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Abstract

The July 1978 deflation of the Krafla volcano in the volcanic rift zone of NE-Iceland was in most respects typical of the many deflation events that have occurred at Krafla since December 1975. Separated by periods of slow inflation, the deflation events are characterized by rapid subsidence in the caldera region, volcanic tremor and extensive rifting in the fault swarm that transsects the volcano. Earthquakes increase in the caldera region shortly after deflation starts and propagate along the fault swarm away from the central part of the volcano, sometimes as far as 65 km. The deflation events are interpreted as the result of subsurface magmatic movements, when magma from the Krafla reservoir is injected laterally into the fault swarm to form a dyke. In the July 1978 event magma was injected a total distance of 30 km into the northern fault swarm. The dyke tip propagated with the velocity of 0.4-0.5 m/sec during the first 9 hours, but the velocity decreased as the length of the dyke increased. Combined with surface deformation data, these data can be used to estimate the cross sectional area of the dyke and the driving pressure of the magma. The cross sectional area is variable along the dyke and is largest in the regions of maximum earthquake activity. The average value is about 1200 m^2. The pressure difference between the magma reservoir and the dyke tip was of the order of 10-40 bars and did not change much during the injection.

Key words: Deflation - Krafla volcano - Iceland - Rifting - Earthquakes - Lateral magma injection - Dyke.

Introduction

On July 10 1978 rapid subsidence started in the caldera region of the Krafla central volcano in the volcanic rift zone of NE-Iceland, and during the following 3 days the central part of the caldera subsided about 60 cm. This deflation event was the nineth in a series of such events that has been in progress since December 20, 1975. The tectonic setting and the course of events of this activity have previously been described in some detail in the literature (e.g. Björnsson et al., 1977 and 1979; Einarsson, 1978; Brandsdóttir and Einarsson, 1979) and will not be repeated here.

In the time intervals between the deflation events, the Krafla volcano inflates at a relatively constant rate. The inflation has been interpreted as the result of constant inflow of about 5 m^3/sec of magma into a magma reservoir at the depth of about 3 km under the central part of the caldera. This interpretation is supported by levelling, tilt, and gravity measurements (Björnsson et al., 1979; Tryggvason, 1978a), and is further strengthened by the geological association with the central volcano and the existence of a zone of high S-wave attenuation near the center of inflation (Einarsson, 1978).

Deflation events are characterized by rapid subsidence of the caldera region, continuous volcanic tremor, extensive rifting and earthquakes along the Krafla fault swarm that crosses the volcano from north to south. Earthquakes begin within or near the caldera and then migrate along the fault swarm away from the caldera, sometimes as far as 65 km. The largest events of the earthquake swarm are confined within a well defined but each time different section of the fault swarm. Rifting, often exceeding 1 m, may occur in the fault swarm, and the area of maximum rifting generally coincides with the area of maximum earthquake activity. Three of the deflation events have been accompanied by a small basaltic

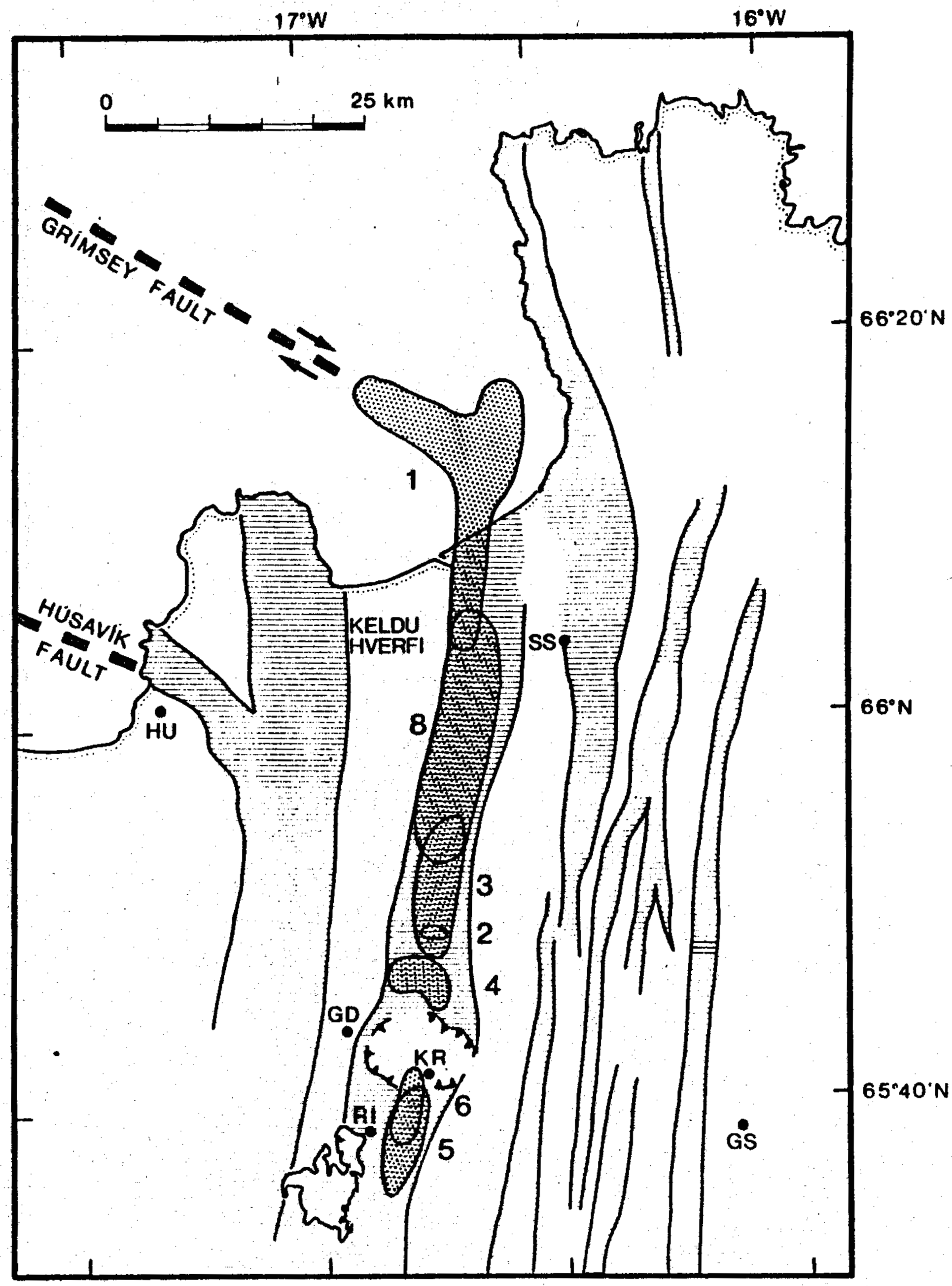

Fig. 1. Index map of the northern part of the volcanic rift zone in NE-Iceland. Dots mark permanent seismograph stations. The hatched areas are fault swarms as mapped by Kristján Saemundsson in Björnsson et al. 1977. The Krafla caldera is located within the Krafla fault swarm. The stippled areas are the areas of maximum earthquake activity during the different deflation events at Krafla. Area no. 1 is the epicentral area of the first earthquake swarm of Dec. 1975-Feb 1976 ; no 2, a small swarm of Oct. 1-2 1976 ; no. 3, swarm of Oct. 31 - Nov. 1 1976 ; no. 4, swarm of Jan. 1977 ; no. 5, swarm of April 1977 ; no. 6, swarm of Sept. 1977 ; a small swarm in Nov. 1977 could not be located and is not shown ; no. 8, swarm of Jan. 1978. The Grímsey fault was delineated by seismicity (Einarsson, 1976) and the sense of fault displacement was derived from focal mechanism solutions (Einarsson, 1979).

eruption in the caldera region. The spatial relationship between the Krafla caldera, the Krafla fault swarm and the epicentral areas of the different earthquake swarms is shown in Fig. 1 together with some of the major tectonic elements of the active zones in NE-Iceland.

The deflation events were interpreted by Björnsson et al. (1977) as the result of lateral migration of magma away from the reservoir under Krafla. This interpretation has since been supported by gravity and crustal deformation data (Björnsson et al., 1979; Tryggvason, 1978 a,b), seismological data (Brandsdóttir and Einarsson, 1979), and petrochemical data (Grönvold and Mäkipää, 1978). During the deflation event of September 1977, e.g., the propagation of the magma could be followed by the seismic events. The hypocenters migrated horizontally from the reservoir area southwards along the Krafla fault swarm to the Námafjall geothermal area, where small amounts of magma were subsequently erupted through a drill hole (Brandsdóttir and Einarsson, 1979). The maximum speed of migration was 0.5 m/sec.

In the present paper another case of lateral migration of seismic activity associated with deflation of the Krafla volcano is documented. In this deflation event of July 1978 magma was injected into the fault swarm to the north of the caldera, a total distance of 30 km.

The course of events

After the deflation event of January 1978 the Krafla volcano inflated at the normal rate. Towards the end of June the elevation of the caldera region was approaching the level it had before the January deflation. A new deflation event, possibly associated with an eruption, was anticipated soon thereafter. Monitoring of the area was therefore intensified and several portable seismographs were set up in the caldera region and south of it. The seismic activity of the area was low, of the order of 1-2 locatable microearthquakes per day.

Slow deflation started on July 10 at about 11 h. (all times are UTC) according to the continuously recording tiltmeter near the Krafla power house that is located within the caldera (Tryggvason, 1978c). The rate of tilting increased and when continuous tremor appeared on the seismograph station GD (Fig. 1) shortly before 17 h. it was clear that a deflation event had started. Earthquakes were small in the beginning but increased gradually in magnitude and number. They clearly originated in the northern part of the caldera and north of it, which indicated that the magma was injecting into the northern fault swarm. The tilt rate and the volcanic tremor reached a maximum at about 20 h. and then slowly decreased. The earthquake activity increased markedly after about 22 h., and a magnitude 4.1 earthquake occurred in the fault swarm at 22:44 (Fig. 2).

A helicopter was made available by the Icelandic Coast Guard, which made it possible to move a seismograph into the epicentral area. A station was set up at Snagi (SN) at 23:20. While the seismograph was being set up earthquakes could be heard every minute and many earthquakes were felt. The magnification of the seismograph had to be set 42 dB lower than normal because of the high activity. Nothing else unusual could be observed in the Snagi area. Steam fields that were formed near Snagi in previous deflation events (see e.g. Björnsson et al., 1979) were not noticeably changed, and the sheep were grazing quietly in the bright, Icelandic summer night.

The earthquake activity decreased significantly after 2 h. in the morning (Fig. 2), but about 6 h. it increased again and was very intensive for 12 hours. The activity had now moved 10-15 km farther north. The earthquakes were not large but very frequent. Only two earthquakes were felt in the inhabited areas. The earthquake of July 11, 12:28 (magnitude 3.9) was vaguely felt in the Kelduhverfi district at the epicentral distance of

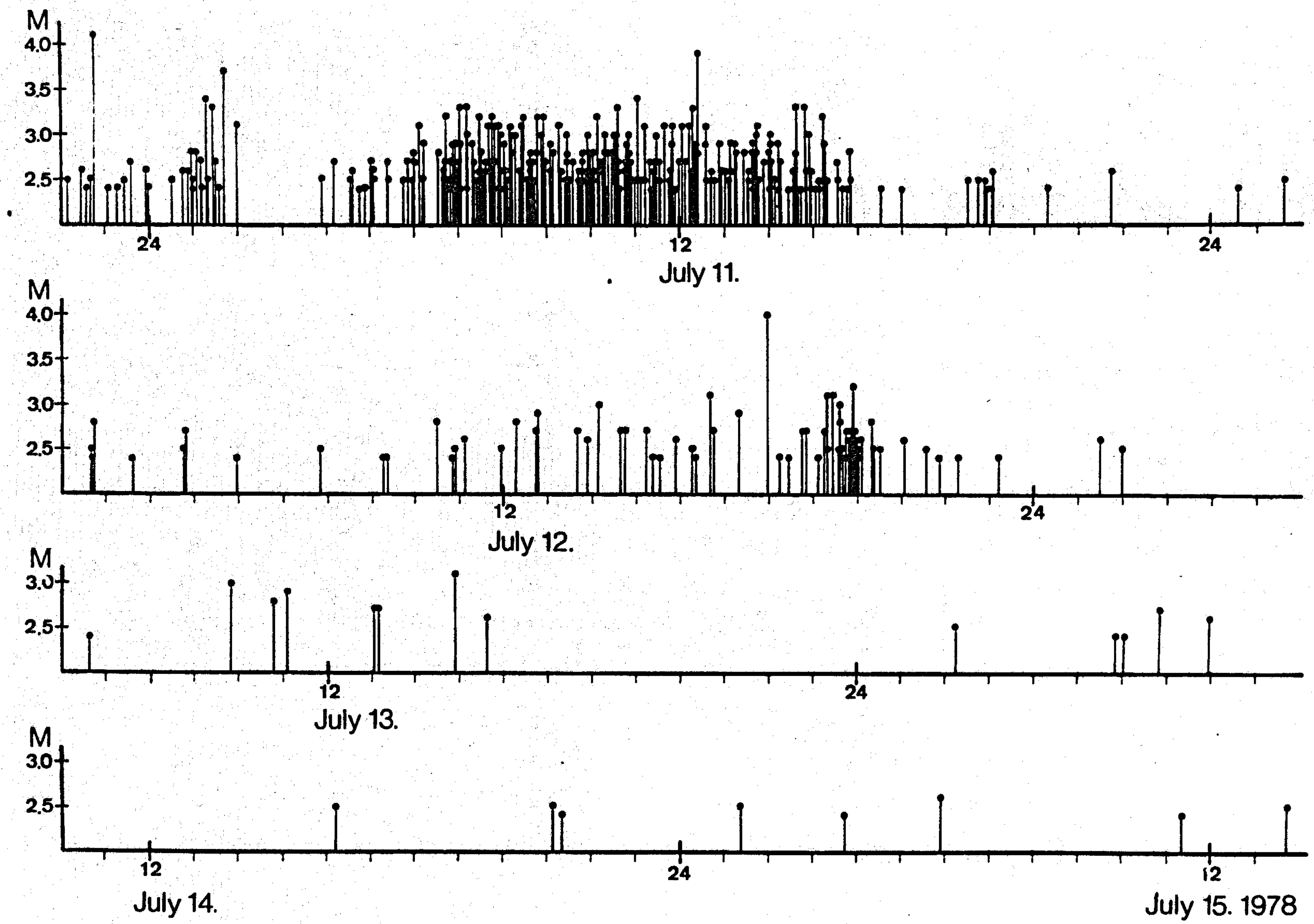

Fig. 2. The time sequence of earthquakes. The magnitude is plotted as a function of time.

5-10 km and the earthquake of July 12, 17:59 (magnitude 4.0) was widely felt in the Kelduhverfi district.

It was noticed in the morning of July 11 that a column of steam was rising from the area north of Krafla. The steam emission started between 03:04 and 03:14 according to photographs taken by a time lapse camera (Oddur Sigurdsson, pers. comm.). Subsequent inspection revealed that the steam came from a steam field located about 1.5 km SE of Snagi. This steam field was formed during one of the injection events in the fall of 1976, but now the steam emission had increased about an order of magnitude. During the following days the emission decreased slowly.

Several new steam fields have formed in the Snagi area in previous deflation events and most of them are aligned along one prominent normal fault. This fault moved a few tens of centimeters during the July 1978 event and was the easternmost fault that moved in this part of the fault swarm in that event. A total movement of 50-150 cm was estimated from the widening of the fissures in this area. The movement was distributed on numerous parallel fissures, some of which had moved in previous events. The estimate is therefore inaccurate. Much more accurate measurements were obtained with a geodimeter. In the Snagi area the extension across the fault swarm was found to be 95-111 cm (Tryggvason, 1978c). Compression occurred immediately outside of the faulted zone, which is a pattern also found in previous rifting events in other parts of the fault swarm (Tryggvason, 1978b,c; Gerke et al., 1978; Björnsson et al., 1979).

Fault movements were also observed in the northern part of the epicentral zone. On a profile across the fault swarm near 65°56'N a total of 80 cm extension was estimated from the widening of the individual fissures. About 50 cm of this movement were estimated to be due to the July rifting. The rest of the movement took place in a previous event, most likely the January 1978 event.

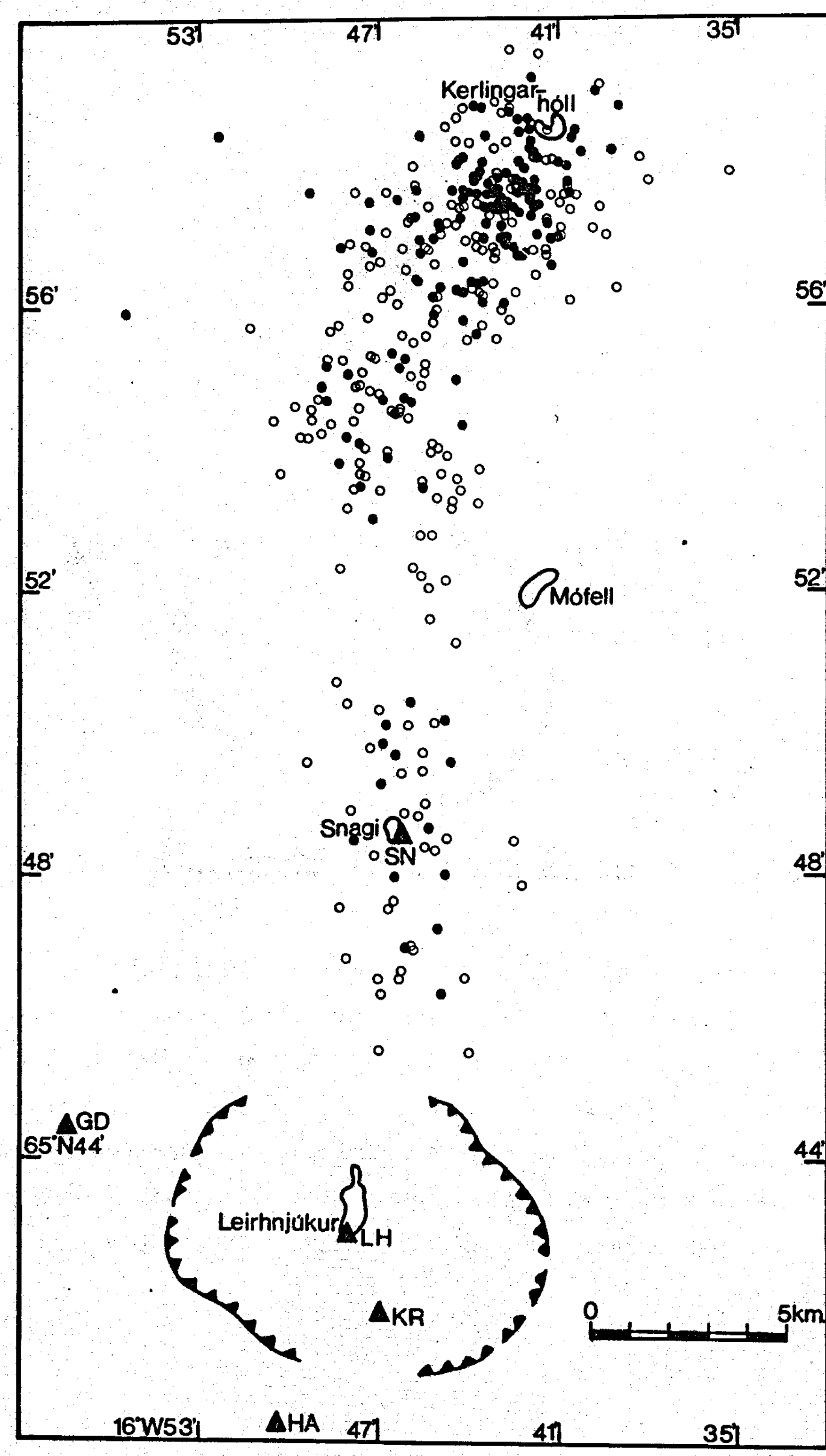

Fig. 3. Epicentral map of the July 1978 earthquake swarm. Dots mark epicenters located with horizontal standard error of 1 km and less, circles denote epicenters with errors between 1 and 2 km. Seismograph stations are shown with triangles. The stations HA, LH and SN were temporary stations. Several stations outside of this map were also used in the locations.

No geodimeter measurements are available for this part of the fault swarm.

The deflation stopped and inflation resumed at Krafla on July 13. The total volume removed from the magma reservoir during this deflation event is estimated to be $37 \times 10^6\ m^3$ (Tryggvason, 1978c).

The hypocentral zone

Hypocenters have been calculated for 397 earthquakes from the swarm. Generally the largest earthquakes were selected for analysis because they were recorded by the largest number of stations. Sometimes large earthquakes had to be omitted, however, because the first arrivals were obscured by tremor or small earthquakes, especially during periods of high activity. The set of located earthquakes may be regarded as nearly complete above magnitude 2.5. Many smaller earthquakes were located as well.

The computer program HYPOELLIPSE (Lahr and Ward, 1975) was used for the locations. P-wave arrival times and, where possible, also S-wave arrival times are used. The location procedure has been discribed in some detail by Einarsson (1978) and Brandsdóttir and Einarsson (1979) and will not be repeated here.

The epicenters are plotted on a map in Fig. 3. The epicentral zone is about 4-7 km wide, 25 km long, and extends from the northern rim of the caldera to Kerlingarhóll, which is a flat lava hill built up around the northernmost eruptive fissure in the Krafla fault swarm (Saemundsson, 1977). Nearly all the epicenters are located within the Krafla fault swarm. The zone is divided in two by a gap near the hyaloclastite hill Mófell. A large concentration of activity occurs in the northern end near Kerlingarhóll.

Depths of hypocenters could be determined with fair confidence in the area around Snagi because of the seismic

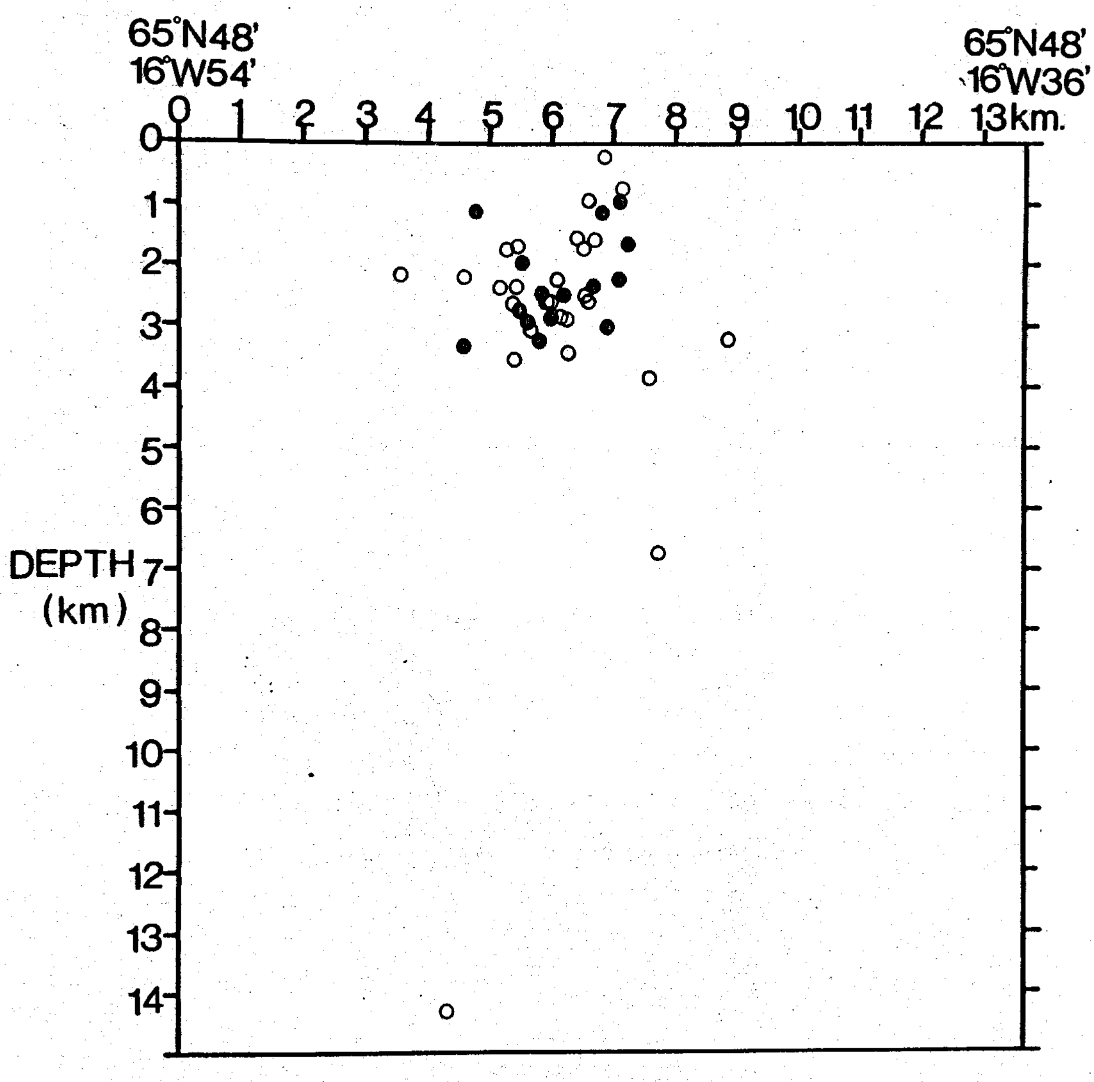

Fig. 4. Depth of hypocenters in the Snagi area. Hypocenters within 4 km horizontal distance from the station SN are projected on a vertical E-W plane. Dots are hypocenters determined with horizontal and vertical standard error of 1 km and less, circles denote hypocenters with errors between 1 and 2 km.

station located there. In the northern part of the epicentral zone the depths are not considered to be reliable because of the relatively large distance to the nearest seismograph station. A cross section of the hypocenters near Snagi is shown in Fig. 4. Hypocenters within 4 km distance from Snagi are projected on a vertical, E-W plane. Most of the hypocenters are at the depth of 1-4 km. An isolated event occurs at 14 km depth.

The migration of seismic activity

One of the most remarkable characteristics of the Krafla fault swarm earthquakes is the propagation of epicenters away from the caldera region. The July 1978 event provided one of the best examples of this. The latitude of the epicenters is plotted as a function of the time of occurrence in Fig. 5. The continuous tremor started in the caldera region at 17 h., the first earthquakes were there also but were too small to be located accurately. The earthquakes between 22 h. on July 10 and 2 h. on July 11 were located in the Snagi area. It is interesting to note that the gap in the epicentral zone near Mófell is also a gap in time. One may say that the seismic activity "jumped" across the gap and continued propagating on the other side with similar velocity as before. If one assumes that the activity started in the center of the caldera, the total distance is 30 km. The speed of propagation during the first 9 hours is 1.6 km/hour or 0.4-0.5 m/sec.. The speed decreases only slightly during the next 8 hours, but during the time period 11 h. - 19 h. on July 11 the average speed is only about 0.1 m/sec..

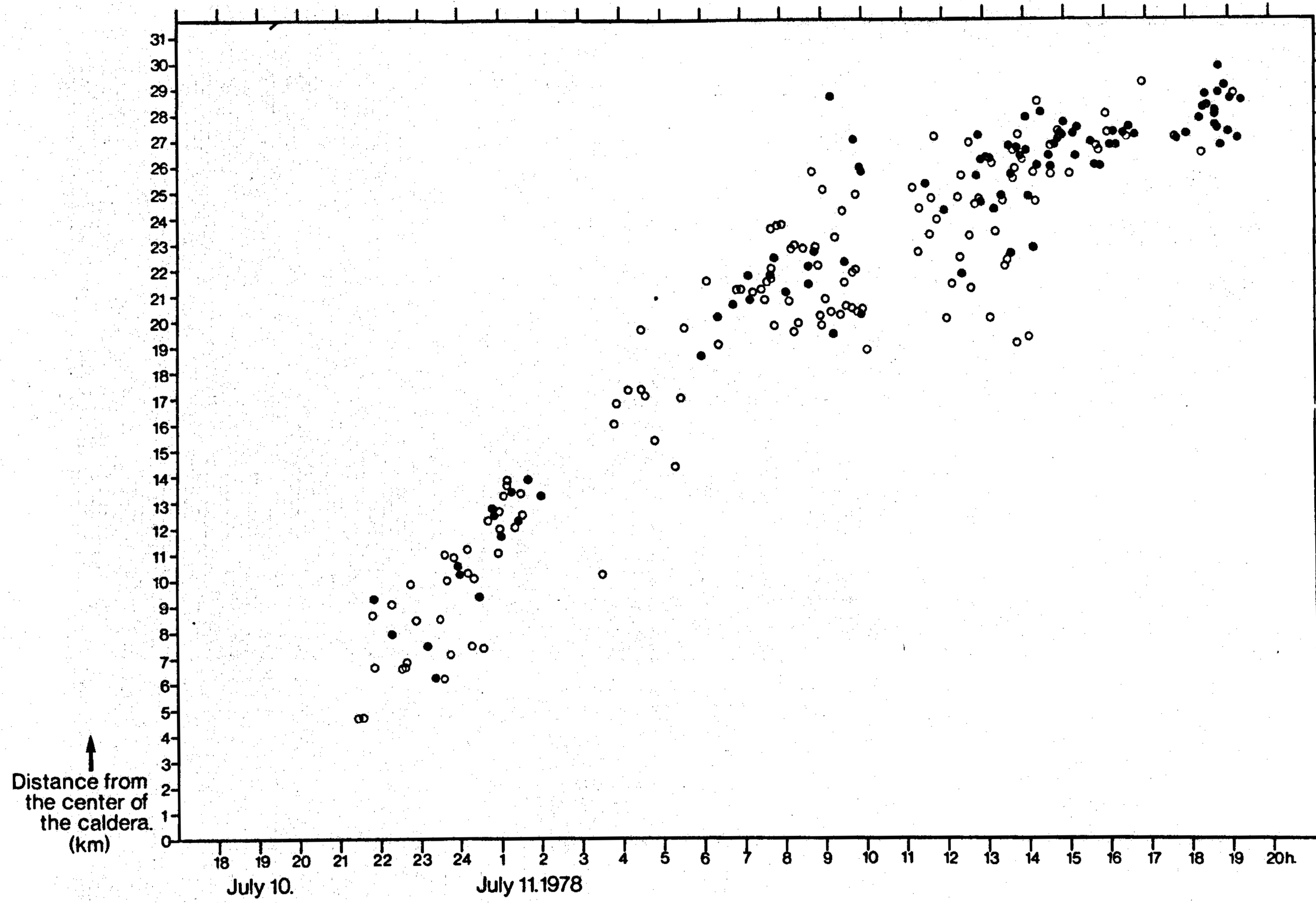

Fig. 5. The migration of seismic activity. The distance of epicenters from the center of the Krafla caldera is plotted as a function of time. Dots and circles have the same meaning as in Fig. 3. The apparent gap in the activity between 10 h and 11 h on July 11 is caused by a time signal failure.

The magnitude-frequency distribution

Magnitudes were obtained for 354 earthquakes from the maximum trace amplitude on the short period seismograms of the WWSSN station AKU at the distance of 65-75 km from the epicentral area. The magnitude data were supplied by Thórunn Skaftadóttir at the Icelandic Meteorological Office. The magnitude distribution is shown in Fig. 6. The log N vs. M distribution is reasonably linear. The negative slope of the curve, or the b-value of the swarm, is 1.7 ± 0.2 assuming a linear relationship. These are a maximum likelihood estimate and 95% confidence limits, respectively.

The main deviation from linearty is a dip in the curve near $M = 3.5$. Similar dip, only more pronounced, was found in the magnitude-frequency curve for the September 1977 earthquakes (Brandsdóttir and Einarsson, 1979), where it was interpreted as the result of mixing two earthquake sequences with different b-values. This interpretation is hardly justified in the present case because of the small number of events larger than magnitude 3.5.

Discussion

It seems to be reasonable to assume that the front of earthquake activity that migrates away from the subsiding Krafla area marks the tip of the dyke that is injected into the fault swarm. Thus the length of the dyke can be found as a function of time. Tryggvason (1978c) has given the tilt at the Krafla power house and the rate of outflow of magma as functions of time for the different subsidence events. With these data it is possible to make some further quantitative estimates of the dimensions of the dyke. The total volume is $37 \times 10^6\ m^3$ and the final length of the dyke is 30 km. The average cross sectional area is thus $1.2 \times 10^3\ m^2$. If the width and the height of

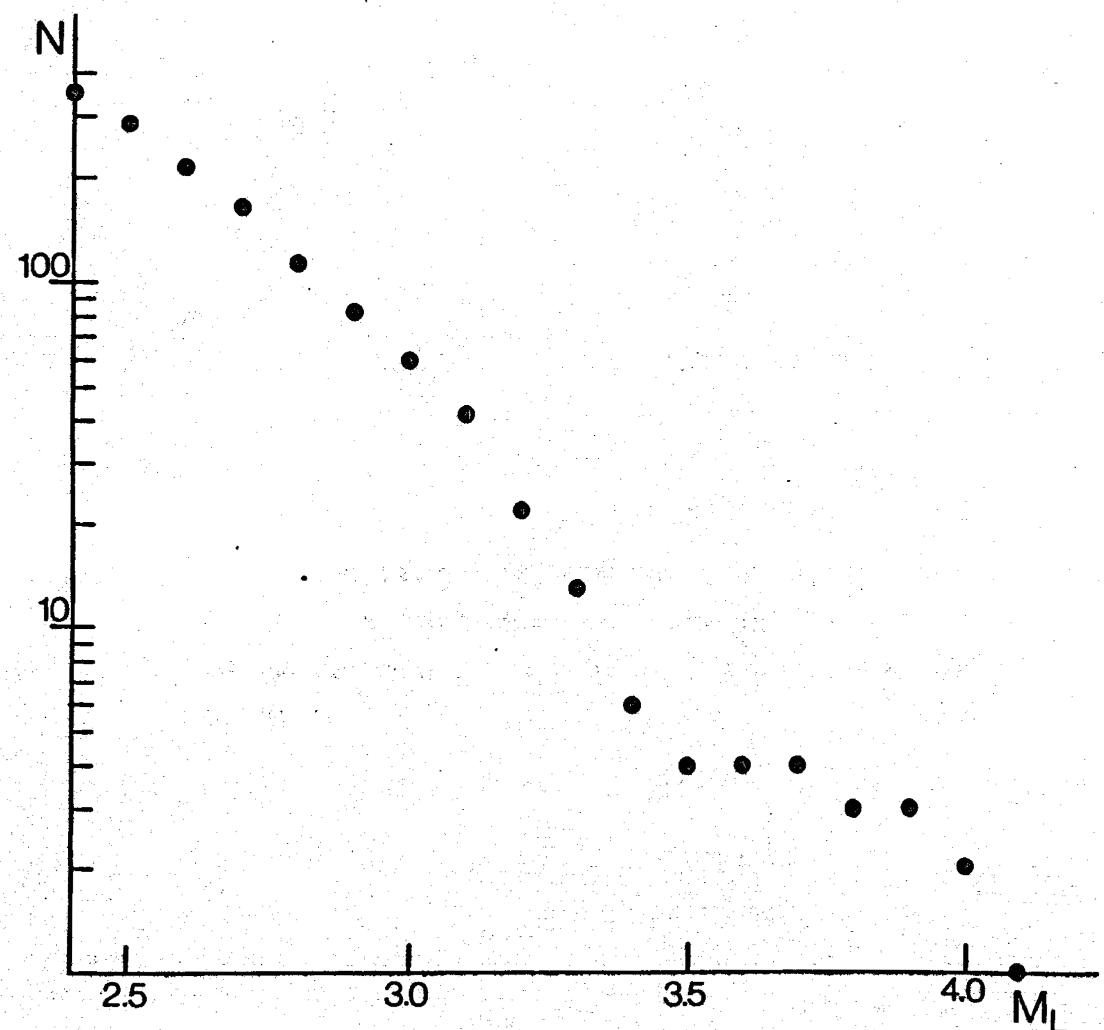

Fig. 6. The frequency - magnitude relationship of the July 1978 earthquakes. The number (N) of earthquakes of magnitude M_L and larger is plotted as a function of M_L.

the dyke are uniform, for example, a 0.5 m wide and 2400 m high or a 1.0 m wide and 1200 m high dyke would have the required cross sectional area. There are indications, however, that the dimensions of the dyke are not uniform along its length. The seismic energy release is clearly not uniform, and the rifting measured in the Snagi area decreases towards the south (Tryggvason, 1978c). Areas of maximum earthquake activity and areas of maximum surface faulting usually coincide. These may be areas where the top of the dyke reaches the smallest depths or where the width of the dyke is largest.

It is possible to show that the average cross sectional area changed as the dyke became longer. From the tilt record of Tryggvason (1978c) the volume of the dyke can be estimated, and thus the average cross sectional area can be found for a given length of the dyke. This is shown in Table 1.

Table 1.

length of dyke (km)	volume of dyke (10^6 m^3)	average cross sectional area (10^3 m^2)
10	12.3	1.23
15	15.6	1.04
20	18.8	0.94
25	22.5	0.90
27	25.2	0.93
30	37.0	1.23

The differences between the numbers are significant, even though the volume estimate is associated with large uncertainties. These changes in the average cross sectional area can be interpreted in a number of ways. If one assumes, for example, that the dyke attains its final cross sectional area immediately behind the tip of the propagating dyke, one can calculate the cross sectional area as a function of distance from the reservoir. Without going into detailed calculations one may conclude that the dyke reached the largest cross sectional area near its northern end and in the area south of Snagi. These are also the areas were the largest earthquakes occurred. It is not possible on basis of these data to say, whether the increased cross sectional area and seismicity are the result of larger height or width of the dyke in these areas.

With the knowledge of the dimensions of the dyke, the rate of flow and a few simplifying assumptions it is possible to derive the relationship between the viscosity and the pressure drop in the dyke. Let us assume that the dyke is a rectangular plate of uniform thickness b, length l and height a. This is a simplifying assumption, since we know that the cross sectional area is not uniform. Let us further assume that the dyke changes its volume by changing its length only. Then only l is a function of time, and $\frac{dl}{dt} = \frac{W}{ab}$ where W is the rate of flow into the dyke. For a Newtonian fluid of viscosity η flowing through a rectangular box of dimensions $a \times b \times l$ ($a \gg b$) it is easy to show that

$$W = \frac{ab^3 \Delta p}{12 \eta l} \quad \text{or} \quad \frac{\eta}{\Delta p} = \frac{ab^3}{12\, l\, W}$$

where Δp is the difference between the pressure in the reservoir and that near the tip of the dyke. We can now tabulate the values of $\eta/\Delta p$ for different lengths of the dyke and different assumptions for the width (Table 2). The values of W are taken from Tryggvason (1978c). The cross sectional area ab used in the calculation is $1200 m^2$.

Table 2.

		$\eta/\Delta p$ (poise/bar)	
l (km)	W (m^3/s)	b = 0.5 m	b = 1 m
10	460	5.4	22
15	300	5.6	22
27	200	4.6	19

From table 2 we see that the change in the pressure difference is relatively small assuming that the viscosity is constant. The rate of flow is therefore primarily governed by the length of the dyke. If a, b, η and Δp are constant, one can solve the differential equation

$$\frac{dl}{dt} = \frac{b^2 \Delta p}{12 \eta l}$$

and find that l is proportional to $\sqrt{t}$, where t is the time from the beginning of the intrusion. This relationship does not fit very well to Fig. 5, which is hardly surprising in the light of our many simplifying assumptions. The fit is not very bad either, and can be improved by assuming that the lateral intrusion started a few hours after the onset of deflation and volcanic tremor.

With the values in Table 2 the pressure difference Δp can be estimated if the viscosity is known. Grönvold and Mäkipää (1978) estimated the viscosity of the Krafla magma to be 200 poise from the chemical composition of erupted material using the method of Shaw (1972). In the calculation the magma was assumed to contain 1% water. The viscosity decreases with increasing pressure and temperature (Kushiro et al., 1976), and the estimate is more likely to be too high than too low. With a viscosity of 200 poise it takes a pressure difference of 40 bar to bring the magma through a 0.5 m wide fissure, and if the fissure is 1 m wide the pressure difference is only 10 bar.

In the model of the Krafla deflation presented in this paper a dyke is formed when the tip of a fluid filled crack propagates through a prestressed medium. The primary driving force of this process is the tectonic stress that has been accumulating on the plate boundary since the last major rifting episode. The mode of strain release depends on the availability of magma. If a magma reservoir is located on the plate boundary a dyke starts propagating away from it when the pressure in the reservoir and/or the regional tectonic stress reach a critical level. The data presented here seem to indicate that the pressure drop in the reservoir associated with the deflation is small compared with the pressure difference between the reservoir and the leading edge of the dyke. The pressure in the magma reservoir probably plays the role of a trigger to initiate the propagation of the dyke. The direction of propagation and the orientation of the intrusion is governed by the regional stress field. The tectonic part of the stress field at the diverging plate boundary in Iceland is likely to be characterized by horizontal tensional stress parallel to the direction of relative plate motion. In this stress field the direction of propagation will be horizontal, and the resulting intrusive body is a dyke oriented perpendicularly to the axis of minimum compressive stress or the maximum tensional tectonic stress.

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