Tectonophysics, 55 (1979) 127—153

SEISMICITY AND EARTHQUAKE FOCAL MECHANISMS ALONG THE MID-ATLANTIC PLATE BOUNDARY BETWEEN ICELAND AND THE AZORES

PÁLL EINARSSON

Science Institute, University of Iceland, Reykjavík (Iceland)

(Submitted December 19, 1977; accepted for publication May 31, 1978)

ABSTRACT

Einarsson, P., 1979. Seismicity and earthquake focal mechanisms along the Mid-Atlantic plate boundary between Iceland and the Azores. In: J. Francheteau (Editor), Processes at Mid-Ocean Ridges. Tectonophysics, 55: 127—153.

The seismicity along the mid-Atlantic plate boundary in the North Atlantic is in a general way consistent with the concept of sea floor spreading. Some detailed studies of the seismicity of Iceland show, however, that the simple picture of an orthogonal system of spreading ridges and transform faults is an oversimplification. On the Reykjanes Peninsula, for example, the plate boundary has both transform fault and spreading ridge characteristics. The Tjörnes fracture zone may be described as a multiple transform fault with at least two parallel strike-slip faults taking up the transform motion. The largest earthquakes within the volcanic rift zones of Iceland appear to be associated with central volcanoes and are not directly related to the rifting process. The Charlie-Gibbs fracture zone and the southern part of the Reykjanes Ridge are examples of a typical transform fault and of a spreading ridge, respectively. The northern part of the Reykjanes Ridge, on the other hand, appears to be spreading obliquely. South of the Charlie-Gibbs fracture zone focal mechanism solutions of two earthquakes on or near the plate boundary have a substantial component of thrust faulting. Crustal compression in this area may be related to mantle plume activity under Iceland and the Azores.

INTRODUCTION

Earthquake focal mechanisms and spatial distribution of epicenters play an important role in the concept of plate tectonics. Seismic zones delineate plate boundaries and from the focal mechanisms of earthquakes one can infer the relative plate motions at the present time. From a scientific point of view, it is unfortunate that a large proportion of all plate boundaries is hidden under the ocean and is not well accessible for direct studies.

In the study of the seismicity of the mid-ocean ridges in particular one is usually limited by the large distances between the source area of the earthquakes and the nearest seismograph stations. Epicenters can only be determined for earthquakes larger than 4—4.5 and uncertainty in the locations is high, of the order of 10—20 km or more, which precludes any detailed tectonic interpretation. Focal mechanism can usually not be determined for earthquakes of magnitude smaller than 5.5, but larger earthquakes are rare in some parts of the mid-ocean ridge system. Therefore, long time will pass until sufficient density of data will be available.

In a study of focal mechanism of earthquakes along the world rift system Sykes (1967) made the important observations that earthquakes along fracture zones were predominantly associated with strike-slip faulting in the sense predicted by the transform fault hypothesis (Wilson, 1965), and earthquakes in other parts of the rift system had focal mechanisms characterized by normal faulting. Since then several authors have studied focal mechanisms and the seismicity of the mid-ocean ridges, and have in general confirmed the results of Sykes. The purpose of this paper is to summarize the seismological data for the mid-Atlantic plate boundary between 40° and 67°N and try to correlate them with the available data on topography and geological structure. As more seismological and structural data accumulate such regional studies will be increasingly important for a better understanding of the processes taking place at the diverging plate boundaries. This particular section of the plate boundary was chosen for the following reasons:

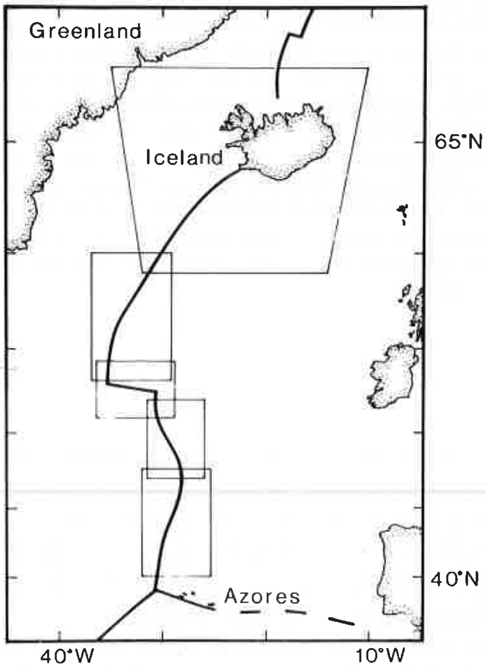


Fig. 1. An index map of the studied area. The submarine plate boundary is marked with a heavy line. The map frames of Figs. 4, 5, 7, 9 and 10 are shown.

(1) Within this section the plate boundary crosses Iceland, where the structure and the seismicity are known in some detail.

(2) This section is relatively well located with respect to the dense seismic networks of Europe and North America.

(3) Relatively many detailed geophysical surveys have been done on the submarine part of this plate boundary.

The study area is bounded by the Iceland hot spot in the north and by the Azores hot spot in the south (Fig. 1). It contains three major fracture zones; the Tjörnes fracture zone, the south Iceland seismic zone and the Charlie-Gibbs fracture zone which is thought of as a typical oceanic fracture zone. It furthermore contains the Reykjanes Ridge that has well developed magnetic anomalies of the sea-floor spreading type indicating a half spreading rate of 1 cm/yr (Talwani et al. 1971).

The results of this study confirm the main results of Sykes (1967) of primarily normal faulting on ridge crests and strike-slip faulting in fracture zones. In addition, a few peculiarities in the tectonic process appear, such as multiple transform faulting, oblique spreading, and thrust faulting in or near the rift valley south of the Charlie-Gibbs fracture zone.

FOCAL MECHANISM DATA

Several of the focal mechanism solutions used in this study were obtained by various other authors, other solutions are reported here for the first time. The parameters of the solutions are listed in Table I. Details of solutions derived independently in this paper are shown in Fig. 2.

The focal mechanism solutions are mostly based on the first motion pattern of the P waves recorded at teleseismic distances. S waves are used whenever possible. Long period records of the WWSSN and the Canadian seismic network are used almost exclusively. Short period P waves are used only when they are impulsive and a long period record is not available.

The seismic rays are projected onto the focal sphere by conventional methods (see, e.g., Sykes 1967), and mapped on an equal area projection. A standard earth model is used and a shallow focus is assumed for the earthquakes. The focal mechanism solutions are not very sensitive for small changes in the depth of focus or the crust and upper mantle structure as long as it is assumed to be laterally homogeneous.

ICELAND

Several authors have studied the connection between the seismicity and the structure of Iceland, both in general (see, e.g., Ward, 1971; Tryggvason, 1973; Björnsson and Einarsson, 1974) and some of the details (see, e.g., Klein et al., 1973, 1977; Einarsson, 1976; Einarsson et al., 1977). In this paper only few features will be mentioned that might be of importance when compared to the submarine plate boundary.

Within the Icelandic insular shelf the mid-Atlantic plate boundary is dis-

TABLE I

Summary of focal mechanism solutions available for the mid-Atlantic seismic zone between 40° and 68°N [a]

No.	Date			Time			Epicenter		m_b	Pole of first nodal plane trend/plunge
	y	m	d	h	m	s	N	W		
1	1969	05	05	21	47	31.6	66.8	18.2	5.1	22/8
2	1976	01	13	13	29	19.5	66.2	16.7	6.0	219/4
3	1963	03	28	00	15	46.2	66.3	19.8		197/3
4	1974	06	12	17	55	08.7	64.8	21.2	5.5	132/34
5	1974	July					64.8	21.4		0/45
6	1967	07	27	05	17	54.0	64.0	20.7	5.0	0/0
7	1968	12	05	09	44	11.0	63.9	21.7	5.5	357/0
8	1973	09	15	01	45	57.7	63.9	22.2	5.3	180/12
9	1972	September					63.8	22.7		153/42
10	1969	09	20	05	08	57.5	58.3	32.2	5.5	104/19
11	1970	04	24	01	23	12.0	55.7	35.0	5.3	no nodal
12	1972	04	03	18	52	59.2	54.3	35.1	5.3	88/32
13	1972	04	03	20	36	22.1	54.3	35.1	5.1	88/32
14	1965	07	05	08	31	58.9	52.9	34.2	5.7	16/10
15	1967	02	13	23	14	23.5	52.8	34.1	5.5	10/0
16	1966	04	08	05	52	40.0	52.7	33.3	5.2	(12/0)
17	1969	09	24	03	58	56.5	52.5	31.9	5.1	4/0
18	1964	08	26	03	18	44.0	52.1	30.1	5.3	214/39
19	1970	09	18	16	12	07.0	51.1	29.6	5.1	6/10
20	1973	01	05	01	44	25.7	49.5	28.2	5.3	55/32
21	1965	09	29	23	20	17.8	45.2	28.2	5.3	113/16
22	1964	09	17	15	02	00.8	44.5	31.3	5.5	263/59

[a] Magnitudes and epicentral data are taken from the PDE listings of USCGS, later NOAA and USGS, unless otherwise noted. Depths given for the teleseismic locations are 33 km or less for all the earthquakes. The depths of the earthquakes used for solution no. *5* and *9* were less than 6 km. The details of solution no. *2* will be given in a forthcoming publication on the Krafla earthquakes.

placed to the east by two fracture zones, the Tjörnes fracture zone in the north and the Reykjanes Peninsula and the south Iceland seismic zone in the south (Fig. 3). Neither of these fracture zones has a clear expression in the topography, both of them are defined primarily by their seismicity. Many readers might find the term "transform fault" an oversimplification for the complex structure of these zones. The plate boundary between the fracture zones is marked by prominent rift zones. In north Iceland the rift zone is only one, but in south Iceland rifting occurs in two parallel zones 100 km apart. Let us now consider each of these active zones separately.

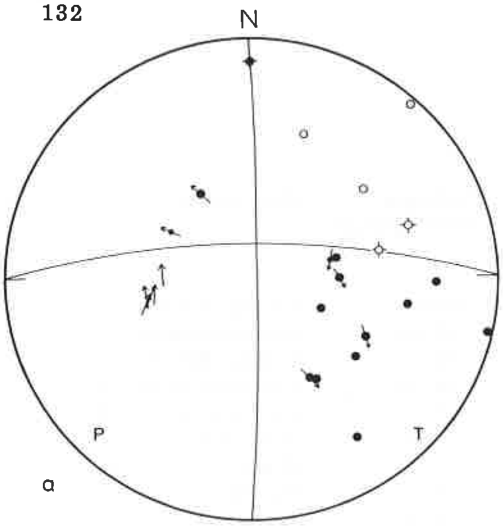
The Tjörnes fracture zone

The transform motion between the submarine Kolbeinsey ridge and the volcanic rift zone in north Iceland is taken up by the Tjörnes fracture zone,

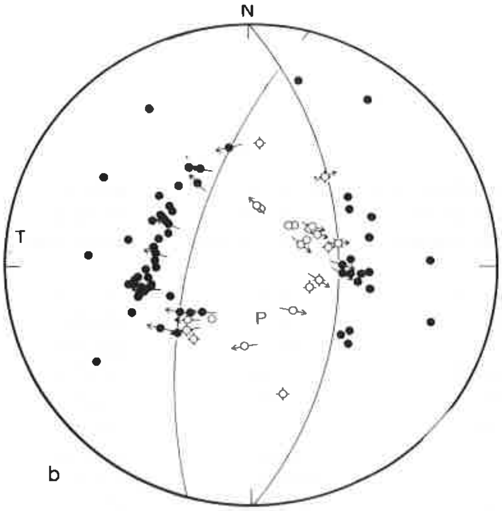
Pole of second nodal plane trend/plunge	Axis of compression trend/plunge	Axis of tension trend/plunge	Null axis trend/plunge	Data source
115/18	159/8	67/19	270/70	Conant, 1972
308/4	174/0	263/5	83/84	to be published
287/12	333/7	241/10	88/77	Sykes, 1967 [b]
268/48	187/64	313/11	28/24	Einarsson et al., 1977
180/45	—/90	0/0	90/0	Einarsson et al., 1977 [c]
90/0	45/0	135/0	—/90	Ward, 1971
267/15	221/9	313/10	87/75	Ward, 1971
271/2	226/10	134/8	9/78	Fig. 2a
296/42	225/70	135/0	45/20	Klein et al., 1977 [c]
270/30	172/72	278/5	9/15	Fig. 2b [d,e]
planes found	near vertical	—	—	Fig. 2c
264/31	218/85	85/4	355/4	Fig. 2d [d]
264/31	218/85	85/4	355/4	Fig. 2d [d]
107/8	331/1	61/12	235/77	Sykes, 1970
100/6	146/4	56/4	280/84	Fig. 2e [f]
(102/0)	(147/0)	(57/0)	(—/90)	Fig. 2f
94/0	139/0	49/0	—/90	Fig. 2g
116/9	157/33	261/20	16/50	Fig. 2h
270/32	224/16	323/30	108/56	Fig. 2i
183/45	212/6	109/61	305/29	Fig. 2j
(293/60)	(113/68)	(293/22)	(23/0)	Fig. 2k [d]
132/22	296/21	166/20	34/21	Sykes and Sbar (1974)

[b] Similar solutions by Stéfansson (1966).
[c] This is an average solution for a number of small earthquakes.
[d] Non-orthogonal nodal planes.
[e] Similar solution by Solomon and Julian (1974).
[f] Similar solution by Solomon (1973).

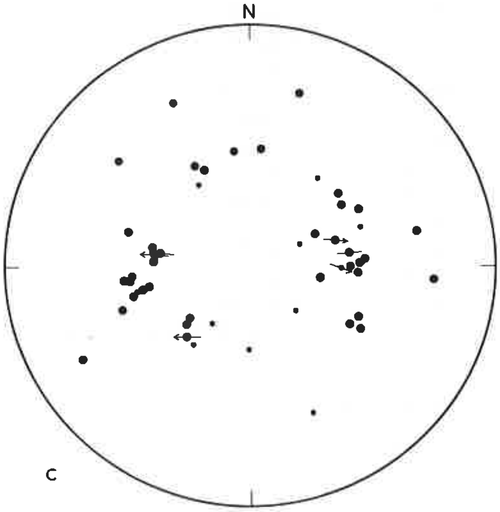
which appears on the epicentral map (Fig. 4) as a zone of seismicity, 80 km broad and 150 km long from east to west. The trend of the epicentral belt and the focal mechanism of a magnitude 7 earthquake (event no. *3* in Table I) near its western end led Sykes (1967) to conclude that it was a transform fault. The large width of the zone suggested to Ward (1971) and Tryggvason (1973) that the transform motion might be taken up by several faults. The NW striking Húsavík fault was identified as a part of this fault system (Tryggvason, 1973; Saemundsson, 1974). Saemundsson (1974) pointed out that the overall structure of the fracture zone was characterized by a series of parallel ridges and troughs with a northerly trend, i.e., transverse to the trend of the seismic belt. He suggested that some of these might have acted as spreading centers at one time or another. Einarsson (1976) used teleseismic data to relocate individual earthquakes in the eastern part of



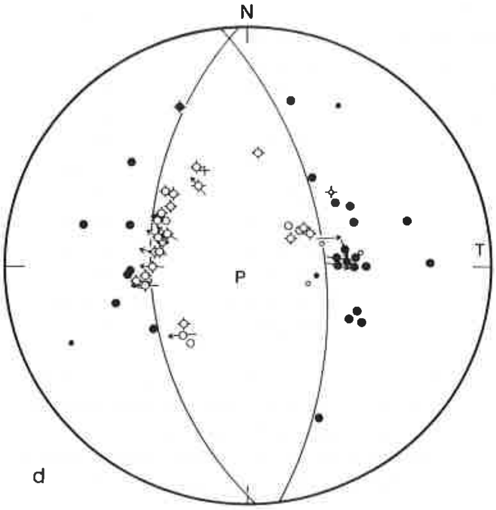
a



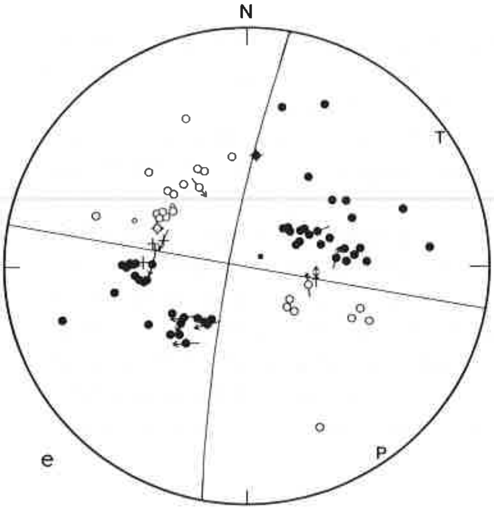
b



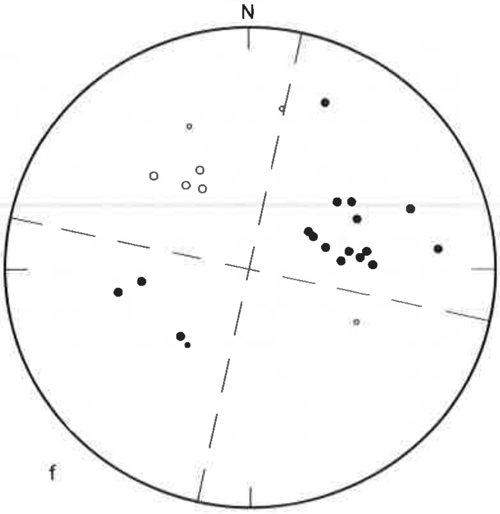
c



d



e



f

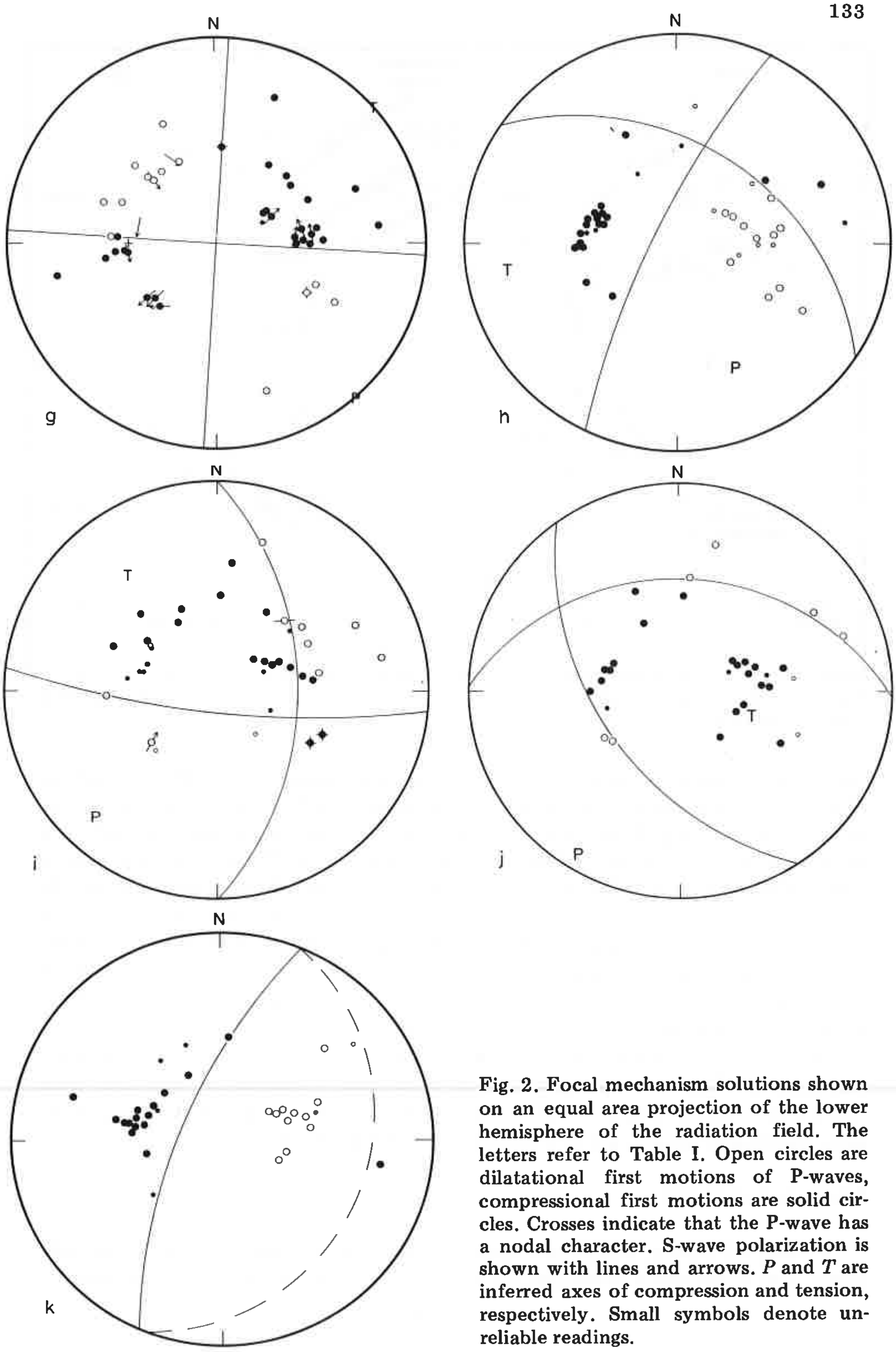


Fig. 2. Focal mechanism solutions shown on an equal area projection of the lower hemisphere of the radiation field. The letters refer to Table I. Open circles are dilatational first motions of P-waves, compressional first motions are solid circles. Crosses indicate that the P-wave has a nodal character. S-wave polarization is shown with lines and arrows. *P* and *T* are inferred axes of compression and tension, respectively. Small symbols denote unreliable readings.

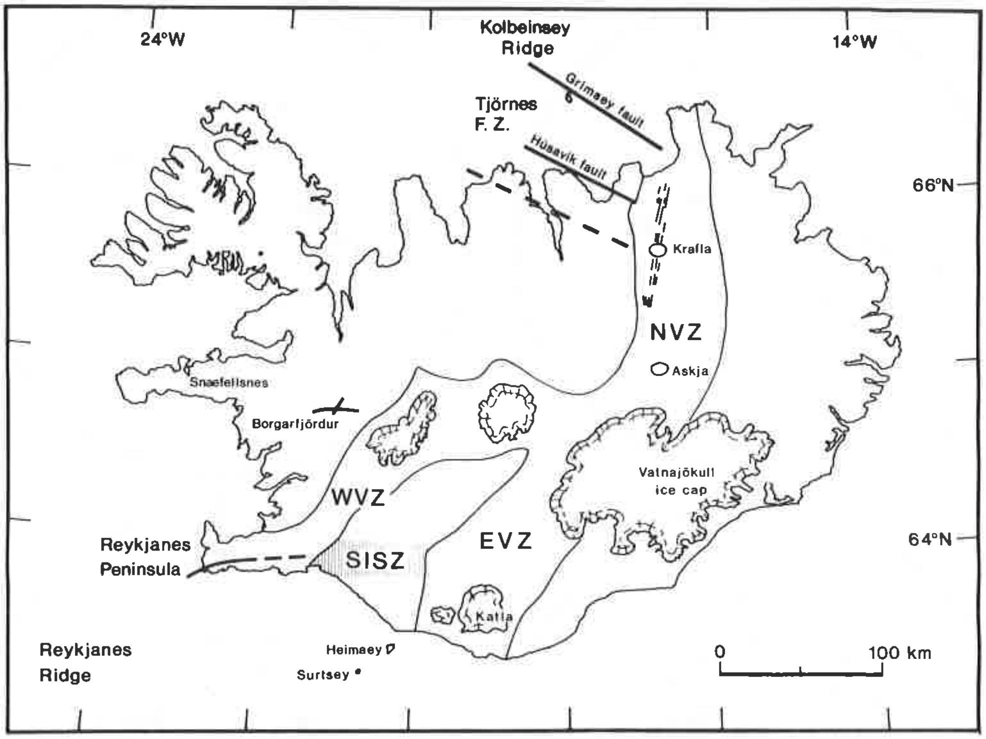


Fig. 3. Index map of Iceland showing the main tectonic elements, ice caps and geographic names mentioned in the text.

the Tjörnes fracture zone with respect to one master event. Thus a narrow line of seismicity was identified near the island Grímsey and was interpreted as a WNW striking fault (Fig. 3). The sense of motion along the fault was derived from a focal mechanism solution (event *1* in Table I) as right-lateral strike-slip. It was also found that a substantial part of the known seismicity of the Tjörnes fracture zone could be accounted for by slip along this Grímsey fault, the Húsavík fault and a third suggested fault farther to the south. All three faults run parallel to each other, 40 km and 30 km apart, respectively. These conclusions have been supported by data from a local short period seismograph system that was installed recently.

The pattern of parallel WNW striking faults is not revealed in the epicentral map in Fig. 4 which is based on epicentral locations of USCGS, later NOAA and USGS. Errors in these locations may be as large as 40 km (Einarsson, 1976) and do not allow tectonic interpretation on a fine scale.

The volcanic zones

A zone of young volcanism and rifting crosses Iceland from NE to SW. There are three main branches of this zone, the northern zone, the eastern

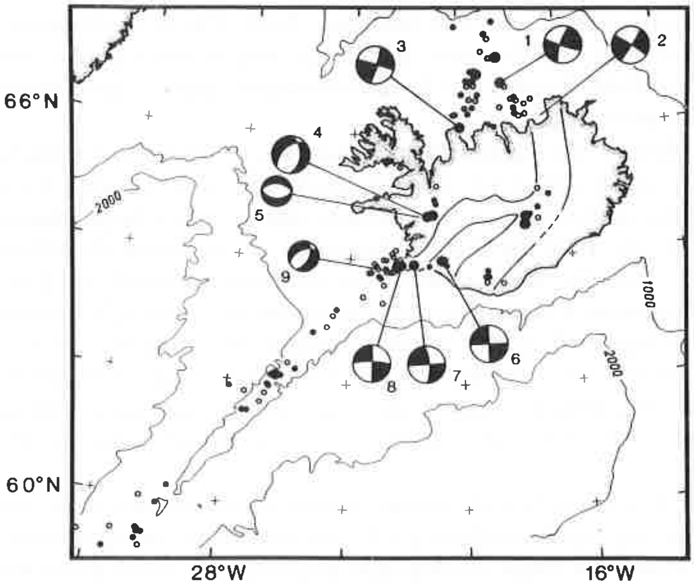


Fig. 4. Epicenters and focal mechanisms of earthquakes in Iceland and the northern part of the Reykjanes Ridge. Epicenters are taken from the PDE lists of USCGS, later NOAA and USGS, for the time period 1962—1974. Open circles denote epicenters determined with fewer than ten P-wave readings or epicenters of earthquakes smaller than $m_b = 4.5$. Solid circles are epicenters of events of $m_b = 4.5$ and larger that are determined with ten or more readings. Large circles are epicenters of events of $m_b = 5.0$ and larger. The focal mechanism solutions are shown schematically as lower hemisphere equal-area projections. The compressional quadrants are shown black. The bathymetry is taken from a map by the Icelandic Hydrographic Service, Reykjavik, 1975. Depths are given in meters. The volcanic rift zones of Iceland are shown.

zone and the western zone. The characteristic features of these zones are normal faults, open fissures and eruptive fissures. Often the faults and fissures are arranged in swarms, many of which pass trough a central volcano, where acidic volcanism, geothermal activity and sometimes caldera structures can be found (Saemundsson, 1974).

The best known of the central volcanoes is the Krafla volcano in the northern zone. Since 1975 this volcano and its associated fault swarm have been continuously active (Björnsson et al., 1977). In December 1975 a small basaltic eruption occurred within the Krafla caldera accompanied by a deflation of up to 2 m of the caldera region, rifting of the fault swarm and a large earthquake swarm that lasted about eight weeks. (These earthquakes are not shown in the epicentral map of Fig. 4.) Most of the earthquakes were confined to two separate areas. One area was within the caldera and the earthquakes there appeared to be related to the deflation process. Depth of the hypocenters was mostly 0—4 km. The largest earthquakes there were of $m_b = 5.0$. The other area was about 60 km north of caldera near the junction between the Krafla fault swarm and the Grímsey fault. These earthquakes were largely associated with the Grímsey fault. The largest earthquake was of $m_b = 6.0$ (event *2* in Table I) and the focal mechanism shows right lateral strike-slip along the Grímsey fault.

Since the initial deflation the Krafla caldera has been inflating at a more or less constant rate, interrupted by several short deflation events. Each deflation event has been accompanied by volcanic tremor and a large swarm of small earthquakes in the fault swarm to the north or to the south of the caldera. At the same time the fault swarm has widened as much as 2 m. Small eruptions accompanied a few of the deflation events. The inflation of the caldera region is interpreted as the result of magma accumulation at the depth of about 3 km under the caldera. The deflation events are then interpreted as the result of the pressure drop in the magma reservoir when a dyke is injected horizontally into the fault swarm to the north or to the south (Björnsson et al., 1977).

For the purpose of this paper the important lesson to be learned from the Krafla events is that an aggregate widening of 3 m can occur along a diverging plate boundary without a single earthquake large enough to be detected by seismic stations outside of Iceland. Almost all the earthquakes that were detected by the WWSSN during the Krafla events so far were associated either with deflation of a central volcano or with a transform fault.

In the epicentral map in Fig. 4 two groups of epicenters appear within the volcanic zones. One group is located in central Iceland. Two of the earthquakes occurred in the caldera of the Askja central volcano and were related to an eruption there in 1961. The rest of the epicenters are in an area that is covered by the ice cap Vatnajökull, where the structure of the volcanic zone is poorly known. Recent studies of ERTS images of this area seem to indicate that the structure is dominated by a group of central volcanoes (Thorarinsson et al., 1973).

The other group of epicenters is located near the southern end of the eastern volcanic zone south of its junction with the south Iceland seismic zone. The structure in this part of the volcanic zone is dominated by several central volcanoes, rifting structures are less significant. The earthquakes shown in Fig. 4 are related to the Katla central volcano. Results of a recent study of the seismicity near Katla shows that earthquakes occur down to a depth of 30 km in this area. The earthquakes were located with data from a local array of seismographs. The standard error of the depth determination is less than 5 km.

The eruptions of Surtsey 1963—1967 and Heimaey 1973 occurred near the extreme southern end of the eastern volcanic zone, about 60—80 km WSW of Katla. Both eruptions were accompanied by seismic activity, but no earthquake reached magnitude 4. The earthquakes during the Heimaey eruption occurred at depth of about 20 km (Björnsson and Einarsson, 1974) and were probably associated with the magma source.

The depth of the Heimaey and Katla earthquakes is much larger than observed elsewhere in Iceland. In these areas the boundary between layer 3 ($v_p = 6.5$ km s^{-1}) and layer 4 ($v_p = 7.2$ km s^{-1}), interpreted to be the crust-mantle boundary by Pálmason (1971), is at the depth of 12—15 km. Earthquakes at the depth of 20—30 km may be taken to imply brittle failure in the mantle where creep or ductile behaviour is normally assumed. In these volcanic regions it is possible, however, that high strain rates associated with magmatic processes may cause brittle failure in material that would be ductile at lower strain rates.

The following conclusions can be drawn about the earthquake activity in the volcanic rift zones of Iceland:

(1) Earthquakes larger than magnitude 4 are usually associated with central volcano complexes.

(2) Earthquakes associated with typical rifting structures such as normal faults, open fissures and volcanic fissures are usually smaller than magnitude 4.

(3) In some areas earthquakes can occur at depths as large as 30 km, i.e., well below the crust.

The south Iceland seismic zone

A zone of historic, destructive earthquakes bridges the gap between the two volcanic zones in south Iceland. The epicenters of the earthquakes line up in the E—W direction near 64° and the size of the largest shocks is estimated as exceeding magnitude 7 (Tryggvason, 1973), Stefánsson (1967) interpreted this zone as a shear zone and Ward et al. (1969) and Ward (1971) used the term fracture zone for this feature. In spite of the clear E—W orientation of the seismic zone there is no indication on the surface of a major E—W fault in this area. Faulting has been seen on the surface during some of the large earthquakes, but the faults had a northerly strike and movement was right-lateral strike-slip (Tryggvason, 1973). This sense of motion implies a least compressive stress in a horizontal NW—SE direction and a maximum compressive stress in a NE—SW direction. This stress field is consistent with the focal mechanism solution of event no. *6* in Table I, which is not well constrained (Ward, 1971) but is the only one available in this area. This is also the same stress orientation that would be predicted by a transform fault interpretation of the seismic zone. For some reason the crust seems to respond to this stress field by movement along a series of northerly striking faults instead of one major easterly striking fault.

The most recent major earthquakes in the south Iceland seismic zone occurred in 1896 (magnitudes 6.5—7.5) and in 1912 (magnitude 7, Tryggvason, 1973). At present the zone is very quiet, even at the microearthquake level. If one considers the previous history of seismic activity in this region and estimates the strain rate from the spreading rate in the Iceland area, one can conclude that major earthquakes are to be expected in this zone in the next decades.

The Reykjanes Peninsula

The Reykjanes Peninsula forms a transition between the Reykjanes Ridge to the west and the western volcanic zone and the south Iceland seismic zone to the east. The peninsula is characterized by high seismicity, recent volcanism and rifting structures, that occupy an elongated area with a trend of N70°E. The rifting structures, however, have a strike of N45°E and are arranged en échelon within the active zone. This structural relationship was interpreted by Tryggvason (1968) as the expression of a deep seated strike-slip fault possibly with a component of opening. This interpretation is supported by the two teleseismic focal mechanism solutions (events *7* and *8* in Table I) available for the central and eastern part of the peninsula.

Detailed studies of the seismicity of the central and western part of the Reykjanes Peninsula revealed that most of the earthquakes occur within a narrow zone. This zone runs along the peninsula, parallel to the zone of rifting and volcanism but is much narrower, less than 2 km wide in most places (Klein et al., 1973, 1977). The large scatter in the epicenters shown in Fig. 4 for this area is therefore caused entirely by location errors. The depths of the earthquake sources were shown to be mostly between 2 and 5 km, but some were near the surface and others as deep as 10 km.

This seismic zone is interpreted as the plate boundary. It is not a single fault, the earthquakes are caused by deformation of the brittle crust above a deeper seated and aseismic deformation zone. Small scale structures could be resolved in the seismicity within the zone. Thus several seismic lineations or faults could be identified that had a strike oblique to the main zone.

Focal mechanism solutions were determined for more than 400 earthquakes on the peninsula. Most of them belong to a swarm of more than 17000 recorded events that ruptured about 12 km segment of the plate boundary in September 1972 (Klein et al., 1977). A common feature of a large majority of all the mechanism solutions was the horizontal NW—SE orientation of the minimum compressive stress axes. The maximum compressive stress axes were vertical (causing normal faulting), inclined (oblique faulting) or horizontal (strike-slip faulting). Different types of faulting may occur close together, but on the whole there seems to be a systematic variation along the seismic zone. Thus strike-slip faulting seems to be more common in the central and eastern part of the peninsula, but normal and oblique faulting dominate towards the western tip of the peninsula. The focal mechanism shown schematically near the western tip of the peninsula in Fig. 4 is a typical mechanism for that part of the zone but does not represent a single earthquake.

Similar regional variation can also be seen in the orientation of the seismic zone and in the mode of strain release. Near the middle of the peninsula the seismic zone trends nearly E—W, but towards the west the orientation becomes more southerly and is N60°E near the tip of the peninsula.

The strain release in the eastern part of the peninsula seems to occur in mainshock—aftershock sequences, where the mainshock may reach magnitude 6 or more. The 1968 earthquake (event *7* in Table I) may be a typical example. Towards the west earthquake swarms become more common. A large earthquake sequence in September 1973 in the middle part of the peninsula consisted essentially of two earthquakes and their aftershocks. The first and largest event of the sequence (event *8* in Table I) had a strike-slip mechanism. The other earthquake occurred the following day and was located farther to the west. A mechanism solution could not be determined, but there were indications of a component of normal faulting. Finally, the earthquake sequence of September 1972 near the tip of the peninsula was an example of a typical earthquake swarm with no principal shock.

In conclusion, the Reykjanes Peninsula appears to be a zone of gradual transition from a transform fault zone in the east to an obliquely spreading ridge to the west. This transition is reflected in the orientation of the plate boundary, earthquake focal mechanisms and the mode of strain release.

Intraplate earthquakes

A long sequence of earthquakes occurred in the Borgarfjördur district of west Iceland in 1974. This sequence was a peculiar combination of an earthquake swarm and a mainshock—aftershock sequence with the largest event of $m_b = 5.5$.

The earthquakes occurred west of the western volcanic zone in a two million years old crust, i.e. they were associated with deformation of the American plate. In Fig. 4 it appears as if the epicenters lie in a N—S trending zone. This apparent trend is caused by location errors. A detailed study of these events revealed that the earthquakes occurred mainly within an E—W trending zone (Einarsson et al., 1977). Another less prominent zone with a NE—SW trend intersected the E—W zone in the middle. The hypocentral depths were mostly between 0—7 km, but a few events were as deep as 10 km. These depths are comparable to the depths of earthquakes on the Reykjanes Peninsula.

Focal mechanism solutions were found for the mainshock (event *4* in Table I) and for several small earthquakes in the western part of the epicentral zone (Einarsson et al., 1977). The mainshock solution was based on teleseismic data and showed normal faulting with the tensional axis in the WNW—ESE direction. The earthquake was probably associated with faulting along the NE—SW trending epicentral zone. The mechanism solutions for the western part of the epicentral zone were obtained using portable seismographs and were all characterized by normal faulting on easterly striking faults. The tensional axes clustered around the N—S direction, and the pressure axes were vertical. Faulting seen at the surface in this part of the zone is in agreement with these solutions. A typical solution is shown schematically in Fig. 4. It is concluded that the crust in the Borgarfjördur area is undergoing horizontal extension.

Even though the focal mechanism solutions of earthquakes in the seismic zones in north and south Iceland are in a general way consistent with a transform zone interpretation, there appears to be a significant discrepancy. In southwest Iceland the easterly striking nodal planes of the focal mechanisms have a strike close to 90°. In the Tjörnes fracture zone, on the other hand, the easterly striking nodal planes strike between 107° and 129°. If we assume that the plates on either side of the plate boundary are rigid we must conclude that the spreading pole is south of Iceland, which grossly contradicts the results of other investigators (e.g., Pitman and Talwani, 1972; Minster et al., 1974) and is very unlikely. We therefore conclude that the plates are not rigid, and that there must either be a net contraction east of the volcanic zones in Iceland or a net extension in central and west-central Iceland. The occurrence of normal faulting in Borgarfjördur and volcanism on the Snaefellsnes peninsula in west-central Iceland is consistent with crustal extension in these areas. The apparent westward divergence of the Icelandic transform zones can thus be explained by crustal extension in the area between them.

This crustal extension may be caused by subcrustal flow radially away from a mantle plume under Iceland (Morgan, 1971; Sigvaldason et al., 1974), but could also be the result of gravitational stresses induced by the regional topographic high of Iceland.

THE REYKJANES RIDGE

The Reykjanes Ridge extends from Iceland in the north to the Charlie-Gibbs fracture zone in the south. Only few sections of the mid-oceanic ridge system have been studied in as much detail as this one (e.g., Heirtzler et al., 1966; Talwani et al., 1971; Fleischer, 1974). The Reykjanes Ridge is often thought of as a type example of a spreading mid-oceanic ridge. The linear magnetic anomalies are well developed and suggest a spreading rate of 1.0 cm/year (Talwani et al., 1971). The characteristics of the ridge, however, are not homogeneous along its length. The topography changes from being relatively smooth in the northern part to being rough in the southern part, where a pronounced central rift valley is also present (Fleischer, 1974). The northern part has a trend of N35°E, but south of 56°N the trend changes to a nearly N—S direction. If the spreading direction is assumed to be parallel to the Charlie-Gibbs fracture zone we must conclude that oblique spreading occurs along a large part of the Reykjanes Ridge.

The transition from the plate boundary on the Reykjanes Peninsula to the crest of the Reykjanes Ridge appears to be gradual. The seismic zone makes a gentle curve and the en échelon structures on the peninsula continue off shore (Jakobsson, 1974). The en échelon pattern is probably a result of the oblique spreading. The change in seismicity along the peninsula noted earlier also continues off shore.

Large variations in the seismic activity along the ridge have been noted.

North of 60° N the seismicity is low, to the south the activity is considerably higher. This pattern has persisted for the past fifty years at least (Francis, 1973). In addition to the large scale variation in seismic activity there also appears to be a small scale variation. In the epicentral maps of Figs. 4 and 5 epicenters cluster near 54° N, 56° N, 57° N and 58° N. These clusters are not related to transform faulting. Three focal mechanism solutions within the centers of activity clearly show that crustal extension perpendicular to the ridge is the fundamental process. Seismicity gaps also exist, the most prominent ones near 60°—61.5° N and 54.5°—55.5° N.

Similar seismicity pattern can be found within the volcanic rift zones of Iceland and it is tempting to draw an analogy. In Iceland the centers of seismic activity are related to groups of central volcanoes, whereas the typical

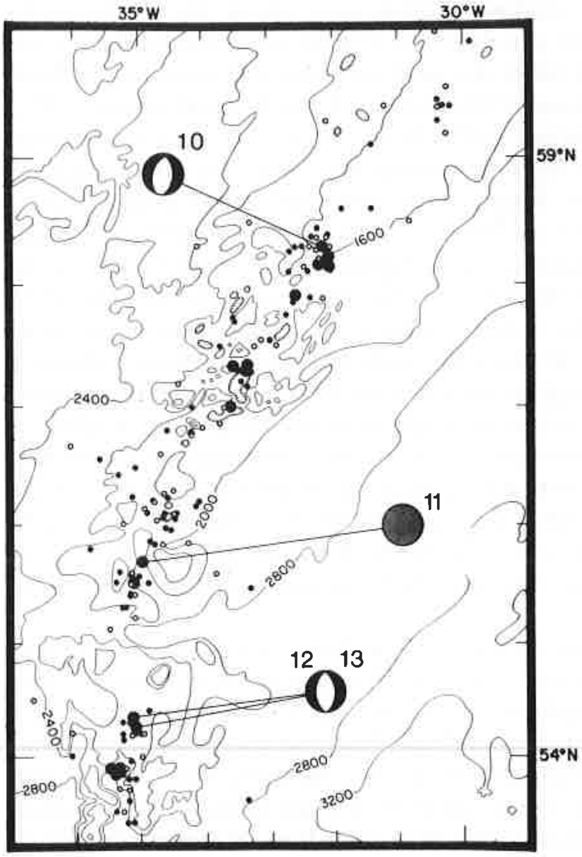


Fig. 5. Epicenters and focal mechanisms of earthquakes on the southern part of the Reykjanes Ridge. Epicentral data source and symbols are the same as in Fig. 3. The bathymetry is taken from an unpublished Bathymetric Atlas of the Atlantic Ocean, Caribbean and Gulf of Mexico, compiled at the Woods Hole Oceangraphic Institution by E. Uchupi. Depths are given in meters.

rift zones are practically void of earthquakes larger than magnitude 4. Perhaps the high seismic activity of the southern part of the Reykjanes Ridge can be related to central volcanoes that might be particularly well developed in this area. Perhaps the low seismic activity on fast spreading ridges such as the East Pacific rise could be explained, if it turned out that central volcanoes were absent there. It would be interesting to test these suggestions by combining detailed structural and petrological studies with seismograph observations on the ocean bottom.

An interesting corollary to the analogy with Iceland is that earthquakes might be expected to occur down to depths of 20—30 km in some areas. No evidence has been found sofar for earthquakes at such depths. Weidner and Aki (1973) found the hypocenters of two dip-slip earthquakes on the Mid-Atlantic Ridge to be shallower than 5 km.

Several authors who have studied earthquakes along the mid-oceanic ridges have found that the nodal planes of the P-wave radiation field of some earthquakes are not orthogonal when the rays are projected back to the focal sphere in the standard way (Sykes, 1967, 1970a; Thatcher and Brune, 1971; Conant, 1972; Solomon and Julian, 1974; Klein et al., 1977). This unusual behaviour is found for normal faulting earthquakes only. On the Reykjanes Ridge three earthquakes were found with non-orthogonal P-wave nodal planes (events *10*, *12*, and *13* in Table I). The angle between the nodal planes is 52° for event *10*. Solomon and Julian (1974) studied this event with similar results. Events *12* and *13* occurred close to each other both in space and time and had the same P-wave pattern. Fig. 2 shows only the focal mechanism solution for event *12*. Here the angle between the nodal planes is 58°. The S-wave polarization suggests that the strike of the nodal planes may be more easterly, which would make the angle between them even smaller.

A common feature of all these earthquakes is the emergent appearance of a large part of the P-waves in the dilatational quadrant of the focal sphere (see Fig. 2). The first motion often shows evidence of being close to a nodal plane even though the ray clearly intersects the focal sphere well within the dilatational quadrant. The compressional first motions, on the other hand, are mostly impulsive.

A most unusual P-wave radiation pattern was found for event *11*. For this event no dilatational first motions could be found and compressional first motions almost fill the focal sphere. If one accepts a few inconsistencies a solution of the thrust faulting type could be fitted to the P-wave data, but the available S-wave data seem to exclude such a solution. The S-waves suggest a normal faulting solution, but it appears as if the dilatational quadrant has shrunk down to almost nothing. We may here have an extreme case of non-orthogonal nodal planes.

A few explanations of the apparent non-orthogonality of the nodal planes have been given in the literature:

(1) The apparent non-orthogonality is an artefact of the projection used to map the earth's surface back to the focal sphere (Solomon and Julian,

1974). The earthquake is caused by a usual shear failure and the orthogonal nodal planes are distorted when the rays pass through a laterally inhomogeneous mantle.

(2) The earthquake source has an explosive component superposed on the double couple (Solomon and Julian, 1974). The nodal surfaces are no longer planar but could be approximated by two non-orthogonal nodal planes. An angle of 60° between the approximate nodal planes would be produced if the explosive component has a P-wave amplitude of 0.5 times the maximum P-wave amplitude of the double couple component. An amplitude ratio of 1 would produce an all-compressive focal sphere.

(3) The earthquake is produced by an extension failure of a porous, fluid-saturated medium (Robson et al., 1968). The pore fluid is considered to be magma. The material is brought to extensional failure by lowering the minimum compressive stress or by increasing the pore pressure. When failure occurs it is accompanied by a sudden drop in pore pressure. The nodal surfaces are not planar, but under certain conditions they may be approximated by two planes.

It does not seem to be easy to exclude any of these explanations on the basis of existing data. Any or all of them may be valid. In the first explanation, which is favoured by Solomon and Julian (1974), it is assumed that the earthquake is of purely tectonic origin. The other explanations would imply a closer connection with volcanic processes such as explosions and dyke formation. If there is similarity between the earthquakes of the Reykjanes Ridge and the earthquake in the volcanic zones of Iceland the latter two explanations might be more appropriate.

On the Reykjanes Peninsula Klein et al. (1977) found earthquakes with non-orthogonal nodal planes in a close proximity to earthquakes were the angle between the nodal planes was constrained to 90°. If the non-orthogonality is caused by inhomogeneities in the crust they must occur on a scale of a few hundred meters.

Solomon and Julian (1974) showed very convincingly how a low velocity lens in the mantle under the mid-oceanic ridge can bend the rays going downwards from the earthquake source towards the vertical. Thus the dilatational quadrant will appear smaller. This focussing would be expected to result in larger P-wave amplitudes within the dilatational sector. This is not confirmed by observations. On the contrary, P-waves in the dilatational quadrant have an emergent character. If the non-orthogonality is caused by a low velocity lens, we must therefore conclude that the low velocity material has a relatively low Q.

Seismic activity on the Reykjanes Ridge changes with time as well as in space. In Fig. 6 the earthquakes of 1962—1974 are plotted on a space—time diagram. The difference in seismicity between the northern and the southern part of the ridge can be seen very clearly. Temporal variations can be seen in the area between 53° and 58°N. Before 1967 the activity was high. This active period ended with a large earthquake swarm near 56°N in 1967. Dur-

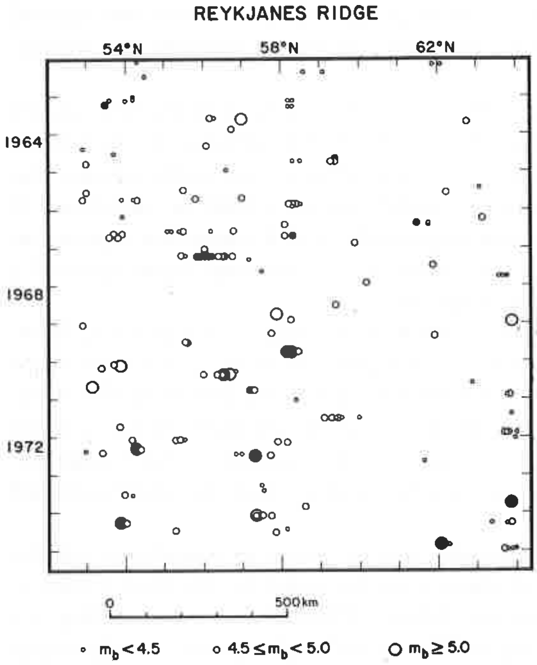


Fig. 6. Space—time diagram of earthquakes along the Reykjanes Ridge. The time of occurrence is plotted as a function of latitude of the epicenter. Solid symbols denote more than one earthquake.

ing the following two years the activity was very low, but since 1970 it has been high again. On the Reykjanes Peninsula (near 64°N) the activity was low until 1967, then an active period started with an earthquake swarm near the southwest tip of the peninsula.

THE CHARLIE-GIBBS FRACTURE ZONE

South of 53°N the crest of the Mid-Atlantic Ridge is offset 350 km to the east by the Charlie-Gibbs fracture zone, one of the major fracture zones of the Atlantic Ocean. The structure of this zone has been described in considerable detail by Johnson (1967), Fleming et al. (1970) and Olivet et al. (1974). The topographic expression of the fracture zone is very pronounced. The structure is dominated by two parallel troughs with a trend of 95°. The troughs are typically V-shaped, about 20 km wide and 1000—3000 m deep, measured from the top of the ridges to the bottom of the trough. In the western part of the transform fault section the northern trough is more pronounced, but in the eastern part the southern trough seems to be better developed. The northern trough is connected to the rift valley of the Reykjanes Ridge, but the southern trough is connected to the rift valley to the south.

The seismic activity is almost entirely confined to the transform fault part of the fracture zone, and most of it seems to be associated with the northern trough (Fig. 7). From the topography, however, one might conclude that both troughs have been active recently. The forteen years period studied here does not seem to be long enough to reveal the full seismicity pattern. This is hardly surprising in the light of the experience from the Tjörnes fracture zone, where two or more parallel faults exist. The full cycle of activity may take many decades, and each single fault may be seismically quiet for a considerable part of that time. In the Charlie-Gibbs fracture zone the transform motion could be taken up by two parallel transform faults, but it is also possible that the transform motion is confined to the western part of the northern trough and the eastern part of the southern trough. In the latter case one might expect to find a short spreading center connecting the two troughs, probably near 32°W. Clear evidence for such a spreading center has not been found yet.

Focal mechanism solutions have been found for four earthquakes in the northern trough (events *14*, *15*, *16* and *17* in Table I). All solutions are remarkably similar and indicate right-lateral strike-slip along the trough in full accord with its transform fault nature.

Near the intersection of the southern trough and the rift valley to the south one focal mechanism solution is characterized by oblique normal faulting (event *18* in Table I). This part of the plate boundary may be transitional

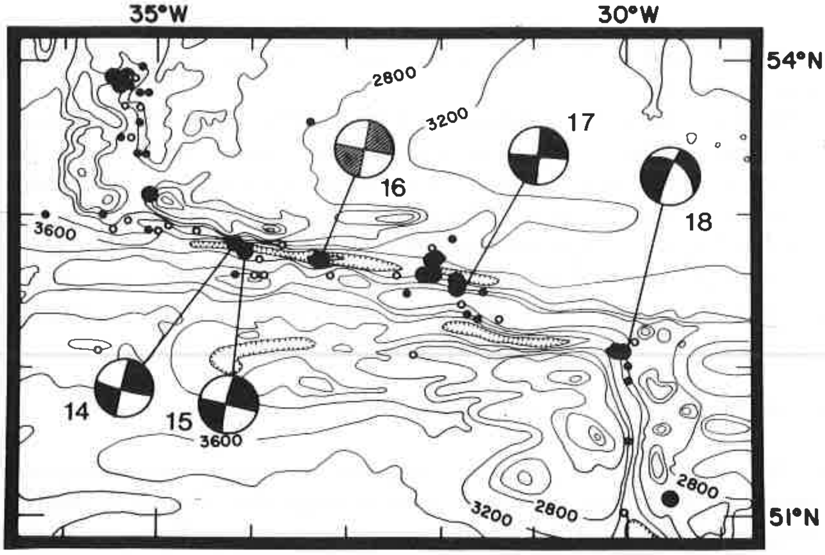


Fig. 7. Epicenters and focal mechanisms of earthquakes in the Charlie-Gibbs fracture zone. Symbols and data sources are the same as in Fig. 5.

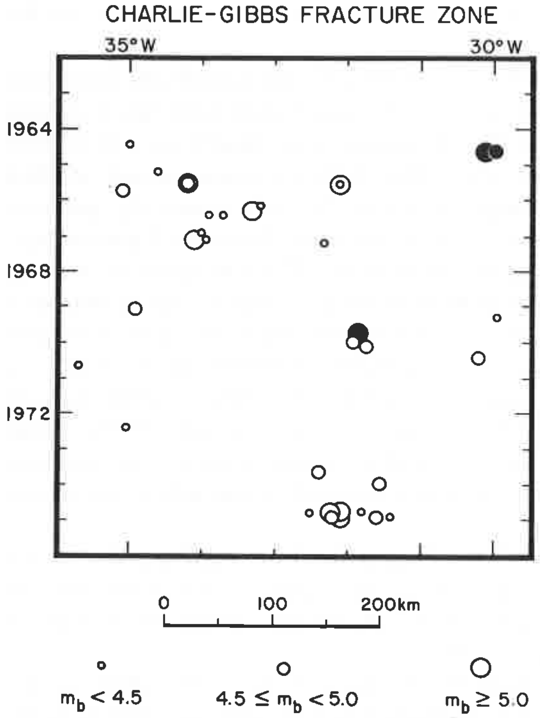


Fig. 8. Space—time diagram of earthquakes in the Charlie-Gibbs fracture zone. The time of occurrence is plotted as a function of longitude of the epicenter. Solid symbols denote more than one earthquake.

between a transform fault and a ridge crest, possibly similar to the Reykjanes Peninsula plate boundary.

Temporal fluctuations in the seismicity of the Charlie-Gibbs fracture zone are large. The period 1965—1967 was a particularly active period (Fig. 8). During one interval of six and a half days in 1965, for example, five earthquakes occurred distributed over a 230 km long segment of the fault (Sykes, 1970b). Since 1967 most of the seismic activity has been limited to a short segment of the fault near 32°W. It is interesting to note that the period 1965—1967 was also a very active period on the Reykjanes Ridge immediately to the north of the Charlie-Gibbs fracture zone (see Fig. 6).

THE MID-ATLANTIC RIDGE BETWEEN 40° AND 52°N

South of the Charlie-Gibbs fracture zone the seismicity belt continues without a major offset to 40°N where it joins with the Azores—Gibraltar plate boundary (Figs. 9 and 10). Small fracture zones have been described near 41°N (the Kurchatov fracture zone, see, e.g., Searle and Laughton, 1977) and suggested near 51°20′N (Johnson, 1967). A focal mechanism

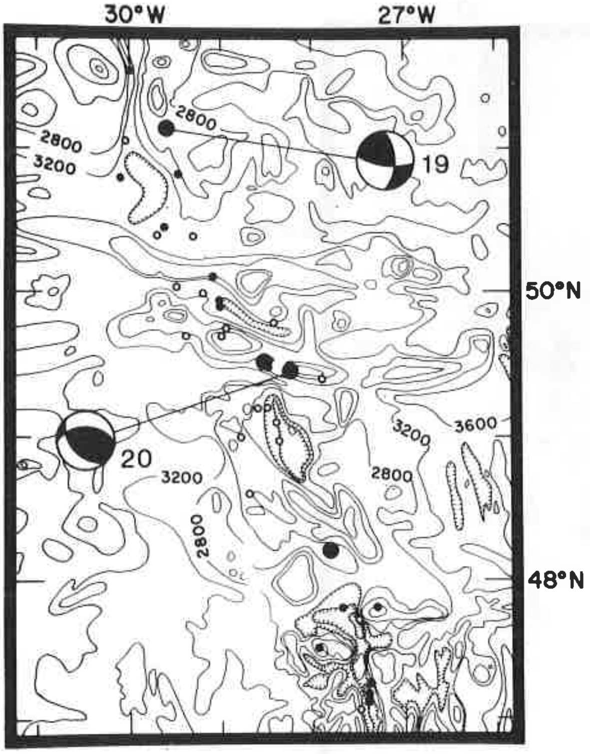


Fig. 9. Epicenters and focal mechanisms of earthquakes along the Mid-Atlantic Ridge between 47° and 52°N. Symbols and data sources are the same as in Fig. 5.

solution for an earthquake in the latter area (event *19* in Table I) has a large component of left-lateral strike-slip faulting consistent with a transform fault. But the mechanism also has a significant component of thrust faulting which is rather atypical of transform fault earthquakes. There is no other seismological evidence for a transform fault along this section of the Mid-Atlantic Ridge.

Between 48° and 51°N the seismic zone has a general trend of NNW. Oblique spreading must therefore occur along this plate boundary if the spreading direction is assumed to be parallel to the Charlie-Gibbs fracture zone. According to Johnson and Vogt (1973) the structure of this part of the Mid-Atlantic Ridge is characterized by alternating N—S trending and oblique spreading axes. The N—S axes are associated with transverse basement ridges, that trend slightly north of the spreading direction on both sides of the plate boundary. This "herringbone" pattern in the topography was interpreted by Johnson and Vogt (1973) as the result of asthenospheric flow southwards from the Iceland hot spot.

Focal mechanism studies in this region show some unexpected results.

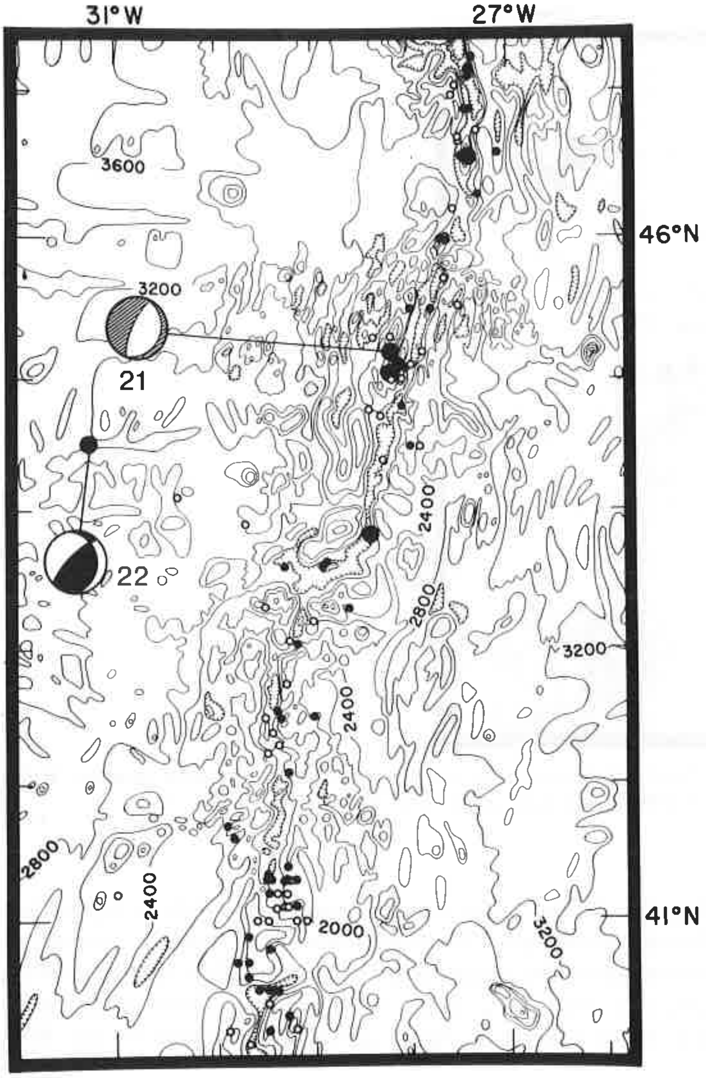


Fig. 10. Epicenters and focal mechanisms of earthquakes along the Mid-Atlantic Ridge between 40° and 47°N. Symbols and data sources are the same as in Fig. 5.

Event no. *20* is located on one of the transverse ridges near a N—S spreading axis, but the focal mechanism is clearly of the thrust faulting type. The axis of maximum compression is horizontal and has a NNE—SSW trend, perpendicular to the basement ridge.

Earthquakes mechanisms with a large component of thrusting such as those for events *19* and *20* are very rare along the mid-oceanic ridges.

It is possible that some fraction of all earthquakes on spreading mid-oceanic ridges is caused by some other processes than normal faulting, but more likely these results suggest that there may be some unusual conditions in the crust in this area. Possible explanations include the following:

(1) Thrust faulting occurs when fault blocks are uplifted to form the rift mountains. Within an extensional regime, however, this uplift is more likely to be associated with normal faulting in the way described by Atwater and Mudie (1968) for the Gorda rise.

(2) Forcible intrusions within central volcano complexes may result in thrust faulting in the adjacent regions. The transverse ridges described by Johnson and Vogt (1973) may be the traces of such complexes.

(3) There is a regional stress field characterized by horizontal compressive stresses superimposed on the usual mid-oceanic ridge stress field. The origin of these stresses may be sought in the regional topography. This part of the ridge system is the topographically lowest between Iceland and the Azores. There is a gradual shoaling of the crest of the Mid-Atlantic Ridge both towards Iceland in the north and the Azores triple junction in the south (Johnson and Vogt, 1973; Vogt and Johnson, 1975). Gravitational stresses caused by this topography would be characterized by N—S compression. Similar stresses may also be generated in the crust above a convergence in

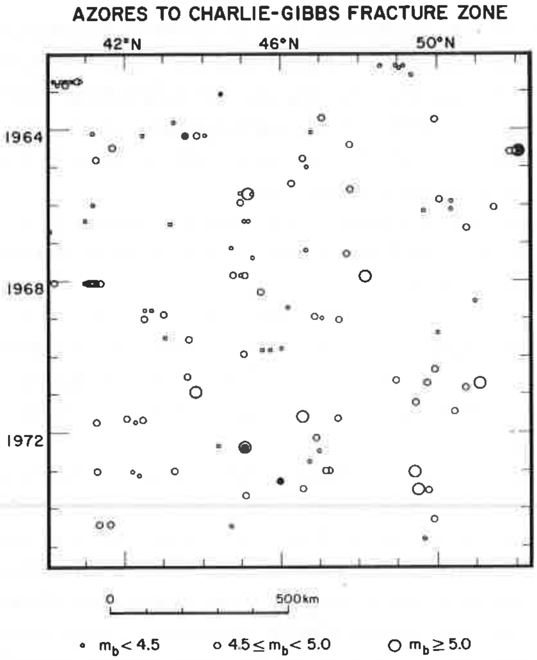


Fig. 11. Space—time diagram of earthquakes along the Mid-Atlantic Ridge between 40° and 52°N. The time of occurrence is plotted as a function of latitude of the epicenter. Solid symbols denote more than one earthquake.

asthenospheric flow. Such convergence may be expected to occur in this area if there is radial flow in the asthenosphere away from the hot spots under Iceland and the Azores. The resulting strain would be concentrated near the plate boundary because of the thin lithosphere there.

South of 48°N the plate boundary is marked by a more or less continuous rift valley. The seismic activity is fairly evenly distributed along the boundary both in time and space (Fig. 11) and earthquakes larger than magnitude 5 are relatively rare. Only one poorly constrained focal mechanism solution was obtained in this area (event *21* in Table I). One nodal plane is fairly well constrained and has a strike parallel to the rift valley and the axial magnetic anomaly in this region as determined by Loncarevic and Parker (1971). If minor inconsistencies are tolerated the other nodal plane can strike anywhere from N—S to E—W. Normal faulting is most likely in light of the local morphology, but the two nodal planes have to be nonorthogonal.

A focal mechanism solution was obtained by Sykes and Sbar (1973, 1974) for an earthquake located about 260 km west of the plate boundary (Fig. 9). The solution is of the thrust faulting type and is consistent with the observation of Sykes and Sbar that thrusting prodominates in oceanic crust older than 20 million years.

CONCLUSIONS

In this paper a summary is given of the seismicity and focal mechanisms of earthquakes along the plate boundary between Iceland and the Azores triple junction for the years 1962—1974. In general the results are consistent with the results of other authors (e.g., Sykes, 1967) that earthquakes along fracture zones are associated with transform faulting and that earthquakes along ridge axes are mainly associated with extensional tectonics.

In detail the relationship between the seismicity and the geological structure may be complex as shown by examples from Iceland. The main observations of this paper are summarized as follows:

(1) The Tjörnes fracture zone is a multiple transform fault, i.e., the transform motion between the Kolbeinsey ridge and the volcanic zones in Iceland is taken up by at least two parallel strike-slip faults. The most prominent structural elements within the zone are ridges and troughs that trend transversely to the main earthquake faults.

(2) Rifting in Iceland takes place within volcanic rift zones where the structure is characterized by swarms of normal faults, open fissures and eruptive fissures. Central volcano complexes occur within the rift zones and are frequently associated with the fault swarms. Magmatic processes within the central volcanoes are intimately related to rifting in the associated fault swarms as demonstrated by the presently active Krafla volcano. A large part of all earthquakes larger than magnitude 4 occurs within the central volcanoes and is apparently not directly associated with the rifting.

(3) In south Iceland the E—W transform zone is marked by epicenters of

large historic earthquakes. Surface rupturing observed during these earthquakes occurs on northerly striking faults. A gradual transition from strike-slip to normal faulting takes place in the western part of the Reykjanes Peninsula.

(4) There is an apparent discrepancy between the relative plate motion as derived from focal mechanisms of earthquakes in the Tjörnes fracture zone and the motion derived from focal mechanisms in southwest Iceland. The discrepancy can be explained if one allows for crustal extension within the American plate in West Iceland as occurred during the Borgarfjördur earthquakes in 1974.

(5) Focal mechanism solutions for earthquakes along the southern part of the Reykjanes Ridge show that maximum compressive stress is vertical and minimum compressive stress is horizontal and perpendicular to the plate boundary. A peculiar feature of these focal solutions is the apparent nonorthogonality of the nodal planes. Possible explanations of this phenomenon include an explosive component in the source function, extensional failure accompanied by a drop in fluid pressure and focusing effects of a low velocity body. An earthquake was found that appears to have an all-compressive P-wave radiation field, probably an extreme case of nonorthogonal nodal planes.

(6) Earthquakes of the Charlie-Gibbs fracture zone are associated with transform faulting. Both troughs of this fracture zone may be active although only the northern one has been seismically active during the last fifteen years.

(7) On the Mid-Atlantic Ridge between 48°N and 51°N the structure is characterized by basement ridges that trend transversely to the plate boundary (the herringbone pattern of Johnson and Vogt, 1973). Two earthquake focal mechanisms on the plate boundary in this area have a significant component of thrust faulting. Thrusting may be associated with forcible intrusion near the plate boundary or with formation of rift mountains but could also be the result of a regional compressive stress field caused by mantle plumes under Iceland and the Azores.

ACKNOWLEDGEMENTS

A large part of this work was done while the author was at Lamont-Doherty Geological Observatory of Columbia University and was partly supported by the National Science Foundation (Grant GA 43382) and the Icelandic Science Fund. The WWSSN record library at Lamont-Doherty was used extensively in this study. The following persons read the manuscript and offered constructive criticism: Sveinbjörn Björnsson, Leonard Johnson, Raul Madariaga, Sean C. Solomon, Eysteinn Tryggvason and an anonymous reviewer. Rósa Karlsdóttir typed the manuscript.

REFERENCES

Atwater, T.M. and Mudie, J.D., 1968. Block faulting on the Gorda rise. Science, 159: 729—731.

Björnsson, S. and Einarsson, P., 1974. Seismicity of Iceland. In: L. Kristjánsson (Editor), Geodynamics of Iceland and the North Atlantic Area. Reidel, Dordrecht, pp. 225—239.

Björnsson, A., Saemundsson, K., Einarsson, P., Tryggvason, E. and Grönvold, K., 1977. Current rifting episode in north Iceland. Nature, 266: 318—323.

Conant, D.A., 1972. Six New Focal Mechanism Solutions for the Arctic and a Center of Rotation for Plate Movements. Thesis, Columbia University, New York.

Einarsson, P., 1976. Relative location of earthquakes within the Tjörnes fracture zone. Soc. Sci. Islandica, Greinar Vol.: 45—60.

Einarsson, P., Klein, F.W. and Björnsson, S., 1977. The Bjorgarfjördur earthquakes of 1974 in West Iceland. Bull. Seismol. Soc. Am., 67: 187—208.

Fleischer, U., 1974. The Reykjanes Ridge — a summary of geophysical data. In: L. Kristjánsson (Editor), Geodynamics of Iceland and the North Atlantic Area. Reidel, Dordrecht, pp. 17—31.

Fleming, H.S., Cherkis, N.Z. and Heirtzler, J.R., 1970. The Gibbs fracture zone: a double fracture zone at 52°30′N in the Atlantic Ocean. Mar. Geophys. Res., 1: 37—45.

Francis, T.J.G., 1973. The seismicity of the Reykjanes Ridge. Earth Planet. Sci. Lett., 18: 119—124.

Heirtzler, J.R., LePichon, X. and Baron, J.G., 1966. Magnetic anomalies over the Reykjanes Ridge. Deep-Sea Res., 13: 427.

Jakobsson, S., 1974. Eldgos vid Eldeyjarboda. (Volcanic eruptions at Eldeyjarbodi, the Reykjanes Ridge; in Icelandic with English abstract.) Náttúrufraedingurinn, 44: 22—40.

Johnson, G.L., 1967. North Atlantic fracture zones near 53°. Earth Planet. Sci. Lett. 2: 445—448.

Johnson, G.L. and Vogt, P.R., 1973. Mid-Atlantic Ridge from 47° to 51°N. Geol. Soc. Am. Bull., 84: 3443—3462.

Klein, F.W., Einarsson, P. and Wyss, M., 1973. Microearthquakes on the mid-Atlantic plate boundary on the Reykjanes Peninsula in Iceland. J. Geophys. Res., 78: 5084—6099.

Klein, F.W., Einarsson, P. and Wyss, M., 1977. The Reykjanes Peninsula, Iceland, earthquake swarm of September 1972 and its tectonic significance. J. Geophys. Res., 82: 865—888.

Loncarevic, B.D. and Parker, R.L., 1971. The Mid-Atlantic Ridge near 45°N. XVII. Magnetic anomalies and ocean floor spreading. Can. J. Earth Sci., 8: 883—898.

Minster, J.B., Jordan, T.H., Molnar, P. and Haines, E., 1974. Numerical modelling of instantaneous plate tectonics. Geophys. J.R. Astron. Soc., 36: 541—576.

Morgan, W.J., 1971. Convection plumes in the lower mantle. Nature, 230: 42.

Olivet, J.-L., LePichon, X., Monti, S. and Sichler, B., 1974. Charlie-Gibbs fracture zone. J. Geophys. Res., 79: 2059—2072.

Pálmason, G., 1971. Crustal Structure of Iceland from Explosion Seismology. Soc. Sci. Islandica, Rit 40, 187 pp.

Pitman, W.C. and Talwani, M., 1972. Sea-floor spreading in the North Atlantic. Geol. Soc. Am. Bull, 83: 619.

Robson, G.R., Barr, K.G. and Luna, L.C., 1968. Extension failure: an earthquake mechanism. Nature, 218: 28—32.

Saemundsson, K., 1974. Evolution of the axial rifting zone in Northern Iceland and the Tjörnes fracture zone. Geol. Soc. Am. Bull., 85: 495.

Searle, R.C. and Laughton, A.S., 1977. Sonar studies of the Mid-Atlantic Ridge and Kurchatov Fracture Zone. J. Geophys. Res., 82: 5313—5328.

Sigvaldason, G.E., Steinthorsson, S., Óskarsson, N. and Imsland, P., 1974. Compositional variation in recent Icelandic tholeiites and the Kverkfjöll hot spot. Nature, 251: 579—582.

Solomon, S.C., 1973. Shear-wave attenuation and melting beneath the Mid-Atlantic Ridge. J. Geophys. Res., 78: 6044—6059.

Solomon, S.C. and Julian, B.R., 1974. Seismic constraints on ocean-ridge mantle structure: anomalous fault-plane solutions from first motions. Geophys. J.R. Astron. Soc. 38: 265—285.

Stefánsson, R., 1966. Methods of focal mechanism studies with application to two Atlantic earthquakes. Tectonophysics, 3: 209.

Stefánsson, R., 1967. Some problems of seismological studies on the Mid-Atlantic Ridge. In: S. Björnsson (Editor), Iceland and Mid-Ocean Ridges. Soc. Sci. Islandica, 38: 80.

Sykes, L.R., 1967. Mechanism of earthquakes and nature of faulting on the mid-ocean ridges. J. Geophys. Res., 72: 2131—2153.

Sykes, L.R., 1970a. Focal mechanism solutions for earthquakes along the world rift system. Bull. Seismol. Soc. Am., 60: 1749.

Sykes, L.R., 1970b. Earthquake swarms and sea-floor spreading. J. Geophys. Res., 75: 6598.

Sykes, L.R. and Sbar, M.L., 1973. Intraplate earthquakes, lithospheric stresses and the driving mechanism of plate tectonics. Nature, 245: 298—302.

Sykes, L.R. and Sbar, M.L., 1974. Focal mechanism solutions of intraplate earthquakes and stresses in the lithosphere. In: L. Kristjánsson (Editor), Geodynamics of Iceland and the North Atlantic Area. Reidel, Dordrecht, pp. 207—224.

Talwani, M., Windisch, C.C. and Langseth, M.G., 1971. Reykjanes Ridge crest: a detailed geophysical study. J. Geophys. Res., 66: 413—417.

Thatcher, W. and Brune, J.N., 1971. Seismic study of an ocean ridge earthquake swarm in the Gulf of California. Geophys. J.R. Astron. Soc., 22: 473.

Thorarinsson, S., Saemundsson, K. and Williams, R.S., 1973. ERTS-1 image of Vatnajökull: analysis of glaciological, structural and volcanic features. Jökull, 23: 7—17.

Tryggvason, E., 1968. Measurement of surface deformation in Iceland by precision levelling. J. Geophys. Res., 73: 7039.

Tryggvason, E., 1973. Seismicity, earthquake swarms and plate boundaries in the Iceland region. Bull. Seismol. Soc. Am. 63: 1327.

Vogt, P.R. and Johnson, G.L., 1975. Transform faults and longitudinal flow below the mid-oceanic ridge. J. Geophys. Res., 78: 1399—1428.

Ward, P.L., 1971. New interpretation of the geology of Iceland. Geol. Soc. Am. Bull, 82: 2991—3012.

Ward, P.L., Pálmason, G. and Drake, C.L., 1969. Microearthquakes and the Mid-Atlantic Ridge in Iceland. J. Geophys. Res., 74: 665—684.

Weidner, D.J. and Aki, K., 1973. Focal depth and mechanism of mid-ocean ridge earthquakes. J. Geophys. Res., 78: 1818—1831.

Wilson, J.T., 1965. A new class of faults and their bearing on continental drift. Nature, 207: 343—347.