Microearthquake Survey and the Mid-Atlantic Ridge in Iceland[1]

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During the summer of 1967, three high-frequency, high-gain, and highly portable seismographs were operated at seventy-eight sites throughout Iceland. Over 99% of the more than 1000 events recorded were found to lie in nine regions with radii of less than about 5 km. Although most of the events were not greater than 4 km deep, six were of the order of 5 to 15 km deep, and one may have been as much as 40 km deep. One large earthquake swarm was recorded from Mýrdalsjökull in south-central Iceland, where four events less than magnitude 5 were reported by the U. S. Coast and Geodetic Survey (U.S.C.G.S.) in early 1967. No earthquake greater than magnitude 4.5 has been reported since 1958 from the other regions of high microseismicity, suggesting that these microearthquakes were not simply aftershocks. Three events of magnitude 4 to 5 did occur, however, in each of two seismic regions after the initial recording period. Thus, some of the microearthquakes may have been foreshocks. A close correspondence was found between areas of major hydrothermal activity and high microearthquake activity. The highest activity recorded was in the Krafla volcanic region in northeastern Iceland, which has not been active since 1746. This activity had a *b* value of 0.83 ± 0.16 over 1½ units of magnitude. The focal mechanisms were consistently similar and gave a solution with one nearly vertical nodal plane striking north–south. Eight of the nine zones of microseismicity lie on an east–west line near 64°N. When considered in relation to adjoining active seismic zones, reported historic seismicity of Iceland, and the location of the mid-Atlantic ridge and areas of active rifting and volcanism, the existence of a transform fault is suggested, following the methods used by Sykes (1967) to outline such faults on the sea floor with larger earthquakes. Magnetic data and some geologic features support this hypothesis. Seismic refraction data are not in disagreement with it. Many of the tectonic features of Iceland do not, however, readily fit into this framework. If sea-floor spreading is active in Iceland, it is more complicated in detail than previously suggested.

INTRODUCTION

Iceland is the largest supramarine part of the worldwide oceanic ridge system. Over 100,000 km² of volcanic formations up to 20 m.y. in age [*Thorarinsson et al.*, 1960] are exposed and cut by numerous rifts, faults, and fractures. The island is situated at the junction of the mid-Atlantic ridge and the Scotland-Greenland ridge. Within Iceland the trend of the mid-Atlantic ridge changes from the NE–SW strike of the Reykjanes ridge on the southwest to the more northerly trend of the Iceland-Jan Mayen ridge north of central Iceland (Figure 4). This Iceland-Jan Mayen ridge should not be confused with the aseismic Jan Mayen ridge [*Johnson and Heezen,* 1967].

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Figure 1 summarizes the available epicentral data for large earthquakes through 1966; the data are primarily taken from an unpublished list by E. Tryggvason of earthquakes from 1912 to 1960. He considers this list complete for all events greater than magnitude 5¾. These data were augmented by summaries by *Tryggvason et al.* [1958], *Sykes* [1965, 1967], and *Stefánsson* [1967], as well as by data from the *Seismological Bulletin* of Vedurstofa Islands and the *Monthly Seismological Bulletin and Earthquake Data Report* of the U.S. Coast and Geodetic Survey. These data are less complete and less accurate for the earlier dates. Epicenters derived from intensity studies, epicenters given by *Sykes* [1965] from 1955 to 1963, and the more recent epicenters given by the U.S.C.G.S. probably are accurate to about 10–20 km.

During the summer of 1967, a survey of microearthquakes was made in Iceland using three high-frequency, high-gain, and highly portable seismographs. This paper describes the new instrumentation and field methods used, discusses the nine zones of major microearthquake ac-

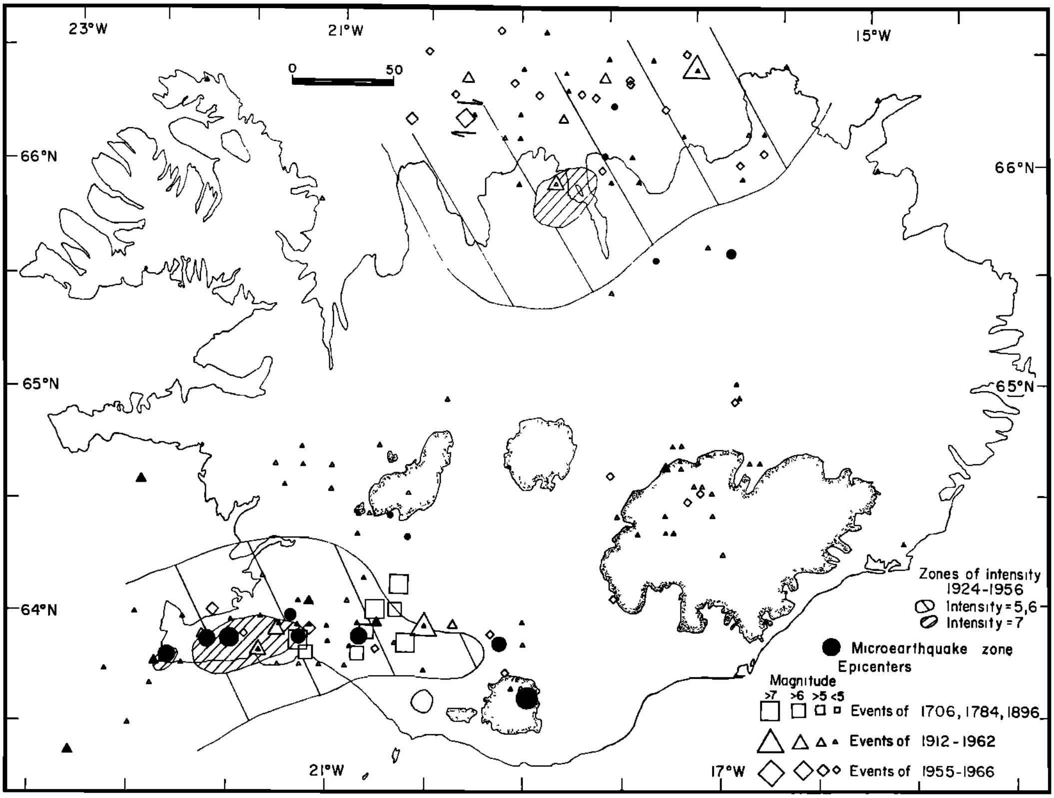

Fig. 1. A summary of the known seismicity of Iceland. Zones of intensity are from *Tryggvason et al.* [1958]. Microearthquake zones discussed in this paper are shown as solid circles. Triangles are epicenters primarily from an unpublished list by Tryggvason. Diamonds are the best determined instrumental epicenters (*Sykes* [1965, 1967] and U.S.C.G.S. determined epicenters). A solid symbol designates an epicenter of several earthquakes. The coverage is by no means complete before 1912 or below magnitude 5¾. Arrows show a focal mechanism given by *Sykes* [1967].

tivity found during the recording period, shows that nearly all the hypocenters are less than 4 km deep, and discusses special properties of these earthquakes. Some larger earthquakes have been reported in most of these zones since 1784. The Mývatn region is found to be a notable exception. In addition, the relative number of microearthquakes from region to region was found to differ from the relative number of larger earthquakes shown in Figure 1. All these seismic data are used to trace the axis of the mid-Atlantic ridge through Iceland. Related geophysical and geological data are summarized in order to point out some problems in a simple application of the hypothesis of sea-floor spreading to the details of Icelandic geology and to make specific suggestions for future research. More general suggestions for geophysical research in Iceland were discussed at the Surtsey Research Conference [*Decker*, 1967].

Field Program

Instrumentation. A high-frequency high-gain seismograph system was developed and packaged under the direction of P. Pomeroy at Lamont. Each system was in an 0.04-m^3 aluminum suitcase mounted on a packframe and weighing only 28 kg. A Dayton 4.5-cps geophone drives an Electrotec SPA-1 amplifier whose output impedance is matched to a Brush pen motor by a small, Lamont-made, 0 gain, power amplifier. The pen writes on a piece of glazed paper 21.6 × 28 cm that is smoked and mounted in a drum recorder built to specification by Sprengnether Instrument Company and driven by a Zener diode regulated dc motor. Time marks (GMT) were added from a Bulova Accutron chronometer, calibrated daily with radio time (BBC London). The whole system draws about 30 ma at 24 volts and is quite suitable for field use both in ease of operation and ease of repair. The amplifier was set with a 30-cps low-pass filter. Figure 2 shows the absolute magnification as a function of frequency calculated from the measured total response of the amplifiers and pen motor multiplied by the manufacturer's sensitivity curve for the geophone. This gives a maximum gain for the system of 78 million at 26 cps. It is believed, however, from tests at Lamont (G. Boucher, personal communication, 1966, and M. L. Sbar, personal communication 1968) that the sensitivity of the geophone may be lower by a factor of 3, giving a maximum gain of about 26 million.

The time resolution, as discussed below, is not adequate for distinguishing high frequencies. If however, it is assumed that very local earthquakes contain sufficient high-frequency spectral

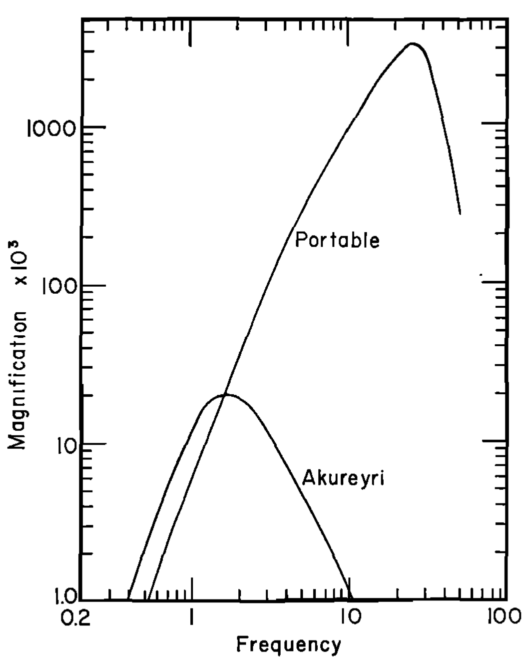

Fig. 2. Ground amplitude magnification of the suitcase seismograph used in this study assuming the lower geophone sensitivity (see text). The magnification of the World-Wide Standardized Seismic Network short-period instrument at Akureyri is shown for comparison.

components, the *Richter* [1958, p. 342] magnitude can be calculated. Earthquakes recorded with 2-mm amplitudes when the maximum gain is attenuated 30 db with *S–P* times less than 2.5 sec would then have a magnitude of approximately −0.7 if the suggested lower geophone sensitivity is used, or −1.0 if the manufacturer's specifications are used. The smallest earthquakes in this study would then have a magnitude of the order of −2.

The recorder was normally operated with a drum speed of 25 mm/min and could run unattended for one day. Drum speeds of 50 and 100 mm/min were also used when greater time resolution was required. With care, records could be read to 0.1 mm, giving timing precisions of 0.24, 0.12, and 0.06 sec for drum speeds of 25, 50 and 100 mm/min, respectively.

Data collection. To survey the general microseismicity of Iceland and to locate regions of activity, three seismographs were normally placed 20 to 30 km apart at the apexes of a triangle. Each day either one or two units were moved, so that each site was normally occupied for two days. Visible records were essential to the operation of the field program. Each day the records were examined, events roughly located, and new sites for the instruments were then selected. When an area of interesting activity was located, more closely spaced sites were occupied.

Data analysis. *P* arrival time, *S–P* time, and *P* and *S* amplitudes were read for all events recorded. Although records with a few events could be completely analyzed in the field, records with several hundred events were read later with the aid of a digital computer. The record was taped to a tabletop digitizer and, by placing the pointer at the appropriate point, *x,y* coordinates were automatically punched for the edges of the record, key minute marks, *P* and *S* arrivals, and *P* and *S* amplitudes. The computer was then programmed using basic analytical geometry to find the arrival time, including the clock correction, the *S-P* time, and the amplitudes. These data could be read accurately to 0.1 mm. In this way 300 events could be read completely in an hour, and the interpreter could concentrate on the types and shapes of the earthquakes without wasting time measuring, interpolating, and recording. The data were then in a form that could readily be used for computing statistical parameters, hypocentral locations, etc.

The number of events recorded at each site was counted. To compare sites, amplitudes were roughly normalized to 30 db by multiplying 2 raised to the (attenuation −30)/6 power. It is hard to miss events with amplitude 1 mm, and, since 36 db was generally the lowest gain used, events with amplitudes 2 mm at 30 db were considered to form a complete set. The number of such events was then divided by the time that the records at the appropriate site had noise <0.5 mm to obtain the number of earthquakes per day recorded at each site. Table 1 summarizes the data from the seventy-eight sites, and Figure 3 shows the location of the stations and the number of local events per day properly normalized. It is important to note that more than half of the events recorded had amplitudes less than 2 mm at 30 db (Table 1) and are therefore not shown in Figure 3.

The number of events per day calculated in this manner was found to be quite characteristic of each site. Any number over 0.5 event/day

TABLE 1
Events per day were calculated with an amplitude ≥2 mm at 30 db.

Station	Events/Day: S–$P \leq 2.5$ sec	Events/Day: S–$P > 2.5$ sec	Events Recorded	Hours with Noise <0.5 mm	Attenuation, db	Date First Occupied
1	0	0	0	3.0	30	June 15
2	0	2.7	9	62.5	30	June 15
3	0	0	0	38.5	36	June 15
4	2.4	1.0	12	70.5	42, 30, 36	June 15
5	0	0	2	65.6	30, 24	June 18
6	0	3.3	5	36.0	30	June 18
7	0	0	0	24.5	36	June 19
8	1.4	1.4	13	33.6	18, 24	June 20
9	0.7	1.4	3	34.0	36	June 21
10	0.6	0.6	7	41.9	30	June 21
11	12.7	1.2	41	39.5	30	June 22
11	15.5	0	27	17.0	24	July 16
12	22.9	0	149	56.5	18, 24	June 23
12	13.6	8.0*	72	30.0	18, 24	July 24
13	5.3	2.7	17	18.0	36	June 23
14	22.6	4.0	30	18.5	36	June 24
15	0.6	0.0	10	38.0	30	June 24
16	2.0	2.7	7	36.0	36	June 25
17†	···	···	0	2.0	60	June 26
18	0	1.7	3	14.5	30	June 29
19	0	0	1	5.0	12	June 30
20	0	0	6	20.0	12, 18	July 1
21	0	0	1	16.5	18, 24	July 1
22	0	0	4	21.5	12	July 1
23	5.1	0	77	84.5	12, 18, 24	July 2
24	0	0	21	67.5	30, 18	July 2
25	0	0	3	32.0	18	July 3
26	86.0	5.1	69	9.5	24	July 5
27	0	0	0	10.0	24	July 6
28	0	0	1	8.0	18	July 6
29	0	0	3	26.0	24	July 6
30	0	0	0	13.5	12	July 7
31	0	0	0	13.5	12	July 7
32	0	0	0	9.0	18	July 7
33	0	0	0	12.0	12	July 9
34	0	0	0	6.0	18	July 9
35	0	0	0	7.5	12	July 9
36	2.8	4.5‡	81	59.0	18	July 10
37	0	1.8	3	13.0	24	July 11
38	0	0	0	5.0	18	July 11

TABLE 1 (*continued*)

Station	Events/Day $S–P \leq 2.5$ sec	Events/Day $S–P > 2.5$ sec	Events Recorded	Hours with Noise <0.5 min	Attenuation, db	Date First Occupied
39	0	0	2	16.5	30	July 11
40	0	9.0‡	20	32.2	18	July 12
41	0	3.9	7	18.5	30	July 12
42	0	0	0	20.8	24	July 13
43	0	0	0	17.4	30	July 13
44	0	0	0	8.0	24	July 13
45	0	1.9	10	25.0	12	July 14
46	1.3	0	7	37.8	30	July 14
47	0.7	0.7	11	36.5	12	July 14
50§	64.3	0.6	216	40.3	30	July 19
51	0	0	1	21.0	30	July 19
52	0	1.8	6	13.0	24	July 19
54‖	191.0	0.0	1192	49.7	24	July 20
60	12.0	9.0	40	32.0	30	July 27
61	10.4	5.2	47	37.0	30, 24	July 29
62	2.6	0.9	11	54.8	30	July 31
63	5.3	1.0	34	49.9	24, 30	Aug. 3
64	10.3	2.0	41	36.0	30	Aug. 5
65	9.6	2.1	36	35.0	24	Aug. 8
66	12.8	6.1	55	39.5	24	Aug. 10
67	5.3	0.0	10	45.4	36	Aug. 12
68	14.7	3.9	108	44.5	30, 24	Aug. 15
69	5.2	1.9	21	37.0	30	Aug. 17
70	13.4	0.6	77	39.4	30	Aug. 19
71	5.8	5.8	53	40.9	30	Aug. 22
72	0.0	1.1	6	43.2	30	Aug. 29
73	0.8	5.1	11	28.0	30, 36	Aug. 31
74	16.9	2.8	84	42.5	30	Sept. 2
75	22.0	6.0	86	36.0	30	Sept. 4
76¶	25.0	3.1	44	23.0	30	Sept. 6
77	15.4	0	45	31.5	30, 36	Sept. 8
78	14.3	2.6	16	18.5	36	Sept. 10

* Hestfjall earthquakes.
† Too noisy.
‡ Swarm only.
§ Stations 48 and 49 at geothermal well in Hveragerdi.
‖ Stations 53 and 55–59; see text.
¶ Small swarm, 11 events in 1 hour.

did not vary from record to record by more than a factor of 3 and usually much less over a period of at least a month. Site 11 was occupied on two occasions 24 days apart; 12.8 and 15.5 events/day ($S–P \leq 2.5$) were recorded. Site 12 similarly occupied 30 days later gave 14.3 as opposed to the original value 22.9. Swarms occur, as in the case of events with $S–P > 2.5$ at stations 36 and 40 or the local events at sites 67 and 76. Nevertheless, these swarms can easily be identified as the recording interval increases.

The number of events per day normally increases as one approaches the source of events, so that these values plotted on a map give a rough idea of the areal seismicity, at least during the month involved. One must use this approach with care because some sites are more sensitive to local activity than others. Stations 23 and 26 were both 9 km from the earthquake source at Krafla in northern Iceland, but during the same period 2.4 events per day were recorded at station 23, while 84.0 were received at station 26. Site 23 was at the top edge of a 20-meter scarp facing across a graben toward the source, and site 26 was on a fresh 200-year-old lava

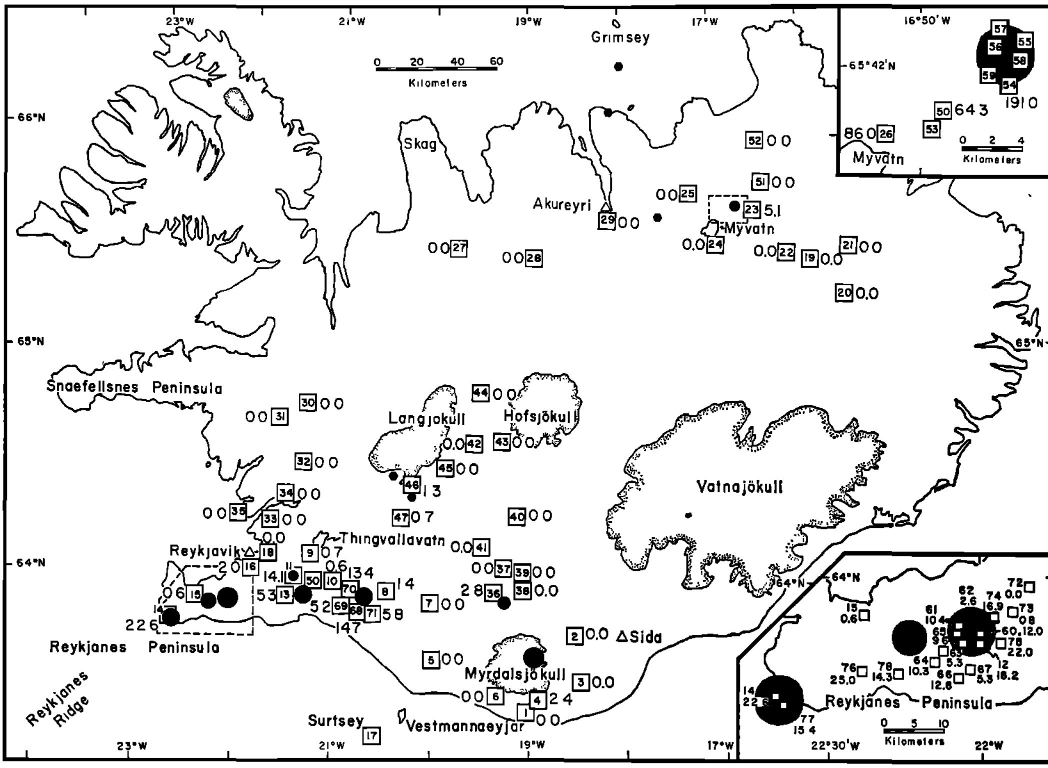

Fig. 3. Location of sites occupied by portable seismographs. Each site number is plotted inside the appropriate square. The number of events per day with amplitudes ≥ 2 mm at 30 db and *S–P* times ≤ 2.5 sec is plotted beside each square. Inserts show data for the more closely spaced stations on the Reykjanes Peninsula and the Mývatn area. Jökull at the end of a name means glacier in Icelandic.

flow that came from a vent near the earthquake source. In all other cases sites at equivalent distances had the same number of events within a factor of 3.

When three instruments were used simultaneously between June 15 and July 23, *S–P* times were used to locate as many events as possible, assuming an *S–P* velocity of 8 km/sec. *S–P* times were usually less than a few seconds, so that the observed variations in *S–P* velocities have little effect on the locations. These very short *S–P* times also place a limit on the maximum depth of the events. All but 9 events with amplitudes ≥ 2 mm at 30 db appear to lie in the nine regions shown in Figures 1, 3, and 4. The size of the regions is primarily a function of the uncertainty in locations due to station distribution.

No consistent source of earthquakes larger than magnitude 1 is likely to have been missed throughout this survey of Iceland. Sporadic or low-activity sources of earthquakes similar to many of those discussed below may, however, have been overlooked in central, eastern, and northwestern Iceland. All the microearthquakes fall in the zone of active volcanism and rifting. However, there is a notable lack of seismic activity in the northern part of the western limb and in the less well surveyed central part of the eastern limb of this zone.

Zones of Microearthquakes

Region 1 Mýrdalsjökull. Events were located in Mýrdalsjökull (Figure 4) from stations 1 through 8 and 37 through 41. The activity seems to fall within a circle centered at 18°56.2′W, 63°41.4′N. Because of the distance of the recording sites, the radius of the circle could be as large as 5 km. The center of this circle lies 10 km north of the volcano Katla (Figure 4). A magnitude 4.8 event with a depth of 2 km was reported from this area on April 1, 1967, by the U.S.C.G.S. In addition, magnitude 4.2 to 4.5 events were reported on March 11, May 16, June 7, and October 4, 1967, near this region. Although only 7 to 16 distant stations reported these particular events, the relative arrival times show that these events probably should also fall within the given circle. Figure 1 shows that this region has been an important source of earthquakes in the past. Preliminary readings of the Icelandic seismic stations in recent years (provided through the kindness of R. Stefánsson of the Icelandic Meteorological Service, Reykjavik) show several events per month recorded from this area, particularly since September 1966.

On July 12 from 1407 to 1422 hours, ten events from this region were recorded at stations 36 and 40. Eight were large enough to be recorded at Reykjavik. These events, plus one smaller event (at 2234 hours), were the only events from the Mýrdalsjökull area recorded during 59 hours at station 36. Swarms have often been associated with volcanic activity [*Eaton and Murata*, 1960; *Minakami*, 1960]. *Tryggvason* [1960] has discussed earthquake swarms from this region that are apparently related to 'jökulhlaups' or glacial bursts. These bursts are flash floods released from under a glacier and are believed in some cases to be caused by melting of the ice by volcanic activity.

Region 2 Torfajökull. The events from this area were recorded only at station 36 with *S–P* times of 0.7 to 1.0 sec. All 48 events were small and lie within a circle 8 km in radius centered about station 36. However, since none of the events were recorded at stations 37 through 39, they are interpreted as coming from south of site 37 and probably from the Hrafntinnusker thermal area, which is centered near 19°12.6′W, 63°56.0′N, and is surrounded by a circle with a radius of about 3 km.

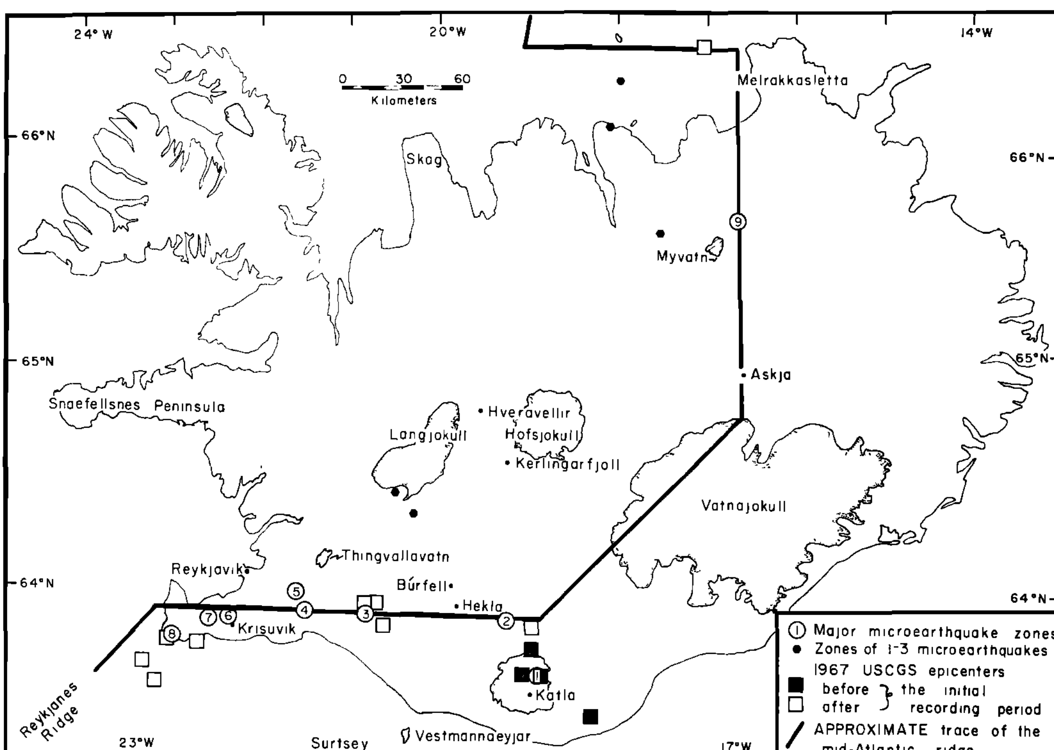

Fig. 4. The nine zones of microearthquake activity numbered as in the text. The solid line shows the *approximate* center of the active zone of the mid-Atlantic ridge as proposed in this paper. (Vatn is the word for lake in Icelandic.)

Region 3 Hestfjall. Several small events were recorded from this region in June from stations 8 and 10. On July 27 a magnitude 5 event was reported from here by the U.S.C.G.S. (Figure 4). This event was followed by magnitude 4.6 and 4.7 events on July 28 and 29, respectively. Stations 68 through 71 were operated from August 15 to August 23. The events recorded at sites 69 and 8 had the same *S–P* time, but the increase in activity after the larger events was only slightly more than a factor of 3, which as shown above is no more than expected from recording at different times and stations. Figure 1 shows some large events from this region, primarily prior to 1912. Only one earthquake of magnitude 3.1 was reported in 1967 (R. Stefánsson, personal communication, 1967). Therefore, the microearthquakes recorded in June might be interpreted as foreshocks of the larger events or as the first sign of a general increase of activity in this area. At least nine foreshocks from this area were recorded at site 12 on the day before the main earthquake on July 26. Numerous aftershocks were recorded but an increase in the background noise limits the accuracy of any count.

All the microearthquakes recorded from this region could be from within a circle 4 km in radius centered at about 20°39.6′W, 63°58.2′N. The June events appear to be primarily from the eastern half of the circle, whereas the July events are apparently from the western hemicircle. *S–P* times show that all depths must be less than 4 km. The three teleseismically located earthquakes lie within 5 km of the edge of the circle, the best located one lying on the northwest perimeter.

R. Stefánsson (personal communication, 1967), using the Icelandic stations, has pointed out a curious interplay between seismic activity in this area and the activity just south of Grímsey, off the coast of north-central Iceland. These two regions are interpreted below as transform faults offsetting the ridge crest in Iceland. Between July 26 and July 30 four bursts of earthquakes in Hestfjall were preceded from 1 to 6 hours by a shorter burst of slightly smaller earthquakes south of Grímsey. No exception was noted.

Region 4 Hjalli. Several events were recorded at stations 8 through 13 from an area of 4-km radius centered about 21°18.1′W, 63°57.6′N. Figure 1 shows one or two events less than magnitude 4 between 1956 and 1962 and two magnitude 5 to 7 events before 1910. A major scarp runs from this region northeast several kilometers to Hveragerdi. Prominent rifting and fresh tension fractures are exposed a few kilometers west and northwest of this area.

Region 5 Skardsmýrarfjall. Several events were recorded within 2.5 km of station 11 (21°22.8′W, 64°02.8′N). This station is on the southwest edge of the Hengill thermal area.

Regions 6 and 7 Trölladyngja. The second most active source of earthquakes recorded was northwest of Krisuvik on the Reykjanes Peninsula. These events were generally less than 4 km distant from sites 60, 61, and 65 and seem to come from an area 4 km in radius and centered on the surface near 22°03.6′W, 63°56.8′N. In addition, most events recorded at sites 15, 76, 78, and 64 had *S–P* times that gave a distance of 7 to 10 km, suggesting another source west of *Keilir* within a 3-km circle centered about 22°15.6′W, 63°56.0′N. Although from the *S–P* times it does not appear likely, it is nevertheless possible that these two sources are, in fact, one elongated source.

Major high-temperature thermal activity occurs at Trölladyngja within the first of these microearthquake zones, as well as at Krisuvik just south of this zone.

Region 8 Reykjanes. Another zone of high activity was near the lighthouse on Reykjanes at the southwest tip of the Reykjanes Peninsula. The events from this area lie within a hemispheres, 4 km in radius, and centered about 22°41.4′W, 63°49.4′N. Historically, this region has been noted for its seismic activity. A high-temperature thermal area occurs within this zone. Recent signs of tectonic and seismic activity have been observed here. In 1966, J. Jónsson (personal communication, 1967) noted an increase in this thermal activity. Infrared imagery was obtained in the summer of 1966 over this area by the Air Force Cambridge Research Laboratories and the U.S. Geological Survey in collaboration with the Icelandic National Energy Authority [*Friedman et al.*, 1968]. The imagery indicated that some faults in the area may have been warmer than expected from surficial observation at that time.

Starting late on September 28, 1967, and continuing for 14 hours, more than a hundred earthquakes were recorded in Reykjavik from the area northwest of Krisuvik (R. Stefánsson, personal communication, 1967). The largest event had a magnitude of 4.2 and occurred at 2222 on September 28, but several shocks were felt throughout the central part of the peninsula. Early on September 30, 16 strong shocks were felt for the first time at the lighthouse, cracking the tower and breaking dishes. Four poorly located 4.3 and 4.4 events were reported during this time. The western side of a fault near the lighthouse (Figure 3) dropped 5–8 cm relative to the eastern side; R. Stefánsson (personal communication, 1968) detected an equivalent amount of horizontal movement. Many new cracks formed and emitted steam. Old and new hot springs erupted water up to 15 meters high.

The thermal activity remained higher than normal for several months and slowly new springs formed to the northeast. Thus, there was an apparent progression of earthquake activity from northeast to southwest, culminating in visible fracturing, changes in the thermal activity, and later a slow progression of thermal activity northeast along the newly formed fracture. The microearthquake survey before these events indicated high activity in this region.

It should be noted, although it may be coincidental, that the larger earthquakes in 1967 began in Mýrdalsjökull, progressed westward to Hestfjall and finally to the Reykjanes Peninsula.

Region 9 Krafla. The source of earthquakes near Krafla was the only consistent source found in northern Iceland. It had the highest activity recorded during the summer by a factor of 8. For these reasons, these events were studied in greater detail.

Figure 1 shows that no large earthquakes have been reported in the literature from this region. Only one magnitude 3.8 event in 1953 is known to have come from the Mývatn area (E. Tryggvason, personal communication, 1968). Here, as at Hestfjall, it is obvious that these microearthquakes are not simply aftershocks, as suggested by *Oliver et al.* [1966], but are instead a unique measure of seismic activity.

Three instruments were run within 2 km of the suspected source area for a day. Fourteen events were clearly recorded on all three instruments. Their *S–P* times were each read several times, but all readings of the same event agreed within 0.1 mm, or 0.06 sec. The hypocenters were then calculated. Figure 5 shows that epicenters falling within the triangle of stations group quite well, whereas the epicenters outside the triangle are more scattered. If the *S–P* time for these latter events at one particular station was changed by the possible reading error (0.06 sec) in the direction most suitable for clustering the events, all epicenters would be within the plotted circle. It is evident that, although the precision of the epicentral computations for events within the triangle is about ±0.5 km, the solutions become very unstable for epicenters outside the triangle of stations. However, it is clear that all these events came from an area encompassed by a circle probably much less than 2 km in radius with an origin at 16°44.1′W, 65°42.2′N.

The hypocenters were calculated assuming an *S–P* velocity of 8 km/sec. Since the maximum *S–P* time was 0.4 sec, a velocity of 6 km/sec would change the length of each ray no more than 0.8 km. Using the 8-km/sec velocity, depths of the events ranged from 0.7 to 2.1 km below sea level (surface elevation 0.5 km). An *S–P* velocity of 6 km/sec gives depths from 0.0 to 1.0 km. Assuming all events from the Krafla region occurred at the center of the circle defined above, the *S–P* times measured at different

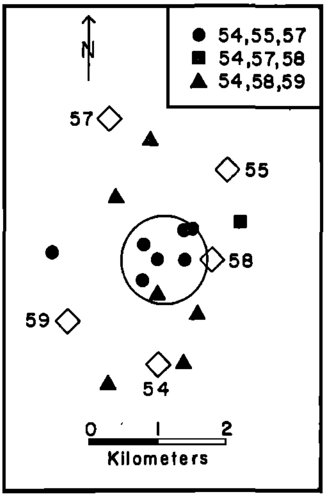

Fig. 5. Location of stations (diamonds) and epicenters used for the focal mechanism study around Krafla. The legend shows the three stations used for locating different events.

stations in northern Iceland gave a velocity of about 8 km/sec up to 65 km away.

Figure 6 shows a compilation of the three first motions recorded for each of the fourteen events located, as described above, in the Krafla region. Assuming a straight ray path for the maximum of 3 km from source to receiver and assuming that all events have the same radiation pattern, we plotted the first motions on the upper focal sphere. One nodal plane dipping about 80° to the east and striking N 5°E is fairly well defined. The other nodal plane could then dip 40° to the north and strike N 85°E. The first nodal plane would imply thrust faulting with some strike-slip component, whereas the second and less well defined plane gives east–west strike-slip motion with some thrusting. The numerous N 10°E striking fissures and grabens in the region favor the first nodal plane. The intermediate stress would be along the rift zone.

About 180 meters east of station 56 at Krafla and within the epicentral zone is a steam vent and small solfatara field. The vapor discharge is apparently continuous and can be seen from at least 5 km away. There is an explosion crater on the west flank of Krafla that was formed on May 17, 1724 [*Thorarinsson et al.*, 1960]. This

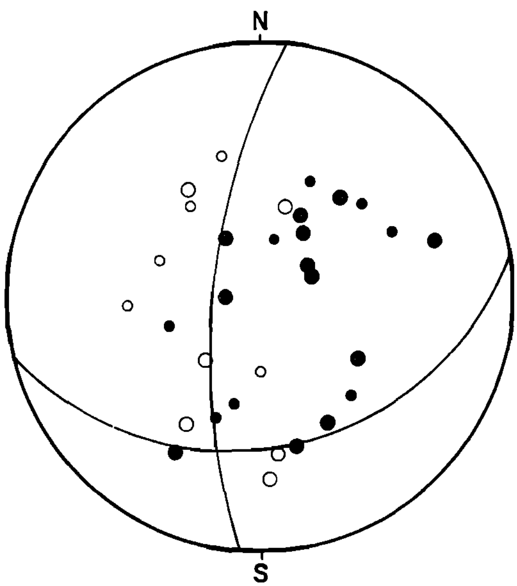

Fig. 6. First motion of the microearthquakes located in Figure 5 and plotted on the upper focal sphere of an equal-area projection. Three points correspond to one event. Solid circles are compressions; open circles are dilatations.

event heralded the beginning of 5 years of volcanic activity characterized primarily by effusive eruptions from a 12-km-long north–south fissure 3 km west of Krafla. No extrusive volcanic activity has been detected in this area since 1746.

Figure 7 shows a record from station 54 on July 21, 1967. Besides the numerous earthquakes and the noise in the middle of the record from an automobile, there is a sporadic background noise with predominant frequency around 3 cps. Such noise was recorded at a few sites in the Mývatn area only and bears much similarity to very small sporadic bursts of harmonic tremor recorded with the same instruments on Kilauea volcano in Hawaii in 1967.

Other local earthquakes recorded. Six events were well recorded at stations 45, 46, and 47 south of Langjökull. Three had the same *S–P* times and were located 3 km south-southeast of station 46 at a depth of 5 km. The other three events also had consistent *S–P* times and were located 11.5 km northwest of station 46 at 14-km depth. From the distribution of stations used, the depth of the first group is considered far more accurate than that of the second.

One event recorded at stations 23, 24, and 25 was located 31 km west of site 25 at a depth of 40 km. At this distance, however, the epicenter and particularly the depth are very unreliable. Two other events were roughly located south of Grímsey. All these events are shown in Figure 3.

Distribution of amplitudes. On July 23 three seismographs were run side by side at location 54 with their attenuations set at 18, 30, and 42 db. In 9 hours over 300 events were recorded. The number of events with P and S amplitudes of a given value were counted in integer amplitude classes from $n - 0.5$ to $n + 0.4$ mm. The S waves normally had the maximum trace amplitude. The minimum amplitude used was 2.5 mm, so that there could be no question of not counting any event of this size or larger. A maximum amplitude of 8.4 mm was used because the amplitude response of the seismograph was not linear above this value. For normalization all amplitudes recorded at 30 db were multiplied by 4; those at 42 db, by 16. Thus by using three instruments simultaneously at the same site, the dynamic range of recording was extended to cover 1.6 units of Richter mag-

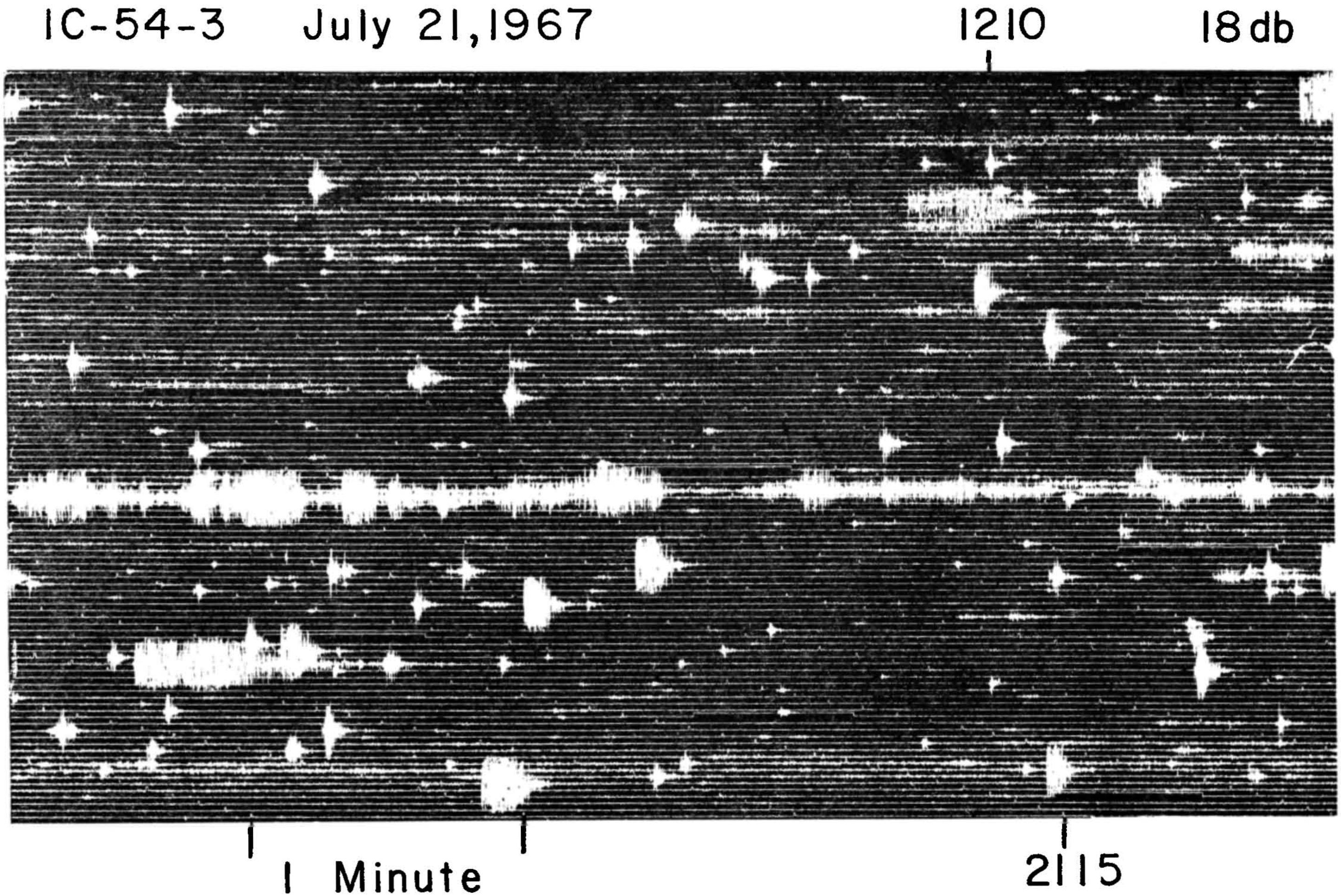

Fig. 7. Record from station 54 near Krafla. The noise for several minutes in the center of the record is from an automobile. The microearthquakes are generally distinguished by their impulsive onsets. *S–P* times are about 0.3 sec. Many short bursts of noise that were peculiar to this region of Iceland can be seen.

nitude. The logarithm of the number of events greater than, or equal to, a given amplitude is plotted against the logarithm of the amplitude in Figure 8.

Page [1967] has discussed in detail the problems involved in fitting a straight line to these points to obtain the b value in the *Gutenberg and Richter* [1949] relationship modified by *Suzuki* [1952].

$$\log N = -b \log A + C$$

Page derived a computer program to find the maximum likelihood estimate of the slope (b) and to assign 95% confidence limits. Applying his method to the data in Figure 8 gives $b = 0.84 \pm 0.29$ for P amplitudes and $b = 0.83 \pm 0.16$ for S or maximum amplitudes. The lines with these slopes are plotted in Figure 8. All other data recorded were examined in this manner. In these cases, however, the dynamic range of the recordings was small and fewer events were recorded, so that the 95% confidence limits were sometimes as large as ± 1.0. Nevertheless, no b value was found that did not contain 0.83 within the confidence limits.

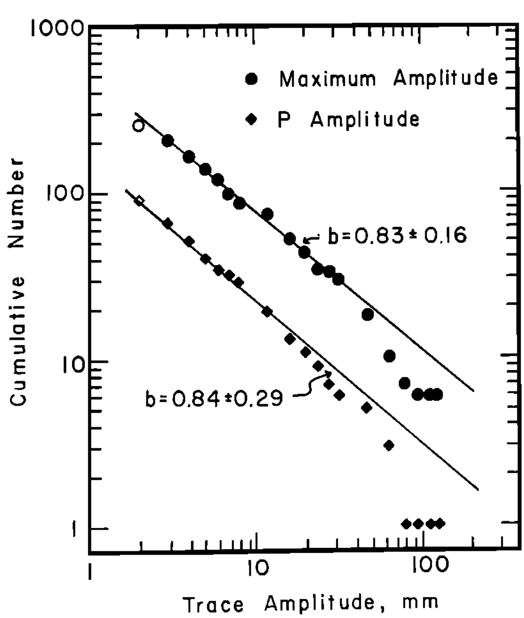

Fig. 8. Amplitude-versus-number relation of microearthquakes at Krafla. The maximum amplitude is normally the S amplitude. The lines are fit by a method of maximum likelihood [*Page*, 1967].

Isacks and Oliver [1964] and *Page* [1967] have discussed variations in b values reported by different authors for different regions. They found that b lay between 0.8 and 0.9 and that any differences reported were not distinguishable from errors in measurement over a magnitude range from -2.0 to $+8.5$. Our data, which are in the lowest magnitudes of this range, agree with their conclusions. *Francis* [1968] has discussed evidence that b values of this order are related to fracture zones and that significantly higher values (up to 3.6) are related to median rifts. The data from Krafla would not agree. We have no well determined b values from fracture zones.

Microearthquakes and hydrothermal activity. A comparison of Figures 1, 4, and 9 shows a fair correlation between major thermal areas and regions of high seismicity. The high-temperature areas in southwestern Iceland, which are nearest to the center of population, have been noted for centuries for their seismicity. Frequent small shocks have been felt by the inhabitants, but they rarely have affected the thermal activity. There have also been many shocks greater than magnitude 4, similar to the shocks at the Reykjanes lighthouse in 1967. These events were accompanied by the opening of fissures, steam explosions that formed mud craters, and profound changes in the loci of thermal activity.

Thermal activity is widespread in Iceland. On the basis of measured or estimated subsurface temperature, the thermal areas have been divided into two groups, the high-temperature areas and the low-temperature areas [*Bodvarsson*, 1961]. The high-temperature areas are characterized by large areas of hot ground and steam vents and a high degree of thermal alteration. The subsurface temperature exceeds 200°C at a depth of a few hundred meters. The low-temperature areas issue water at temperatures up to the boiling point. Subsurface temperature in such areas has been found to be up to 150°C.

The devision into high- and low-temperature areas is by no means unambigous. It is estimated that about fourteen thermal areas can be classified as major high-temperature areas. They are all confined to the neovolcanic zone or zone of active volcanism and rifting. Low-temperature activity is found mainly outside this zone in northern, western, and southern Iceland.

Microearthquake zones coincide with high-temperature areas at Reykjanes, Krisuvik, Hengill, and Krafla. On the other hand, the high-temperature areas at Kerlingarfjöll and Hveravellir did not show microearthquake activity as far as can be judged from stations 43 and 44. The microearthquakes are generally of small focal depth, less than 4 km. The hot-water circulation is usually considered to take place in the uppermost few kilometers of the crust [*Bodvarsson*, 1961; *Pálmason*, 1967*b*]. These observations, together with the areal coincidence of microearthquake zones and thermal areas, suggest a close correlation between these two phenomena. Fracturing, related to the microearthquakes, may increase the rock permeability so as to promote water circulation and thus give rise to the thermal areas. Other correlations can also be envisaged. The present survey indicates, however, that not all high-temperature thermal areas had microearthquake activity associated with them. It should be kept in mind that the thermal areas are by their nature very slowly changing phenomena but that there are not yet sufficient data available to state the same about the microearthquake activity. When averaged over a long period of time, the correlation may be closer than the present results indicate.

Studies of microearthquakes in California [*Brune and Allen*, 1967] likewise indicate a possible correlation between thermal areas and microearthquake zones.

Summary of data collected. The use of high-gain, high-frequency, highly portable seismographs has proved an effective way of studying the seismicity of Iceland. Large numbers of events undetected by permanent stations may be recorded. In this study we have shown that over 99% of the microearthquakes in the parts of Iceland surveyed can be located in nine zones less than 5 km in radius. Some minor source areas may have been missed in the unsurveyed areas. Although the precise location and number of microearthquake zones may in time change, the distribution of microearthquakes during these months was found to agree in a general way with the historic seismicity (Figure 1). There were, however, several important differences.

1. Eight microearthquake zones define an east–west line in southern Iceland, which is much narrower than the region defined by the less well located historic seismicity.

2. The most active microearthquake region at Krafla (region 9, Figure 4) had no reported large earthquakes and therefore had no previously appeared active.

3. Although no large earthquakes had been reported from the Hestfjall area (region 3) since the nineteenth century, microearthquake activity was recorded a few weeks before three such events. For these events at Hestfjall, as well as at Krafla, it is clear that not all microearthquakes are simply aftershocks of larger events; many are a unique measure of seismic activity.

4. This study has shown that nearly all microearthquakes recorded in Iceland were less than 4 km deep; only a few south of Langjökull are well documented as being from 5 to 15 km deep. It would therefore appear that most small earthquakes in Iceland are at a very shallow depth, as has been previously assumed for larger earthquakes [*Stefánsson*, 1967].

5. We have documented an apparently close but not direct relationship of major thermal areas to microseismicity.

Trace of Mid-Atlantic Ridge through Iceland

Earthquakes and transform faults. *Sykes* [1967] has clearly demonstrated that the strike-slip focal mechanisms of earthquakes in fracture zones of the mid-oceanic ridges are consistent with the motion predicted by the transform-fault hypothesis [*Wilson*, 1965] and are opposite to the motion expected for a simple offset of the ridge. Furthermore, he found that focal mechanisms of events on the ridge but away from the fracture zones are characterized by a large component of normal faulting with an inferred axis of tension approximately perpendicular to the ridge. Finally, he demonstrated that seismic activity on fracture zones is confined almost exclusively to the region between the two crests of the ridge. The epicenter data in Figures 1 and 4 can be interpreted, following Sykes' observations, to trace the mid-Atlantic ridge through Iceland.

Sykes [1965, 1967] had pointed out the east–west striking epicentral zone north of central Iceland near 66.4°N. With the aid of one strike-slip focal mechanism solution and the relationship of known ridges and rifts, he has interpreted this as a transform fault. *Tryggvason et al.* [1958] have already pointed out that the larger historic earthquakes fall in this zone and a similar zone in southern Iceland. The microearthquake data clearly fill in some of the gaps in Tryggvason's data and show that this zone may be much narrower than previously recognized. *Stefánsson* [1967] further related these zones of larger events with the idea that larger stresses can develop in shear zones than can develop in zones of tension. With all these observations, we also interpret the southern zone as a transform fault. The eastern zone of active rifting and volcanism is then interpreted as the present crest of the ridge, and it is seen that all the seismic activity on the proposed fracture zones is limited to the zones between the ridge crests. Furthermore, the focal mechanism for the Krafla earthquakes gives the intermediate stress along the axis of the ridge, which is consistent with *Sykes'* [1967] results. No focal mechanism is available yet from the proposed transform fault.

There is no earthquake evidence for a fracture zone near 65° east of Langjökull, as might be implied from *Walker*'s [1964] map of the neovolcanic zone and Figure 9 in this paper. Microearthquake stations 42 to 44 (Figure 3) were specifically occupied to tests this implication.

Although other interpretations of the structure of Iceland and its tectonics have been given [*Bodvarsson and Walker*, 1964; *T. Einarsson*, 1965, 1967; *Th. Einarsson*, 1967], the identification of two transform faults connecting zones of rifting is an important step in interpreting the geology of Iceland in terms of sea-floor spreading. By such an interpretation of the only substantial area on the ocean ridge above sea level and by using techniques commonly accepted in less accessible areas, we should be able to shed light or doubt on these techniques and indeed on the spreading hypothesis itself. Supplementary and conflicting evidence will now be discussed to show the complexities of such an interpretation.

Magnetic data. T. Einarsson [1967] and *Talwani et al.* [1968] have both traced the central magnetic anomaly on the Reykjanes ridge [*Heirtzler et al.*, 1966] northward into the western end of the Reykjanes Peninsula. Although they disagree in details, they both clearly show the mid-Atlantic ridge extending up to southwestern Iceland. *Avery et al.* [1968] have mapped the magnetic anomalies in the Norwegian Sea, east and north of Iceland. They find disturbances in the magnetic lineations characteristic of old fracture zones north of 66°N and just south of 64°N. These zones strike approximately N 80°W and clearly line up with the transform faults discussed above. *Sigurgeirsson* [1967] presented a compilation of twenty-one aeromagnetic profiles flown by Serson, of the Dominion Observatory, Canada, NW–SE over Iceland at 3000–4300 meters, spaced 36 km apart. Although more closely spaced flight lines will be needed to correlate the profiles, which are far less regular than those recorded by *Talwani et al.* [1968] and *Heirtzler et al.* [1966] on the Reykjanes ridge, zones of high intensity can be traced along both sections of the neovolcanic zone up to approximately 65°N. In addition, the central anomaly of the Reykjanes ridge cannot readily be traced on an uninterrupted NE trend north of the western tip of the Reykjanes Peninsula and therefore may well be offset by a fracture zone near 64°N.

Seismic refraction data. The upper part of the crust in Iceland has been studied extensively by refraction measurements [*Pálmason*, 1963, 1967*a*]. A characteristic layering has been found. In the neovolcanic zone a surface layer with an average *P* velocity of 2.8 km/sec (layer O) is interpreted as Quaternary volcanic rocks. This is underlain by the Tertiary flood basalts, which form the surface rocks on both sides of the neovolcanic zone. The upper part of the Tertiary basalts (layer 1) has an average *P* velocity of 4.2 km/sec, whereas the lower part (layer 2) has a velocity of 5.0 km/sec. At a depth varying between approximately 2 and 4 km, the velocity increases to form 6.0 to 6.7 km/sec (layer 3). One noteworthy irregularity in the upper crustal structure of Iceland is the apparent absence of layer 2 on the Reykjanes Peninsula [*Pálmason*, 1967*a*]. Furthermore, there appears to be a lateral change in layer 3 near the proposed transform fault from 6.2 km/sec on the Reykjanes Peninsula and in southern Iceland to 6.5 km/sec in western Iceland.

Geodetic surveys. Geodetic measurements across the active volcanic belt in northeast Iceland [*Gerke*, 1967] made in 1938 and in 1964 and 1965 have failed as yet to reveal any horizontal changes in distances in this zone. *Tryggvason* [1967, 1968], *Decker* [1968], and R. G.

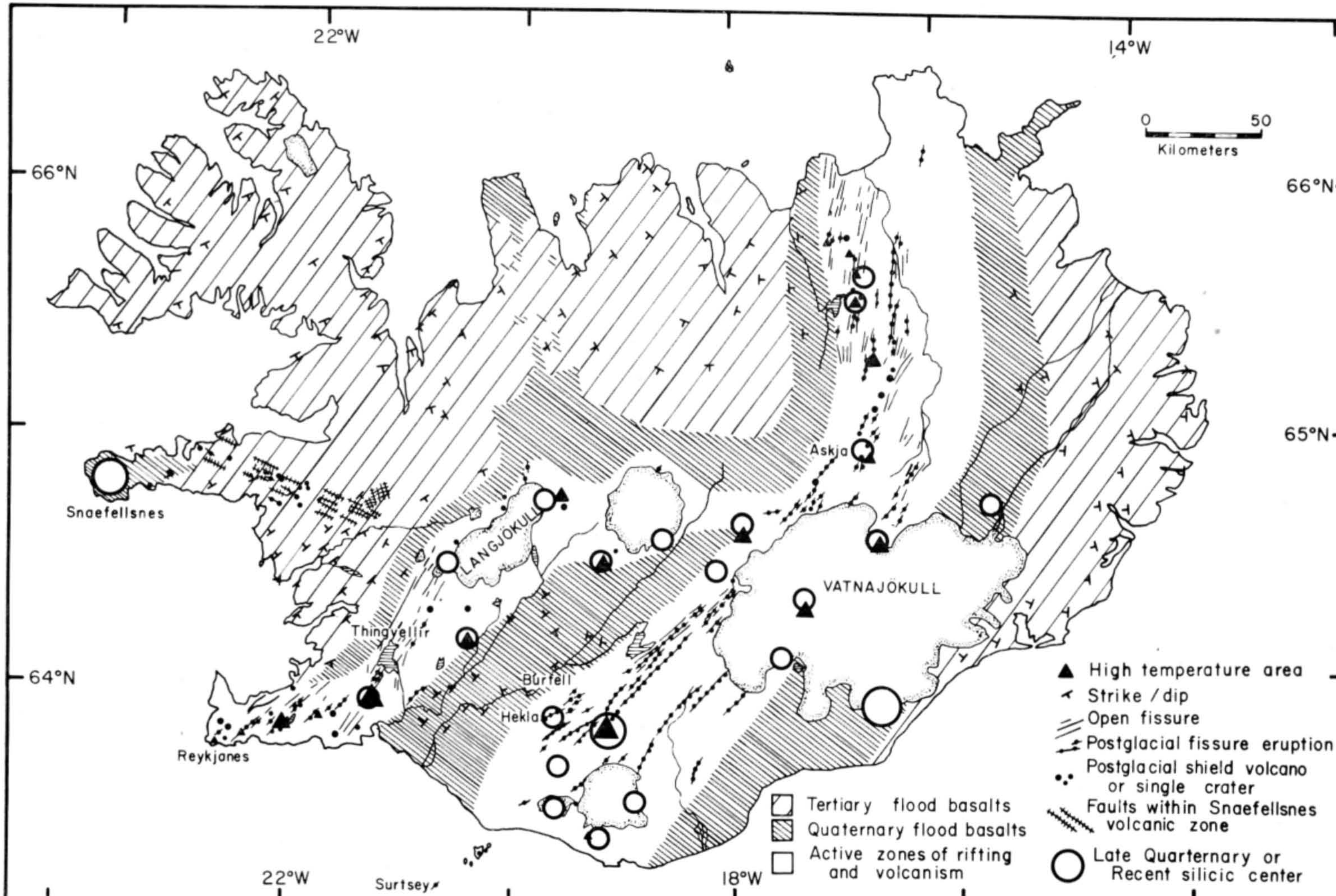

Fig. 9. A preliminary map of the main tectonic features of Iceland. The number of data available varies greatly from region to region (see text).

Mason (Imperial College, London), as well as Gerke, are pursuing geodetic measurements of rifting in different parts of Iceland.

Topography and faulting. The topographic evidence for the transform faults is not very compelling. Topography similar to that of ocean fracture zones or large strike-slip faults is not found in southern Iceland. The relative motion on a transform fault between two ridges would be twice the spreading rate and therefore would average several centimeters per year. One would expect that evidence of such movement would be visible. A postglacial lava flow, Thjórsárhraun, flowed due west from somewhere east of Hekla near 64°N and finally turned south to the sea south of Thingvallavatn, and it might conceal more evidence of fracturing. There is a clear east–west topographic and stratigraphic line (64°12.6′) northeast of Reykjavik that could be related to a transform fault. Although the two major rivers of south-central Iceland swing sharply westward near 64°N, a sense opposite to that predicted by the transform fault hypothesis, their course seems primarily influenced by recent lava flows. The postglacial fissure eruptions shown in Figure 9 diverge in the eastern active rift zone. Near Hekla the fissures turn westward in line with the proposed transform fault. Northeast of Mýrdalsjökull they turn southward near 64°N.

The predominance of en echelon fractures in southwestern Iceland, striking approximately N 30°E (Figure 9), suggests the possibility that they may reflect faulting at some shallow depth causing a rotation of the surface layers, as has been demonstrated in clay models [*Nadai*, 1931; *Cloos*, 1932]. *Tryggvason* [1967, 1968] has described the motion observed on faults at shallow depth. The sense of displacement is left-lateral and is generally consistent with the transform fault hypothesis and is clearly inconsistent with a simple right-lateral transcurrent offset of the ridge. The only fault that he has described that is inconsistent with the transform fault is a 30-km-long, *N–S* striking, right-lateral fracture formed in 1912 near Hekla. Strike-slip faults trending WSW–ENE have also been reported in the Búrfell area NW of Hekla [*Tómasson*, 1967]. The motion on these faults is dominantly left-lateral, and the offset on single faults is of the order of 500–800 meters.

T. Einarsson [1967] has described seven sets of en echelon fractures, five of which lie in the proposed transform fault zone from Hekla to the western Reykjanes Peninsula. The other two zones are close together in the central part of the proposed crest of the ridge, 20 or 30 km south of the proposed transform fault north of Iceland. He cites them as examples of a larger NE–SW trending fault. In experimental work, the angle of the en echelon tension fractures to the principal strike-slip fault was initially 45°–47°, but, as deformation continued, it approached 60° [*Hills,* 1963, p. 171]. If surface tension was sufficient, shear fractures formed at an angle of 12°–17°. The width of the fracture zone was found to be a function of the depth of the fault. This observation is hard to apply since any motion at depth may not be along one fault but in a zone. The fact that the en echelon fracture *zone,* of the order of 20 km wide, in southwestern Iceland strikes nearly east–west is, however, considered significant. During the 1966 Parkfield, California, earthquakes en echelon fractures formed with strikes 30°–45° east of the strike of the San Andreas fault [*Oakeshott,* 1966]. Many of these fractures opened 5–10 cm during this short sequence of earthquakes, the largest of which was magnitude 5.5.

Main structural features of Iceland. Figure 9 summarizes the main geological and structural features of Iceland. This map was compiled for this paper by K. Saemundsson from many sources and the available data vary from region to region. Not all known data can be given on one map. The area south of Vatnajökull, the NW peninsula, and the northeastern part of the country are the least well known. Eastern Iceland was drawn according to *Walker* [1964] and *Wensink* [1964], the flood basalt areas of northern Iceland and the NW peninsula were drawn mainly according to *T. Einarsson* [1960] with some minor additions from other sources. Dips in Snaefellsnes are drawn according to *Sigurdsson* [1967]. A geological map of Iceland edited by Kjartansson, with sheets covering southwest, south-central, and central Iceland, (printed in 1960, 1962, and 1965, respectively) provided information on the distribution of rocks and rifting in these sections of the country. Information on high-temperature areas and on the occurrences of silicic rocks in the zones of rifting was taken from various papers and reports by the personnel of the National Energy Authority, as was the information on dips on both sides of the western limb of the active volcanic zone in SW Iceland.

The dominant active zones of rifting and volcanism are seen to cut through central Iceland with a bend near Askja. If the eastern zone is interpreted as the crest of the ridge offset by two transform faults as suggested above, the western rift zone from Thingvellir northward is left unexplained. The western zone has generally more normal faults, less fissure volcanism, and more open fissures than the southern half of the eastern zone. *Walker* [1965] has estimated 0.5-cm/yr spreading during the last 5000–10,000 years in this zone. Also, similar groups of rocks of similar age (2.5 to 3.3 m.y.) (Grasty, discussion, p. 159 in *Björnsson* [1967]) occurring on both sides of this zone and dipping toward it suggest spreading for several million years. No major unconformities are seen lower down in these lava sequences [*Saemundsson,* 1967].

Although the western zone may connect to a transform fault at 64°, if it is actively spreading and not simply opening at one end, the other end must connect with a ridge or fault. No seismic evidence for a fault at 65°N was found, as discussed above. Little geologic evidence is available, but the fact that the eastern zone of volcanism and rifting is approximately of the same width north and south of 65°N seriously weakens such an hypothesis.

Saemundsson [1967] has pointed out a NNW trending structural zone north of Langjökull (Figure 9) on the basis of prominent faulting and volcanism at either end of this zone. It might be suggested that a structural weakness extends northward along this zone to join more or less directly with the ridge north of Iceland. *Saemundsson* [1967] has also noted that NNW–SSE trending hyaloclastite ridges and postglacial eruptive fissures related to the zone north of Langjökull extend southeastward into the young volcanic zone, which suggests a slight rejuvenation of activity in the northern zone.

The possibility of a shift in the central zone of rifting at one or more times in the past, possibly after pauses in spreading, must be carefully considered. Synclinal structures indicated by the dip west of the active volcanic zones (Figure 9) have been interpreted by *Saemundsson* [1967] as former zones of rifting and volcanism, which became extinct. *Avery et al.* [1968] and *Le Pichon* [1968] have suggested changes in the axis of spreading both north and east of Iceland. *Ewing and Ewing* [1967] have shown that a worldwide pause in sea-floor spreading may have occurred approximately 10 m.y. ago. Such a pause might explain a 5° to 15° dip discordance between the so-called Tertiary plateau basalts and the Upper Pliocene sediments on Tjörnes at the western edge of the neovolcanic zone in northern Iceland [*Th. Einarsson,* 1967].

The Snaefellsnes volcanic zone may be related to the apparent end of the western section of the neovolcanic zone. This WNW trending zone is a prominent feature of faulting and volcanism of youngest Quaternary to recent age [*Saemundsson,* 1967]. It is discordantly imposed on eroded Tertiary flood basalts, whose anticlinal axis runs parallel to the Thingvellir rift zone and the Snaefellsnes syncline. Until more data are available it might be tentatively suggested that the Snaefellsnes volcanic zone resulted from recent active spreading south of Langjökull and no spreading to the north. An analysis of dykes, fractures, and folds in western Iceland by *Sigurdsson* [1967] shows a fanning of NE–SW folds and structural trends near the Snaefellsnes Peninsula consistent with this hypothesis. This interpretation suggests that a detailed analysis of the Snaefellsnes zone might show whether the crest of the ridge is pushing itself apart or being pulled apart.

The currently active volcanic zone extending from Mýrdalsjökull to the Vestmannaeyjar and Surtsey falls to the south of the eastern end of the proposed transform fault and thus outside of the proposed zone of spreading. In 1964, a 1565-meter-deep borehole was drilled in the Vestmannaeyjar [*Pálmason et al.,* 1965]. About 180 meters of late Quaternary volcanic breccia was found to overlie 640 meters of marine sediments. The rest of the well was in hydrothermally altered basalts that are regarded as Tertiary. Thus the ages of the basement rocks of this region are consistent with the transform fault proposed. The recent volcanism may be attributed to a recent transform fault that produced a wedge-shaped zone to the southwest. This zone does not appear to extend southwest of Surtsey.

All other historical volcanism with the exception of some activity on the eastern end of the Snaefellsnes Peninsula [*Thorarinsson,* 1967] lies on the proposed transform fault or crest of the ridge.

There is no good geologic evidence of a transform fault north of Iceland.

Summary and Conclusions

By following the reasoning applied elsewhere [*Sykes,* 1967], two transform faults have been proposed in northern and southern Iceland, thus suggesting a connection of the mid-Atlantic ridge through Iceland. A brief discussion of the geological and geophysical data bearing on this interpretation shows many complexities that are not readily reconciled. The possibility of detailed examination of all facets of a mid-ocean ridge in Iceland provides an unprecedented opportunity to shed light or doubt on the hypothesis of sea-floor spreading. Therefore, we have pursued the interpretation of the available data far enough to focus attention on future research that would be most relevant to the spreading hypothesis.

1. The structural and geologic relationship of the Thingvellir rift zone to the Snaefellsnes volcanic zone and the NNW trending zone north of Langjökull should be carefully examined.

2. Geologic and structural details of the proposed transform fault areas should be examined carefully to show whether there are faults at the surface or at depth and indeed to show if the geophysical interpretations used to identify these zones are valid.

3. A careful study of the intersection of the Reykjanes ridge with the proposed transform fault should better define the relation of the magnetic lineations to fractures, thermal areas, and earthquakes and might show why the ridge should be offset instead of going straight through Iceland.

4. Geodetic work should be aimed at establishing the relative spreading rates of the Thingvellir and eastern volcanic zones. Also, the deformation along the transform faults and their intersections with the ridge should be measured. Some of this work is in progress.

5. The detailed study of the magnetics over Iceland will need to be completed to verify or correct our interpretation of the location of the ridge and to allow the detailed geological and geophysical studies of Iceland to be applied to ridges in general.

6. Finally, focal mechanisms and a more detailed understanding of the spatial and temporal relations of earthquakes along the proposed transform faults should and will be sought by the authors. Focal mechanisms at different depths should show whether faulting at some depth is causing the fractures at the surface.

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