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The geometry and growth of dykes

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ABSTRACT: Field studies of dykes and volcanic fissures (feeder-dykes) in Iceland indicate that the length and thickness size distributions of dykes are commonly power laws. Many dykes are composed of several columnar rows or consist of two or more dykes separated by chilled margins (are multiple). These observations suggest that the final geometry of a dyke is commonly the result of many magma injections in the dyke-fracture. All dimensions of a dyke may increase during its growth. It is proposed that the most common thickness of dykes in a swarm is normally reached by a single magma injection from the source chamber. Thermal and mechanical considerations indicate that thick, coarse-grained regional dykes with one to three columnar rows form when the time between successive magma injections is of the order of months or less. If the dyke is composed of many columnar rows, the time between successive magma injections is at least several months and may be several years. For a multiple dyke, however, the time between successive magma injections is at least many years and more commonly hundreds or thousands of years. These results compare well with the feeder-dyke growth during the Krafla rifting episode in North Iceland 1975-1984. During nine eruptions, the feeder-dyke gradually increased its surface length to 11 km and its maximum thickness to 9 m. During each eruption, the dyke thickness increased by about 1 m. The dyke-growth model presented here indicates that the final shape of a thin dyke may be reached in several hours, but that of a thick dyke in periods of time that commonly range from months or years to hundreds or thousands of years.

1 INTRODUCTION

The author's observations of more than 5000 dykes in Iceland and elsewhere indicate that the final geometry of many dykes is the result of long-term growth. During its growth, the dyke commonly increases its length, height and thickness in a self-affine (statistically self-similar) way.

The feeder-dyke growth during the 1975-1984 rifting episode in the Krafla volcanic system in North Iceland (Figure 1) was monitored by geophysical measurements and direct observations (e.g., Björnsson et al. 1977, Tryggvason 1984, Einarsson 1991). This feeder-dyke, emplaced in nine eruptions and reaching a maximum thickness of 9 m, largely coincides with the one that was emplaced in the previous rifting episode, the Myvatn fires 1724-1729. Consequently, the dyke formed in the Krafla fires 1975-1984 is presumably a part of a multiple dyke formed over a period of at least 250 years.

This paper presents the results of the author's field studies of the geometry of mafic dykes in the extinct Tertiary and Pleistocene volcanic systems Iceland.

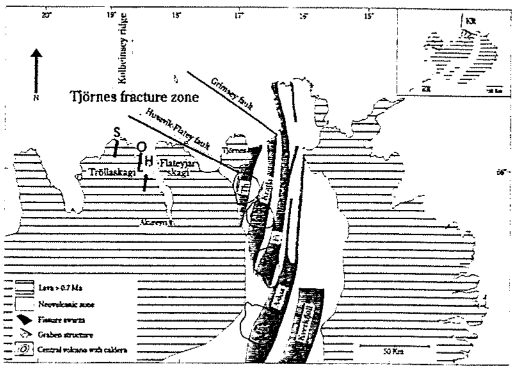

Fig. 1. General map of the geology of Northeast Iceland. The dyke swarms of Siglufjördur (S), Olafsfjördur (O) and Hrisey (H), as well as the Krafla volcanic system, the Krafla central volcano, and its caldera, are indicated. The other Holocene volcanic systems of the rift zone in North Iceland are Theistareykir (Th), Fremri-Namur (Fr), Askja and Kverkfjöll. The Tjörnes fracture zone, with its two main seismic lineaments (the Grimsey fault and the Husavik-Flatey fault), is a transform fault (active for 7-9 Ma) that connects the Kolbeinsey ridge and the rift zone in North Iceland. On the inset figure of Iceland, the Kolbeinsey ridge (KR), the Reykjanes ridge (RR), and the neovolcanic zone (not shaded) are indicated. Modified from Fjäder et al. (1994).

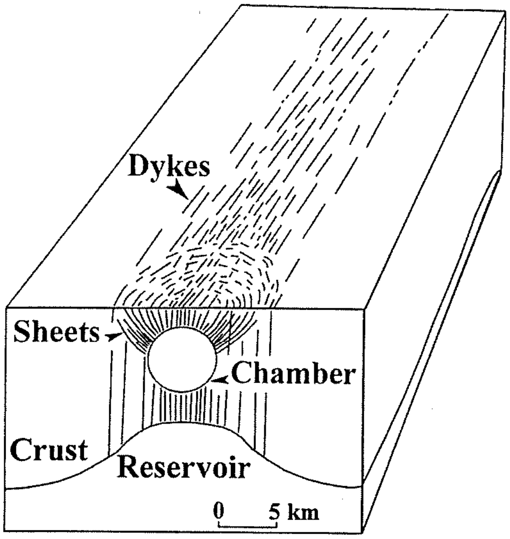

Fig. 2. Schematic illustration of the two basic types of dyke swarms in Iceland: local sheet swarms and regional dyke swarms. Most of the sheets are injected from a small, crustal magma chamber, whereas many of the regional dykes are injected from a much larger magma reservoir that underlies the whole volcanic system and is located at the bottom of the crust. During a particular volcano-tectonic rifting episode, such as in Krafla 1975-1984, many sheets may form even if only one regional dyke is emplaced. Modified from Gautneb & Gudmundsson (1992).

These results, with simple thermal and mechanical considerations, are used to provide a model of dyke growth. The model predictions are compared with, and constrained by, the observations of the feeder-dyke growth during the Krafla fires 1975-1984.

2 GEOMETRY OF DYKES IN ICELAND

Dykes in Iceland are of two main types (Figure 2): inclined (cone) sheets and regional dykes (Gudmundsson 1990). The sheet swarms are confined to central volcanoes, are circular or slightly elliptical in shape and several kilometres in radius. Most sheets dip towards the centre of the shallow source magma chamber. Commonly, the dip distribution has two peaks; one at 20°-50°, the other at 75°-90°. The most common (mode) sheet thickness is normally less than half a metre; the thinnest sheets are a few centimetres, the thickest ones rarely exceed ten metres.

In regional swarms, most dykes dip steeply and strike subparallel to each other. The swarms are elongate, tens of kilometres long and several kilometres wide, their dimensions being comparable to those of the active volcanic systems. The regional dykes range in thickness from a few centimetres to sixty metres; the mode thickness in individual swarms is commonly one or two metres.

2.1 Dyke geometry in subhorizontal exposures

Four basaltic dykes have been followed along their length in the field in East Iceland. Their lengths are from 4 km to 22 km (Gudmundsson 1983). The estimated lengths of 12 basaltic dykes in Northwest Iceland are from 3 km to 12 km (Gudmundsson 1984a). The lengths of these 16 dykes are presumably minimum values because the lateral ends are not seen. The mean length of these 16 dykes is 8 km and their mean thickness is 9.5 m. The length/thickness ratios of these dykes range from 300 to 1500 (for comparison, that of the Krafla feeder at the surface is 1375), with a mean value of 926, suggesting that at depths of 0.5-1.3 km below the initial top of the basaltic lava pile (where these dykes were measured) this ratio is commonly around 1000.

The lengths of volcanic fissures (feeder-dykes at the surface) are from less than one hundred metres to tens of kilometres. The longest Holocene volcanic fissure in Iceland is the 6000-8000-year-old Sveinar-Randarholar crater row in Northeast Iceland (Bäckström & Gudmundsson 1989); its length is at least 50 km and may be as much as 70 km. The Laki 1783 volcanic fissure in South Iceland is at least 27 km long (Thordarson & Self 1993). The volcanic fissure that formed in the Krafla fires 1975-1984 is 11 km long (Saemundsson 1991).

The most detailed data on the length distribution of Holocene volcanic fissures in Iceland is from the Reykjanes Peninsula in Southwest Iceland. The

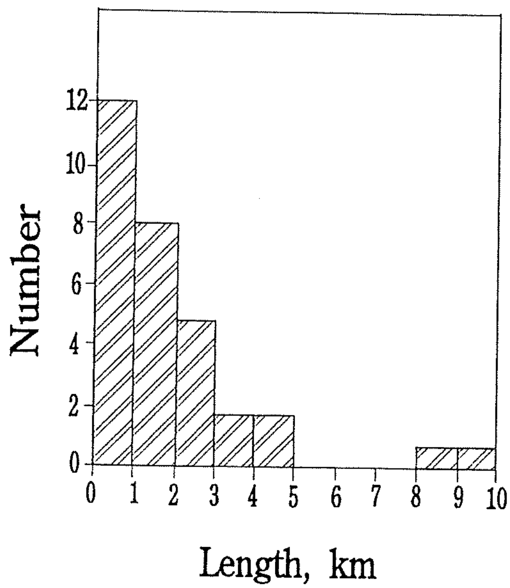

Fig. 3. Length size distribution of 31 Holocene volcanic fissures on the Reykjanes Peninsula in Southwest Iceland. Data from Jonsson (1978).

longest fissure is a nearly 10-km-long crater row, containing over 100 craters. The length size distribution is a power law (Figure 3), suggesting that length size distributions of feeder-dykes, and by implication non-feeders as well, are commonly power laws.

Most volcanic fissures (feeder-dykes) are discontinuous in lateral sections. Some parts of the longest volcanic fissures, such as the Sveinar-Randarholar crater row, may have erupted at different times. The whole of the Laki 1783 volcanic fissure, however, was active at the same time, but it remained discontinuous throughout the eruption (Thordarsson & Self 1993). The volcanic fissure formed in the Krafla fires is discontinuous (Saemundsson 1991).

Most regional Tertiary and Pleistocene dykes are discontinuous and consist of segments that are commonly offset by distances that exceed their thicknesses. The arrangement of some dyke-segments is en echelon; more often it is irregular (cf. Hoek 1991). When the distance between the nearby ends of two segments is small compared with the lengths of the segments, they are normally connected by thin veins or other igneous structures. When the distance is large, however, the segments are normally

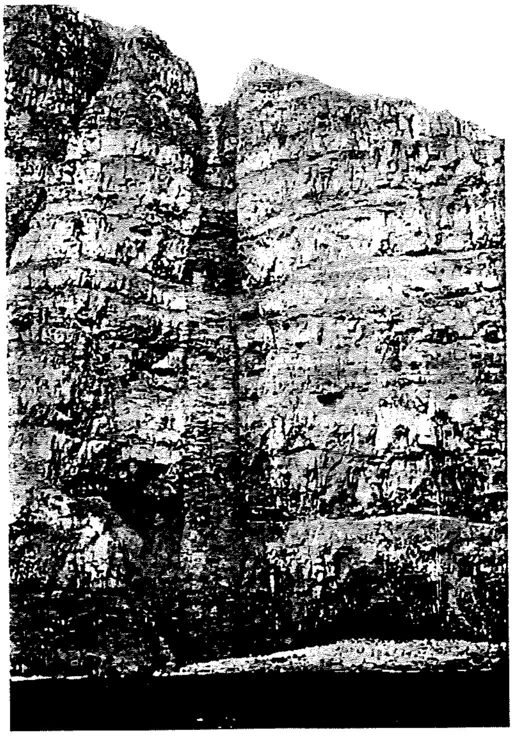

Fig. 4. Straight dyke in the lava pile on the eastern coast of the Tröllaskagi peninsula in North Iceland (Fig. 1). View SW, the dyke consists of one or two columnar rows, strikes N37°E, dips 83°E, and is around 6 m thick.

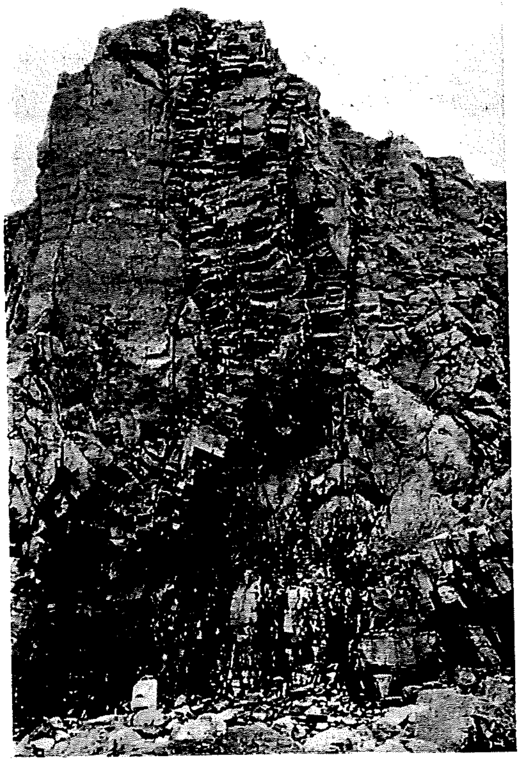

Fig. 5. Sinuous, thin dyke on the eastern coast of the Tröllaskagi peninsula in North Iceland (Fig. 1). View S, the dyke strikes N10°W, dips on average 83°E, is 2 m thick, consists of only one main columnar row and has chilled selvage at its margins.

not visibly connected. Most long dykes are laterally discontinuous at any exposure depth (below the initial surface) and may be so to their sources.

2.2 Dyke geometry in subvertical exposures

Some dykes, particularly thick ones, are quite straight in subvertical sections (Figure 4). More commonly, however, the dykes are sinuous (Figure 5) or show various degrees of irregularities. Many dykes consist of segments where the thickness varies as in an ellipse; some segments are in contact and overlap slightly (Figure 6), others are only connected by thin igneous veins. Polygonal joints in the dykes form columns (Figures 4 & 5) that occur in subvertical parallel sheets or rows.

In the Tertiary and Pleistocene lava pile in Iceland, tens of dykes have been seen to end upwards. The dykes approach their ends in different ways. Some expand somewhat and then thin out, others end bluntly, and a few ones gradate into a number of small igneous veins or dykelets. Most dykes, however, end vertically by tapering away (Figure 6). With the possible exception of one 6-m-thick, blunt-ended and faulted dyke (Gudmundsson 1984a), there

Fig. 6. The vertical end of a basaltic dyke, consisting of slightly offset and partly overlapping segments, in the Tertiary lava pile of East Iceland. View S, the dyke-rock is fine-grained basalt with sparsely distributed vesicles and amygdales, 1-2 mm in diameter. Columnar joints are well developed near the margins of the dyke, but less so in its central part. The dyke does not occupy a fault. The dyke terminates in a fine-grained basaltic lava flow where vesicles are mostly absent. Near the base of the photograph the dyke is 2 m thick.

are no major fractures ahead of the dyke's upper end. It is not clear, however, if the 6-m-thick dyke generated the associated normal faults, was captured by them, or if they were formed later. Dykes commonly follow normal-fault planes for a while and then leave the fault plane higher up in the lava pile (Forslund & Gudmundsson 1991).

Most dykes that are observed to end upwards in the Tertiary lava pile are exposed at depths of 0.5-2 km below the surface at the time of dyke emplacement. In the Pleistocene lava pile, however, there are dykes that end vertically at depths of less than a few hundred metres below the surface. Some dykes appear to have their lower ends (commonly thick and of round shape) exposed in the lava pile. All lower ends observed by the author, however, are connected by thin igneous veins with other nearby dyke segments, commonly at a somewhat greater crustal depth, suggesting that they are not the real bottoms of the dykes.

2.3 Dyke thickness

The thickness size distributions in dyke swarms in Iceland are commonly power laws. As an example, consider the dyke swarms in the 9-12 Ma basaltic lava pile of the Tröllaskagi peninsula in North Iceland (Figure 1). Apart from a few andesitic dykes, all the dykes in these swarms are basaltic (tholeiitic) and exposed at depths of 1-1.4 km below the initial top of the lava pile. The mode dyke thickness in all these swarms is around 1 m (Figure 7). The thin dykes are fine-grained and normally consist of a single or a few columnar rows (Figure 5). Several thick, coarse-grained (microgabbroic) dykes contain only one or a few columnar rows (Figure 4). The thickest single dykes in North Iceland are around 30 m (Fjäder et al. 1994); the thickest (apparently) single (basaltic and course-grained) dyke in Iceland occurs in the Pleistocene lava pile of South Iceland and exceeds 60 m. Most thick dykes, however, are either composed of several columnar rows (Figure 8) or are multiple, that is, consist of two or more dykes separated by chilled margins (Figure 9).

3 DYKE GROWTH DURING THE KRAFLA FIRES

3.1 The Krafla fires

The Krafla volcanic system (spreading centre) in North Iceland is a nearly 100 km long and 4 to 10 km

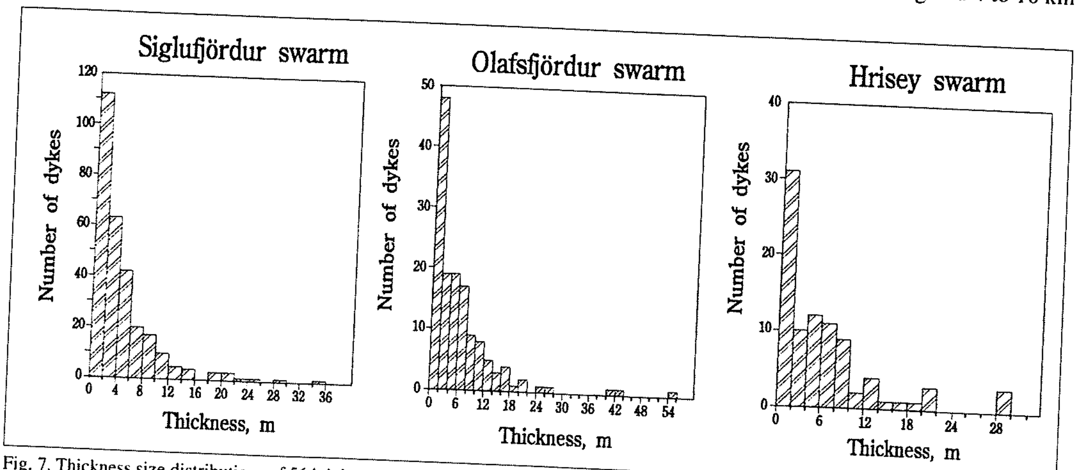

Fig. 7. Thickness size distributions of 564 dykes in three dyke swarms on the Tröllaskagi peninsula in North Iceland (Fig. 1). The arithmetic average thickness of all dykes in these swarms is 5.9 m, but all these thickness distributions are power laws so that the average is not a good indication of the mode. Around one-third the dykes are less than 2 m thick.

Fig. 8. Dyke consisting of many columnar rows on the east coast of the Tröllaskagi peninsula (Fig. 1). View N, the dyke strikes N25°E, dips 81°E, is 20 m thick and consists of at least 7 columnar rows (indicated by arrows). The person provides a scale.

wide swarm of tension fractures, normal faults, and volcanic fissures. The Krafla central volcano (Figure 10) is a typical rift-zone volcano: it erupts frequently, produces intermediate and acid in addition to basaltic rocks, and includes a collapse caldera with an E-W diameter of 10 km and a N-S diameter of 8 km. A rift-zone volcanic system in Iceland is normally active for 0.5-1 Ma; the Krafla volcanic system has been active for at least 0.2 Ma (Saemundsson 1991).

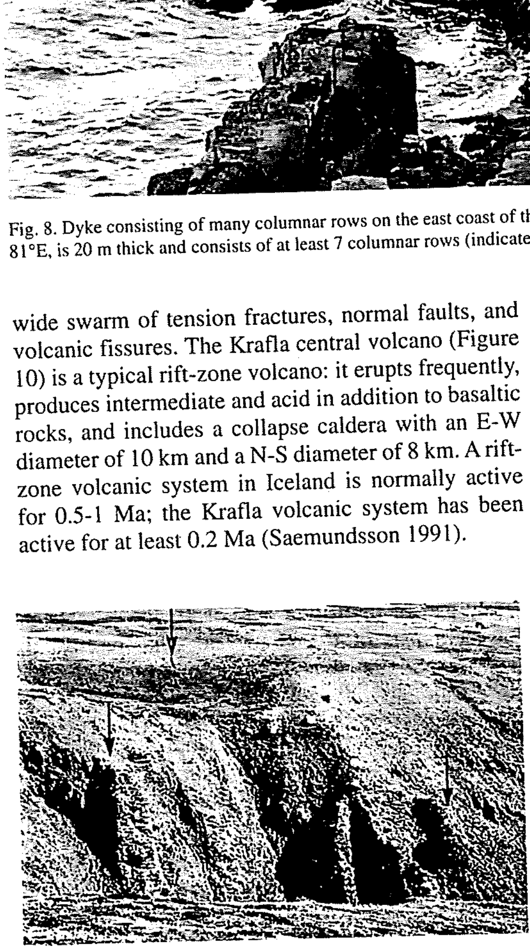

Fig. 9. Multiple dyke, consisting of three separate dykes, dissecting basaltic lava flows near the town Dalvik on the Tröllaskagi peninsula (Fig. 1). View N, the approximate attitude of the dyke is N16°E, 80°E. Its total thickness is 54 m, making it the second thickest basaltic dyke in Iceland. The arrows indicate the outer margins of the multiple dyke and a person that provides a scale.

During the Holocene there have been many rifting episodes in the Krafla volcanic system (Saemundsson 1991). In the period 10,000-8000 B.P. eruptions were frequent, leading to the formation of basaltic lava flows and lava shields. During the next 5000 years, up to 3000 B.P., there was much faulting and rifting activity, but only one eruption. During the past 3000 years there have been six rifting episodes with volcanic activity, the last two being the Myvatn fires 1724-1729 and the Krafla fires 1975-1984.

The Krafla fires began in the morning of 20 December 1975 with a very intense earthquake swarm and, 15 minutes later, a small basaltic eruption near the central (Leirhnjukur) part of the Krafla caldera (Björnsson et al. 1977, Einarsson 1991). Initially, the earthquakes were mostly located within the Krafla caldera, but in the next two hours they migrated by at least 40 km to the north along the Krafla fissure swarm. Subsequently, the earthquakes continued for nearly three months in the Krafla caldera and in the Kelduhverfi-Axarfjördur area: these areas are about 40 km apart (Figure 10). The magnitudes of the largest earthquakes continued to increase until mid-January 1976; in the Krafla caldera the largest earthquakes, of magnitude 5, occurred on January 6 and 19, whereas in the Axarfjördur area the largest earthquake, of magnitude 6.4, occurred at depth of 12 km on the dextral strike-slip Grimsey fault on 13 January.

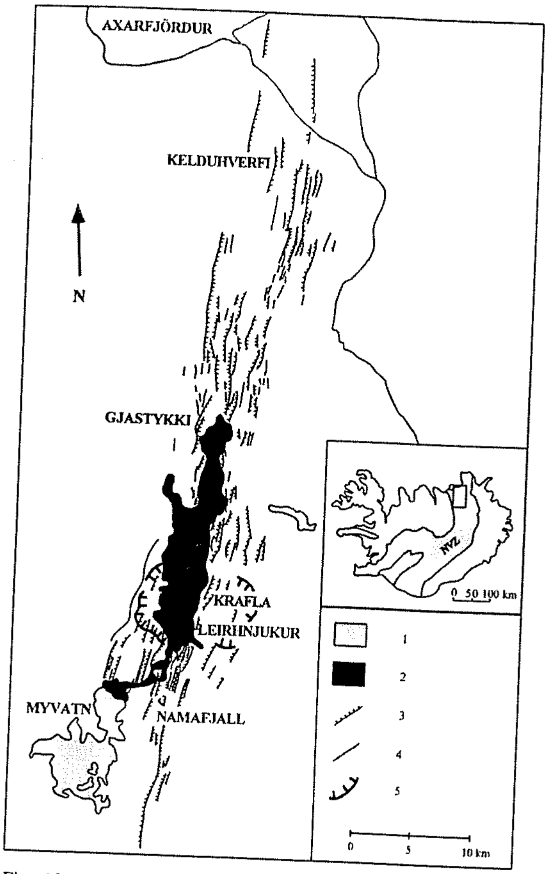

Fig. 10. Schematic illustration of the Krafla volcanic system, including the Krafla caldera. Leirhnjukur in the Krafla caldera was the centre of inflation/deflation events during magma-injections from the Krafla magma chamber. 1: Lake; 2: basaltic lava flows from the Krafla fires 1975-1984 and the Myvatn fires 1724-1729; 3: normal fault; 4: tension fracture; 5: ring-fault of the Krafla caldera. On the inset map of Iceland, the neovolcanic zone (NVZ) is indicated. Modified from maps by Opheim & Gudmundsson (1989) and Saemundsson (1991).

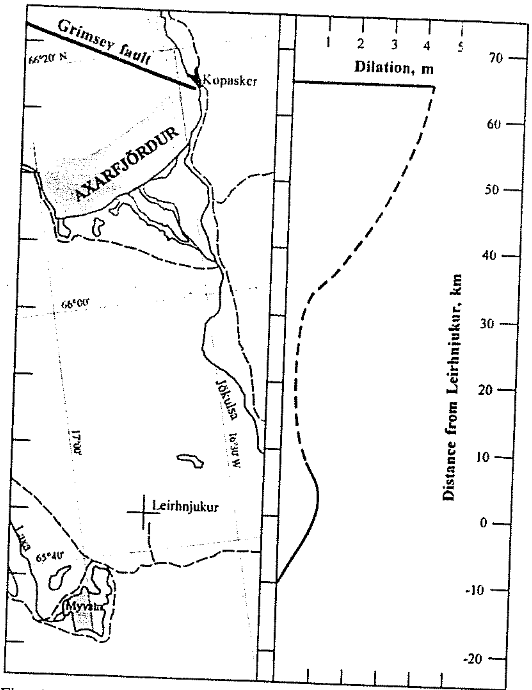

Fig. 11. Dilation of the Krafla volcanic system during the first rifting event. The left half of the figure is a location map (the thin, broken line being the main road), the right half shows the measured (solid line) or estimated (broken line) crustal dilation. Lake Myvatn and the sea are shaded. Modified from Tryggvason (1983, 1984).

3.2 Dyke growth

In the first rifting event, the average dilation associated with the feeder-dyke in the Krafla volcano was 1 m (Figure 11). South of Krafla, the crustal dilation gradually decreased to zero, but from Krafla north to Kelduhverfi, a distance of nearly 30 km, there was hardly any dilation. In Kelduhverfi, significant crustal dilation began around 15 km south of the main road, reached 2-3 m on that road, 3-4 m at the coast of Axarfjördur (Tryggvason 1983, 1984), and probably increased somewhat out to the Grimsey fault (Figures 10 and 11). The subsidence of this same part of the Krafla fissure swarm was zero 15 km south of the main road, 1 m on the road, 2 m at the coast of Axarfjördur and probably continued as a couple of metres north to the Grimsey fault. The flanks of this same part of the fissure swarm were uplifted by a few tens of centimetres (Björnsson et al. 1977, Sigurdsson 1980).

The crustal dilation along the Krafla fissure swarm in individual rifting events was variable during the early part of the rifting episode (Figure 12). In the northern part of the swarm no dyke reached the surface and the dilation was at least partly non-magmatic and directly attributable to the plate-tectonic stress field. Inside, and in the vicinity of, the Krafla caldera, however, the measured crustal dilation is entirely attributable to the feeder-dyke. All the nine eruptions occurred within 7 km of the ring-fault of the Krafla caldera, and in this part the dilation was measured in greater detail than in other parts (Tryggvason 1983, 1984). While these measurements do not resolve the geometry of individual segments of the feeder-dyke at the surface, they indicate its general growth.

The growth of the feeder-dyke emplaced during the Krafla fires can be inferred from the crustal dilation in the 18-km-long profile measured in the vicinity of, and including, the Krafla caldera (Figure 13). The profile extends from the northernmost part of the

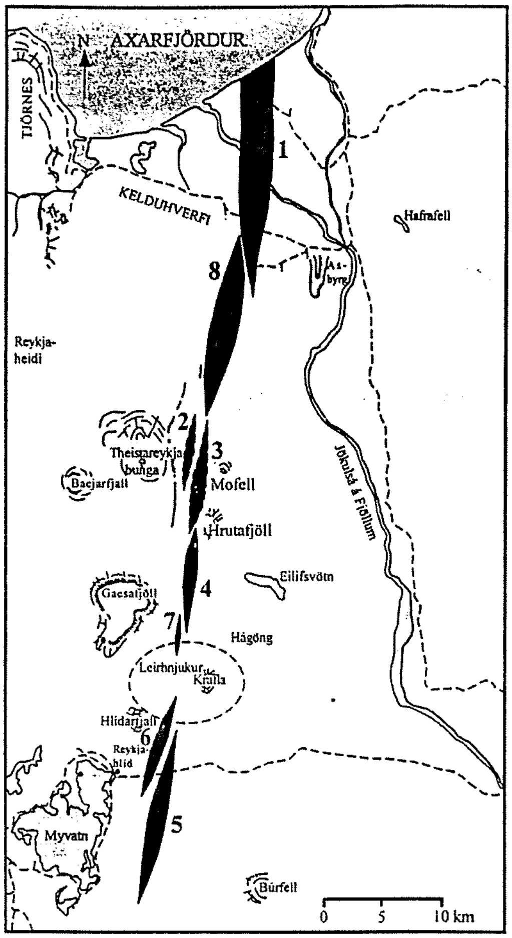

Fig. 12. Size and location of the areas (black) where the main surface fracturing was observed during 8 rifting events up to January 1978 in the Krafla volcanic system. The area associated with the earliest rifting event is number 1; that associated with the latest event is number 8. No fracturing was observed in two rifting events during this period. Modified from Sigurdsson (1977).

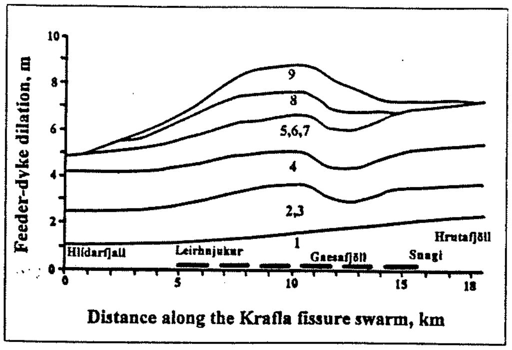

Fig. 13. Growth of the feeder-dyke in the 9 eruptions of the Krafla fires 1975-1984. The dilation profile is 18 km long and extends south and north of the Krafla caldera. Also shown schematically is the final length and location of the volcanic fissure (thick broken line between Leirhnjukur and Snagi). A general increase in the length of the fissures in the later eruptions suggests a gradual increase in the length of the feeder-dyke. The profiles measured after several of the early eruptions represent the cumulative crustal dilation associated with more than one magma injection (volcanic fissure), but the last two eruptions (8th and 9th) represent only single magma injections. The thickness increase during each magma injection was around 1 m. The final length of the feeder-dyke at the surface (the volcanic fissure) is 11 km and its maximum thickness is 9 m. (Based on unpublished data by Eysteinn Tryggvason).

volcanic fissure, at the mountain Hrutafjöll, to the mountain Hlidarfjall in the south (Figure 12). The total length of the discontinuous volcanic fissure that erupted during the rifting episode 1975-1984 is 11 km (Saemundsson 1991); the profile in Figure 13 extends beyond the southern end of the volcanic fissure (the feeder-dyke at the surface).

In the early eruptions, the volcanic fissures were short, but longer in subsequent eruptions; in the last two eruptions the volcanic fissures were 8-9 km long (Saemundsson 1991). The fissures that formed in the later eruptions commonly coincided with those that formed in the earlier eruptions (e.g., Einarsson 1991). Thus, the feeder-dyke gradually increased its length, as well as its thickness, during the rifting episode (Figure 13). Some of the early dilation profiles represent several eruptions, and are therefore not a very clear indication of the growth of the feeder-dyke during individual magma injections. The dilation profiles measured after the 8th and 9th eruptions, however, represent single magma injections. Both are elliptical displacement profiles, 12-15 km long, and are associated with feeder-dyke injections that at the surface reached lengths of 8-9 km. The maximum thickness of both feeder-dyke injections was just over 1 m, and the average thickness around 0.8 m.

None of the volcanic fissures (feeder-dykes) was continuous in a lateral section. Each eruption initiated on short segments, sometimes small vents (Figure 14), often far apart. Subsequently, the segments propagated laterally and approached each other, some eventually nearly or quite coalescing. During the initial stages of the September 1977 eruption a segment of the volcanic fissure increased its length at the rate of 0.3-0.4 m/s (Tryggvason 1977). Also, the volcanic fissure of the July 1980 eruption initiated from a small circular glow on the ground, and then changed into a several-metre-long fissure-segment from which lava issued (Figure 14). This segment increased its length (laterally) at the rate of 0.1-0.2 m/s. For comparison, the rate of lateral subsurface migration of earthquakes in several rifting events during the Krafla rifting episode was generally in the range of 0.4-1.2 m/s (Einarsson 1991).

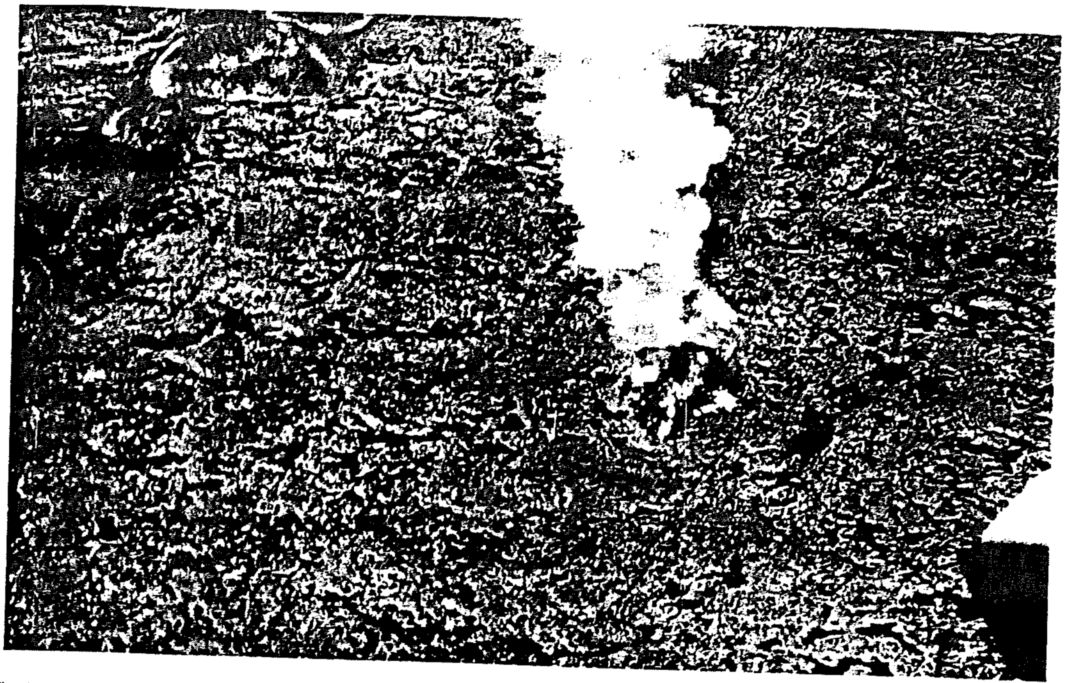

Fig. 14. Aerial view of the initiation of the July 1980 eruption of Krafla. No surface fractures were generated by the magmatic overpressure of the subsurface part of the dyke (the nearby faults are old). Photograph: Aevar Johannsson.

4 A MODEL OF DYKE GROWTH

4.1 Dyke injection

The general condition for the rupture of a magma chamber is:

$$p_L + p_e = \sigma_3 + T_0 \quad (1)$$

where p_L is the lithostatic pressure at the rupture, p_e is the magmatic excess pressure in the chamber, σ_3 is the minimum compressive principal stress (maximum tensile stress), and T_0 is the tensile strength of the rock hosting the magma chamber. When condition (1) is met, a magma pulse becomes injected from the chamber to form a dyke (or sheet). When magma flows out of the chamber, its volume and magmatic excess pressure decrease. When the magmatic excess pressure becomes zero, the dyke closes at its intersection with the surface of the magma chamber and the magma flow out of the chamber stops.

If a dyke or sheet grows as a statistically self-similar (self-affine) fracture, then its thickness is related to its volume; thin self-affine dykes in a particular swarm are normally of less volume than thick dykes in that swarm. I propose that the most common volume of the non-feeder dykes or sheets in a swarm (as indicated by the mode of the thickness size distribution) is reached in a single magma injection from the source chamber or reservoir.

The most common thickness of dykes in a typical sheet swarm is several tens of centimetres, but in a regional dyke swarm it is one or two metres (Figure 7; cf Gudmundsson 1983, 1984a, 1990). For self-similar dykes, thickness difference of, say, a factor 4 would imply a volume difference of 64. Many regional dykes are presumably injected from magma reservoirs of volumes up to one hundred times that of the crustal source chambers of the sheet swarms (Gudmundsson 1990). The author's field studies, and the results from Krafla (Tryggvason 1984, Saemundsson 1991), indicate that during a particular rifting episode, many small-volume sheets may be emplaced, each consisting of one or a few magma pulses, but only one regional dyke, consisting of many magma pulses.

4.2 Thermal aspects of dyke growth

The model presented here makes the following predictions: Most dykes in a particular swarm that are thicker than the mode thickness in that swarm are the result of many magma injections. When magma is injected into a dyke-fracture in a rapid succession of magma pulses, the resulting dyke consists of one to three columnar rows. If the time between successive magma injections is somewhat longer, the dyke is still single but composed of many columnar rows. When the time between magma injections is still longer, the resulting dyke is multiple. As the time between successive magma injections increases, there is thus a gradation from a single to a multiple dyke.

The time limits for the formation of a single dyke, a dyke with columnar rows, and a true multiple dyke can be estimated crudely as follows. The most common dyke thickness in swarms (Figure 7) suggests that a single magma injection leads to the emplacement of a dyke of the order of 1 m thick. For such a thin dyke, the conduction formulas of Jaeger (1964, 1968) yield valid order-of-magnitude results for the cooling of the dyke. Jaeger (1968) concluded that the time t_s for complete solidification of a dyke is:

$$t_s = w^2/(4\kappa\lambda^2) \quad (2)$$

where w is the half-thickness of the dyke, κ is the thermal diffusivity of the solidified magma (assumed to be equal to that of the host rock), and λ is the root of the equation:

$$\lambda(1 + \operatorname{erf}\lambda)e^{\lambda^2} = cT_m/(L\pi^{1/2}) \quad (3)$$

where c is the specific heat capacity of the host rock, T_m is the melting-point temperature of the magma, and L is the latent heat of solidification.

Columnar joints start to form at above 60% of the initial temperature of the magma (Jaeger 1961). The temperature of basaltic tholeiite magma is in the range of 1150-1225°C (Williams & McBirney 1979), suggesting that columnar rows in a basaltic dyke start to form at around 800°C. For a typical basaltic dyke, the range of solidification until columnar joints start to form is thus 1200-800°C. For this range and using $L = 0.42$ MJ/kg (Jaeger 1968, Fukuyama, 1985), the value of λ is 0.108.

The thermal diffusivity κ is obtained from the formula (Carslaw & Jaeger 1959):

$$\kappa = K/(\rho_r c) \quad (4)$$

where K is the thermal conductivity and ρ_r is the density of the host rock. The average thermal conductivity of basalt is 2.1 W m^{-1} K^{-1} (Oxburgh 1980), the specific heat capacity is 835 J kg^{-1} (Jumikis 1979), and the density of the uppermost 1.5 km of the Icelandic crust, where most of the dykes (including the Krafla feeder-dyke) are observed, is 2500 kg m^{-3} (Gudmundsson 1992). Substituting these values in eqn. (4) yields $\kappa = 1\cdot10^{-6}$ m^{-2} s^{-1}. Inserting this value for κ and 0.108 for λ in eqn. (2), the approximate time for the initiation of columnar joints in a dyke becomes $t_s \cong 2.14\cdot10^7 s\cdot w^2$, or in years ($t_y$):

$$t_y \cong 0.680 \cdot w^2 \quad (5)$$

Gudmundsson (1984b) suggested that between successive magma injections (pulses) the outer parts of the dyke cool and form columnar joints, but that the next injection splits the previous one in two, using its central part where the dyke material may be partly molten and thus with a reduced tensile strength (Shaw 1980). If the magma of the previous injection had cooled down to the temperature of the host rock before the next injection, the tensile strength of the central part of the previous injection would not be less that that of the host rock and, therefore, no tendency for the new injection to follow the central part of the previous one.

The columnar rows in the dykes in Iceland are commonly 0.5 to 2 m thick (Figure 8). If they split in two by the next injection, the corresponding magma injection is two times as thick, or 1 to 4 m (Gudmundsson 1984b). From eqn. (5) it follows that columnar joints start to develop in a 0.5 m thick row (1 m thick injection) after 0.17 yrs (2 months) and in a 1 m thick row after 0.68 yrs (8 months). The increases in thickness (commonly just over one metre) during the growth of the Krafla feeder-dyke (Figure 13) indicate that the resulting columnar rows started to develop after several months. For comparison, the time before the initiation of columnar joints in a 2 m thick row (4 m thick injection) is nearly 3 years.

These approximate results indicate that thick coarse-grained regional dykes with a single or a few columnar rows form when the time between successive magma injections is of the order of months or less, in which case the magma cannot solidify completely between successive injections. Then the parameter w in eqn. (2) is the half-thickness of the final dyke. Solidification of a thick dyke therefore takes a long time, which may partly explain its large grain size. For a dyke consisting of many columnar rows, the time between successive magma injections is at least several months and may be several years. The results of Jaeger (1964) indicate that, for basaltic magma with solidus temperature of 1000°C (Williams & McBirney 1979), some remelting may take place at the contacts between columnar rows unless the first magma injection cools down to temperatures of around 600°C before the next one is injected. The time between successive magma injections in a multiple dyke is at least many years and may be hundreds or thousands of years.

4.3 Segmentation during dyke growth

The results from the Krafla fires show that the feeder-dyke does not propagate as a single sheet but rather as a set of discontinuous segments, each with a

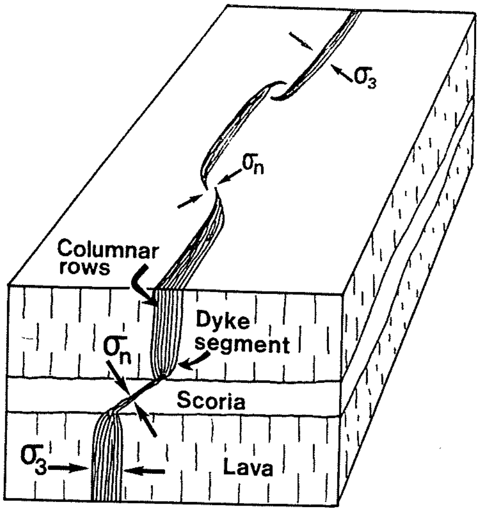

Fig. 15. Segmentation of a dyke in vertical and horizontal sections is partly attributable to the dyke using pathways that are subject to a normal compressive stress (σ_n) that is higher than the minimum principal compressive stress (σ_3). In this schematic illustration the segmentation in the vertical section is largely attributable to the competence contrast between the basaltic lava flows (here subhorizontal) and the scoria layer and the associated stress changes at their contacts. The columnar rows in the dyke are indicated.

propagation front. Similar results have been obtained from other fissure eruptions (e.g., Thordarson & Self 1993) and dyke studies in general (e.g., Gudmundsson 1983, 1984b). At any instant, the tip of a vertically propagating dyke reaches shallowest crustal depths at one or several points; these parts of the dyke are the first to intersect the surface during an eruption (Figure 14). If the vertical propagation is to continue (at any particular crustal level) from these points there must be simultaneous lateral propagation. As the dyke-segments propagate vertically and laterally, some eventually coalesce, others continue as individual segments.

Part of the segmentation of dykes in horizontal and vertical sections is because dyke parts that are not normal to the minimum compressive stress tend to be thin or nearly closed (Figure 15). The magmatic overpressure in a dyke-segment is the total magmatic pressure minus the tectonic compressive stress normal to that segment (Gudmundsson 1990). The normal stress σ_n on a dyke-segment is (cf. Gudmundsson 1984b, Delaney et al. 1986):

$$\sigma_n = \tfrac{1}{2}(\sigma_1 + \sigma_3) - \tfrac{1}{2}(\sigma_1 - \sigma_3)\cos 2\alpha \qquad (6)$$

where σ_1 and σ_3 are the maximum and minimum principal compressive stresses, respectively, and α is the angle between the dyke-segment and the direction of the maximum principal stress (assumed to be vertical). Normally during dyke injection there is a difference between the principal stresses (the differential stress is not zero), so that, from eqn. (6), the normal stress is larger than the minimum principal stress. Consequently, a dyke-segment that is oblique to the minimum principal stress has less overpressure than segments that are normal (perpendicular) to that stress. It follows that, other things being equal, the oblique segments will be narrower than the nearby segments (normal to the minimum principal stress) of the same dyke.

One reason why dyke-segments are commonly oblique to the minimum compressive stress is that most dykes follow the pathway of least resistance and propagate along planes where the term $\sigma_3 + T_0$ in eqn. (1) is minimum (Gudmundsson 1984b). In Iceland there is a general tilting of the lava pile at the rate of 1° for every 150 m depth in the crust. Thus, the columnar joints (which are perpendicular to the contacts of the lava flows) gradually become tilted with increasing depth in the lava pile and therefore oblique to the (horizontal) orientation of the minimum compressive stress (Gudmundsson 1992). Field observations show that dykes commonly use these joints as pathways (e.g., Gudmundsson 1984a,b) so that that many dyke-segments are oblique to the minimum compressive stress. Thus, the dykes become segmented, with thin or veinlike oblique segments alternating with thick segments that are perpendicular to the minimum compressive stress.

Commonly, a dyke-segment becomes thin where it dissects layers of competent rocks that are subject to higher compressive stresses than the surrounding layers. This happens, for example, where dykes cut through thick, competent basaltic lava flows that alternate with less competent lava flows or sedimentary layers.

Segmentation is a universal feature of dykes (e.g., Delaney & Pollard 1981, Rickwood 1990, Hoek 1991). Commonly, the segmentation is partly attributable to the pre-existing cracks in the crust (such as the columnar joints in Iceland) that are oblique to the minimum compressive stress but, nevertheless, used by the dykes as pathways (e.g., Baer & Beyth 1990).

5 DISCUSSION

If a dyke with dimensions similar to those of a typical regional dyke in Iceland is emplaced in an instant at a shallow depth below the free surface of the earth, some elastic crack models indicate that the dyke generates a graben at the surface above it (e.g., Pollard & Holzhausen 1979, Rubin & Pollard 1988).

Nevertheless, field observations show that even a feeder-dyke does not necessarily generate a graben (Figure 14). Because partial or complete relaxation of the dyke-induced stresses can occur during the dyke's growth (e.g., between successive magma injections), crack models based on instantaneous emplacement of dykes may overestimate their surface effects in general and their potential of generating grabens in particular.

An experimental model that partly simulates the surface effects of the dyke-growth model presented here was provided by Mastin & Pollard (1988). The results indicate that, to generate surface fractures, the top of a dyke with a blunt end must be at a depth less than about ten times the top thickness; the surface fractures start to develop into normal faults when this depth is less than about five times the top thickness. A one-metre thick magma-injection in Krafla (Figure 13) would have to be within 10 m of the surface to generate tension fractures, and within 5 m to generate a graben. Because the width of the graben is equal to the depth to the top of the dyke that generates it (Mastin & Pollard 1988), the resulting "graben" would be 5 m wide. Because most dykes do not have blunt ends, these depths would have to be even shallower and the dykes would normally be feeders. Furthermore, if a vertically propagating dyke enters a graben, the magmatic overpressure associated with the dyke would tend to lock the graben faults adjacent to the dyke.

Although the average dyke thickness in regional swarms is 2 to 6 m, and some dykes are tens of metres thick, on approaching their vertical ends most dykes in Iceland are thin (Figure 6). The author's field observations show that the radius of curvature of a typical dyke tip in Iceland is of order of centimetres or less. The dyke-growth model indicates that, normally, only one magma injection (columnar row) reaches the shallowest level in the crust; this row is commonly 0.5-2 m thick. This may partly explain why dykes are thin immediately below their vertical ends. The tips of the dykes are very thin, however, because dykes become arrested on entering crustal layers where the dyke-normal compressive stress exceeds the magma pressure in the dyke.

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