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Weak Earthquakes in the Northern Part of the Rift Zone of Iceland

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Abstract. More than 900 microearthquakes were recorded during 2.5 months in the summers of 1972–73 by a network of 7 automatic seismic stations in the northern part of the rift zone of Iceland.

The more permanent activity was found in the rift zone near the shore of Axarfiördur and on the oblique Husavik fault zone, as well as under the sea north-west of the land rift zone. The focal depths range from 3 to 8–13 km under the land rift zone and Husavik fault and increase to 20–30 km under the sea. Most of the foci in the rift zone are situated within the blocks between the vast fissure swarms. The focal solutions here reveal the dominance of horizontal compressive stresses oriented E – W.

These results are discussed together with other data from Iceland. In particular, the focal solution is given for the strong earthquake ($M=6.4$) which took place in the region of our investigation in January 1976. The orientation of compression for this and other strong earthquakes is approximately perpendicular to that of the microearthquakes of this study and coincides with the stress orientation measured by Hast (1969) in rocks in shallow boreholes. The compressive stresses are oriented in general perpendicular to the isolines of the Bouguer anomaly (Einarsson, 1954). This probably shows some interrelation between the origin of the gravity field and that of the stresses, which are released through strong earthquakes. The discharge of the main stress occured in weak places and is often accompanied by the revival of the fissure swarms. The weak seismicity during the time between the strong shocks is probably connected with the redistribution of the stresses in smaller areas. In particular, our results in the north of Iceland could be explained by the development of the depression on the shelf between the shore and the continuation of the submerged Kolbeinsey ridge.

Key words: Iceland rift – Foci – Section mechanism structure.

Introduction

Iceland and the North Atlantic have lately become vast geodynamic test areas, where the key problems of modern geology are being unfolded. Recent hopes

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that Iceland is just an uplifted part of the Mid-Atlantic Ridge were, however, not justified, and it is now generally acknowledged that Iceland and the adjacent regions have a specific, anomalous structure. As the problems of plate tectonics are elaborated in and around Iceland in variants and the details grow in complexity (see e.g., Kristjansson, 1974), it becomes evident that it is necessary to break through the limitations of the standard set of experimental data and to pass over to a new and higher level of geological-geophysical observation. An effort in this direction has been made by the Geodynamic Expedition of the Academy of Sciences of the USSR, which has been working in and around Iceland since 1971 (Beloussov and Milanovsky, 1976).

The processing of observed material, obtained by extensive and complex geological-geophysical methods, is still incomplete and, therefore, as yet not available to the greater part of the scientific community.

The present paper discusses the basic results of seismological studies carried out by the Geodynamic Expedition of the Academy of Sciences of the USSR in Iceland in 1971–73. The choice fell on Northern Iceland as the major subject of research because here we can see clear evidence of modern volcanism and riftogenesis, and a very complicated interrelation between the rift zone of the island and the oceanic Kolbeinsey ridge. Northern Iceland is scarcely studied compared to the southern and south-western part of the island, where extensive seismological researches were undertaken by Icelandic and American scientists. (e.g. Klein et al., 1973.)

Some Geophysical Data on Iceland

Extensive multidisciplinary studies were carried out in Iceland during the last few years. Here we mention some of them as they deal with the earth's crust and mantle and with seismicity.

The Bouguer gravity map is shown in Figure 1. This map is taken from an old paper (T. Einarsson, 1954), but more recent surveys generally confirm it (G. Palmason, personal communication 1975). The axial symmetry and the Bouguer minimum in the centre of Iceland have been explained as being caused by the low density mantle column existing to depths of 200 km (Long and Mitchell, 1970). Deep seismic sounding shows a thick crust under north-eastern Iceland (Zverev et al., 1976), and combined interpretation of seismic and gravity data led us to the conclusion that the Bouguer minimum is connected with the depression of normal density mantle under Iceland to depths of 50–60 km (Zverev et al., in press). A somewhat different interpretation has been advanced by Bott (in Kristjansson, 1974).

The map of the earthquake epicenters with $m_b \geqq 4.0$, determined by the world wide standard seismograph network is shown in Figure 1. Open circles represent foci between 1954 and 1963 (Sykes, 1965); shaded circles show ISC and NEIS data for 1964–1976.

Data summaries of the seismicity of Iceland were published by Sykes (1965), Ward et al. (1969), Ward and Björnsson (1971), Ward (1971), Palmason and Saemundsson (1974), Bjornsson and Einarsson (1974), Einarsson (1976). The

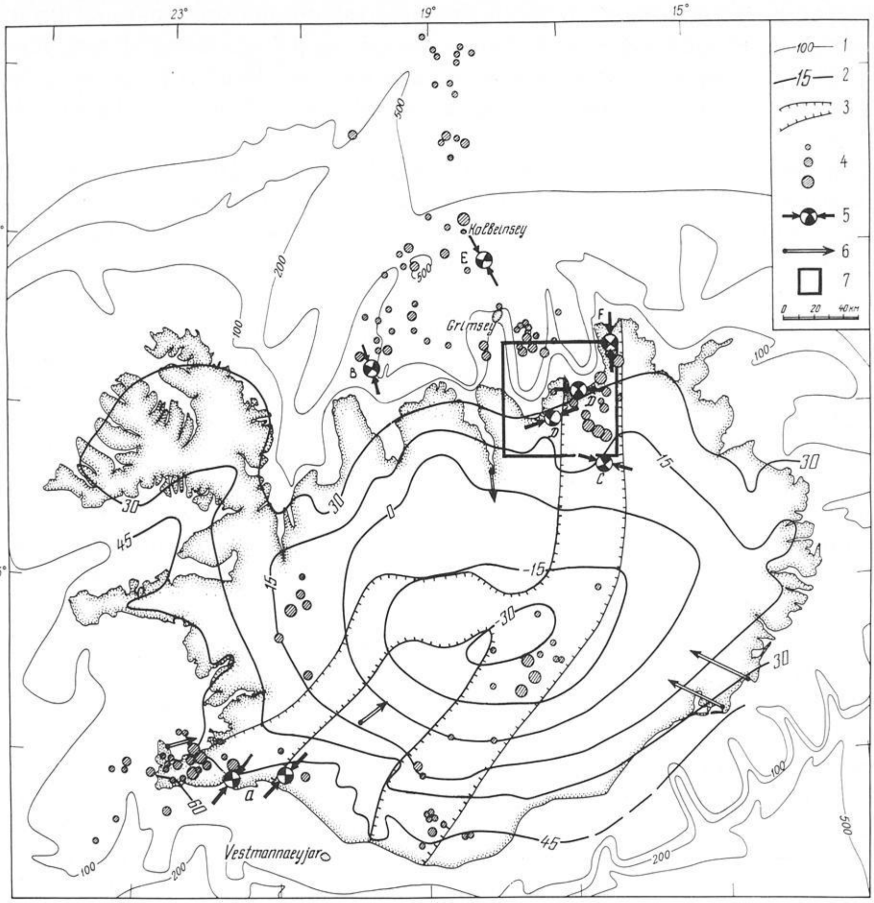

Fig. 1. Some geophysical data for the region of Iceland. 1: depth contours (m); 2: gravity Bouguer anomaly (mgal) (Einarsson, 1954); 3: active zones of rifting and volcanism (Palmason and Saemundsson, 1974); 4: epicenters of earthquakes recorded by the world network for 1954–1976: $m_b<4.5$; $4.5 \leqq m_b \leqq 5.5$; $m_b>5.5$; 5: focal mechanism solutions; the arrows indicate the directions of the principal compressive stresses: *A*: $m_b \sim 3.5$–4.0 (Ward, 1971); *B*: $m_b=6.8$ (Sykes, 1967); *C*: $m_b \sim 1.0$ (Ward et al., 1969); *D*: $m_b \sim -0.5$–1.5 (this paper); *E*: $m_b=5.2$ (Einarsson, 1976) *F*: $m_b=6.4$ (this paper); 6: the vectors of the compressive stresses measured in rocks (Hast, 1969); 7: region of detailed investigations of the Soviet expedition in 1972–73

geographical distribution of earthquake foci and their deviation from the general strike of the Mid-Atlantic Ridge was an important reason for the suggestion of the presence of two transverse (transform) faults south and north of Iceland (Sykes, 1967; Ward, 1971).

The analysis of epicenters indicates that transverse faults in the southwest of the island and along the northern coast are lacking the distinction of transform faults. We prefer to suggest the existence of two seismoactive zones on the island approximately corresponding to two branches of riftogenesis. These two zones are joined north of Iceland in the Kolbeinsey region, and continue northward along the submerged ridge. The data collected on the seismicity during more than 20 years (see Fig. 1) testify that the seismoactive zones are traced across Iceland with significant gaps, and this distinguishes the island from the adjacent parts of the Mid-Atlantic ridge.

The microseismicity of Iceland was studied by Icelandic and American scientists. Thirteen local zones of microearthquakes were established, all of them situated within the modern rift zone (Ward et al., 1969). These data can be compared with the results of bottom seismic observations in the rift valleys of submarine mid-ocean ridges. Such observations were carried out in 1971 off the coasts of Iceland on the submarine Reykjanes and Kolbeinsey ridges by the staff of the Soviet Geodynamic Expedition on the research vessel "Akademik Kurchatov". Weak local earthquakes, specific for rift valleys or their slopes, were discovered everywhere on submarine ridges. Observations on Carlsberg Ridge in the Indian Ocean (Neprochnov et al., 1969) and on the Mid-Atlantic Ridge in two regions (Francis and Porter, 1972; Spindel et al., 1974) were similar. On the other hand, seismicity studies in Iceland revealed that microearthquake occurrence is not universal but local, which in general indicates a weaker manifestation of seismic activity on the island than on submarine ridges.

The available mechanism solutions for some of the earthquakes and for groups of microearthquakes are also presented in Figure 1. The solid arrows indicate the direction of the principal compressive stresses. These compressive stresses are always close to horizontal. For the two strongest earthquakes on the north shore of Iceland ($M > 6.0$) it is possible to observe the same feature. For the earthquake in the western branch of the seismoactive zone (B see Fig. 1) (Sykes, 1967), one of the nodal planes is with good accuracy parallel to the strike of this branch (N 25°). For the earthquake of 1976, situated in the eastern branch (F, Fig. 1), the nodal planes are well concordant to the strike of this eastern branch (N 315°). For the weaker earthquake north of the eastern branch (E, Fig. 1) (Einarsson, 1976), the distribution of the signs of P-wave arrivals does not contradict such a feature.

The data on the orientation of compressive stresses in the foci of strong earthquakes agree comparatively well with stress vectors measured in rocks (Hast, 1969) (see Fig. 1). The compressive stresses at depth (from earthquakes data) and at the surface (Hast, 1969) are approximately perpendicular to the Bouguer isolines and they are oriented toward the center of Iceland. Exceptions are two examples of microearthquakes: C (Ward et al., 1969) and D (this paper) (see Fig. 1).

Region of Research Purpose and System of Observations

In 1971 our expedition carried out a reconnaissance survey of seismicity of Northern Iceland, as the result of which a high seismicity region was determined near the Tjörnes peninsula where a network of stations was set up for temporary observations (Fig. 2).

It was the purpose of the observations to determine the peculiarities of the

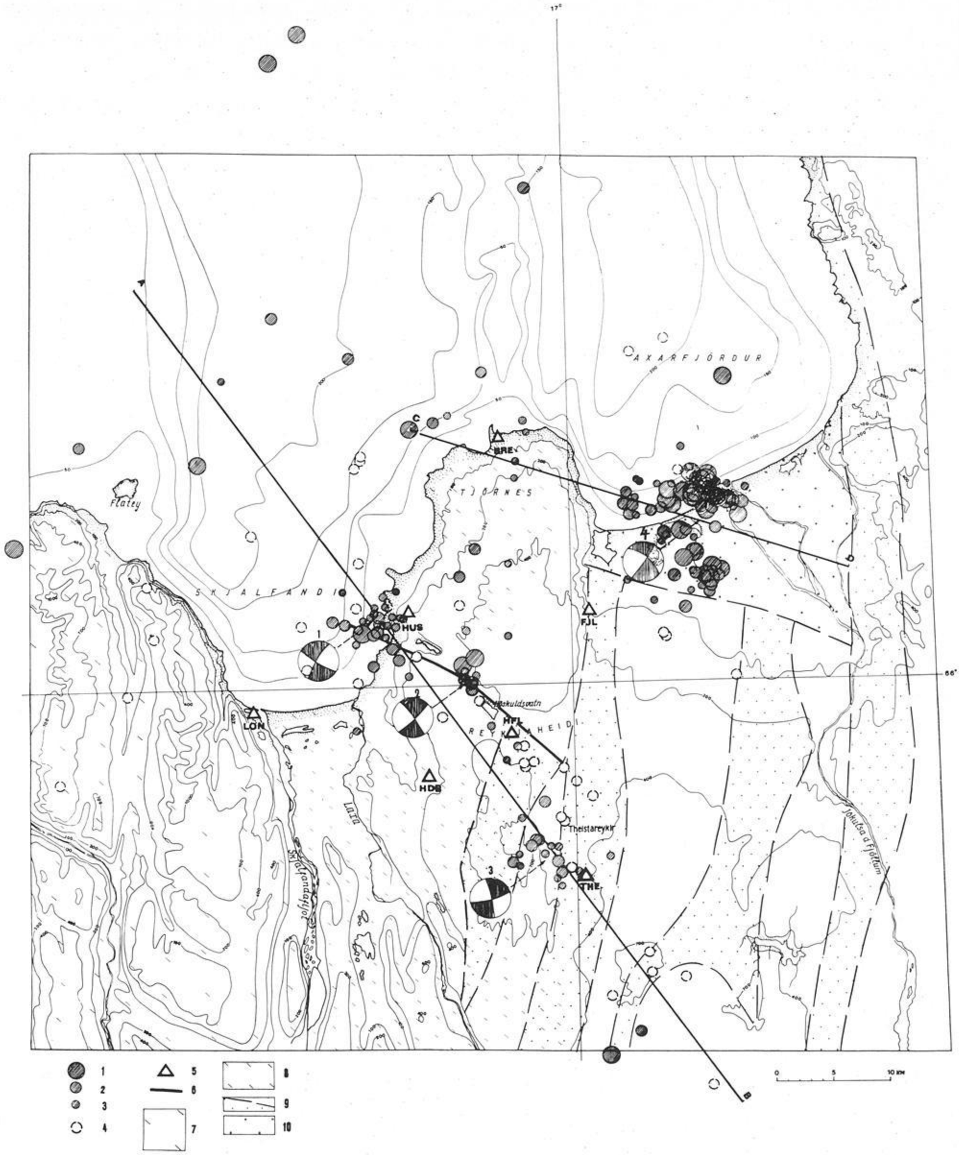

Fig. 2. Map of earthquake epicentres located by our stations in 1972–1973. 1: $m_{SH} \geqq 0.9$ ($K_{SH} \geqq 6.0$) 1973; 2: $0.3 \leqq m_{SH} \leqq 0.9$ ($5.0 \leqq K_{SH} \leqq 6.0$) 1973; 3: $m_{SH} \leqq 0.3$ ($K_{SH} \leqq 5.0$) 1973; 4: earthquakes 1972; 5: recording stations; 6: Husavik fault zone, 7 and 8: Tertiary and Quaternary flood basalts respectively; 9: fissure swarms, 10: Jökulsá river gravel plain. 6 to 9: according to Saemundsson (1974)

seismic activity in space and time, i.e. to determinate the focal coordinates, the energy characteristics, focal mechanisms and, finally, to establish the relationships between modern seismicity and tectonics and to compile the distribution model of elastic stresses in the region studied.

The Tjörnes peninsula and its environs were the major object of seismological research. As stated above, this region is located on the eastern branch of the seismoactive zone (see Fig. 1) associated with the Mid-Atlantic Ridge.

According to geolociacl data, Tjörnes peninsula is a block of the Tertiary and Quaternary basalts, joined to the western edge of the rift zone of Iceland and bounded in the south by the large Husavik fault (Saemundsson, 1974). The structures of the peninsula, as those of the adjoining bays Skjalfandi and Axarfjördur in the west and east, can be traced northward below sea level and are then cut off by the shelf edge.

During the reconnaissance in 1971 the larger part of the epicenters was fixed in the Husavik fault area and, therefore, in 1972 the network of stations was set up in view of the best determination of the focal parameters of these earthquakes (see Fig. 2). In 1973 the area of study was extended and the network of stations had the form of a triangle 30×40 km.

Equipment

Seismic observations in Iceland and at sea around Iceland were carried out with the aid of automatic seismic stations, designed at the Institute of Physics of the Earth, Academy of Sciences of the USSR (Zverev et al., 1971). In 1972 and 1973 the three-component stations had identical instruments; the seismograph NSP-3 with $T_o = 0.7$ s was used as the detector. The signals were automatically recorded on magnetic tape. A tape lasts 5–7 days. The speed of the tape was 0.47 mm/s and the frequency band of the recording channel was 5–13 Hz. The lower sensitivity level of the band was given by the microseismic background of displacement, on the average, $\sim 0.8 \cdot 10^{-9}$ m; the upper limit of undistorted recording was determined by the dynamic range of the recording channel being $\sim 70 \cdot 10^{-9}$ m for $f = 10$ Hz. Beside the three seismic channels, the magnetic tape also recorded the coded signals of quartz clocks, thus allowing the absolute arrival time of the earthquake waves to be established with an accuracy of a few hundredths of a second. Sensitivity control was achieved with the aid of the calibrating generator, transmitted to the input of the recording amplifiers, while time control was obtained by recording the signals of the British radiostation "Rugby" when switching on and off the automatic stations.

General Description of Observation Material

The conditions of installation and the microseismic background allowed us to record a large number of shocks. The most favourable conditions were at stations HFL and HDB (see Fig. 2), which recorded up to 80 earthquakes per diem, and more than 300 earthquakes in total during 1.5 months of observation in 1973. The least favourable conditions were at station LON, which recorded only 20 events. The bulk of earthquake records had $t_{s-p} = 0.7$–5.0 s. The lag time t_{s-p} of the transverse wave relative to the longitudinal P-wave is in this case the most convenient and objective function of distance. Since for most of the seven stations of 1973 the distribution of t_{s-p} records was approximately the same, the recording might be presumed to be optimal.

For most earthquake records the times of arrivals of P- and S-waves were determined, as well as the maximum amplitude and the dominant oscillation period for longitudinal and transverse waves and the sign of the first P-wave displacement and the direction to the epicenter. However, not all earthquakes were recorded by the entire station network. Rather often very weak oscillations close to the level of the noise background were observed. Therefore, the coordinates and the source depths were determined for only 41 earthquakes of 1972 and for 167 earthquakes of 1973.

Processing Technique of Observation Material

In the first stage, the coordinates of the earthquake foci were determined with the aid of a computer program based only on t_{s-p} times recorded by three and more stations. In this case the velocity model was used, which was obtained for the surrounding region by explosion seismology (Palmason, 1971). This cross-section was transformed into an apparent velocity $V_{s-p}=\dfrac{V_p}{\dfrac{V_p}{V_s}-1}$ (Table 1). The coefficient $K=V_p/V_s$ was determined by the sum of records of many earthquakes at our stations with a precision of up to ± 0.02.

Table 1. Velocity section assumed for determination of hypocenters of Icelandic earthquakes

H (km)	V_p (km/s)	V_p/V_s	V_{s-p} (km/s)
0–1.0	2.5	1.72	3.5
1.0–3.5	4.6	1.72	6.4
more than 3.5	6.4	1.72	8.9

The standard error of the instrumental hypocenter determinations within the perimeter of the station network is 0.1–0.6 km; for remote earthquakes it increases to 1.0–3.0 km.

The magnitude of the earthquakes m_{SH} was evaluated on the basis of the body waves SH according to the relationship $m_{SH}=\lg\,(A/T)_{max}+\sigma(\mathrm{R})$. As the calibrating function $\sigma(\mathrm{R})$ we took the one for European earthquakes at small epicentral distances (Christoskov, 1969). For our purposes this function was approximated by logarithmic curves:

$$\sigma(\mathrm{R})=1.98\cdot\lg\mathrm{R}-1.05.$$

Instead of the epicentral distances Δ in degrees we used R in kilometers. The standard deviation of the experimental points from the approximating curve is about 0.1. The accuracy of magnitude determination is evaluated to be 0.3. Beside the magnitudes the energy classes K (energy $E=10^K$ joules) were determined by means of nomograms (Fedotov, 1972) for the longitudinal (K_p) and shear (K_s) waves.

Map of Epicenters and Cross-Sections

The positions of the instrumental hypocenters are shown on the map (Fig. 2) and on two cross-sections (Fig. 3). Figure 2 shows that there are two groups of epicenters. One of them, in the form of a band, follows the Husavik fault and its continuation in the rift zone. Here the most stable manifestation of seismicity has been noted during the two observation seasons in 1972 and 1973. Within the band the greatest concentration of epicenters was observed in the region of the HUS station. The second group of epicenters is compact; it is located on the coast of Axarfjördur. This group lies on the continuation of the northern branch of the rift zone. Here the activity is quite variable with time; in the summer of 1972 only a few single foci were observed, while in 1973 it was a dense group of epicenters. Other earthquakes are scattered mostly in the sea area. A few single epicenters were detected on the Tjörnes peninsula and on land west of Skjalfandi Bay.

The focal depths distributed along the Husavik fault, are from 2–3 to 10–12 km (Fig. 3). Submarine earthquakes in the Skjalfandi Bay are much deeper: (up to 20–25 km). Such depths were confirmed here by additional observations in 1975, when one of the recording stations was placed on Grimsey island. The maximum focal depth in general obviously increases in the north-western direction. In the region of Axarfjördur (Fig. 3) the earthquakes occur mainly at the depth interval from 3–5 to 17–20 km. Here the focal depths also appear to increase toward N–W. On the eastern shore of Axarfjördur where the swarm of strong earthquakes took place in late 1975 – early 1976, no event of $m_{SH} > -1.0$ was fixed during our observations in the summers 1972, 73 and 75.

The positions of the fissure swarms are shown in Figure 3 after K. Saemundsson (1974). These fissures trace vast tension zones. Most of the foci are located outside of fissure swarms, within blocks between them.

All foci are located in the seismic layer with the P-wave velocity of 6.4 km/s.

Earthquake Mechanisms

From the signs of the P-wave arrivals the mechanisms active in the foci have been determined for various groups of earthquakes. The entire epicentral zone was divided into four regions according to the temporal stability of the signs of the first arrivals. For each of these regions a single mechanism for a group of epicenters has been defined. The distribution of signs of the P-wave arrivals is presented in Figure 4. The most reliable nodal planes could be drawn for the first three regions along the Husavik faults. In the fourth region corresponding to the group of foci at Axarfjördur the one-sided location of the stations prevents a reliable determination of one of the nodal planes. On the whole, the good correlation of data for different earthquakes indicates the stable character of the focal mechanisms of the selected groups of earthquakes. In all cases the main compressive stresses are close to horizontal mainly in the E–W direction. The schematic focal spheres presented in Figure 2 show a gradual turning of one of the nodal planes along the Husavik fault from west-north-west in the north to

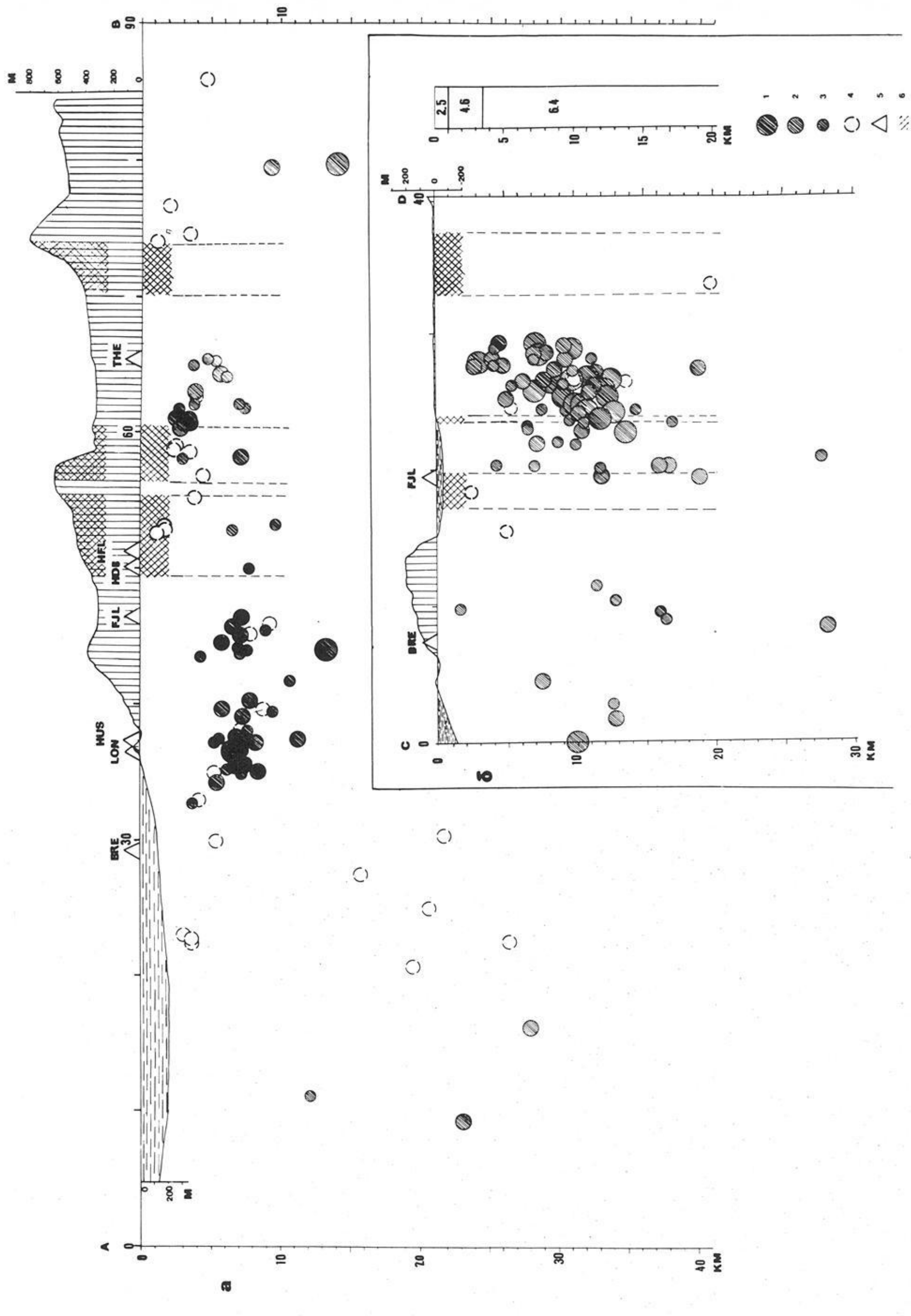

Fig. 3. Seismic cross-sections along the lines AB and CD (see Fig. 2) and the velocity model adopted for locating the earthquakes. Legend 1–5 identical to that of Figure 2. 6-fissure zones (Saemundsson, 1974)

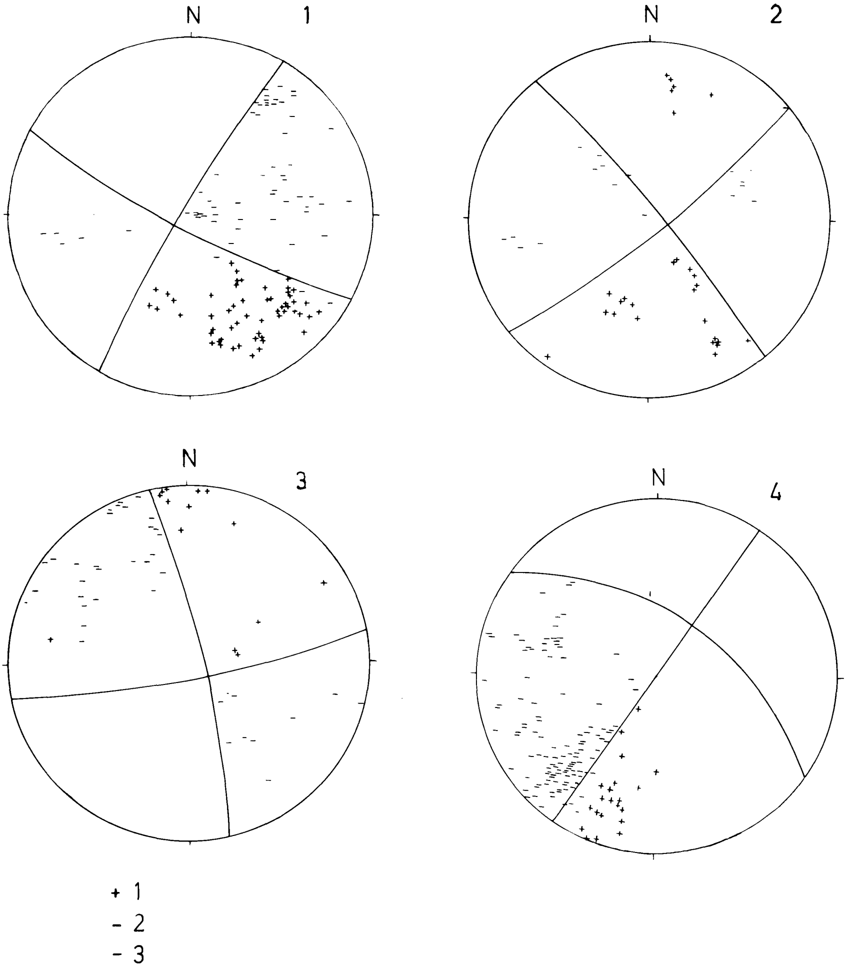

Fig. 4. Distribution of signs of first motion of P-wave onsets on the Wulff grid (upper hemisphere) for four epicentral regions (see Fig. 2); 1: compression; 2: dilatation; 3: nodal lines

north-north-west in the south. This tendency seems likely to continue further to the south to the Krafla region, where according to the data by Ward et al. (1969), the strike of one of the nodal planes is almost north-south (see Fig. 1).

The maximum compression axes for the microearthquakes are approximately perpendicular to the strike of the tension fissures (see Fig. 2) and to the directions of the compressive stress in the foci of the strong earthquakes (see Fig. 1), in particular, the earthquakes of 13 January 1976. In Figure 5 the distribution of the signs of the P-wave onsets for these earthquakes is shown on the lower hemisphere. The data of more than 100 of the world network stations were used. The main compressive axis is horizontal and has an azimuth of 355°. If we suppose that the slip plane follows the strike of the eastern branch of the seismoactive zone, the displacement is north for the western side of the fault. It

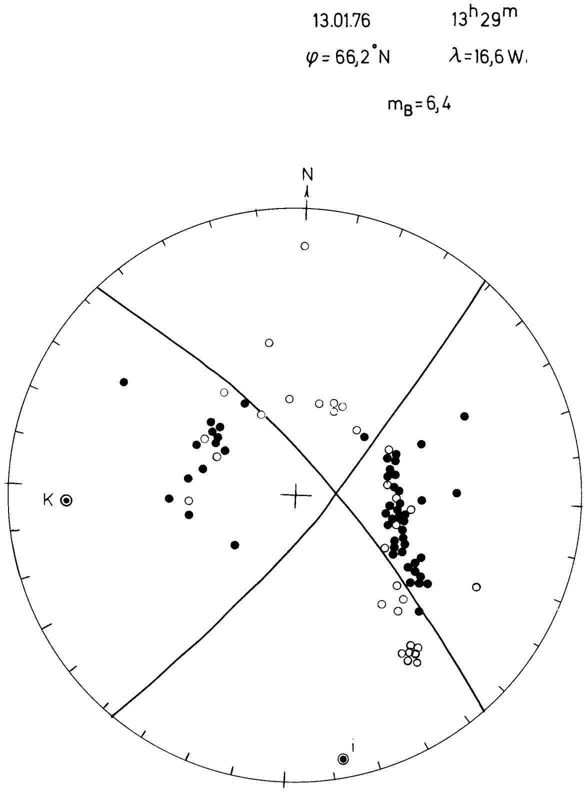

Fig. 5. Distribution of signs of first motion of *P*-wave onsets the Wulff grid (lower hemisphere) for the earthquake of 13 January 1976, $m_b = 6.4$. Solid circles denote compression; open circles dilatation

may be mentioned that the swarm of strong ($m_b = 4.0$–6.4) shocks 1975–76 on the eastern part of the neovolcanic zone took place in the vicinity of Kopasker village. These shocks reveal the revival of the known fissure swarms in the rift zone.

Seismic Regime of the Area

The analysis of the time variations of activity for the given area (Fig. 6) demonstrates that on particular days the number of recorded earthquakes with $m_{SH} \geqq 0.5$ changes from 0 to 10 and more. These variations are even stronger for weaker earthquakes ($m_{SH} \geqq -0.3$) which reach 40 per day.

Activation covered the entire research area simultaneously. Figure 6 represents the accumulated and released energy (more exactly; ΣK_{SH} in earthquakes with $K_{SH} \geqq 4.0$, i.e. $m_{SH} \geqq -0.3$). These plots show that in the two most active sites, the Axarfjördur region (a) and the Husavik fault zone (b), the activation periods are displaced by 1–3 days, but they invariably have the same succession with stable, quiescent periods in between. The shocks started always in the Husavik fault zone and then continued in the Axarfjördur region. For the group

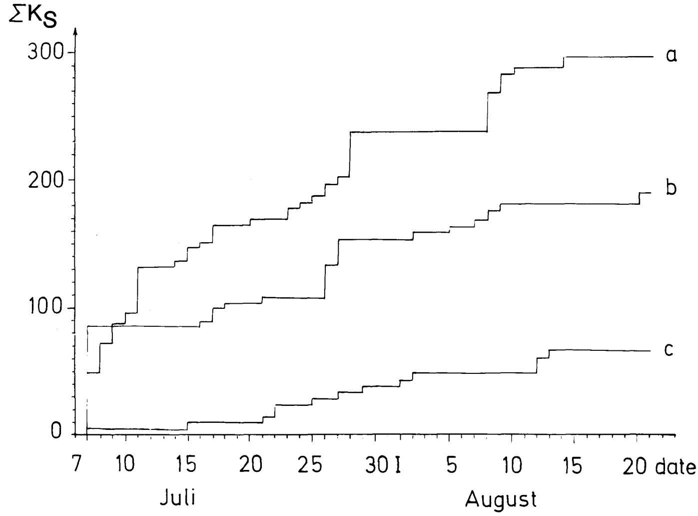

Fig. 6. Plots of ΣK_S accumulation for earthquakes with $K_S \geqq 4.0$ ($m_{SH} \geqq -0.3$) for the Axarfjördur region (*a*), the Husavik fault region (*b*), and the north-western marine region (*c*)

of submarine earthquakes (c), where the probability of omissions is high, the accumulation curve ΣK does not have such a good correlation with the first two regions. Apparently, the earthquakes of the most active regions, where 90% of all shocks occurred, separated by an aseismic block, are to a considerable extent connected with each other. The similarity of focal mechanisms for both regions and for the adjacent Krafla region supports this supposition.

Discussion of Results

The seismic activity in Iceland considerably differs from most of the submarine regions of mid-ocean ridges. In the area of Iceland far fewer earthquakes are recorded than on the ridges to the south and to the north of Iceland, and besides, the earthquakes themselves are much weaker. The microearthquakes are also localed only in small zones.

Most of the strong earthquakes in Iceland were observed near south-western and northern shores. The largest number of earthquakes both strong and weak occured as a result of the release of the near horizontal compressive stresses. In all the strongest earthquakes and part of the microearthquakes these compressive stresses are oriented approximately toward the centre of Iceland. Their directions agree with the orientation of the stress axes in rocks determined by overcoring (Hast, 1969) and they are approximately perpendicular to the Bouguer anomaly isolines.

The focal depths under Iceland do not exceed 10–12 km, but reach 17–20 km under the northern shore and sometimes 20–25 km under the shelf. It may be mentioned that some deep foci at 20–30 km depth were observed on the southern side of Iceland by Björnsson and Einarsson (1974).

We think that all the above data reflect large scale geophysical phenomena in Iceland. Probably the shape of Iceland and its shelf, the gravity anomalies, the generation and accumulation of the compressive stresses, revealed in the rocks and in the earthquake focal mechanisms have a common origin at depth.

The stresses are first released in strong earthquakes in the weakest places, i.e. in the rift zones. The submerged rift zone runs from the north toward Iceland as the single Kolbeinsey ridge. The seismic zone follows the ridge, but south of Kolbeinsey it forks into two branches. The two seismoactive branches in Iceland generally coincide with the modern and ancient zones of rifting and volcanism. The nodal planes for the strongest shocks coincide with the strike of these two branches.

The attenuation of the rate of seismicity from the borders toward the middle of Iceland should be discussed together with the crustal thickness. The interpretation of explosion seismology investigations (Zverev et al., 1976) shows that in northeastern Iceland the normal mantle dips to depths of more than 40–50 km toward the centre of the island, and the combined interpretation of seismic and gravity data led us to suggest a cap shaped depression of the mantle under Iceland filled by the low density crust (Zverev et al., 1977). Taking this into account, we can tie the attenuation of seismicity in with the growing thickness of the crust and its absorption properties.

The weak earthquakes give us another more local scale of geophysical phenomena. In northern Iceland most of the microearthquakes fall in between the vast tension zones of the fissure swarms (see Fig. 3). The vast fissures were absolutely inactive in seismically quiet periods. The compressive stresses in the foci were oriented normally to the strike of the fissures, i.e. the blocks were compressed between the tension zones. The orientation of one of the nodal planes for the different groups of foci very closely follows the variable strike of the Husavik fault zone.

The above observations of focal depths increasing toward N–W, the surface and bottom relief, the gravity and magnetic anomalies (Johnsson, 1974) could be most easily understood by supposing an important role of the depression of the shelf between the shore of Iceland and Kolbeinsey. The shape of this depression is triangular with Kolbeinsey at the top and the base following the north shore of Iceland. Its sides may be traced from the epicentres of the strong earthquakes from positive magnetic anomalies, from outcrops of fresh basalts, and from some bathymetric contours (Johnsson, 1974).

The base of the triangle is characterized by a large step in surface and bottom relief, and by a local negative gravity anomaly (Palmason, 1974). The active modern development of this depression, probably bordered by large fault zones, can explain the above phenomena better than the suppositions of transform faulting and large scale horizontal movement in this region (Saemundsson, 1974).

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