Journal of Structural Geology, Vol. 13, No. 4, pp. 443 to 457, 1991
Printed in Great Britain

0191-8141/91 $03.00+0.00

Crustal spreading due to dikes and faults in southwest Iceland

TROND FORSLUND and AGUST GUDMUNDSSON

Nordic Volcanological Institute, University of Iceland, IS-101 Reykjavik, Iceland

(*Received* 20 *March* 1990; *accepted in revised form* 7 *September* 1990)

Abstract—A 30 km long profile, extending from 3 million year-old Pleistocene rocks to the Holocene rocks of the rift zone in southwest Iceland, is dissected by 443 dikes and 156 normal faults. Nearly all the dikes are basaltic, with an average strike of N26°E, which is subparallel to the trend of fractures in the nearby rift zone. Their average dip is 69° and average thickness 1.37 m, both values much lower than averages for regional Tertiary dike swarms in Iceland. Many dikes are seen to change into sills. The great majority of the faults are pure dip-slip, with an average throw of 10.2 m. Their average strike, N37°E, is different from the average strike of the dikes in this area and the fractures in the nearby rift zone. Their average dip is 75.2°, which is somewhat steeper than in the Tertiary areas, but the dips are less in the interbedded sedimentary layers than in the lava flows. Growth faults are very rare in the area and most of the faults appear to have completed their growth in less than 10,000 years. In subsections, crustal dilation due to dikes is 0.4–3.7%, that due to faults is 0.6–5.3%, and that due to both dikes and faults is 1.04–6.44%. In these subsections, average fault spacing is one fault per 170–563 m. Spacing decreases toward the active rift zone. It is concluded that in typical fissure swarms of the rift zone, the dilation at the surface is almost entirely due to extensional fractures and gaping vertical normal faults. In the depth range of a couple of hundred metres to perhaps 1 km, the non-vertical faults contribute most to the dilation. At deeper crustal levels, however, dilation due to dikes becomes dominant.

INTRODUCTION

CRUSTAL spreading at divergent plate boundaries has frequently been attributed to dikes. In the first sea-floor spreading model of Iceland, for example, the rifting of the crust was considered to be the result of dike injection into a static volcanic zone (Bodvarsson & Walker 1964). This model was later modified by Gibson & Piper (1972) and Daignieres *et al.* (1975), and quantified by Palmason (1973, 1980, 1986). All these models predict that the number of dikes increases rapidly with depth in the crust. For example, Palmason (1973) suggested a rate of increase of dike fraction by volume of 15% per km.

The contribution of normal faults to crustal spreading at divergent plate boundaries has received less attention. In most models of crustal accretion in Iceland the contribution of normal faults to spreading is assumed to be minor. For instance, Bodvarsson & Walker (1964) concluded that the extension due to faults was only some 10% of that due to dikes. Daignieres *et al.* (1975) attributed 90% of the extension to feeder-dikes (emissive fissures) and gaping fissures, but only 10% to normal faults. Similarly, Palmason (1973) did not consider normal faults in the early version of his model, and in a later version (Palmason 1980) normal faults contribute much less than dikes to crustal spreading.

In order to provide accurate data on crustal extension due to dike emplacement and faulting, we have measured several hundred dikes and faults in a profile in southwest Iceland. This profile is about 30 km long, extends from the oldest Quaternary rocks (about 3 Ma, Kristjansson *et al.* 1980) to the Holocene lavas of the active rift zone at Thingvellir (Fig. 1), and is here referred to as the Eyrarfjall–Karastadir profile (Fig. 2). In addition to obtaining geometric data, the purpose of this study was to observe if the faults were growth faults, and to infer the infrastructure of a typical fissure swarm, exemplified by the Thingvellir fissure swarm north of Lake Thingvallavatn (Fig. 2).

GEOLOGIC SETTING

The surface of Iceland is almost entirely made of volcanic rocks, with basalts being 80–85% of the volcanic pile, and acid and intermediate rocks 10%. Sedimentary rocks and volcanic breccias are 5–10% of the Tertiary lava pile but locally may attain 50% of the Quaternary lava pile (Saemundsson 1986).

The neovolcanic zones of Iceland are defined by rocks younger than 0.7 Ma (that is rocks belonging to the Brunhes magnetic epoch) and by seismic activity. They correspond to the magnetic plate boundary (anomaly 1) at the mid-ocean ridges and consist of the axial rift zone and two or three off-rift flank zones (Fig. 1). The rift zone marks the divergent plate boundary in Iceland. It is characterized by well-developed extensional structures such as tension fractures, normal faults and grabens. The flank zones, by contrast, lack such well-developed extensional structural elements.

Our research area is located at the western margin of the rift zone in southwest Iceland (Fig. 1). Apart from a study of joints (Jefferis & Voight 1981) and minor faults (Bergerat *et al.* 1990), no detailed tectonic studies have been made in this area before. Such studies have, however, been carried out in the adjacent Thingvellir fissure swarm (Gudmundsson 1987a) as well as on the Reykjanes Peninsula (Gudmundsson 1987b) to the south of the research area. General geological mapping has been made in parts of the research area (Rutten

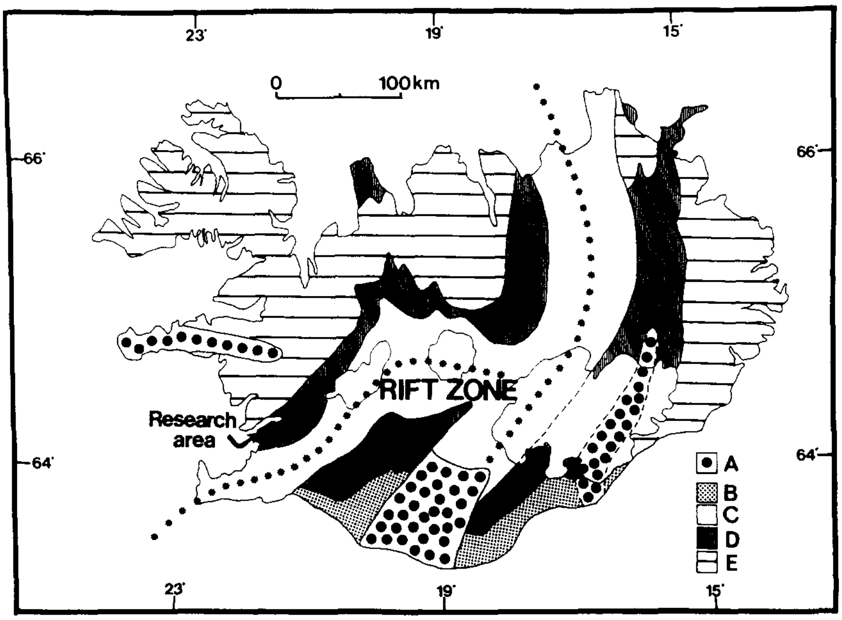

Fig. 1. Simplified geologic map of Iceland. The research area is shown in black. A, off-rift flank zones; B, outwash and lava apron; C, Upper Pleistocene–Holocene rocks, i.e. rocks younger than 0.7 Ma in age; D, Lower Pleistocene rocks, 0.7–3.1 Ma in age; E, Tertiary lava pile, 3.1–16 Ma in age. The trace of the axis of the rift zone is indicated by stars. Based on maps by Saemundsson (1979).

1958) and in nearby areas (Fridleifsson 1973, Franzson 1978). Several works have focused on paleomagnetism (Einarsson 1962, Kristjansson *et al.* 1980). General mapping of the stratigraphy of parts of our profile is to be found in unpublished dissertations (Gunnlaugsson *et al.* 1972, Hjartarson *et al.* 1973, Jonasson *et al.* 1973, Helgason 1976, Snorrason 1976). A general geological map of southwest Iceland at the scale 1:250,000 has been compiled. It includes our profile (Saemundsson & Einarsson 1980).

The rocks along the profile are mostly basaltic lava flows, with interbedded layers of basaltic breccias and tillites. Rhyolite and microgabbro occur at a few places. The proportion of basaltic breccias (hyaloclastites) gradually increases eastward along the profile toward the rift zone. The maximum relief is about 500 m. The distribution of amygdales indicates that the coastal areas of the main fjord, Hvalfjördur (Fig. 2), are as much as 1100 m below the original surface of the lava pile (Jonasson *et al.* 1973).

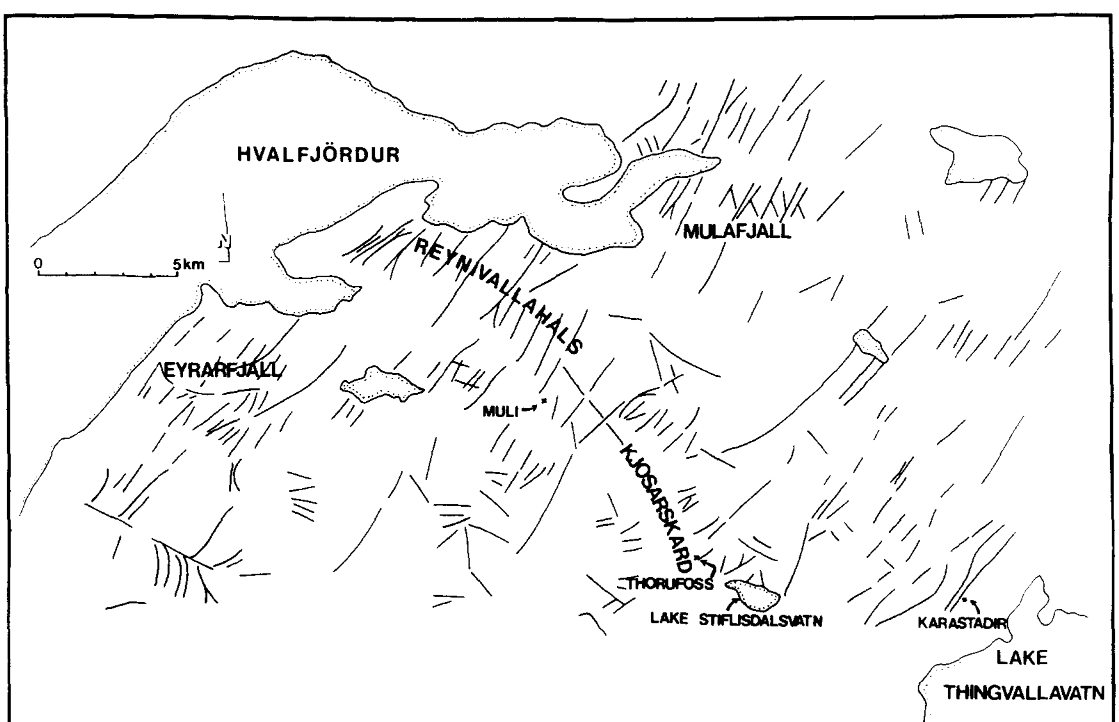

Fig. 2. Lineament map of the research area.

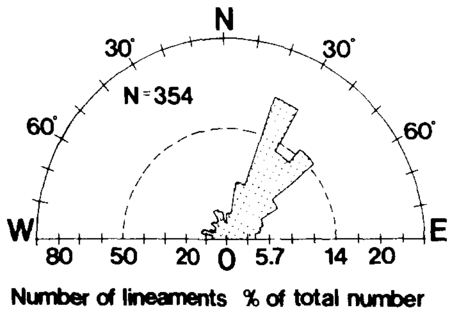

Fig. 3. Rose diagram for lineaments. The average (arithmetic mean) strike is N21°E.

LINEAMENTS

The lineaments were studied on aerial photographs at the scale of 1:34,000. The strike was taken as the azimuth of the line joining the endpoints of each lineament. On a regional scale, the most prominent lineaments have NE–SW and WNW–ESE trends, the latter being parallel to many of the main valleys in the area (Fig. 2). Lineament interpretation using aerial photographs reveals two main lineament directions, N25°E and N45°E (Fig. 3), coinciding with the two most prominent fault directions (see below). Many lineaments are discontinuous and most of them can be traced for only 1–3 km. Field observations indicate that many of the lineaments are normal faults, but some have been identified as dikes and others may be joints. Most of the dikes, however, are too thin to be recognized on the aerial photographs. Some clear-cut lineaments show no evidence of being either dikes or faults and we interpret these as joints. The average strike of all lineaments, N21°E, does not represent the dominant strike very well, but is close to the average strike of all dikes measured in the area.

DIKES

Careful field observations were made of 443 dikes, nearly all of which are basaltic. The great majority are tholeiitic, but some are of olivine–tholeiite and others of an intermediate composition. Two, both about 10 m thick, are composite with rhyolite in the centre. Apart from the dikes, many inclined sheets and sills were observed. For convenience these are classified with the dikes. The longest continuous dike visible on the aerial photographs can be traced for 1.5 km, while the thickest dike measured in the field is 25 m and is of coarse-grained dolerite.

Many dikes display circular or elongate vesicles, mostly a few millimetres in diameter. Amygdales are common in dikes at all elevations and include zeolites, chalcedony, opal, jasper and calcite. Rapid cooling of the magma in the dike is evidenced by chilled margins on some of the dikes. The majority of the dikes do not, however, have glassy margins, although the dike-rock is normally finer grained near the margins.

Only one dike has xenoliths of the host rock. This dike is 0.6 m thick but all the xenoliths, which are as much as 10 cm in diameter, occur at the margins of the dike.

Strike

The strike distribution of dikes in the area (Fig. 4) shows a northeasterly trend with a mean strike of N26°E. Figure 5 shows the strike distribution in subareas A–G for both dikes and faults. The dikes in subarea C show an average strike of N10°E (Table 1) which, at both the 0.05 and 0.01 level of significance, is statistically different from the strikes of the dikes in subareas A and B and from dikes measured in the coastal area north of Eyrarfjall. Dikes in subareas D and E show a mean strike of N28°E and N29°E, respectively, which, at a 0.05 level of significance (but not at the 0.01 level), is statistically different from the strike of dikes in subarea C. Subarea F shows two dominant strike directions, WNW–ESE and ENE–WSW, with a mean strike of N9°E. In two cases, cross-cutting relations indicate that the WNW–ESE-trending dikes are younger than the ENE–WSW-trending dikes.

Dip

The dip distribution of dikes in the research area is shown in Fig. 6. The average dip is 69°, and only 33% of the dikes dip within 10° of the vertical, which is a much lower percentage than in the Tertiary dike swarms in east and northwest Iceland (Gudmundsson 1983, 1984a,b). Because 43% of the dikes dip to the east, and because the lavas dip only very gently eastward (2–10°), the deviation of dike dip from vertical cannot be explained solely by subsequent tilting of the lavas. The average dip in each subarea (Table 1) varies significantly but, again, subarea C is distinctive, with an average dike dip of only 50°. There is a statistical difference between the mean dip in subarea C and the mean dips in all other subareas, both at the 0.05 and the 0.01 levels of significance.

The change in dip of the dikes upward in the lava pile was tested by comparing the dip of dikes east of Muli (Fig. 2) with the dip of dikes along the coast north of Eyrarfjall. The difference in altitude between these two sections is about 450 m and the difference in mean dip is about 8°, the dips of the coastal dikes being higher. However, at the 0.05 level of significance there is no

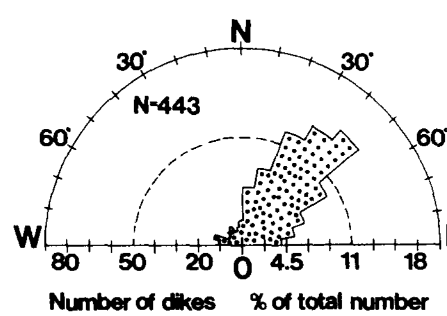

Fig. 4. Rose diagram for dikes. The average strike is N26°E.

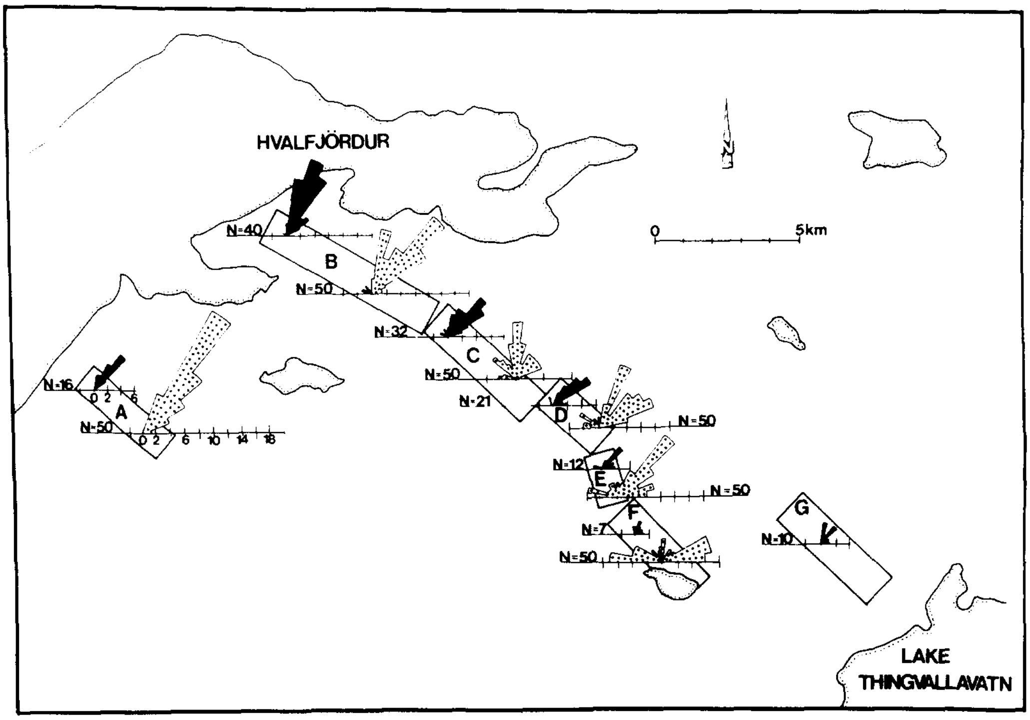

Fig. 5. Rose diagram for dikes (dotted) and faults (black) in subareas A–G. Each subarea is chosen so as to include 50 dikes, but the number of faults in each area varies. The horizontal scale (shown calibrated in subarea A) refers to the number of dikes and faults.

statistically significant difference between the mean dip of dikes in these two sections.

Thickness (width)

Figure 7 shows the thickness distribution of all dikes in the research area. The average thickness is 1.37 m, varying from 1.78 m for the dikes at the coast north of Eyrarfjall to 0.86 m in subarea E (Table 1). Although the average thickness is about 1.4 m, 57% of all dikes are less than 1 m thick. A prominent decrease in dike thickness occurs eastward in the profile, in particular between subarea B (1.52 m) and subarea C (0.92 m). There is a statistically significant difference in the mean dike thickness in these two subareas, at both 0.05 and 0.01 levels.

There is no statistical correlation between thickness of the steeply dipping (>70°) dikes and their strike, although dikes trending both N–S and E–W tend to be thinner than NE–SW-trending dikes. There is a tendency for shallowly dipping dikes to be thinner than

Table 1. Average strikes and dips of dikes and faults, and the average dike thickness, in the coastal area and in subareas A–G

	Average strike (°) dikes/faults	Average dip (°) dikes/faults	Average thickness (m) dikes
Coastal area	27/25	75/74	1.79
Subarea A	42/39	80/77	1.66
Subarea B	37/32	80/74	1.52
Subarea C	10/41	50/74	0.92
Subarea D	28/49	62/81	1.08
Subarea E	29/52	57/73	0.86
Subarea F	9/	69/77	1.17
Subarea G	/40	/76	

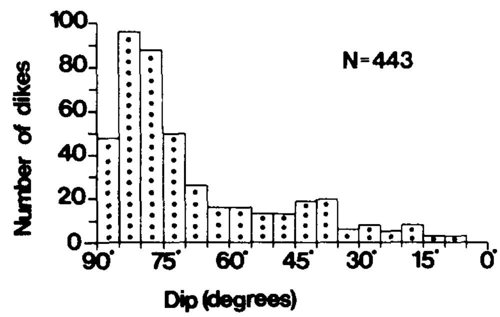

Fig. 6. Dip distribution of dikes. The average dip is 69°, the minimum is 7° and the maximum is 90°.

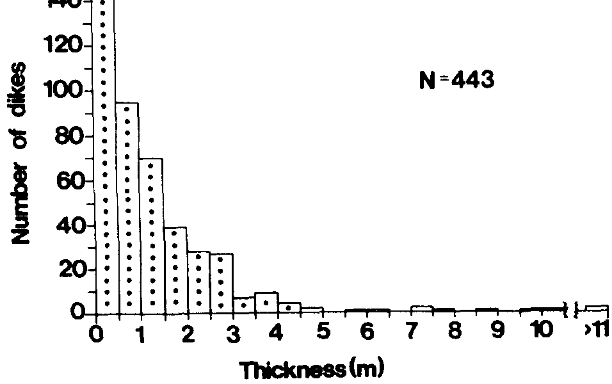

Fig. 7. Thickness distribution of dikes. The average thickness is 1.37 m, the maximum is 25 m and the minimum is 0.02 m.

steeply dipping dikes, but this difference is not significant in all the subareas.

Geometry

In a vertical section, many of the dikes are nearly straight, with parallel edges (Fig. 8a), but some dikes, although continuous, occur as slightly offset segments. In lateral sections, the dikes are commonly offset, and some are apparently discontinuous. Neither in vertical nor in lateral sections do the dike segments show preferred offset, that is the dike segments do not have an en échelon arrangement.

Where the dikes are offset they sometimes develop horns (Fig. 8b). Such horns may be 'blind' pathways, where the dike tried a certain pathway that turned out to be less favourable than the one finally selected. Alternatively, horns may develop where the segments combine in the same way as Mode I tension cracks grow (Gudmundsson 1987b). Some dikes contain up to 10 columnar layers, which suggests that they are 'multiple' in the sense that they were generated in a rapid succession of magma pulses (Gudmundsson 1984c).

Many dikes were seen to end vertically, but not a single one was seen to connect to a lava flow. This suggests that many of the dikes, perhaps the majority, were non-feeders, in agreement with results obtained elsewhere in Iceland (Gudmundsson 1984c). Several dikes, however, were feeders for sills, and low-dipping sills are much more common in this area than in the Tertiary areas of Iceland. The common occurrence of sills can be partly explained by temporary stress barriers (such that the horizontal compressive stress is higher than the vertical stress) that developed much more easily in the multilayered Quarternary lava pile than in the more homogeneous Tertiary lava pile (Gudmundsson 1986, 1990).

FAULTS

All 156 studied faults in the field are normal faults (Fig. 9a), and the great majority are pure dip-slip. Only in two cases did we observe a strike-slip component. Most of the faults are steeply dipping, but the dips tend to be less steep in the interbedded sedimentary layers than in the lava flows (Fig. 9b).

Strike and dip

The majority of the faults strike N20°–50°E, with an average of N37°E (Fig. 10). This differs significantly from the mean strike of dikes measured in the area and the average trend, N29°E, of fractures in the Thingvellir fissure swarm (Gudmundsson 1987a) at both the 0.05 and the 0.01 levels of significance.

Fault dips range from 53° to 89°, with an average of 75.2° (Fig. 11). This is a somewhat higher figure than for normal faults in the areas of Tertiary rocks where the average dip is 69° (Gudmundsson 1984a). Where the faults dissect lavas, fault planes are normally steeply dipping, but they show lower dips where they dissect sedimentary layers (Fig. 9b). In a vertical section, many of the faults are composed of segments that have different dips. We attribute this segmentation to the growth of faults by coalescence of tensile (Mode I) crack arrays, as demonstrated by experiments (Peng & Johnson 1972) and field observations (Gudmundsson 1987a) and supported by theoretical considerations (Scholz 1989).

About 51% of the faults dip to the west, that is away from the axial rift zone. The largest difference, 20°, in mean strike is between subareas B and E (Fig. 5), the mean strike being N32°E in subarea B and N52°E in subarea E. The faults clearly become more easterly-trending eastward along the profile (Table 1). Prominent fault scarps are easily observed on the aerial photographs. Few of the normal faults are occupied by dikes, and there is no evidence that any fault was formed by a propagating dike. In spite of nearby suitably orientated faults, dikes tend to generate their own fractures, nearly all of which are purely extensional.

Throw

The maximum estimated throw on a single fault is 200 m, but the maximum measured throw is 150 m (Fig. 12). The average throw is 10.2 m. Fifty-eight per cent of the faults have throws of less than 5 m and 90% have throws of less than 20 m. Some faults were followed upward in the lava pile to see if any noticeable change in throw occurred. Over an altitude difference of up to 150 m no difference in throw was observed except for one fault. In that case, however, the throw decreased upward by 6 m over a vertical distance of only 15 m, thus indicating a growth fault (Fig. 9a).

Breccia

Rocks along many of the faults have been severely fractured, brecciated and altered. The fault-rock products observed include fault breccias and variably developed cataclasites. The most typical fault breccia consists of angular and slightly rotated fragments (up to 10 cm in diameter) of the host rock set in a very fine- to medium-grained grey-brown matrix. A brief thin section inspection reveals a typical cataclastic microtexture with angular to subrounded mineral fragments in a dense- to fine-grained matrix of fragmented and granulated rock material, zeolites and calcite. Late, pervasive, extensional veins, showing no preferred orientation and commonly filled with calcite and zeolites, seem to represent the latest deformational episodes in the fault breccias.

Geometry

Striations on slickensided surfaces are rare, but where observed they are normally down-dip (80–90°), indicating dominantly dip-slip movements. At two localities, however, striae, pitching 5–35°W and 19–23°N, respectively, were observed on E–W- and N–S-trending fault planes, possibly indicating a strike-slip component of motion along E–W and N–S structures. Subvertical extensional veins, mainly filled with zeolites, quartz and calcite, strike N30–40°E, subparallel to dikes and faults in the area, but several ENE–WSW-striking veins were also observed.

Field evidence suggests that extension in the fault zones is mostly accommodated by sets of parallel synthetic normal faults. Although several antithetic faults were observed, they are thought to play subordinate roles as adjustment faults.

The geometry of faulting in the area is consistent with dominant NW–SE extension involving slip on NE–SW-trending structures. Striations on a few fault planes suggest the possibility of strike-slip faulting on E–W and N–S structures, but this is clearly of minor deformational significance.

JOINTS

Excluding the nearly vertical columnar joints in the lava flows, joints and tensional fractures of tectonic origin were measured in one subarea (subarea C). The strike distribution is characterized by four modes (Fig. 13), with an average strike of N44°E. The most common and best represented mode is N50–70°E, but the N30–40°E, E–W and the N–S modes are also clear. Joints occur singly or in small clusters. Their spacing is highly variable, and they are subvertical to vertical. Many joints are open with widths of commonly less than 1 cm, but widths of tens of centimetres are attained in places. The joints are probably the downward continuation of shallow tension fractures. They are also parallel to dikes, faults and extensional veins in the area and they are all interpreted as extensional features, formed perpendicular to the axis of the minimum compressive (maximum tensile) stress.

LACK OF GROWTH FAULTS

The conclusion that growth faults are rare in the research area is in agreement with results from the Tertiary areas of Iceland. In a study of 68 normal faults in northwest Iceland, Gudmundsson (1984a) followed several faults upward in the pile to see if there was any noticeable change in throw, but none was found.

During the current rifting episode in the Krafla fissure swarm in northeast Iceland vertical displacement of as much as 3 m has occurred at a single point of measurement (Sigurdsson 1980). Measurement points within the main part of the Krafla fissure swarm have been displaced by a couple of metres over the period 1977–1986 (Tryggvason 1986). During the Askja-Sveinagja eruption of 1875, the strip of land north of the then-formed Nyjahraun lava subsided some 3–6 m (Thoroddsen 1958, Bäckström & Gudmundsson 1989). During an earthquake in 1789, the vertical displacement on the major faults of the Thingvellir fissure swarm in southwest Iceland was 1–2 m (K. Saemundsson personal communication 1990). Similarly, in his analysis of historic surface faulting, Bonilla (1979) found that slips on faults of all kinds are commonly 0.5–3.0 m during earthquakes, and Speed (1979) obtained similar results for normal faults in the Great Basin of the western United States.

Using the figure of 1 m as a typical average slip on normal faults during rifting episodes in Iceland, it follows that the number of slips on a particular fault is equal to the measured vertical displacement in metres. A rough estimate is that rifting episodes occur every 500 years in the Krafla fissure swarm (Opheim & Gudmundsson 1989). That part of the rift zone, however, consists of three or four partly overlapping en échelon fissure swarms, each of which takes up part of the tensile strain associated with spreading. The volcanic rift zone in southwest Iceland consists of only one swarm, the Thingvellir fissure swarm, so that rifting might be expected to be more frequent in that swarm. The volcanic zone in the southern part of Iceland is, however, split into two parts, and the eastern part contains three parallel volcanic systems. The frequency of rifting in southwest Iceland may thus be similar to that of the Krafla fissure swarm. In the absence of more accurate data, we therefore take the time between rifting events in the southwest rift zone as 500 years.

Available evidence indicates that a particular rifting episode may affect all faults in a fissure swarm. The current rifting episode in the Krafla fissure swarm has affected the whole swarm, even if the main ground deformation has been confined to a 1–2 km wide central zone (Bjornsson *et al.* 1979, Sigurdsson 1980, Tryggvason 1984, Opheim & Gudmundsson 1989). The maximum throw on some Holocene faults is several tens of metres, whereas the average throw is only a few metres (Gudmundsson 1987a,b, Bäckström & Gudmundsson 1989, Opheim & Gudmundsson 1989). This can partly be explained if the faults with small throws became active later in the Holocene than the faults with large throws.

Data from the Tertiary lava pile as a whole outside the central volcanoes (McDougall *et al.* 1976a,b, 1977, 1984, Saemundsson *et al.* 1980, Watkins & Walker 1977) suggest that, on average, one lava flow was erupted every 15,000 years. Data from the western part of our profile indicate that the average time between eruptions was 6000–12,000 years (Gunnlaugsson *et al.* 1972, Jonasson *et al.* 1973). For comparison, no lava has erupted within the fissure swarms of Vogar and Thingvellir during the past 9000–10,000 years (Gudmundsson 1987a,b).

If a lava flow is erupted once every 10,000 years, and an average slip of 1 m occurs on a fault every 500 years, the great majority of the faults in our profile would have completed their growth before a new lava flow buried them. A throw of 20 m would be attained in only 10,000 years. Consequently, one would rarely see changes in throw with altitude on faults with throws of less than

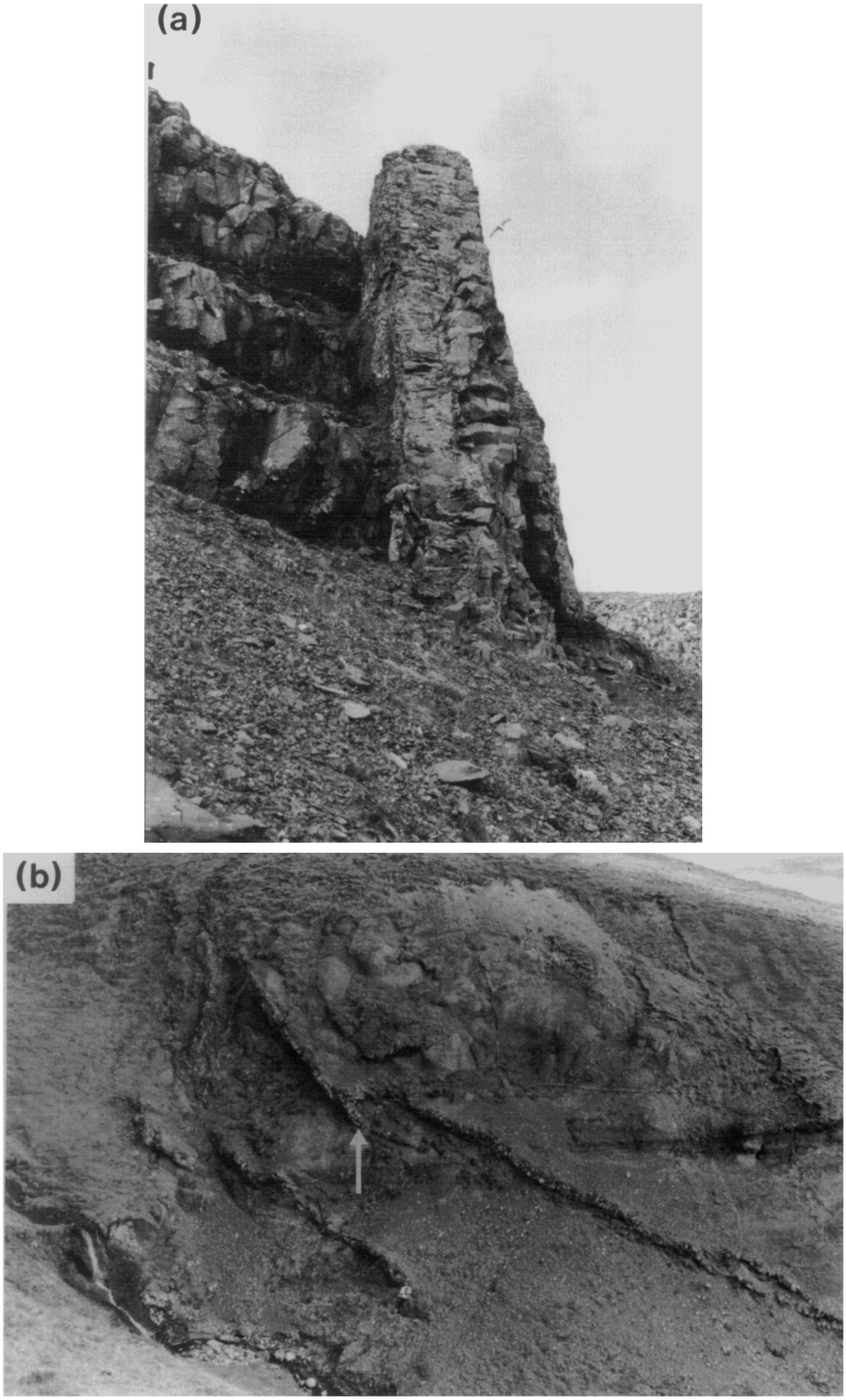

Fig. 8. (a) A 2.5 m thick straight basaltic dike, dipping 85°SE. (b) Basaltic sheet with horns (indicated by the arrow). The upper sheet (with the horns) is 0.8 m thick and dips 28°S, whereas the lower sheet is 0.9 m thick and dips 52°S. The host rock is basaltic breccia (hyaloclastite). The person below and to the right of the arrow provides a scale.

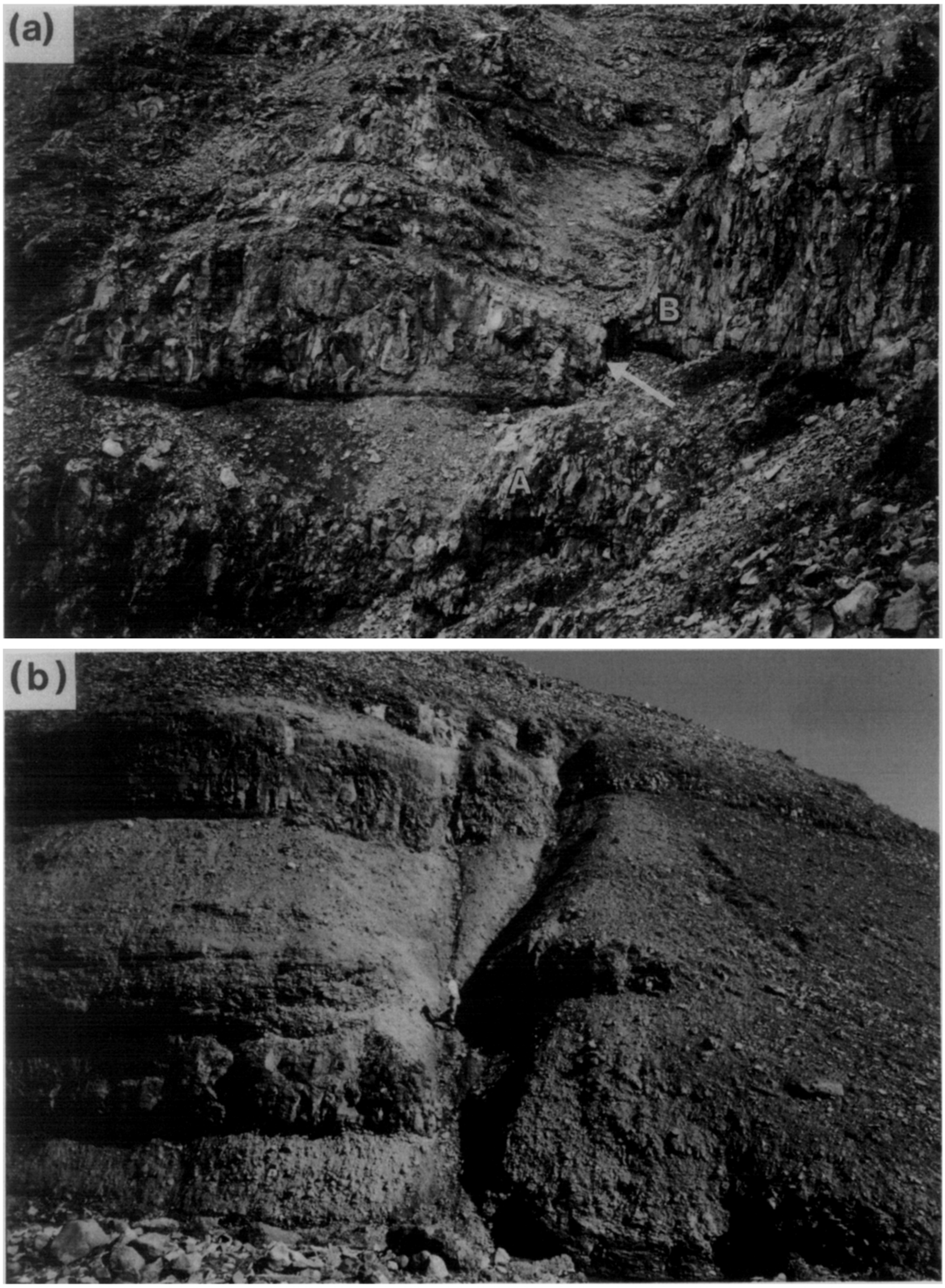

Fig. 9. (a) Pure dip-slip normal fault (located by the arrow), dipping 82°W. This is the only growth fault found in the area. The throw decreases from 9 m in the lava flow at the bottom (A) of the figure to 3 m in the lava flow in the centre (B) of the figure. (b) Normal fault dipping up to 73° in the lava flows and down to 58° in the conglomerates, with an average dip of 71°NW. The total throw is 4 m, but the upper part of the fault is split in two, where the right part has a throw of 3 m and the left part a throw of 1 m. The two parts join where the person is standing.

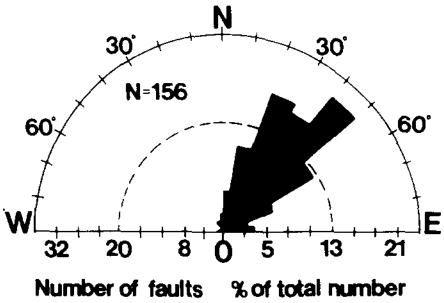

Fig. 10. Rose diagram for faults. The average strike is N26°E.

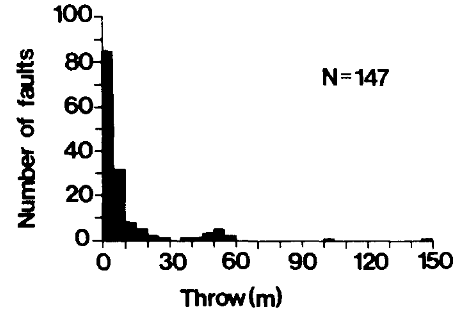

Fig. 12. Throw distribution of faults. The average throw is 10.2 m and 90% of the faults have throws of less than 20 m. The maximum measured throw is 150 m, and the minimum 0.5 m. One fault has an estimated throw of 200 m, but it is omitted from this figure.

20 m, which would apply to 90% of the faults (Fig. 12). The largest faults in the area, with throws of 150–200 m, would, at a constant slip rate, need 75,000–100,000 years to reach their present throws, and might be expected to show variation in fault displacement with depth. Because of the lack of good markers and suitable exposures we were not able to test if the largest faults were growth faults.

CRUSTAL DILATION

The amount of crustal extension due to faulting has not been widely discussed in the literature, except in papers on the Basin and Range Province, where province-wide extension is estimated to range from 64% to over 100% (Wernicke & Burchfiel 1982). Across 40–70 km-wide rift basins of the western rift of the East African rift system, Ebinger (1989) estimated the crustal extension in 13 basins as 2–15%. On the Mid-Atlantic Ridge at 37°N, Macdonald (1986) estimated the extension due to faults and fissures as 11% to the west and 18% to the east of the ridge axis, the difference reflecting the asymmetric spreading rate.

A brief discussion of the method used here is appropriate before presenting the results. For calculating the crustal dilation due to faulting the most important parameter, besides fault throw, is fault dip. At some localities only fault throw could be measured accurately, and for these the average dip of 10 major faults was used. The following procedure is in accordance with the method described for dikes by Gudmundsson (1984a,b).

Dikes and faults measured in the field are located on an aerial photograph. A line is drawn perpendicular to their average orientation, on to which all dikes and faults are projected. The numbers of both dikes and faults (N_d, N_f), the cumulative thickness (width) of dikes (T), and the cumulative horizontal displacement of faults (D) are then calculated for a specified interval (L) on the line (Figs. 14–17). The interval chosen on the aerial photographs is 0.5 cm, which corresponds to about 170 m in the field. The cumulative horizontal fault displacement (cumulative heave) is calculated according to the following equation

$$D = \cot\theta \cdot u, \qquad (1)$$

where u is the fault throw and θ is the fault dip. The percentage dilation in each interval is then calculated from the following relationships,

$$\% \text{ dilation due to dikes: } \frac{T\cdot 100}{L}$$

$$\% \text{ dilation due to faults: } \frac{D\cdot 100}{L}.$$

The Eyrarfjall profile (Fig. 14) is 4 km long and the other profiles (Figs. 15–17) are 8–9 km long. The combined length of all the profiles (i.e. the Eyrarfjall–Karastadir profile) is about 30 km.

The greatest dilation in one 170 m interval exceeds 40%, and this occurs in the Reynivallahals–Muli profile (Fig. 15). This high figure is the result of a high D value due to a fault dipping 71° with an estimated throw of 200 m. The greatest dilation due to dikes is 19%, and this occurs in one interval in the Eyrarfjall profile (Fig. 14) where there are 18 dikes with a cumulative dike thickness of 32 m. The dike distribution in the Eyrarfjall profile is typical for a narrow dike swarm. The average dilation of the 2.2 km wide Eyrarfjall dike swarm is

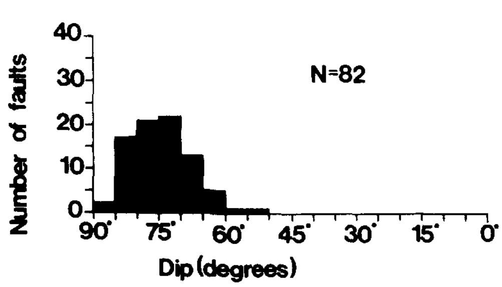

Fig. 11. Dip distribution of faults. The average dip is 75.2°, the minimum is 53° and the maximum is 89°.

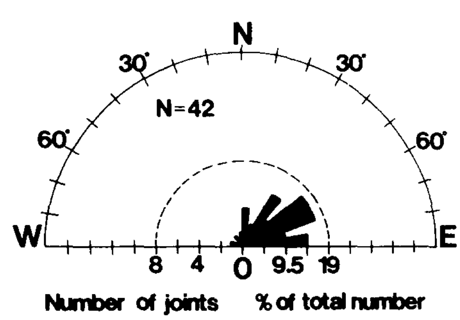

Fig. 13. Rose diagram for joints in subarea C. The average strike is N44°E.

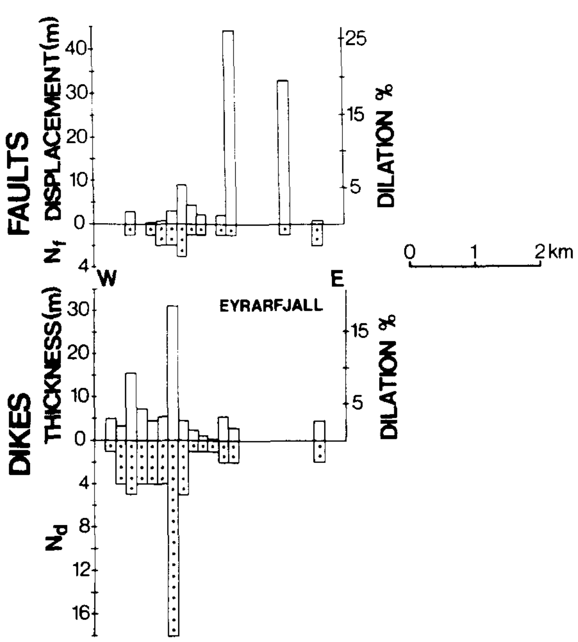

Fig. 14. Crustal dilation due to dikes and faults in the Eyrarfjall profile. The figure shows the number of dikes (N_d) and their cumulative thickness, the number of faults (N_f) and their cumulative horizontal displacement, and the percentage dilation of the crust due to dikes (lower profile) and due to faults (upper profile). Dots represent the numbers of dikes or faults in each interval (i.e. one dot represents one dike). See also Table 2.

4.0%. In general, the average dilation due to dikes varies from a maximum of about 3.7% in the coastal area (Table 2) to a minimum of only 0.4% in the easternmost (Thorufoss–Karastadir) profile (Table 2).

The average dilation due to faults ranges from a maximum of 5.3% in the Reynivallahals–Muli profile to a minimum of 0.6% in the Thorufoss–Karastadir profile (Table 2). This latter profile has the lowest values for all parameters, partly because of limited exposure, but also because the dike intensity is clearly lower in the shallowly eroded areas nearest to the active rift zone. The average crustal dilation in the four profiles combined, that is from Eyrarfjall toward the rift zone (the coastal area not included), is 3.6%, of which the dikes and faults contribute 1.3 and 2.3%, respectively.

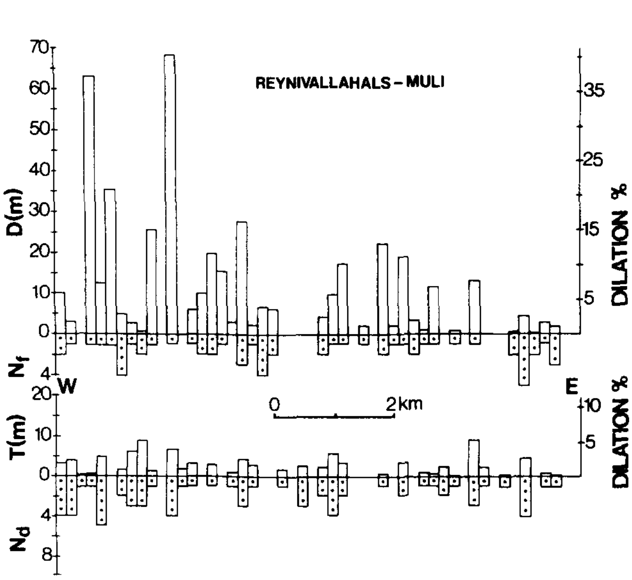

Fig. 15. Crustal dilation due to dikes and faults in the Reynivallahals–Muli profile. D, cumulative horizontal displacement (in metres); T, cumulative dike thickness (in metres); other symbols as in Fig. 14.

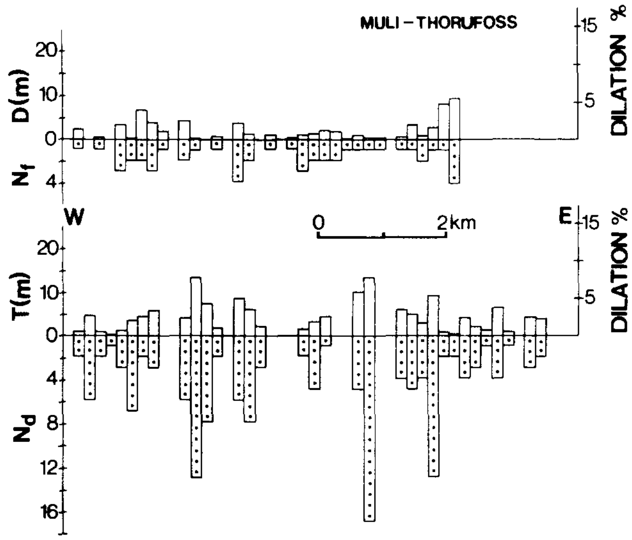

Fig. 16. Crustal dilation due to dikes and faults in the Muli–Thorufoss profile. Same symbols as in Fig. 15.

COMPARISON WITH DIKE SWARMS AND SHEET SWARMS

Dilation

Using the same method as here, Gudmundsson (1984b) calculated the dilation, due to dikes only, in two regional dike swarms in east Iceland. In a 5.1 km long profile across the Breiddalur dike swarm (Gudmundsson 1984b) there were 59 dikes, giving an average crustal dilation of 5.2% and a maximum dilation (in a 223 m long part of the profile) of about 12%. In a 9.3 km long profile across the nearby Alftafjördur dike swarm there were 95 dikes, giving an average crustal dilation of 5.5% and a maximum dilation (in a 250 m long part of the profile) of about 16%. Using a somewhat different method, Walker (1974) obtained an average dilation of as much as 8% in the central part of the Alftafjördur dike swarm.

In a similar study of the dike swarms of northwest Iceland (Gudmundsson 1984a), the average dilation in 1–2 km long profiles was 5–6%, with a maximum dilation

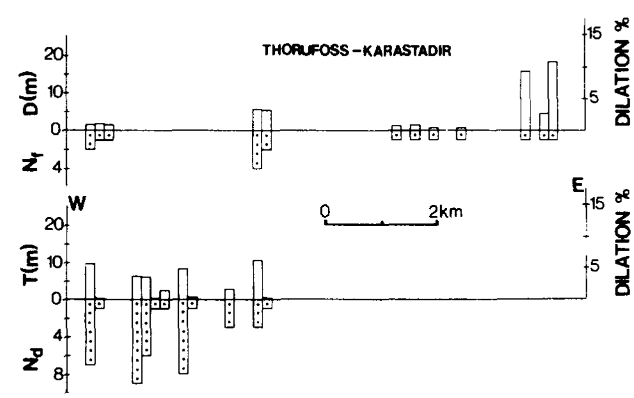

Fig. 17. Crustal dilation due to dikes and faults in the Thorufoss–Karastadir profile. Same symbols as in Fig. 15.

(in a 185 m long part of the profile) of about 17%. Continuous profiles across whole dike swarms are, however, lacking in this area. Nonetheless the average dilation across whole swarms in northwest Iceland is probably on the order of a few per cent (Gudmundsson 1984a). In a similar study of regional dikes in west Iceland, Gautneb (1989) obtained an average dilation of 0.8–1.8% in 3.0–10.5 km long profiles.

Crustal dilation has also been calculated for swarms of inclined (cone) sheets in several Icelandic central volcanoes. In all cases normal faults have been ignored so that the dilation is solely due to the sheets. In five 540 m long profiles in the Hafnarfjall volcano in west Iceland, the percentage of sheets ranges from 6.1 to 10.9% (Gautneb *et al.* 1989). In shorter traverses much higher (and also much lower) percentages of sheets were obtained, the maximum being 60%. Similarly, in 1–1.5 km long profiles in the sheet swarm of the Reykjadalur volcano in west Iceland, Gautneb (1989) obtained dilations of 5.8 and 7.9%. In traverses less than 100 m long, however, the percentage of sheets was as much as 90%.

Considering only the dilation attributable to dikes, the figures from our research area are similar to those obtained from the regional dike swarms in other areas of Iceland. The average dilation in the 2.2 km wide Eyrarfjall dike swarm is 4% and the maximum dilation in a 170 m long part of the profile is 19%, which is similar to comparable figures from the Tertiary swarms in east and northwest Iceland. Even in the much longer coastal profile, the average dilation due to dikes is 3.7%, which is only slightly lower than in east Iceland and similar to the dilation in northwest Iceland. The Eyrarfjall swarm (which is also observed in the coastal profile) is, however, the only true dike swarm in the area, as the other dikes do not group into clear-cut swarms. Accordingly, the dilation due to dikes in the profiles east of Eyrarfjall is low, only 0.4–1.8%.

Dike thickness

Although the dilation is not dissimilar from what is found in the Tertiary dike swarms, the average thickness of the dikes is much less in our area. The average dike thicknesses in the coastal area and the Eyrarfjall profile (subarea A, Table 1) are only 1.8 and 1.7 m, respectively. By contrast, the average thickness of regional dikes in east Iceland is 4.1 m (Gudmundsson 1983), 4.3 m in northwest Iceland (Gudmundsson 1984a) and 3.2 m in west Iceland (Gautneb 1989). Because the dilation is similar in all cases, it follows that the number of dikes per kilometre is much higher in our profile than in most profiles of equivalent length in the Tertiary areas. For example, the 9.3 km long (coastal) profile across the Alftafjördur swarm in east Iceland transects only 95 dikes whereas the 6.3 km long coastal profile in the research area transects 135 dikes.

There are several possible explanations why the dikes in the Eyrarfjall–Karastadir profile are thinner than dikes in the Tertiary dike swarms of Iceland. For a dike that is analogous to a through-going crack (Kassir & Sih 1975), which is a good model for many dikes exposed at shallow depths in volcanic rift zones (Gudmundsson 1990), the maximum thickness (width), W, of the dike in a horizontal section is (Sneddon 1951, Gudmundsson 1990)

$$W = \frac{2L(1-\nu^2)\Delta p_h}{E}, \tag{2}$$

where L is the horizontal length of the dike, ν is Poisson's ratio, E is Young's modulus and Δp_h is the magmatic overpressure as compared to the horizontal compressive stress in the crust (it is considered uniform along the length of the dike). Because the Quaternary lava pile is a multilayer of basaltic breccias and lava flows, its average Young's modulus should be lower (Gudmundsson 1986), thus favouring thicker dikes according to equation (2), than the average Young's modulus of the Tertiary lava pile. Thus, high Young's modulus cannot explain the thin dikes and, according to equation (2), the explanation is likely to be magmatic overpressure, shortness of the dikes, or both.

The horizontal magmatic overpressure is given by (Gudmundsson 1990)

$$\Delta p_h = \Delta p_v + \sigma_v - \sigma_h, \tag{3}$$

where σ_v is the vertical compressive stress (lithostatic pressure), σ_h is the horizontal compressive stress perpendicular to the dike, and Δp_v is the static magmatic overpressure as compared with the vertical stress. The latter is given by (Gudmundsson 1990)

$$\Delta p_v = (\rho_r - \rho_m)\mathbf{g}h + p_e, \tag{4}$$

where ρ_r is the average density of the crust, ρ_m is the density of the magma (assumed constant), $\mathbf{g}$ is the acceleration of gravity, h is the height of the dike and p_e is the excess magmatic pressure in the source reservoir before intrusion of the dike.

Because of the low-density basaltic breccias, the lava pile of the Eyrarfjall–Karastadir profile is likely to be of less average density, at a given depth, than the lava pile of the Tertiary dike swarms. From equations (3) and (4), the horizontal magmatic overpressure in dikes might thus be expected to be less, for magma of given density, in the research area than in the Tertiary swarms. Other things being equal, the dikes in the Eyrarfjall–Karastadir profile would, from equation (2), be expected to be thinner than the Tertiary dikes. In addition, they would be expected to become somewhat thinner with increasing altitude, as the magma was flowing upward through a crust of density lower than that of the magma.

The prediction that the dikes thin upward in the research area is supported by observations. For example, the coastal dikes are thicker, on average, than the dikes in subarea A. Both areas cover the same dike swarm, but subarea A is 200–300 m higher up in the lava pile. Furthermore, the dikes in the east part of the Eyrarfjall–Karastadir profile, that is subareas C–F, are significantly thinner than the dikes in the west part of the

Table 2. Crustal dilation due to dikes and faults and fault spacing in the research area

	Coastal area	Eyrarfjall profile	Reynivallahals–Muli profile	Muli–Thorufoss profile	Thorufoss–Karastadir profile
Length of profile (m)	6290	3740	8500	7990	9010
Number of dikes	135	54	67	150	41
Cumulative thickness (m)	34.7	93.4	98.4	146.6	36.2
Dilation due to dikes (%)	3.73	2.50	1.16	1.83	0.4
Number of faults	16	16	41	47	16
Cumulative horizontal displacement (m)	10.7	102.1	448.4	64.7	57.5
Dilation due to faults (%)	0.17	2.73	5.28	0.81	0.6
Fault spacing (m)	393	234	207	170	563
Total dilation (%)	3.90	5.23	6.44	2.64	1.04

profile (Table 2). This is so even though the basaltic breccias with low Young's moduli, favouring thick dikes, are more common in the east part. The depth of erosion and the average density of basaltic breccias are less in the east part than in the west part of our profile. Thus, the dikes appear to become thinner upward in the lava pile.

The dikes in the Eyrarfjall–Karastadir profile might also be thinner because they were shorter, on average, than the dikes of the Tertiary dike swarms. Equation (2) shows that, other things being equal, shorter dikes are thinner than longer dikes. At divergent plate boundaries, dikes tend to become shorter, but also thicker, at shallow crustal levels, but that conclusion applies only at crustal depths below 1–3 km (Gudmundsson 1990). In the uppermost 1–3 km of the crust, the relationships between magmatic overpressure, dike length and dike thickness are complex, and the available theoretical models do not make exact geometric predictions.

There are indications that the dikes in the research area might be much shorter than Tertiary dikes at similar depths in the crust. The length of dikes in a swarm is related to the length of the source reservoirs (Gudmundsson 1990). The number of central volcanoes, and hence the number of source reservoirs, has apparently been higher, per unit area of crust, during Quaternary than during Tertiary time (Gudmundsson 1986). This suggests that the lengths of the Quaternary source reservoirs have been less than the lengths of corresponding reservoirs during the Tertiary. In addition, the Tertiary dike swarms are as much as 50 km long (Walker 1974, Gudmundsson 1983), whereas the spacing of central volcanoes indicates that the dike swarms in our area are 10–20 km long. The source reservoirs would be similar in length to the dike swarms, and the maximum length of the dikes would rarely exceed the length of the source reservoir. The maximum length of dikes in our area might therefore be one-quarter of the maximum length of the Tertiary dikes. From equation (2) it then follows that the thickness of the dikes in the Eyrarfjall–Karastadir profile might be one-quarter of the thickness of Tertiary dikes, which agrees reasonably well with what is observed.

The possible short average length of the dikes in the research area, combined with lower magmatic overpressures, could explain why these dikes are, on average, much thinner than the dikes of the regional Tertiary swarms. An additional factor might be that a higher proportion of these dikes was derived from shallow magma chambers beneath central volcanoes, because the spacing of central volcanoes is less in Quaternary areas than in Tertiary areas. The overpressure of dikes from shallow chambers is normally less, at a given crustal depth, than that of dikes from a deep-seated magma reservoir (Gudmundsson 1988). This might also contribute to the general thinness of these dikes.

Dilation due to normal faults

Another difference between the region of Tertiary dike swarms and our area is the high contribution of normal faults to crustal dilation in the latter. The contribution of normal faults to the dilation in Tertiary dike swarms appears to be small. The exact figures are not known, but the available data indicate that the contribution is much less than in the Eyrarfjall–Karastadir profile. For example, in many profiles in east Iceland, Gudmundsson (1984b) found a total of only 19 normal faults, with an average throw of 2.7 m, but more than 600 dikes. Similarly, Bodvarsson & Walker (1964), who based their conclusions on Walker's studies in east Iceland north of the area studied by Gudmundsson (1983, 1984b), estimated the dilation due to normal faults as being only 10% of that due to dikes. Even if the contribution of normal faults to dilation is higher in northwest Iceland (Gudmundsson 1984a) than in east Iceland, it is still much less than in the Eyrarfjall–Karastadir profile.

In the present area, the extension due to faults is similar to that attributable to dikes in two profiles, the Eyrarfjall profile and the Thorufoss–Karastadir profile, but dissimilar in others (Table 2). Fault dilation is less than 5% of dike dilation in the coastal profile, 44% of the dike dilation in the Muli–Thorufoss profile, and 455% of the dike dilation in the Reynivallahals–Muli profile. In this latter profile, most crustal extension is due to normal faults.

Even if the lack of suitable exposures may lead to something of an underestimate of the contribution of normal faults in the coastal area, we believe that the above figures are significant. Except for the easternmost profile, Thorufoss–Karastadir, where the exposures are not good and the level of erosion shallow, the results show that where the dilation due to dikes is highest (coastal area) the dilation due to faults is lowest, and vice versa (Reynivallahals–Muli profile). They also show a general eastward fall-off in dilation due to dikes, that is with increasing altitude in the lava pile. By contrast, the dilation due to faults first increases rapidly with altitude (i.e. eastward) in the lava pile and then decreases again in the areas that are least eroded. This decrease upward (eastward) in the pile above a certain level may partly be because joints (tension fractures) become common (at least in subarea C) where the contribution of faults starts to decrease.

COMPARISON WITH FISSURE SWARMS

Fault spacing

A rough estimate of the average fault spacing in the profiles gives a range of from one fault every 563 m to one fault every 170 m. There is a gradual decrease in fault spacing eastward in the research area until the easternmost profile, where the spacing increases again. This increase, however, may be partly due to lack of suitable exposures.

Spacing of fractures, including both normal faults and large tension fractures, in Holocene fissure swarms in Iceland is very variable (Gudmundsson 1987a, Opheim & Gudmundsson 1989). The most accurate measurements are from 10 profiles across the Vogar fissure swarm on the Reykjanes Peninsula (Gudmundsson 1987b), where the average spacing ranges from 967 to 256 m, with an overall average of 456 m. In all our profiles except one, the average spacing is much less than it is in the profiles of the Vogar swarm, but similar to the minimum spacing of fractures in profiles across the Krafla fissure swarm in northeast Iceland (Opheim & Gudmundsson 1989). This suggests that the number of faults increases, and the fault spacing decreases, during the lifetime of a fissure swarm.

Dilation

Crustal dilation across several fissure swarms in Iceland has been measured. In the 5–7 km wide Thingvellir swarm, located north of Lake Thingvallavatn (Fig. 2), the maximum combined fracture width is about 100 m (Gudmundsson 1987a). The maximum crustal dilation is thus 1.4–2%. Similar measurements along a different profile gave 1.25% dilation (Bernauer 1943).

In the 7.5 km wide Vogar swarm, Gudmundsson (1987b) obtained a maximum extension of 15 m. This gives a maximum dilation of only 0.2%, but it must be stressed that the Vogar swarm is only one of four parallel swarms on the Reykjanes Peninsula and the combined dilation across them could be at least four times higher (Gudmundsson 1987a). Opheim & Gudmundsson (1989) measured the extension in five 2–6 km long profiles across the Krafla swarm. The dilation there is from 0.25 to 2.3%, which falls within the range obtained from the swarms at Thingvellir and Vogar.

The dilation due to faults in the present study area is generally similar to the dilation obtained for the Holocene fissure swarms. The lengths of our profiles (Table 2) are also similar to the lengths of the profiles across the fissure swarms. The greatest dilation in our area is only twice the maximum dilation obtained in the fissure swarms (in the Krafla swarm). These figures are, however, not completely comparable because the dilation measured in the fissure swarms is almost entirely due to the width of open fractures, whether normal faults or tension fractures, whereas the dilation in the Eyrarfjall–Karastadir profile is the horizontal extension (the heave) calculated from the dip and throw on the normal faults. Furthermore, our profile is measured at depths of a few hundred metres to about one kilometre below the initial top of the lava pile, whereas the dilation across the fissure swarms is measured at the surface.

Infrastructure of fissure swarms

The results obtained in this study, in addition to the results of other studies of fissure swarms and Tertiary dike swarms in Iceland, allow us to infer the infrastructure of a typical fissure swarm (dike swarm) in Iceland (Fig. 18). At the surface the tensile strain is almost entirely accommodated by tension fractures and gaping normal faults. Because the faults are vertical at the surface (Gudmundsson 1987a,b), the heave is normally zero and does not contribute to the dilation. Occasional feeder-dikes may also contribute to the dilation.

At greater depths in the crust the tension fractures gradually taper away and decrease in number, whereas the dip of the normal faults becomes non-vertical and their contribution to the dilation (i.e. the heave) increases. Gudmundsson (1987a,b) concluded that the depth of most of the pure tension fractures is on the order of several hundred metres or less, which is the depth where the tectonic joints in our profile were observed (subarea C). In the depth range from a couple of hundred metres to 1 km or so the normal faults contribute most significantly to the dilation and may be the dominant extensional structures in some areas.

At depths greater than about 1 km, but which may be

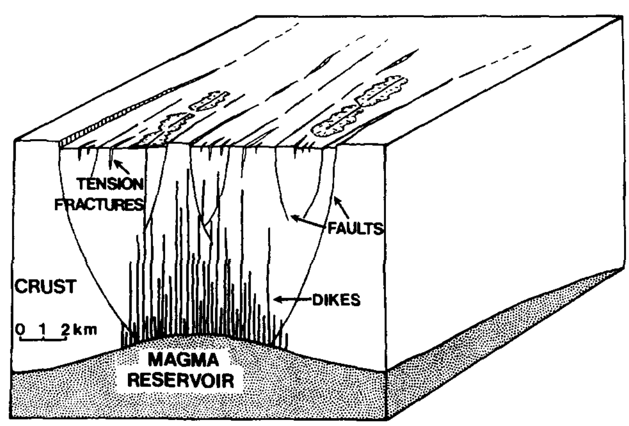

Fig. 18. Schematic illustration of the infrastructure of a typical fissure swarm in the rift zone of Iceland.

lesser or greater in some areas, the contribution of dikes to crustal dilation becomes gradually dominant and continues to be so at deeper crustal levels. At such depths most or all of the tension fractures have disappeared and only the major normal faults continue to the deeper levels. Some of these major faults may continue right to the bottom of the crust, that is down to the magma reservoirs.

Acknowledgements—We thank Peter Hudleston, Kristjan Saemundsson and an anonymous reviewer for helpful comments.

REFERENCES

Bäckström, K. & Gudmundsson, A. 1989. The grabens of Sveinar and Sveinagja, NE Iceland. *Prof. Pap. Nordic Volcanol. Inst.* **8901**.

Bernauer, F. 1943. Junge Tektonik auf Island und ihre Ursachen. In: *Spalten auf Island* (edited by Niemczyk, O.). Wittwer, Stuttgart, 14–63.

Bergerat, F., Angelier, J. & Villemin, T. 1990. Fault systems and stress patterns on emerged oceanic ridges: a case study in Iceland. *Tectonophysics* **179**, 183–197.

Bjornsson, A., Johnsen, G., Sigurdsson, S., Thorbergsson, G. & Tryggvason, E. 1979. Rifting of the plate boundary in northern Iceland, 1975–1978. *J. geophys. Res.* **84**, 3029–3038.

Bodvarsson, G. & Walker, G. P. L. 1964. Crustal drift in Iceland. *Geophys. J. R. astr. Soc.* **8**, 285–300.

Bonilla, M. G. 1979. Historic surface faulting: map patterns, relation to subsurface faulting, and relation to preexisting faults. *U.S. Geol. Surv. Conference on Actual Fault Zones in Bedrock. U.S. geol. Surv. Open-file Rep.* **79–1239**, 36–65.

Daignieres, M., Courtillot, V. & Bayer, R. 1975. A model for the evolution of the axial zone of mid-ocean ridges as suggested by Icelandic tectonics. *Earth Planet. Sci. Lett.* **26**, 222–232.

Ebinger, C. J. 1989. Tectonic development of the western branch of the East African rift system. *Bull. geol. Soc. Am.* **101**, 885–903.

Einarsson, T. 1962. Upper Tertiary and Pleistocene rocks in Iceland. *Soc. Scientiarum Islandica* **36**, 1–196.

Franzson, H. 1978. Structure and petrochemistry of the Hafnarfjall–Skardsheidi central volcano and the surrounding basalt succession, W. Iceland. Unpublished Ph.D. thesis, University of Edinburgh, Edinburgh.

Fridleifsson, I. B. 1973. Petrology and structure of the Esja Quaternary volcanic region, SW Iceland. Unpublished Ph.D. thesis, University of Oxford, Oxford.

Gautneb, H. 1989. Sheet intrusions associated with the Reykjadalur volcano, western Iceland; structure and composition. *Prof. Pap. Nordic Volcanol. Inst.* **8902**.

Gautneb, H., Gudmundsson, A. & Oskarsson, N. 1989. Structure, petrochemistry and evolution of a sheet swarm in an Icelandic central volcano. *Geol. Mag.* **126**, 659–673.

Gibson, I. L. & Piper, J. D. A. 1972. Structure of the Icelandic basalt plateau and the process of drift. *Phil. Trans. R. Soc. Lond.* **A271**, 141–150.

Gudmundsson, A. 1983. Form and dimensions of dykes in eastern Iceland. *Tectonophysics* **95**, 295–307.

Gudmundsson, A. 1984a. Tectonic aspects of dykes in northwestern Iceland. *Jökull* **34**, 81–96.

Gudmundsson, A. 1984b. A study of dykes, fissures and faults in selected areas of Iceland. Unpublished Ph.D. thesis, University of London, London.

Gudmundsson, A. 1984c. Formation of dykes, feeder-dykes, and the intrusion of dykes from magma chambers. *Bull. Volcanol.* **46**, 537–550.

Gudmundsson, A. 1986. Formation of crustal magma chambers in Iceland. *Geology* **14**, 164–166.

Gudmundsson, A. 1987a. Tectonics of the Thingvellir fissure swarm, SW Iceland. *J. Struct. Geol.* **9**, 61–69.

Gudmundsson, A. 1987b. Geometry, formation and development of tectonic fractures on the Reykjanes Peninsula, Southwest Iceland. *Tectonophysics* **139**, 295–308.

Gudmundsson, A. 1988. Effect of tensile stress concentration around magma chambers on intrusion and extrusion frequencies. *J. volcanol. Geotherm. Res.* **35**, 179–194.

Gudmundsson, A. 1990. Emplacement of dikes, sills and crustal magma chambers at divergent plate boundaries. *Tectonophysics* **176**, 257–275.

Gunnlaugsson, G., Gislason, G., Imsland, P. & Hafstad, Th. 1972. Jardfraedi Reynivallahals i Kjos. Unpublished dissertation, University of Iceland, Reykjavik.

Helgason, J. 1976. Jardfraedi Kjosarskards, bergflokkun, ummyndun, utskolun og gangar. Unpublished dissertation, University of Iceland, Reykjavik.

Hjartarson, A., Kaldal, I., Vikingsson, S. & Olafsdottir, Th. 1973. Medalfell og Reynivallahals, jardfraediskyrsla. Unpublished dissertation, University of Iceland, Reykjavik.

Jefferis, R. G. & Voight, B. 1981. Fracture analysis near the mid-ocean plate boundary, Reykjavik–Hvalfjördur area, Iceland. *Tectonophysics* **76**, 171–236.

Jonasson, B., Hallsdottir, M., Fridriksdottir, S., Zophoniasson & Skaftadottir, Th. 1973. Eyrarfjall, jardfraediskyrsla. Unpublished dissertation, University of Iceland, Reykjavik.

Kassir, M. K. & Sih, G. C. 1975. Three-dimensional crack problems. In: *Mechanics of Fracture, Vol.* 3. Noordhoff, Leyden.

Kristjansson, L., Fridleifsson, I. B. & Watkins, N. D. 1980. Stratigraphy and paleomagnetism of the Esja, Eyrarfjall and Akrafjall mountains, SW-Iceland. *J. Geophys.* **47**, 31–42.

Macdonald, K. C. 1986. The crest of the Mid-Atlantic Ridge: models for crustal generation processes and tectonics. In: *The Geology of North America, Volume M, The Western North Atlantic Region* (edited by Vogt, P. R. & Tucholke, B. E.). *Geol. Soc. Am.*, 51–68.

McDougall, I., Watkins, N. D. & Kristjansson, L. 1976a. Geochronology and paleomagnetism of a Miocene–Pliocene lava sequence at Bessastadaa, eastern Iceland. *Am. J. Sci.* **276**, 1078–1095.

McDougall, I., Watkins, N. D., Walker, G. P. L. & Kristjansson, L. 1976b. Potassium–argon and paleomagnetic analysis of Icelandic lava flows: limits on the age of anomaly 5. *J. geophys. Res.* **81**, 1505–1512.

McDougall, I., Saemundsson, K., Johannesson, H., Watkins, N. D. & Kristjansson, L. 1977. Extension of the geomagnetic polarity time scale to 6.5 m.y.: K–Ar dating, geological and paleomagnetic study of a 3,500 m lava succession in western Iceland. *Bull geol. Soc. Am.* **88**, 1–15.

McDougall, I., Kristjansson, L. & Saemundsson, K. 1984. Magnetostratigraphy and geochronology of northwest Iceland. *J. geophys. Res.* **89**, 7029–7060.

Opheim, J. A. & Gudmundsson, A. 1989. Formation and geometry of fractures, and related volcanism, of the Krafla fissure swarm, northeast Iceland. *Bull. geol. Soc. Am.* **101**, 1608–1622.

Palmason, G. 1973. Kinematics and heat flow in a volcanic rift zone, with application to Iceland. *Geophys. J. R. astr. Soc.* **33**, 451–481.

Palmason, G. 1980. A continuum model of crustal generation in Iceland: kinematic aspects. *J. Geophys.* **47**, 7–18.

Palmason, G. 1986. Model of crustal formation in Iceland, and application to submarine mid-ocean ridges. In: *The Geology of North America, Volume M, The Western North Atlantic Region* (edited by Vogt, P. R. & Tucholke, B. E.). *Geol. Soc. Am.*, 87–97.

Peng, S. & Johnson, A. M. 1972. Crack growth and faulting in cylindrical specimens of Chelmsford Granite. *Int. J. Rock. Mech. & Mining Sci. Geomech. Abs.* **9**, 37–86.

Rutten, M. G. 1958. Geological reconnaissance of the Esja–Hvalfjördur–Armannsfell area, southwestern Iceland. *Verh. K. Ned. geol. mijnbouwkd. Genoot.* **17**, 219–298.

Saemundsson, K. 1979. Outline of the geology of Iceland. *Jökull* **29**, 7–28.

Saemundsson, K. 1986. Subaerial volcanism in western North Atlantic. In: *The Geology of North America, Volume M, The Western North Atlantic Region* (edited by Vogt, P. R. & Tucholke, B. E.). Geol. Soc. Am., 69–86.

Saemundsson, K. & Einarsson, S. 1980. *Geological Map of Iceland, Sheet* 3, *SW Iceland* (2nd edn). Museum of Natural History and the Iceland Geodetic Survey, Reykjavik, Iceland.

Saemundsson, K., Kristjansson, L., McDougall, I. & Watkins, N. D. 1980. K–Ar dating, geological and paleomagnetic study of a 5 km lava succession in northern Iceland. *J. geophys. Res.* **85**, 3628–3646.

Scholz, C. H. 1989. Mechanics of faulting. *Annu. Rev. Earth & Planet. Sci.* **17**, 309–334.

Sigurdsson, O. 1980. Surface deformation of the Krafla fissure swarm in two rifting events. *J. Geophys.* **47**, 154–159.

Sneddon, I. N. 1951. *Fourier Transforms*. McGraw-Hill, New York.

Snorrason, S. P. 1976. Jardfraedi Kjosarskards, jardlagaskipan. Unpublished dissertation, University of Iceland, Reykjavik.

Speed, R. 1979. Extension faulting in the Great Basin: kinematics and possible changes with depth. *Conference on Actual Fault Zones in Bedrock. U.S. geol. Surv. Open-file Rep.* **79–1239**, 121–138.

Thoroddsen, Th. 1958. *Ferdabok, Vol.* 1 (2nd edn). Snaebjörn Jonsson, Reykjavik.

Tryggvason, E. 1984. Widening of the Krafla fissure swarm during the 1975–1981 volcano-tectonic episode. *Bull. Volcanol.* **47**, 47–69.

Tryggvason, E. 1986. Vertical ground movement in the Krafla region 1977–1986. *Prof. Pap. Nordic Volcanol. Inst.* **8602**.

Walker, G. P. L. 1974. The structure of eastern Iceland. In: *Geodynamics of Iceland and the North-Atlantic Area* (edited by Kristjansson, L.). Reidel, Dordrecht, 177–188.

Watkins, N. D. & Walker, G. P. L. 1977. Magnetostratigraphy of eastern Iceland. *Am. J. Sci.* **277**, 513–584.

Wernicke, B. & Burchfiel, B. C. 1982. Modes of extensional tectonics. *J. Struct. Geol.* **4**, 105–115.