65°, mostly towards the centre of the associated volcano, and have an average thickness of about 1 m. The regional dykes occur outside central volcanoes in swarms that are commonly 50 km long and 5–10 km wide. In several-kilometre-long traverses, commonly 1–5% of the rock consist of dykes but occasionally as much as 15–28%. Most regional dykes are subparallel and subvertical. The average dyke thickness in the Pleistocene swarms is less than 2 m but 4–6 m in the Tertiary swarms. While active, the volcanic systems in the rift zone are supplied with magma from reservoirs located at the depth of 8–12 km at the boundary between the crust and upper mantle. The reservoirs are partially molten, with totally molten top regions, and of cross-sectional areas similar to those of the volcanic systems that they feed. Some active volcanic systems, especially those that develop central volcanoes, high-temperature areas and calderas, have, in addition to the deep-seated magma reservoirs, shallow crustal magma chambers, located at 1–3 km depth, which, in turn, are fed by the deeper magma reservoirs. It is proposed that the regional dyke swarms are supplied with magma largely from the deep-seated reservoirs, whereas the local sheet swarms are mainly fed from the associated crustal magma chambers. Because the volume of a crustal chamber is less than that of its deeper source reservoir, a single magma flow (dyke intrusion) from the reservoir may trigger tens of magma flows (sheet intrusions) from the chamber, which is one explanation for the enormous number of sheets associated with many central volcanoes. The sheets follow the stress trajectories of the local stress fields around the source chambers, whereas the regional stress field associated with the divergent plate movements controls the emplacement of the regional dykes. It is suggested that many dykes develop as self-affine (some as self-similar) structures. When applied to the Krafla volcanic system in northern Iceland, model calculations suggest that a magma flow from the Krafla reservoir, with regional dyke formation, should occur, on average, once every several hundred years, and

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that tens of sheets might be injected from the shallow Krafla chamber and triggered by a single magma flow from the Krafla reservoir. These results are in broad agreement with the available data.

1. Introduction

Iceland is the only large subaerial part of the mid-ocean ridges and is a place where the processes at divergent plate boundaries can be studied in great detail. The unique feature of Iceland is that one can compare current rifting episodes, with fracture formation, dyke emplacement and volcanic eruptions, with the infrastructure of the nearby extinct Tertiary and Pleistocene volcanoes. These extinct volcanoes are commonly exposed at the depths of 1–2 km beneath the initial top of the rift zone and located within several tens of kilometres from the currently active rift zone. Iceland is thus an excellent natural laboratory for formulating and testing models on the infrastructure and mechanics of volcanoes at divergent plate boundaries in general.

The divergent plate boundary in Iceland consists of more than two dozen systems where most of the volcano-tectonic activity takes place (Fig. 1). At the surface these volcanic systems are characterised by 5–20-km-wide and 40–100-km-long swarms of tension fractures, normal faults and volcanic fissures. In addition to the fissure swarms, many volcanic systems have a central volcano, some of which have developed collapse calderas. In the late Tertiary and Pleistocene lava pile, extinct volcanic systems are represented by local sheet swarms and regional dyke swarms. The sheet swarms are confined to central volcanoes and commonly associated with large plutons, presumably the uppermost parts of extinct crustal magma chambers. The regional dykes occur in elongate swarms outside the central volcanoes.

During the past decade the author and his co-workers have studied more than four thousand dykes and inclined sheets in many extinct and exhumed Pleistocene and Tertiary volcanic systems in Iceland (Fig. 1). Detailed tectonic studies have also been made of several active volcanic systems (Fig. 1). The principal aim of these studies was to obtain a three-dimensional view of the infrastructure of the volcanic systems as well as to provide data for formulating and testing models on the mechanics of volcanoes at divergent plate boundaries.

The first objective of this paper is to provide extensive data on the structures of the active and extinct volcanic systems in Iceland. These include structural data on several thousand tension fractures, normal faults, volcanic fissures, dykes and inclined sheets, in addition to brief descriptions of central volcanoes (and associated calderas) as well as several large plutons (the tops of extinct crustal magma chambers). A second objective is to use these data as a basis for a simple, quantitative model of the mechanics of

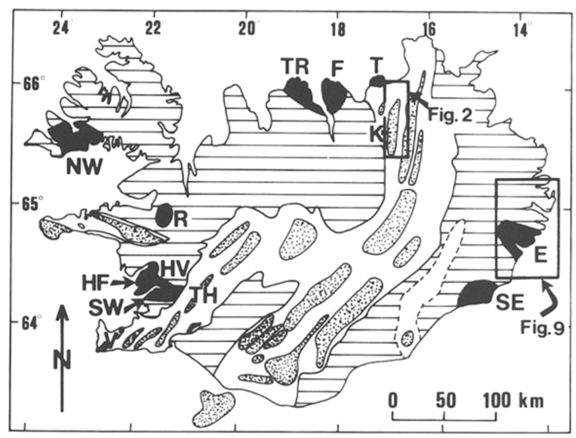

Fig. 1. The main volcanic systems (dotted) in the volcanic zones of Iceland (from Jakobsson, 1979). The main fissure swarms studied by the author are identified. The Krafla fissure swarm (*K*) is associated with the Krafla system, but the Thingvellir fissure swarm (*TH*) is associated with the Hengill volcanic system, and the Vogar fissure swarm (*V*) is associated with the Reykjanes volcanic system. Also located are the main areas (black) where the author and his co-workers have studied dykes, inclined sheets and faults. These are northwestern Iceland (*NW*), Reykjadalur volcano (*R*), Hvalfjördur volcano (*HV*), Hafnarfjall volcano (*HF*), southwestern Iceland (*SW*), southeastern Iceland (*SE*), eastern Iceland (*E*), Tjörnes peninsula (*T*), Flateyjarskagi peninsula (*F*), where the Eyjafjördur profile is located, and Tröllaskagi peninsula (TR). The areas covered by Figs. 2 and 9 are also indicated.

volcanic systems at the divergent plate boundary in Iceland.

2. Structure of active volcanic systems

2.1. Fissure swarms

The Holocene fissure swarms are commonly 5–10 km wide and 40–80 km long, a few ones are as much as 20 km wide and 100 km long, and contain hundreds of tectonic fractures. These swarms are confined to less than 10,000-year-old basaltic lava flows (mostly thick pahoehoe flows) that occur near the axis of the active rift zone. Each swarm consists of tension fractures ($\sim 10^2$ m long), normal faults ($\sim 10^3$ m long) and volcanic fissures ($\sim 10^3$ m long). In any particular swarm the number of tension fractures exceeds that of either normal faults or volcanic fissures (Gudmundsson, 1987a,b; Opheim and Gudmundsson, 1989). The frequency (density) distribution of the trends of the fractures in individual fissure swarms can be approximated by a normal curve.

The most recent rifting episode in Iceland occurred in the Krafla fissure swarm (Fig. 2) from 1975 to 1984. The swarm is 80 km long and 4–10 km wide, with hundreds of tectonic fractures (Opheim and Gudmundsson, 1989). All the fractures are vertical at the surface, indicating that the surface parts were generated by an absolute tensile stress. ring the rifting episode, 9 eruptions occurred and the new lavas partly covered the fissure swarm (Fig. 3). In several cases lava was seen to flow down into the gaping fissures, presumably forming pseudodykes (false dykes) (Fig. 4). During such episodes, many fractures grow, as happened in the Krafla swarm during the 1975–1984 episode. Then the total dilation across the swarm reached a maximum of 9 m and the vertical displacement was of the order of 1–3 m (Tryggvason, 1984; Bjornsson, 1985; Wendt et al., 1985). The maximum dilation is, however, largely attributable to the feeder-dyke(s) emplaced near the caldera rim. The maximum measured throw (accumulated vertical displacement) on a normal fault in the Holocene fissure swarms in Iceland is 32 m and occurs on a normal fault at the western margin of the Krafla fissure swarm (Fig. 5). Throws of 20–30 m are limited to the largest faults in the Holocene swarms, and the mean throw is only a few metres. For example, the mean throw on 35 normal faults of the Vogar fissure swarm in southwestern Iceland (Fig. 1) is only 2.3 m (Gudmundsson, 1987a).

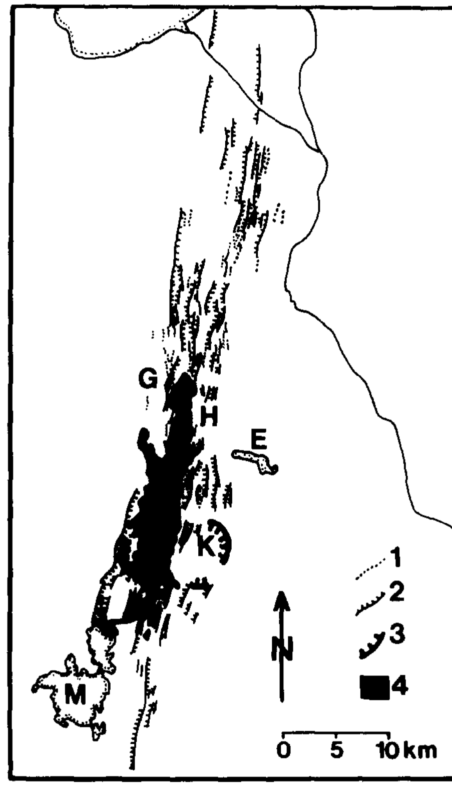

Fig. 2. Main structural elements of the Krafla fissure swarm. These include tension fractures (*1*), normal faults (*2*), the caldera fault (*3*), and the lava flows from the Myvatn fires 1724–1729 and the Krafla fires 1975–1984 shown in black (*4*). Also located are the Gjastykki area (*G*), Hrutafjöll (*H*), and the lakes Eilifsvatn (*E*) and Myvatn (*M*). Based on maps by Opheim and Gudmundsson (1989) and Saemundsson (1991).

In each fissure swarm, most short fractures are tension fractures whereas the long ones are normal faults. The shortest fractures detected on aerial photographs at the scale of 1:34,000 are several tens of metres long and are invariably tension fractures. The lengths of the longest frac-

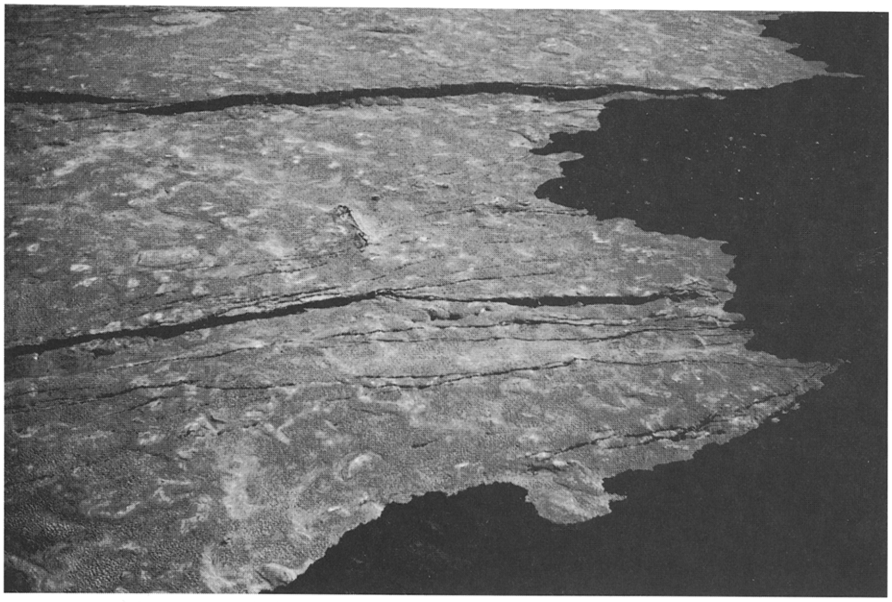

Fig. 3. Aerial view east of the black lava front from the 1984 eruption of Krafla burying tension fractures and normal faults in the Gjastykki area (located in Fig. 2).

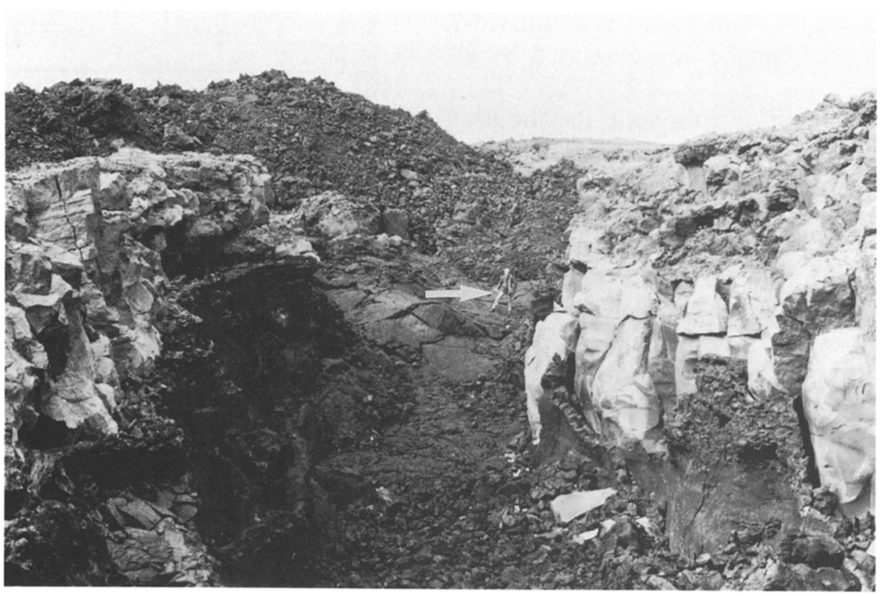

Fig. 4. Lava flows from the Krafla fires fill a large tension fracture near Hrutafjöll (located in Fig. 2) presumably forming a pseudodyke (false dyke) at depth. The person indicated by an arrow provides a scale.

tures are 4–10 km and they are normal faults. The long fractures are normally composed of shorter segments (Fig. 6), some of which are composed of still smaller segments down to the scale of columnar joints. The lower limit of fracture length, as given above, is thus determined by the scale of the aerial photographs used, whereas in the field the lower limit is determined by the lengths

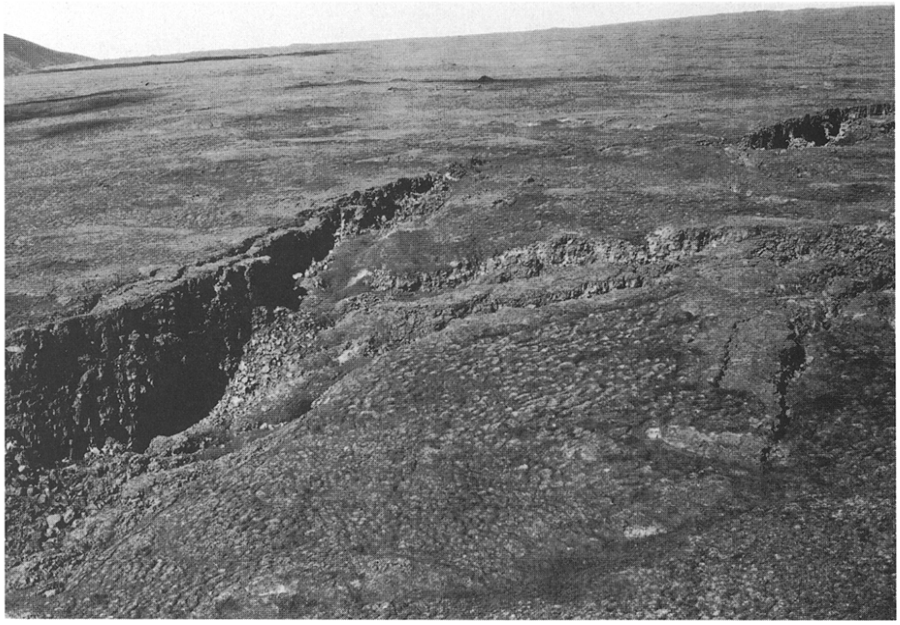

Fig. 5. Aerial view, looking northwest, of the western boundary fault of the main graben in the Gjastykki area of the Krafla fissure swarm (located in Fig. 2). This normal fault dissects a Holocene pahoehoe lava flow, is vertical at the surface and reaches a maximum throw of 32 m (at the left margin of the photograph). Notice how rapidly the throw on this segment decreases where the fault becomes offset.

of the columnar joints (commonly 0.2–0.5 m). The fracture length has a power-law size distribution (Gudmundsson, 1987a,b). The surface widths of these fractures are as much as several tens of metres.

2.2. Central volcanoes

In addition to the fissure swarms, many volcanic systems develop central volcanoes (composite volcanoes). Many, perhaps most, central volcanoes in Iceland are fed by shallow crustal magma chambers. Each chamber acts as a trap for upward propagating dykes (from a deeper reservoir) and channels the magma towards the surface within a limited area that eventually becomes the central volcano. This focusing of the magma towards an area of a limited size, as well as the extra magma added to the chambers and reservoirs through melting (anatexis) of the host rock, is here considered to be the main reason for the formation of specific central volcanoes within the volcanic systems. Melting of the host rock is one major process by which crustal chambers make space for themselves and has therefore an important role in their evolution and growth, but melting occurs also in the roofs of the reservoirs.

The main characteristics of Icelandic central volcanoes can be summarised as follows.

(1) They erupt frequently and have lifetimes of the order of 10^5 years. A typical eruption frequency is one eruption every several hundred years. Over short periods of time, however, the eruption frequency of a particular central volcano may exceed this figure by an order of a magnitude. For example, during the past 1100 years (historical time in Iceland) the Hekla volcano has erupted, on average, once every 55 years (Gudmundsson et al., 1992a). Also, the Grimsvötn volcano, located beneath the Vatnajökull glacier, has erupted once every 11 years during the past 400 years. The average lifetime of a central volcano is around 700 thousand years; some are active for only 300 thousand years whereas others may be (intermittently) active for over 2 million years.

(2) They produce basaltic, intermediate and

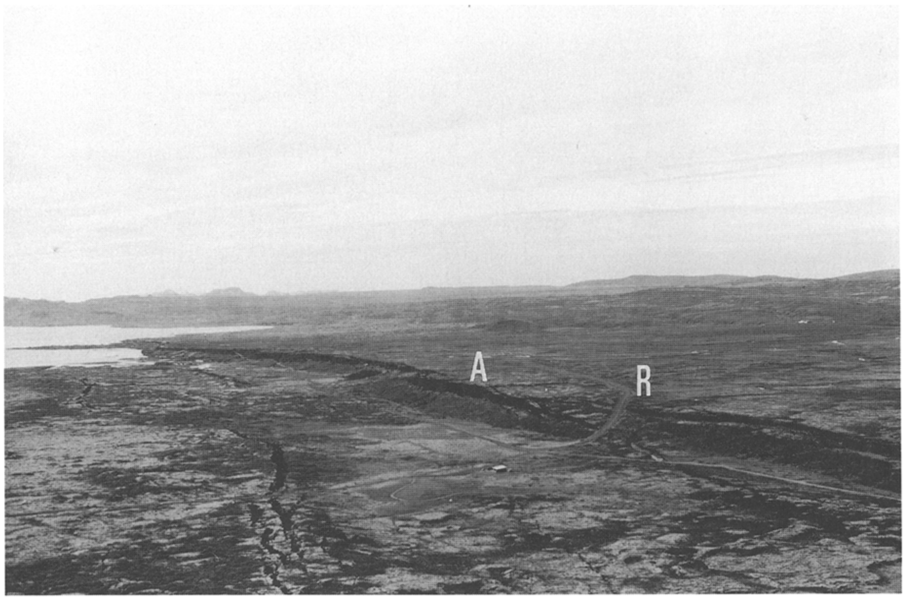

Fig. 6. Aerial view, looking southwest, of offset parts of Almannagja (*A*), the western boundary fault of the main graben of the Thingvellir fissure swarm (located in Fig. 1). The main offset is where the road (*R*) crosses the fault. The vertical displacement is commonly 20–30 m. The fault dissects a 9000-year-old pahoehoe lava flow. Notice the tilting (up to 11°) of the eastern fault wall.

acid rocks. All volcanoes and volcanic systems in Iceland produce basaltic rocks, but the intermediate and, in particular, the acid rocks are mainly produced in the central volcanoes. This restriction of acid extrusives to central volcanoes applies not only to the presently active volcanic systems but also to the extinct Tertiary and Pleistocene systems. The formation of silicic magmas thus appears to be the result of processes that are essentially confined to central volcanoes and are probably often occurring within, or in the vicinity of, crustal magma chambers.

(3) Many central volcanoes are associated with, and fed by, shallow crustal magma chambers. Seismic studies and geodetic measurements indicate that many active central volcanoes, such as Krafla and Askja (Fig. 7), have shallow crustal magma chambers. Gudmundsson (1988a) suggested that all central volcanoes in Iceland that are associated with collapse calderas and high-temperature areas are fed by crustal magma chambers. Evidence that many central volcanoes are supplied with magma from shallow chambers comes also from the study of deeply eroded Tertiary and Pleistocene central volcanoes. At depths of 1–2 km below the original surface of the lava pile, there are gabbro and granophyre plutons, commonly associated with dense swarms of inclined (cone) sheets, which form the uppermost parts of the extinct crustal magma chambers.

(4) They are the largest volcanic structures in Iceland. The main topographic types are volcanic cones and collapse calderas. The former are mainly found outside the rift zone, in the off-rift flank zones (Fig. 7). Examples of active volcanic cones include Snaefellsjökull, Eyjafjallajökull and Öraefajökull (Fig. 7). By contrast, many collapse calderas occur in the rift zone, but they are also found in the off-rift flank zones. Examples of calderas in the rift zone include Krafla, Grimsvötn and Askja (Fig. 7). The most recent caldera collapse occurred in the Askja volcano in 1875, during an associated eruption, and generated the Lake Öskjuvatn caldera.

In addition to the central volcanoes, the Holocene volcanic systems contain basalt volcanoes. As the name implies, these produce only basaltic

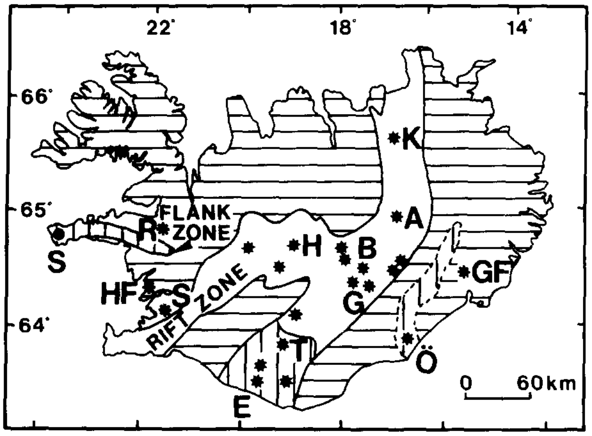

Fig. 7. Location of calderas in Iceland. Within the volcanic zones, the calderas referred to in the text are Krafla (*K*), Askja (including Lake Öskjuvatn) (*A*), Bardarbunga (*B*), Hofsjökull (*H*), Grimsvötn (*G*), Öraefajökull (*Ö*), Torfajökull (*T*), and Eyjafjallajökull (*E*). The Tertiary and Pleistocene calderas indicated are Geitafell (*GF*), Stardalur (*S*), Hafnarfjall (*HF*) and Reykjadalur (*R*). Also shown is the volcano Snaefellsjökull (*S*). Data from Saemundsson (1982) and many other sources.

rocks, normally erupt only once and, with the exception of large shields, do not form mountains. The most common basalt volcanoes in Iceland are crater rows and shield volcanoes (lava shields). There are hundreds of Holocene crater rows in Iceland, some containing over hundred individual craters and reaching the length of tens of kilometres, the best known being Lakagigar (Thordarson and Self, 1993). There are about 40 Holocene shield volcanoes in Iceland, the best known being Skjaldbreidur with a volume of 15 km^3. Other basalt volcanoes, with examples in parenthesis, include lava rings (Eldborg, western Iceland), tephra rings (Hverfjall, northern Iceland), and maars (Graenavatn, southwestern Iceland). In addition to these subaerially formed Holocene basalt volcanoes, many volcanic systems contain tablemountains (Herdubreid, northern Iceland), hyaloclastite ridges (Sveifluhals, southwestern Iceland), and hyaloclastite cones (Keilir, southwestern Iceland) formed in subaquatic (or subglacial) eruptions during the Pleistocene.

2.3. Calderas

About twenty calderas have been identified in the central volcanoes of the neovolcanic zones of Iceland (Fig. 7). These calderas are associated with central volcanoes that are still active or became extinct very recently. Most of the calderas are somewhat elliptic in shape. The lengths of the major semi-axes range from 1 to 9 km, with an arithmetic average of 4 km, whereas the lengths of the minor semi-axes range from 1 to 7 km, with an average of 3 km. The average ratio between the axes is 1.4. The largest caldera has a minor semi-axis of about 6 km and a major semi-axis of about 9 km and is associated with the Torfajökull central volcano (Fig. 7).

The vertical displacement (subsidence) on the caldera faults commonly reaches several hundred metres. During the formation of the Lake Öskjuvatn caldera in 1875, the displacement exceeded 250 m. The Askja caldera itself was formed in early postglacial time (Sigvaldason et al., 1992). The minimum vertical displacement on the caldera fault, as measured from the highest ridges of the flanks of the western part of the caldera down to the caldera floor, is normally 160–180 m. Borhole data from the Krafla caldera indicate a vertical displacement of as much as 370 m (Gudmundsson et al., 1983).

Many active calderas in Iceland are covered by glaciers in which case the only available data on the vertical displacements are from maps of the bedrock topography as revealed by the radio echo-sounding technique (Björnsson, 1988). These maps suggest that the Hofsjökull caldera may have a vertical displacement of 500–600 m and the caldera in Bardarbunga a displacement of 500–700 m. Topographic results, together with seismic studies, indicate that the vertical displacement on the Grimsvötn caldera fault exceeds 300 m and may be as much as 500 m (M.T. Gudmundsson, 1989).

3. Structure of extinct volcanic systems

3.1. Fault swarms

Below the surface of the fissure swarms of the rift zone, the volcanic systems are composed of swarms of faults and dykes. Most surface tension fractures are limited to the uppermost few hundred metres of the crust; if they propagate to greater depths they change into normal faults. Theoretical considerations suggest that the tension fractures commonly become normal faults on reaching crustal depths of about 500 m and, in general, that normal faults initiate on tension fractures and/or sets of en echelon columnar joints at depths of 500–1500 m (Gudmundsson, 1992).

In the late Tertiary and Pleistocene areas studied by the author, the level of erosion ranges from about 500 m to about 1800 m. At these crustal depths faults are much more scarce than dykes. In the numerous profiles studied, nearly 4300 dykes and sheets, but only 500 (mostly normal) faults, have been observed. Most of the faults are restricted to the uppermost one kilometre of the crust; at greater depths the dykes are the main structural elements contributing to the crustal dilation. For example, in the most deeply eroded dyke swarms, such as in eastern Iceland, faults are very rare (Walker, 1959; Gudmundsson, 1983). Also, in the Pleistocene and Tertiary areas in southwestern Iceland, Forslund and Gudmundsson (1991, 1992) observed that where the crustal dilation due to dykes is greatest the dilation due to normal faults is least, and vice versa. Dykes are rarely seen to occupy fault planes, and where this occurs the dyke normally leaves the fault plane after following it upwards for, at most, a few tens of metres. There is hardly any evidence that the internal magmatic overpressure of the dykes is responsible for the fault formation.

The Pleistocene normal faults in southwestern Iceland (Forslund and Gudmundsson, 1991, 1992) are representative of the rift-zone generated fault swarms (Fig. 8). Fault strike follows approximately a normal curve, and the mean strike is similar to that of the nearest segment of the axial rift zone. The faults dip between 53° and 89°, the average dip being 75°, whereas the lavas within which the faults are located dip mostly between 2° and 8°. Most faults dip either towards, or away from, the axis of the rift zone, and the number of faults dipping in either direction is about the same.

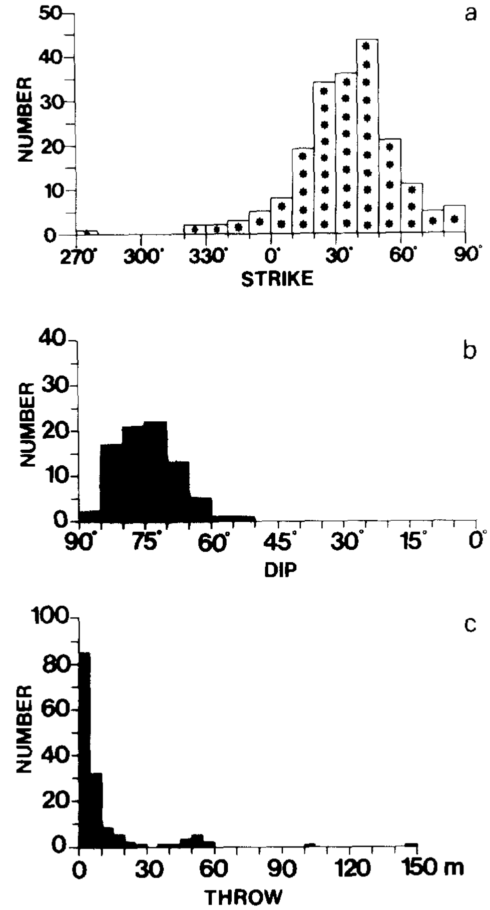

Fig. 8. Normal faults from the Pleistocene area in southwestern Iceland (located in Fig. 1). (a) Strike of 197 faults; (b) dip of 82 faults; (c) throw of 147 faults. Data from Forslund and Gudmundsson (1991, 1992).

The maximum measured fault throw is 150 m, but one Pleistocene normal fault has an estimated throw of 200 m. Surface irregularities of the lava flows are such that throws less than 0.5 m are commonly not significant, so that the minimum measured throw is 0.5 m (fractures with smaller throws are classified as joints or tension fractures). The mean throw of 147 faults is 10.2 m; 90% of the faults have throws that do not exceed 20 m. In the Tertiary areas, which are generally more deeply eroded than the Pleistocene areas, the mean fault dip and throw is less than that in the Pleistocene fault swarms (Forslund and Gudmundsson, 1992). This difference may be related to the general development of normal faults in Iceland (Gudmundsson, 1992).

3.2. Dyke swarms

Outside the central volcanoes, the volcanic systems consist primarily of subvertical and subparallel dykes, here referred to as regional dykes. These are easily distinguishable from the thin, shallowly dipping and variably striking sheets that are associated with the central volcanoes. The dykes occur in well-defined swarms that are 5–10 km wide and may exceed 50 km in length (Fig. 9). At the level of exposure, each swarm contains hundreds of dykes, and in some swarms the number of dykes may exceed one thousand (Gudmundsson, 1990a). The number of dykes, however, increases with depth in the lava pile (Walker, 1960), so that the intensity of a dyke swarm is a function of the depth of exposure below the initial top of the lava pile.

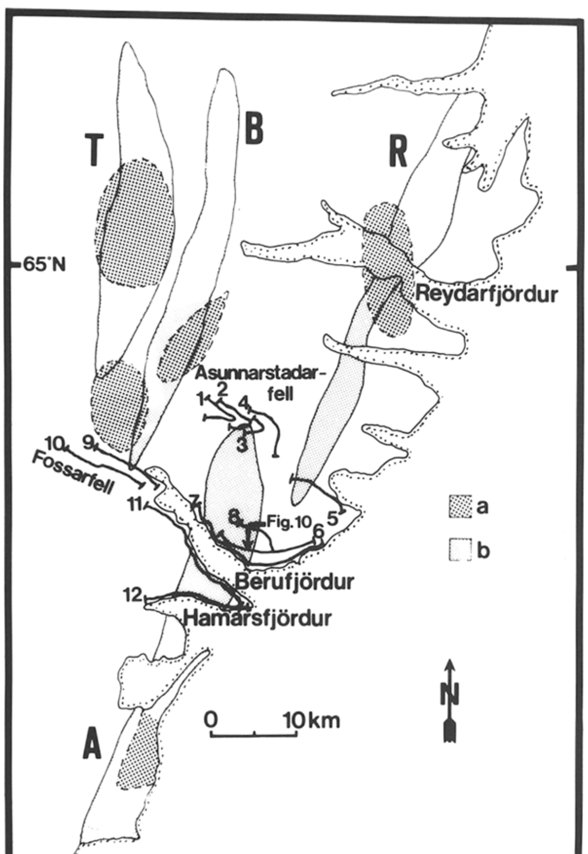

Fig. 9. Regional dyke swarms of eastern Iceland (located in Fig. 1). The author's main profiles (*1–12*) are indicated. The main dyke swarms are the Thingmuli swarm (*T*), the Breiddalur swarm (*B*), the Reydarfjördur swarm (*R*) and the Alftafjördur swarm (*A*). The shadings refer to the associated central volcanoes (*a*) and the high-intensity parts of the dyke swarms (*b*). Data from Walker (1959, 1960) and the author.

As an example of a typical dyke swarm, one may consider the Alftafjordur swarm in eastern Iceland (Fig. 9). A 9-km-long profile across this swarm is dissected by about 90 dykes, resulting in an average crustal dilation of 5.5% at the depth of 1.3 km below the initial top of the lava pile. A part of this swarm is seen in Fig. 10.

The dykes from a 17-km-long profile on the eastern coast of Eyjafjördur in northern Iceland (Fig. 11) may be considered representative of the general strike, dip and thickness distributions in the regional swarms. The dykes are generally subparallel and the frequency distribution of strike follows a normal curve. Field evidence shows that most dykes in Iceland are pure extension fractures so that they should propagate in a direction that is perpendicular to the least principal compressive stress. The spread in dyke strike, as indicated in Fig. 11a and obtained for other dyke populations as well (Gudmundsson, 1990a), may reflect fluctuations in the direction of this stress about a time-averaged mean direction.

Most regional dykes in Iceland are subvertical. About 68% of the dykes in the Eyjafjördur profile (Fig. 11b) dip within 10° of the vertical. This is similar to the results obtained in eastern Iceland (Gudmundsson, 1983) but lower percentage than for the dykes in northwestern Iceland (Gudmundsson, 1984a). In the Eyjafjördur profile the lavas dip 9–12°, which is several degrees steeper than the average lava dip in the profiles in northwestern and eastern Iceland. Many dykes cut the lavas at nearly right angles, so that the lava dip may affect the dip distribution of the dykes. Nevertheless, the subverticality of the dykes indicates that the direction of the least compressive stress was subhorizontal at this crustal level.

The thickness (width) of the regional basaltic dykes is from 0.02 m (2 cm) to 28.5 m. The Eyjafjördur profile (Fig. 11c) includes the thickest single basaltic dyke (28.5 m) observed so far

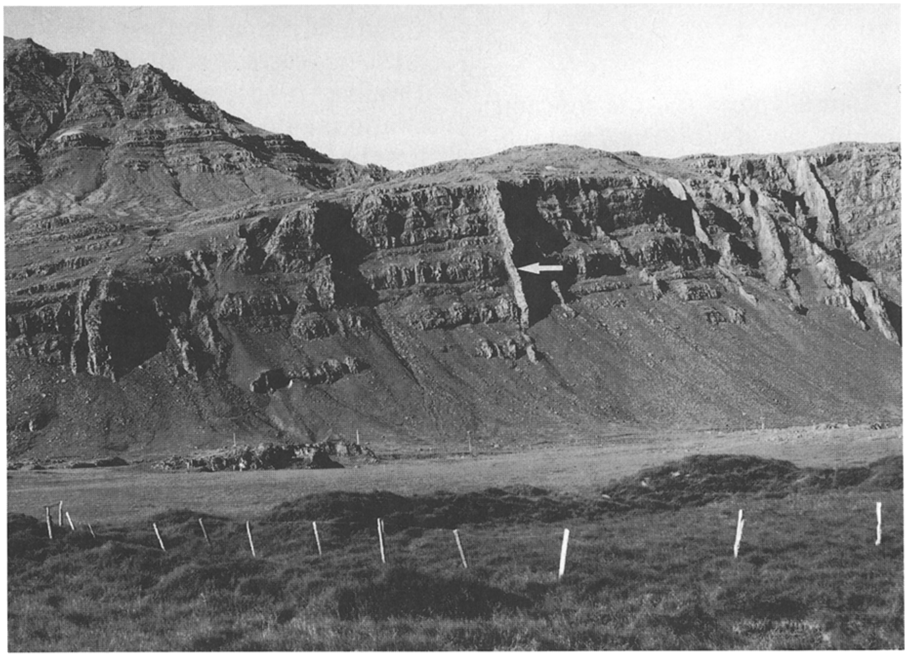

Fig. 10. Dykes from part of profile 6 in the Alftafjördur swarm (located in Fig. 9), looking north. The most conspicuous dykes are 5-10 m thick (one is indicated by an arrow), but the average thickness of dykes in this profile is 5.5 m. The lava pile is tilted 6-8° to the east.

by the author. The thickest multiple (triple) dyke (54 m), however, is on the Tröllaskagi peninsula at the western side of Eyjafjördur (Långbacka and Gudmundsson, 1993). The thickest composite dyke is 35 m and occurs in eastern Iceland (Gudmundsson, 1985). The average thickness of the regional Tertiary dykes in eastern and northwestern Iceland is 4.1 m and 4.3 m, respectively, but only 1.4 m for the Pleistocene dykes in southwestern Iceland. The average thickness of the Tertiary dykes in the Eyjafjördur profile is 5.4 m but on Tröllaskagi it is 5.9 m, which is the greatest average thickness of dykes obtained in Iceland.

The dyke-thickness size distributions in swarms in Iceland are commonly power laws. Thus, the arithmetic average thickness does not represent the most common (mode) dyke thickness (Fig. 11c). It is also known that thick dykes in any particular swarm tend to be longer than the thin dykes in the same swarm (Gudmundsson, 1983, 1984a). Observations by the author indicate that volcanic-fissure (feeder-dyke) length size distribution in the rift zone is a power law, suggesting that the size distribution of dyke lengths in the rift zone is also a power law. In view of these results it is proposed that dykes in the rift-zone volcanic systems in Iceland commonly grow as self-affine (sometimes self-similar) structures. Accordingly, all dyke dimensions (thickness, length and height) should be proportional to the number of magma pulses injected into the dyke-fracture during the formation of the dyke (cf. Gudmundsson, 1984b).

The crustal dilation due to dykes in the regional swarms varies widely (Gudmundsson, 1983, 1984a, 1990a). In the most deeply exposed Tertiary swarms in eastern and northwestern Iceland, the average dilation in 5-10-km-long profiles across the dyke swarms is commonly 5-6%. In some Tertiary swarms, however, the dilation is as low as 1-2% (Gautneb and Gudmundsson, 1992). In the Tertiary swarm in Eyjafjördur, the dilation reaches an average of 15% in a 14-km-long completely continuous part (with 100% exposure) of the 17-km-long profile

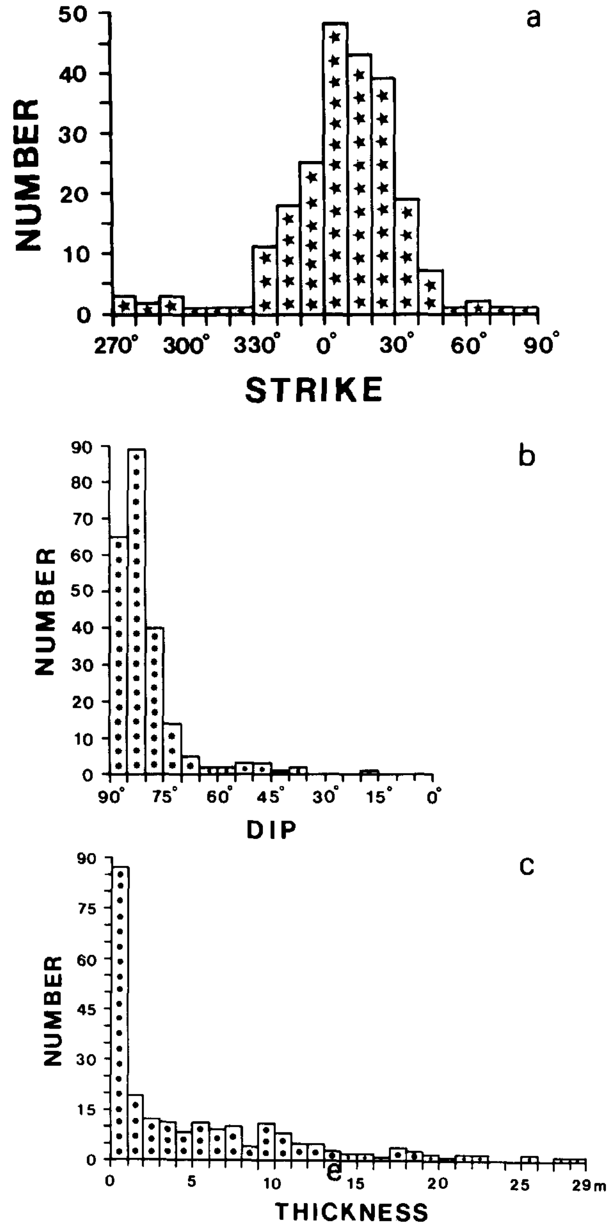

Fig. 11. Strike (a), dip (b) and thickness (c) distributions of 226 dykes from the Eyjafjördur profile (on the western coast of the Flateyjarskagi peninsula, located in Fig. 1) in northern Iceland. The arithmetic mean dip of the dykes is 80.7° and the mean thickness is 5.4 m.

but as much as 28% in a 4.5-km-long profile at the north coast of the Trollaskagi peninsula (Långbacka and Gudmundsson, 1993). In a Pleistocene dyke swarm in southwestern Iceland the crustal dilation due to dykes is 3–4% (Forslund and Gudmundsson, 1991).

Tens of dykes have been seen end vertically in the lava pile. The dykes end in different ways, but the great majority simply taper away (Fig. 12). With one possible exception (Gudmundsson, 1984a), there are no fractures that could be generated by the dykes ahead of their upper ends. Many basaltic Tertiary and, especially, Pleistocene dykes, were feeders to sills, but not a single one has be found that was, beyond reasonable doubt, the feeder to a lava flow. Many Tertiary and Pleistocene dykes have, indeed, been suggested to be feeders (Rickwood, 1990), but none of those observed by the author has been clearly

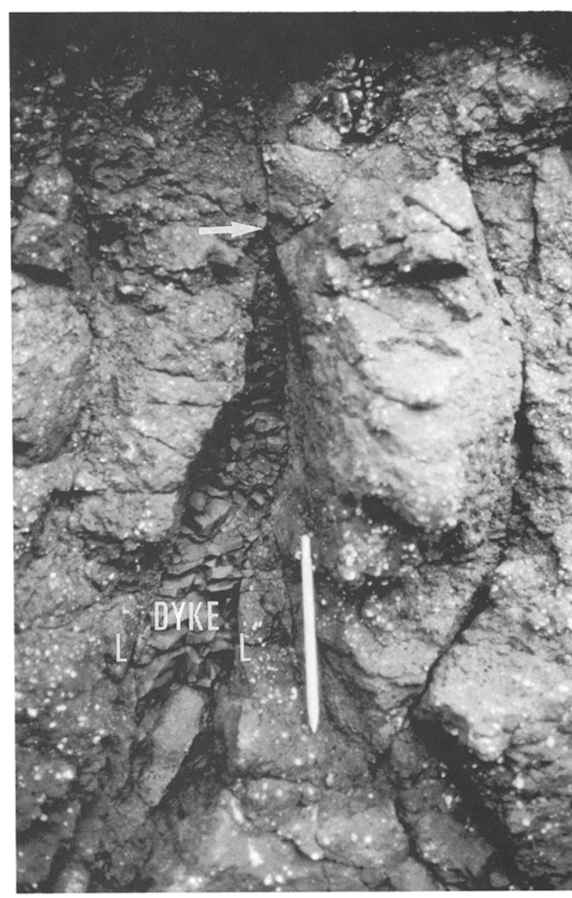

Fig. 12. Typical ending of dykes in Iceland. The dyke dissects a basaltic lava flow (L) and simply tapers away in a vertical section (the arrow indicates the upper end). The pencil provides a scale.

connected to the flow that it is supposed to feed. This does not mean that feeder-dykes do not occur in Iceland, only that the majority of dykes are probably non-feeders. The scarcity of observed feeder-dykes in the Tertiary and Pleistocene lava pile can be explained in terms of a probabilistic model (Gudmundsson, 1984b).

Several feeder-dykes have been reported from the late Pleistocene and Holocene areas of the rift zone (e.g., Saemundsson, 1967; Jonsson, 1978). By far the best example, however, is the feeder to the 6000–8000-year-old Sveinar–Randarholar crater row in northeastern Iceland (Fig. 13). The dyke is best exposed in the eastern wall of a 100-m-deep canyon, of Holocene age, of the river Jökulsa a Fjöllum. The dyke is very variable in thickness. In the eastern wall of the canyon next to the river a segment of the dyke, offset 85 m to the east, is only 2 m thick. The segment that connects with the crater of the Randarholar crater row, however, is 4.5 m thick next to the river, but 6.7 m thick at the foot of the vertical cliffs below the crater of Randarholar. On approaching the crater (and the surface) the dyke becomes thicker and is about 10 m where it connects with the crater and associated lava flow. In the western wall of the canyon there is an offset 2-m-thick segment of this same dyke. The dyke strike is N4°W, the dip is from 82°W to about 90°, and the dyke rock is very fine grained, with well-developed columnar jointing but no visible vesicles.

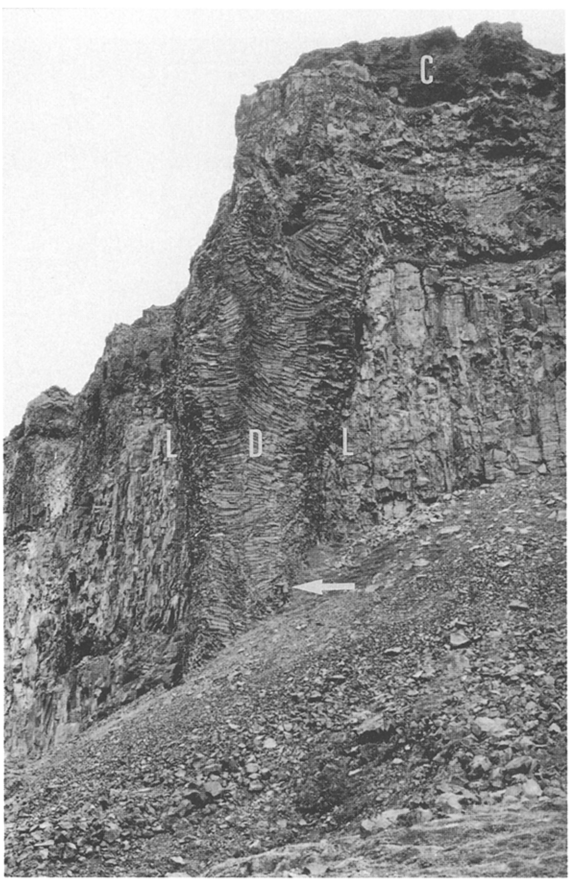

Fig. 13. Looking north at the feeder-dyke to the 6000–8000-year-old Sveinar–Randarholar crater row in northern Iceland. The dyke (*D*) dissects a Pleistocene lava flow (*L*) and feeds a crater (*C*). The person standing at the bottom right margin of the dyke and indicated by an arrow provides a scale.

Like the feeder-dyke described above, dykes are commonly offset in lateral sections and many appear to be discontinuous. The lateral distance (the offset) between the dyke segments is frequently several times the thickness of the segments, whereas in other cases this lateral distance is similar to, or less than, the thickness of the dyke segments. In vertical sections, large offsets of dyke segments are less frequent and the offset parts are commonly connected by very thin dyke parts (igneous veins). Large offsets and discontinuities in lateral sections would normally indicate vertical or inclined flow of magma in the dyke.

3.3. Sheet swarms

In the roots of most deeply eroded Pleistocene and Tertiary central volcanoes in Iceland there are dense swarms of inclined (cone) sheets. Examples include many Tertiary central volcanoes in eastern and southeastern Iceland (Walker, 1975; Torfason, 1979), and several volcanoes in western (Gautneb et al., 1989; Gautneb and Gudmundsson, 1992), northwestern (A. Gudmundsson, unpubl. data) and northern (Annells, 1968) Iceland.

All observed sheet swarms in Iceland are confined to central volcanoes, are commonly circular or slightly elliptic in shape and several kilometres in radius. The strike and dip of the sheets have wide spreads, and abrupt changes in strike and dip, as well as offsets of individual sheets and cross-cutting, are common. The thickness distribution, however, has a narrower range than that of the regional dykes. The sheet thickness is down to a few centimetres but rarely exceeds ten metres, with a mean close to one metre. Most sheets dip towards the centre of the volcano to which they belong. Many sheet swarms are associated with large plutons, the uppermost parts of the extinct crustal chambers that were the sources of the sheets. Inclined sheets are common feeders to lava flows in the active central volcanoes. Examples include many eruptions in the volcanoes Askja and (presumably) Grimsvötn (Fig. 7) during this century.

As an example of a typical sheet swarm one may consider the swarm associated with the Pleistocene Hvalfjördur central volcano in western Iceland (Fig. 1). The strike, dip and thickness distributions of 257 sheets from this swarm are shown in Fig. 14. The sheet strike is spread over the whole circle, but there is a peak in the northeastern sector of the circle, similar to the peaks in the sheet swarms associated with the nearby central volcanoes at Kjalarnes (Gudmundsson et al., 1992b) and Hafnarfjall (Gudmundsson, 1990a). This strike distribution may be explained in terms of the interaction of the local stress fields associated with the shallow source chambers and the regional stress field associated with the rift zone (Gautneb et al., 1989; Gudmundsson et al., 1992b).

The dip-frequency distribution has two peaks, one corresponding to shallowly dipping sheets and the other to steeply dipping sheets (Fig. 14b). Similar dip distributions occur in the nearby sheet swarms, though with less conspicuous peaks (Gudmundsson, 1990a; Gudmundsson et al., 1992b). Again, the distribution can be explained by the steeply dipping sheets being largely controlled by the regional stress field and the shallowly dipping ones being controlled by the local field associated with the shallow source chamber. The dip of the Hvalfjördur sheets

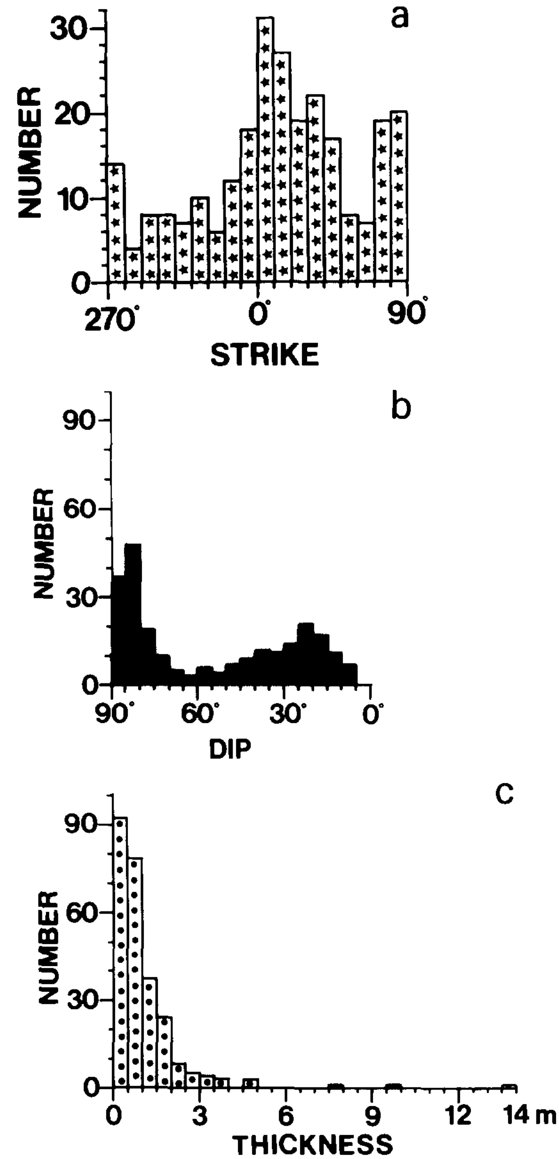

Fig. 14. Strike (a), dip (b) and thickness (c) distributions of inclined sheets in the swarm associated with the Hvalfjördur central volcano in southwestern Iceland (located in Fig. 1). The peak in strike in the northeastern sector of the circle reflects the effects of the regional stress field associated with the rift zone in southwestern Iceland. The interaction between the local and the regional stress fields is also reflected in the two-peaked dip distribution. The majority of the sheets are less than 1 m thick.

ranges from 5° to 90°, which is similar to that in the nearby swarms.

The sheets of the Hvalfjördur swarm range in thickness from several centimetres to fourteen metres (Fig. 14c). For comparison, the thickness range of the Hafnarfjall sheets is 0.03–11 m and that of the Kjalarnes sheets is 0.1–10 m. The thickness has a power-law size distribution with a negative exponent (hyperbolic) as has been obtained for other sheet swarms in Iceland (Gudmundsson, 1990a).

The great majority of the sheets are basaltic (tholeiitic) in composition. Petrological studies of the swarms associated with the volcanoes of Hafnarfjall (Gautneb et al., 1989) and Reykjadalur (Gautneb and Gudmundsson, 1992) indicate that the most evolved sheets occur near the centre of the swarm and that the sheets are more evolved than the dykes of the associated regional swarms.

3.4. Calderas and plutons

Many calderas have been observed in the late Tertiary and Pleistocene central volcanoes (Fig. 7). These are generally of similar dimensions, vertical displacements and shapes as the calderas of the neovolcanic zones. For example, Johannesson (1975) mapped a caldera in the Tertiary Reykjadalur central volcano in western Iceland with a 4-km-long minor and a 5-km-long major semi-axes (cf. Gautneb and Gudmundsson, 1992). Johannesson estimated the maximum vertical displacement on the caldera fault as 800 m. Franzson (1978) studied the Tertiary Hafnarfjall caldera, with 1.3-km-long minor and 3.6-km-long major semi-axes, and estimated the vertical displacement as being of the order of several hundred metres (cf. Gautneb et al., 1989). Fridleifsson and Kristjansson (1972) mapped the Pleistocene Stardalur caldera. According to their maps, the caldera is almost circular with a diameter of 6.5 km, but no estimate of the vertical displacement was made. As a final example, one may mention the Geitafell caldera, studied by Fridleifsson (1983). He estimated the length of the semi-minor axis as 4 km, that of the semi-major axis as 5 km, and the subsidence as being from 200 m to 500 m.

Many large plutons occur in the roots of the most deeply eroded extinct Tertiary and Pleistocene central volcanoes. The largest one is Vesturhorn (Roobol, 1974), with an exposed area of over 19 km^2, the second largest one is Slaufrudalur (Beswick, 1965), with a surface area of 15 km^2, and the third one is Austurhorn (Blake, 1966), with a surface area of 11 km^2. All these large plutons occur in southeastern Iceland. Vesturhorn and Austurhorn are made of basic and acid intrusives, but Slaufrudalur is entirely acid. Most large plutons in Iceland, however, are basic and usually made of gabbro (Fridleifsson, 1977; Torfason, 1979).

Many large plutons are associated with sheet swarms. Magma bodies of tens of cubic kilometres, as these plutons must be, are unlikely to be able to generate space for themselves at depths of only 1–2 km below the surface without the roof failing frequently and magma ascending to the surface. These bodies of magma must, therefore, have given rise to dyke intrusions and, by implication, extrusions and are thus extinct magma chambers.

4. Mechanics

4.1. Tectonic framework

Gudmundsson (1987c) suggested that each volcanic system in the rift zone is supplied with magma from a dome-shaped magma reservoir located at 8–12 km depth, at the boundary between the crust and upper mantle (Fig. 15). The magma reservoirs are partially molten, except for the uppermost parts which are likely to be totally molten. Because of (mostly relative) tensile stress concentration, crustal failure and spreading is most likely to take place where the crust is thinnest and weakest. The reservoirs are thus analogous to notches of half-ellipsoidal shape, with cross-sectional areas similar to those of the volcanic systems that they feed (Fig. 15).

Various geophysical measurements indicate that the upper mantle beneath Iceland is partially molten down to the depth of at least several hundred kilometres (Tryggvason et al., 1983; Gudmundsson, 1987c). Below the axial rift zone, this partially molten mantle is at the depth of 8–

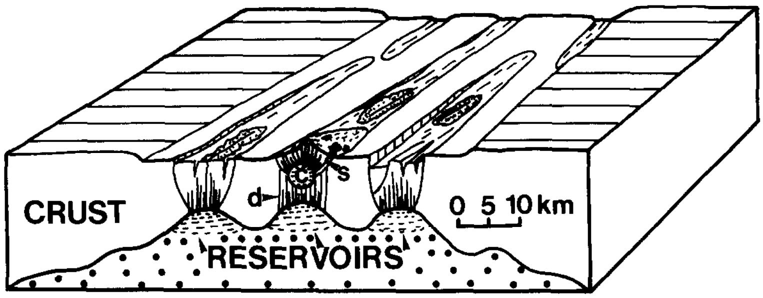

Fig. 15. Schematic illustration of the proposed magma reservoirs below the volcanic systems of the rift zone in Iceland. The reservoirs are semi-ellipsoidal, dome-shaped top regions of the partially molten mantle beneath the rift zone. The reservoirs are separated by regions of thicker crust, so that the magma in each reservoir can develop independently of the magmas in the neighbouring reservoirs. The infrastructure of the volcanic system in the centre is marked by dykes (*d*), inclined sheets (*s*), and a crustal magma chamber (*c*) that feeds a central volcano, whereas the other systems represent the regions outside the central volcanoes. The scale is approximate. (Modified from Gudmundsson, 1987c).

12 km but its depth gradually increases to 20–30 km farthest from this zone (Björnsson, 1985; Eysteinsson and Hermance, 1985). The melt in the mantle migrates upwards, as a results of buoyancy, but becomes temporarily trapped at the bottom of the semi-impervious crust. Inside the rift zone, the magma is normally unable to propagate into the crust (through dykes) except during rifting episodes. According to the model presented here, the magma reservoirs are the melt traps at the bottom of the crust, and there the primitive melt accumulates, changes its composition (through anatexis of the crust and fractional crystallisation) and gradually becomes basaltic magma (olivine tholeiite or tholeiite) or, less frequently, more evolved magma.

The birth of any volcanic system within the rift zone is here thought to coincide with the initiation of a magma reservoir at the bottom of the crust. The proposed semi-ellipsoidal form of the reservoirs reflects the elliptical cross-sections of the associated volcanic systems (Fig. 15). The crust above the reservoir undergoes some uplift and bending (doming), as a result of crustal extension due to plate movements as well as increasing magmatic pressure, that contribute to the reservoir becoming semi-ellipsoidal in form. The pressure increase is partly attributable to the increased volume of magma in the reservoir, due to inflow of magma, and partly to a petrological evolution where the magma in the top region of the reservoir becomes less dense and thus more buoyant.

During the early activity of a volcanic system, there exists no shallow crustal magma chamber so that all intrusions (dykes) and extrusions are fed directly from the reservoir. Some volcanic systems, in particular those where the rate of spreading is relatively low (Gudmundsson, 1990b), apparently never develop crustal magma chambers (Gudmundsson, 1986). The earliest dykes in a developing regional dyke swarm are thus exclusively injected from the deep-seated reservoir, whereas the later-formed dykes may, in some volcanic systems, be partly injected laterally from a subsequently formed crustal magma chamber. The early stages of all volcanic systems are thus characterised by the development of a regional dyke swarm.

4.2. Dyke emplacement

The physics of dyke emplacement has received much attention in recent years. In particular, the fluid dynamics of magma flow in fractures has been considered by many authors (e.g., Spence et al., 1987; Lister and Kerr, 1991). In the schematic model explored below, the focus is on the dyke-injection frequencies in the rift-zone volcanic systems, in which case the details of the magma dynamics are not relevant. The model is quantitative, however, and its predictions are tested on results from dyke and sheet swarms, as well as on the dyke-injection frequency of the Krafla volcanic system.

During rifting episodes in Iceland, the crustal segment that fails is normally much longer than it is thick. For example, the volcanic systems are commonly 40–100 km long whereas the crustal thickness is only 8–12 km. Failure would normally start at one or several points of high tensile stress concentration in the reservoir's roof. From these points, the failure spreads laterally and vertically into the crustal segment above the reservoir, thereby opening a pathway for the vertical flow of magma into the crustal segment. Lateral spreading of failure is often accompanied by earthquake swarms, as occurred during the 1975–1984 rifting episode in the Krafla volcanic system in northern Iceland (Brandsdottir and Einarsson, 1979; Einarsson and Brandsdottir, 1980).

The direction of flow of magma may be at any angle to the direction of spreading of the failure zone into which the magma flows. Nevertheless, due to magma buoyancy, the direction of flow of magma would commonly be perpendicular to the direction of spreading of the failure zone, especially in the lower part of the crust where the density difference between the magma and the crustal layers is greatest (Fig. 16). Because stress is force per unit area, dyke intrusion into part of the reservoir roof increases the tensile stress concentration in the neighboring parts so that the failure, hence the dyke injection, may gradually spread along the whole crustal segment above the reservoir. The lateral spreading of failure, and the propagation of the dyke fracture, are facilitated by the vertical flow of magma into the failure zone, and the magma front would normally be near the failure front at any instant. This may explain the close association of magma flow and seismicity during rifting events in Iceland (Brandsdottir and Einarsson, 1979).

The failure of the crustal segment of a particular volcanic system, and the associated flow of magma out of the reservoir, may take many months or years. Gradual failure of the volcanic system, with dyke emplacement and volcanism, is referred to as a rifting episode, whereas individual events within that episode take only days or weeks and are referred to as rifting events. For example, the 1975–1984 rifting episode in the Krafla volcanic system lasted 9 years and included at least 21 rifting events. During a rifting episode there is flow of magma out of the reservoir, and this flow continues, perhaps intermittently, while the episode lasts.

The frequency of dyke injection from a magma reservoir at a divergent plate boundary can be estimated as follows (Gudmundsson, 1988b). The condition for the flow of magma (through a dyke) from the reservoir into the crust is:

$$p_l + p_e = \sigma_h + T \quad (1)$$

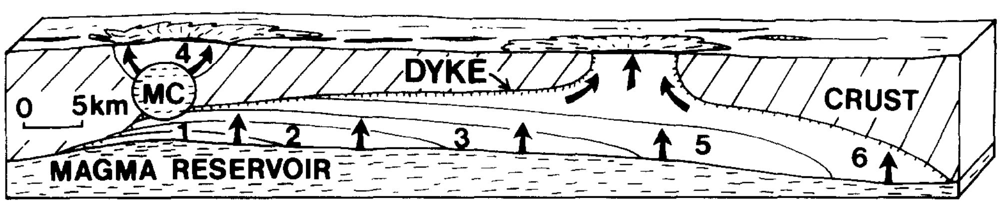

Fig. 16. Proposed model for the emplacement of regional dykes at the divergent plate boundary in Iceland. This schematic illustration applies to a particular rifting event within a longer rifting episode and assumes that the dyke grows as a self-affine structure. The shape of the dyke in this vertical cross section, as a function of time, is indicated by the numbers *1*, *2*, *3*, *5* and *6*, but the number *4* refers to the injection of an inclined sheet from the magma chamber (*MC*) which is triggered by the dyke meeting the chamber during stage 3. The main direction of flow of magma is indicated by black arrows; the magma flow is primarily vertical whereas the dyke fracture propagates essentially laterally at any crustal level. The dyke and the sheet feed lava flows at the surface. Only a part of the volcanic system is shown here, and in subsequent events (repeated flow of magma) the crustal part to the left of this dyke (and chamber) may also fail, resulting in a vertical flow of magma (dyke injection) into that part. A regional dyke is often discontinuous in a lateral section and may also propagate into the crust without meeting a crustal chamber.

where p_l is the lithostatic pressure, p_e is the excess magmatic pressure (in excess of σ_h) before flow of magma out of the reservoir starts, σ_h is the horizontal compressive stress in the roof of the reservoir, perpendicular to the dyke, and T is the tensile strength of the roof (crust). The condition of Eq. (1) is probably largely reached by the buildup of (relative) tensile stress in the roof of the reservoir as a result of divergent plate movements (Gudmundsson, 1988b). It follows from Hooke's law in elasticity that the buildup rate of tensile stress σ in the roof of the reservoir is:

$$\frac{d\sigma}{dt}=\frac{vkE}{u} \quad (2)$$

The corresponding dyke injection (magma flow) frequency I_f is:

$$I_f=\frac{vkE}{Tu} \quad (3)$$

where t is time, v is the spreading rate, E is Young's modulus, T is the tensile strength of the reservoir roof, u is the width of the zone undergoing tensile strain during plate movements (approximately equal to the width of the plate-boundary zone), and k is the tensile stress concentration factor, defined as the ratio of the maximum tensile stress at the boundary of the reservoir to the regional tensile stress associated with spreading.

From Eq. (3) one can calculate roughly the dyke-injection frequency associated with a particular reservoir beneath the rift zone of Iceland. Gudmundsson (1990a) calculated the dyke injection frequency for a typical volcanic system in Iceland and compared the results with the dyke swarms in the Tertiary areas of eastern Iceland. He concluded that for a volcanic system active for 0.3–1.0 Ma, the number of dykes that should reach the crustal level of the presently exposed swarms is 500–1700. The estimated number of dykes in the Alftafjördur swarm (Figs. 9 and 10), at the present level of exposure, is 1000–1500, which agrees with the model predictions.

Here we apply this model to the Krafla volcanic system in northern Iceland. This is a typical rift-zone volcanic system and its last volcano-tectonic episode was in 1975–1984. This rifting episode started 20 December 1975 after a repose of about 250 years (Björnsson, 1985; Opheim and Gudmundsson, 1989). There is a crustal magma chamber, with a top at the depth of 2–3 km and extending down to 6–7 km, located below the Krafla central volcano (Einarsson, 1978; Tryggvason, 1980), but the deep-seated magma reservoir, with which we are concerned here, is located at the depth of 8–12 km below the whole volcanic system of Krafla.

Using Eq. (3) as a basis, one can calculate roughly the magma flow frequency of the Krafla reservoir. The volcanic systems in northeastern Iceland are arranged en echelon (Fig. 1), but they overlap to a great extent so that the factor u (the width of the zone that undergoes tensile strain during plate movements) must be taken as the combined width of these volcanic systems, or roughly 40 km. The static Young's modulus for crustal layer 3 in Iceland is about 50 GPa, the tensile strength is about 3 MPa, and the total spreading rate is 2 cm yr^{-2} (Gudmundsson 1988b). Taking the tensile stress concentration factor k as 1.0, i.e., assuming no stress concentration, which would apply to a flat reservoir, then eqn. (3) gives $M_f=0.0083$ yr^{-1}, or about one magma flow every 120 years.

This figure should be accurate within a factor of two, the main uncertainty being the value of the parameter u. This value may be from about 30 km to about 60 km, in which case there would be one magma flow every 90–180 years. During any particular short period, say a few thousand years, one of the volcanic systems may take up most of the tensile strain during which time the magma flow frequency in that system would be much higher than on average. Over tens of thousands of years, however, such anomalies should even out and magma flows from one of the 3–4 reservoirs beneath the northeastern volcanic zone might occur at intervals of 90–180 years. This agrees very well with the historical evidence, suggesting that volcano-tectonic activity in this zone takes place every 100–150 years (Björnsson, 1985).

For the Krafla reservoir these results suggest that a magma flow should occur once every 300–700 years, the range being related to the poorly constrained factor u as well as to the uncertain number of reservoirs (3 or 4) parallel to the Krafla reservoir (Fig. 1). Not every magma flow from a reservoir gives rise to an eruption. Model calculations suggest that the long-term (i.e., over tens of thousands of years) eruption frequency of a volcanic system may be as little as 10% of the magma flow frequency of the source reservoir (Gudmundsson, 1988b). There is evidence of 7–9 major volcano-tectonic episodes in the Krafla system during the Holocene (Saemundsson, 1991), whereas the corresponding number of magma flows from the reservoir should, according to the model presented here, be in the range 14–34. If all the volcano-tectonic episodes that actually happened in this system during Holocene have been identified, roughly three-fifth to one-fifth of the magma flows led to eruptions, which is in reasonable agreement with the model predictions (Gudmundsson, 1988b).

4.3. Formation of crustal magma chambers

At divergent plate boundaries such as the rift zone of Iceland, the stress field favours dyke emplacement rather than sill emplacement. Nevertheless, sills do occur in the extinct volcanic systems of Iceland and are relatively common in the Pleistocene areas (Forslund and Gudmundsson, 1991). The condition for sill formation is that the horizontal compressive stresses are higher than the vertical stress. At divergent plate boundaries such stress barriers may be temporarily formed, particularly where the crust is a multilayer of competent and incompetent rocks, as a result of rapid injection of dykes with high magmatic overpressures (Gudmundsson, 1986, 1990b; Parsons et al., 1992).

In the rift zone, crustal magma chambers may develop from sills, either when many nearby sills combine into a large one, or when an initial thick sill absorbs the magma of many dykes that enter it in a rapid succession while the sill is liquid (Gudmundsson, 1986, 1990b). The sill grows into a crustal magma chamber partly by elastic-plastic expansion as occurs during the emplacement of laccoliths (Fig. 17). Part of the space needed for the chamber is, however, generated by anatexis and stoping of the host rock.

When the crustal magma chamber grows, its shape may change and become nearly spherical or prolate ellipsoidal instead of the initial oblate ellipsoidal shape of the sill. The local stress field around the chamber then also changes (Gautneb et al., 1989) and, because this stress field controls the dip and strike of the inclined sheets derived from the chamber, later sheets may cut the earlier ones. This is one reason for the very common cross-cutting relationships observed in the local sheet swarms.

4.4. Sheet emplacement

A common scenario for the rift-zone volcanic systems in Iceland could be as follows (Figs. 15 and 16). During a rifting episode, magma from a deep-seated reservoir at the bottom of the crust flows, sometimes repeatedly, vertically into the crust and occasionally right up to the surface. If this vertically flowing magma meets a crustal magma chamber, it partly ponds in the chamber and normally triggers dyke injections from the chamber, some of which lead to volcanic eruptions. The dykes injected from the crustal chamber are mostly inclined sheets or radial dykes and are of less volumes, on average, than the regional dykes from the source reservoir. Because the chamber volume is normally much less than the source-reservoir volume, a single magma flow (dyke injection) from the reservoir may trigger many magma flows (sheet injections) from the chamber.

The number of sheets from a typical magma chamber triggered by a single magma flow (dyke injection) from the source reservoir can be estimated as follows (Gudmundsson, 1990a). Let the partially molten source reservoir be of combined melt and solid matrix volume V_r, melt fraction (porosity) ϕ, pore compressibility β_p, and magma compressibility β_m. Then the volume of magma V_e^r that flows out of the reservoir through a single regional dyke (which may be formed in many rifting events) during a rifting episode is:

Fig. 17. Elastic–plastic expansion and uplift of the lava pile in the vicinity of the Sandfell (S) laccolith (of rhyolite) in eastern Iceland. This is one mechanism by which crustal magma chambers at deeper crustal levels generate space for themselves during their growth. This laccolith is described by Hawkes et al. (1933).

$$V_e^r = \phi p_e (\beta_m + \beta_p) V_r \quad (4)$$

The total volume of magma V_e^c in a sheet injected, during a particular rifting event, from a crustal magma chamber of combined melt and solid matrix volume V_c is:

$$V_e^c = \gamma p_e (\beta_m + \beta_c) V_c \quad (5)$$

where γ denotes the melt fraction (porosity) of the chamber ($\gamma = 1.0$ for a totally molten chamber) and β_c the compressibility of the chamber host rock. Chambers at slow-spreading plate boundaries receive magma infrequently and may be partially molten, but at fast spreading boundaries, and in areas like northern Iceland where one volcanic system may take up the whole spreading in any particular segment of the rift zone, the chambers are likely to be more or less totally molten.

Theoretical considerations, supported by seismic, magnetotelluric and field observations, suggest that the most likely site for crustal magma chambers in Iceland is where the crustal density equals that of typical basaltic magmas (Gudmundsson, 1986, 1988b). Magma buoyancy may thus be neglected in the term p_e in Eq. (5). The tensile strength and other mechanical properties of the crust hosting the magma chambers are essentially the same as those of the reservoirs' roofs, so that the excess pressure p_e in a reservoir needed to inject a dyke is approximately the same as that needed to inject a sheet from the associated crustal chamber. If follows that if the volume of magma received by a chamber from its source reservoir during a rifting episode is bV_e^r ($0.0 \leq b \leq 1.0$), then, from Eqs. (4) and (5), the number of sheets N_s injected from this chamber during the same episode is:

$$N_s = \frac{b\phi V_r(\beta_m + \beta_p)}{\gamma V_c(\beta_m + \beta_c)} = \frac{bV_e{}^r}{V_e{}^c} \quad (6)$$

These results can now be applied to the Krafla volcanic system. The Krafla chamber appears to be totally molten (Einarsson, 1978; Gudmundsson, 1987c) so that $\gamma = 1.0$. One may use $\phi = 0.25$ for the porosity of the reservoir, and $\beta_m = 1.25 \times 10^{-10}$ Pa, $\beta_p = 8.8 \times 10^{-11}$ Pa^{-1}, and $\beta_c = 2.94 \times 10^{-11}$ Pa^{-1} for the compressibilities of the magma, the pores and the host rock of the chamber, respectively (Gudmundsson, 1987c). Shield volcanoes in Iceland are thought to be generated in single eruptions, probably (because of their primitive composition and often large volumes) from magma reservoirs (Gudmundsson, 1990a). The largest Holocene shield volcano in the Krafla system (Gjastykkisbunga) is about 3 km^3 (Saemundsson, 1991), which may be regarded as a rough estimate of V_e^r. By contrast the largest single eruption (from the shallow chamber) during the 1975–1984 episode was 0.08 km^3 and the average volume of the 9 eruptions during this episode is about 0.03 km^3 (Gudmundsson, 1987c). The factor b is not well constrained, but from the rifting episode in Krafla it may be inferred that it is in the range 0.1–0.5.

Substituting these figures in Eq. (6), the number of sheets injected from the Krafla chamber and triggered by a single magma flow from the Krafla reservoir could range from 4 to 50. These calculations, though crude, are in broad agreement with the data from the Krafla rifting episode 1975–1984. During that episode a single magma flow from the Krafla reservoir triggered at least 21 intrusive events (dykes and inclined sheets) from the Krafla chamber, 9 of which resulted in volcanic eruptions.

5. Summary and Conclusions

(1) Most of the volcano-tectonic activity in the rift zone of Iceland takes place in specific volcanic systems that are commonly 40–100 km long and 5–20 km wide. Outside central volcanoes the infrastructure of such systems is as follows. At the surface they consist of tension fractures, normal faults and volcanic fissures. At greater crustal depths the tension fractures gradually decrease in number and most are limited to the uppermost few hundred metres of the crust. In the depth range from a few hundred metres to roughly 1 km, steeply dipping normal faults contribute much to the crustal dilation and are, in that depth range, the dominant extensional structures in some volcanic systems. At crustal depths exceeding about 1 km the dykes gradually become the dominant extensional structures and probably continue to be so down to the bottom of the crust. Some of the main normal faults may continue to the bottom of the crust but most are limited to its upper half.

(2) Many volcanic systems have distinct central volcanoes, some of which have developed collapse calderas. Normally, a central volcano erupts once every few hundred years and has a lifetime of the order of 10^5 years. Production of intermediate and acid rocks is essentially restricted to the central volcanoes, the main topographic types of which are volcanic cones and collapse calderas. The diameters of the calderas are commonly 6–8 km and the vertical displacement (relative subsidence) several hundred metres.

(3) Many volcanic systems are petrologically distinct. It is suggested that they are all supplied with magma from partially molten magma reservoirs located at the depth of 8–12 km at the boundary between the crust and the upper mantle. The reservoirs are thought to be of half-ellipsoidal form, the dome-shape being partly attributable to crustal thinning and uplift, as a result of crustal extension during plate movements, and partly to magma accumulation and pressure increase in the reservoirs. In the absence of shallow crustal magma chambers, dyke emplacement is thought to be primarily by vertical flow of magma from these reservoirs.

(4) Rapid injections of dykes may temporarily alter the normal rift-zone stress field in such a way that subsequent dykes change into sills. A sill can develop into a crustal magma chamber if dykes inject the sill frequently enough while it is liquid. In addition to the deep-seated magma reservoirs, many volcanic systems, especially those that develop central volcanoes and collapse calderas, have shallow crustal magma chambers. The tops of these chambers are commonly 1–3 km below the surface of the rift zone, and these can be observed (as plutons of gabbro or granophyre) in the most deeply eroded extinct volcanic systems in the Tertiary areas of Iceland.

(5) A quantitative model developed in this

paper indicates that many regional dykes are injected from the deep-seated reservoirs at the bottom of the crust, whereas most inclined sheets are injected from crustal magma chambers. During rifting episodes, the loaded segment is normally much longer than it is thick, so that the failure spreads mainly laterally along part of or all the segment (that is, the roof of the reservoir). Lateral migration of earthquakes may be associated with this failure. Magma flows vertically into the failure zone and occasionally right up to the surface. If, however, the magma meets a crustal chamber it partly ponds in the chamber and may trigger injection of sheets from the chamber. Failure may start anywhere along the segment hosting the volcanic system but commonly where the crust is thinnest, particularly if that part of the segment contains a magma chamber that concentrates the tensile stress. Many rifting events, each lasting several days or a few weeks, may be associated with a single rifting episode, commonly lasting several years. Rifting events, and associated dyke and sheet injections, in a volcanic system continue until the (mostly relative) tensile stress is relaxed in that segment of the divergent plate boundary.

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