The Geology of North America
Vol. M, The Western North Atlantic Region
The Geological Society of America, 1986

Chapter 7

Seismicity along the eastern margin of the North American Plate

Páll Einarsson
Science Institute, University of Iceland, Dunhaga 3, 107 Reykjavik, Iceland

INTRODUCTION

The eastern boundary of the North American Plate is marked by a narrow and continuous belt of seismic activity. The earthquake zone follows the crest of the Mid-Atlantic Ridge and its associated fracture zones (F.Z.) (Fig. 1; Plate 8A; Vogt, this volume, Ch. 12.) and is intimately related to the process of plate separation, transform faulting, and generation of oceanic crust. This part of the Mid-Atlantic plate boundary has the advantage of being relatively close to the dense seismograph networks of North America and Europe, and has therefore been the subject of more study than most other parts of the oceanic rift system. For this reason it has sometimes been taken as the type example of an oceanic rift, which can be misleading. Other segments of the ridge system, such as parts of the East Pacific Rise, have been shown to possess a distinctly different seismicity pattern (Stover, 1973), with most earthquakes occurring along the fracture zones and practically no seismicity along the ridge axes.

In this paper we will describe the main features of the seismicity of the mid-oceanic ridge system in the North Atlantic and Arctic Oceans and how it relates to the physical state and processes near the plate boundary. Intraplate earthquakes, that is events within the North American Plate, are dealt with by Zoback and others (this volume).

Because of the enormous amount of literature on the subject no attempt has been made to give complete reference to every paper. For more complete references the reader is referred to recent review papers on different aspects of the subject, for example those by Savostin and Karasik (1981) and Husebye and others (1975) on Arctic seismicity; Einarsson (1979) and Einarsson and Björnsson (1979) on the seismicity of Iceland and the ridge to the south; Duschenes and others (1983) and Whitmarsh and Lilwall (1983) on ocean bottom seismograph studies, Sanford and Einarsson (1982) on the detection of magma chambers in rifts; and Lilwall (1982) on the seismicity of oceanic rifts. Studies of focal mechanisms are summarized by Einarsson (1985).

Early studies of the world's seismicity identified the Atlantic Ocean as a seismically active area even though the distribution of known earthquakes was understandably limited to inhabited areas along the coast and on Atlantic islands. The identification of a continuous seismic zone in the middle of the ocean was first possible with instrumental observation, first by Tams (1922, 1927a, b) and later by Gutenberg and Richter (1941, 1949).

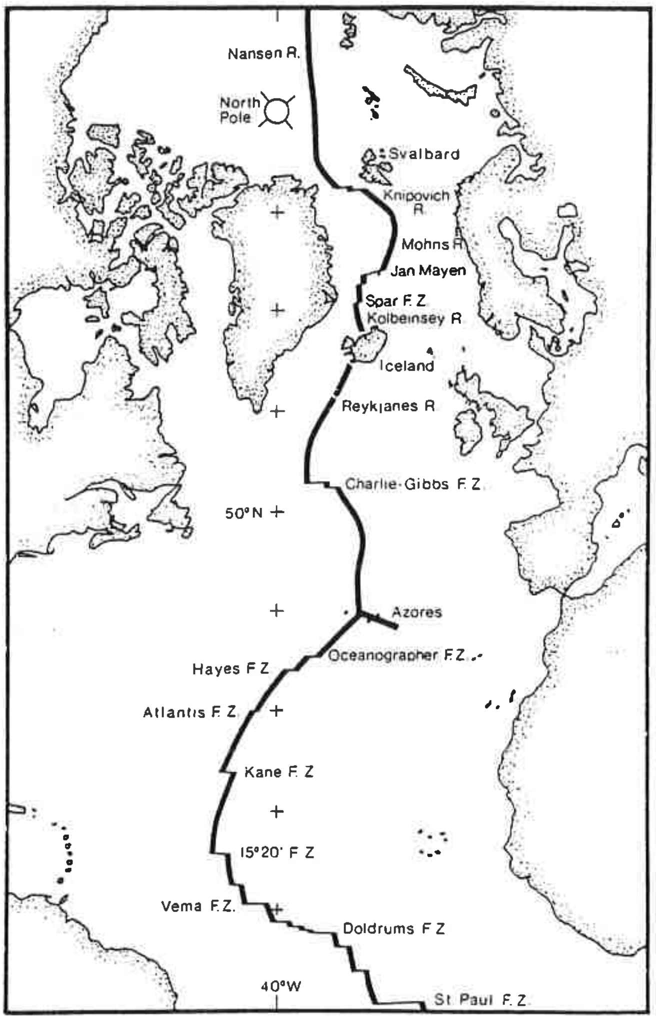

Figure 1. Index map of the plate boundaries in the North Atlantic and Arctic Oceans.

Einarsson, P., 1986 Seismicity along the eastern margin of the North American Plate; *in* Vogt, P. R., and Tucholke, B. E., eds., The Geology of North America, Volume M, The Western North Atlantic Region: Geological Society of America.

A major step in the attempt to define the seismic zones followed the implementation in the early sixties of the World Wide Standardized Seismograph Network (WWSSN), and the use of electronic computers and improved Earth models to locate earthquakes. Sykes (1965) published a map of relocated epicenters of the period 1955–1964 in the Arctic region, which revealed more details than possible before. The continuity and narrowness of the seismic zone was further established, as well as its coincidence with rifted mid-oceanic ridges, an extension of the Mid-Atlantic Ridge into the Arctic.

The increased resolution of the seismic networks coincided with a rapid increase in knowledge of sea floor morphology. It now became possible to correlate the seismic zones with specific topographic features such as fracture zones and rifts along ridge crests. New seismicity maps of the world, such as the one by Barazangi and Dorman (1969) showed that seismic zones form a continuous net encircling the Earth. The global network of long period seismographs offered new opportunities to obtain reliable fault plane solutions for earthquakes. Hypotheses on sea floor spreading and plate tectonics provided a new framework of thought, and were strongly supported by improved seismological data (Isacks and others, 1968). Seismic zones delineate lithospheric plates and fault plane solutions along plate boundaries indicate the relative plate movements. For oceanic plate boundaries the landmark paper was written by Sykes (1967), who demonstrated the main features of mid-oceanic ridge seismicity. Earthquakes along fracture zones were shown to be limited to the section between the adjacent ridge axes, and were accompanied by strike-slip faulting along the fracture zone, in an opposite sense to that expected for a simple offset of the ridge crest. This means that the offset of ridge segments across the fracture zones is not caused by transcurrent motions along the fracture zones. This was an important support to the transform fault hypothesis of Wilson (1965). Earthquakes along the ridge crests were found to be associated with normal faulting, indicating crustal extension in the axial zone. Further differences between ridge crest and fracture zone earthquakes were demonstrated by Francis (1968a,b), who found significantly different magnitude-frequency relationships for the two classes of events. Earthquake swarms were found to be frequent along the ridge axes (Sykes, 1970), whereas earthquakes along the fracture zones tended to occur in mainshock-aftershock sequences. The studies of Sykes and Francis showed that the more or less orthogonal system of ridges and fracture zones is not the expression of horizontal shear along conjugate fault planes, as had been suggested by some earlier authors (e.g. Van Bemmelen, 1964; Tr. Einarsson, 1968). In fact, the theory of sea-floor spreading provided the only consistent explanation for the distribution of earthquake foci and focal mechanisms along the Mid-Atlantic Ridge.

DATA

For obvious reasons one must rely on instrumental data when studying Mid-Atlantic Ridge seismicity, except in Iceland, where considerable information on historical seismicity is available. Instrumental data for this region come from three types of sources.

1. Teleseismic data give relatively homogeneous information over the North-Atlantic area. These data give epicentral location for all earthquakes larger than magnitude 4.5, and fault plane solutions can be obtained for earthquakes of body wave magnitude 5.5 and larger. In time the data are relatively homogeneous since 1963, that is, after the implementation of the WWSSN. Epicentral locations and fault plane solutions of earthquakes since 1963 are shown on Plate 11 (see also Plate 8A).

2. Local seismograph networks in Iceland give detailed information on the seismicity of that part of the ridge system. Four stations were in operation in the late sixties, but the number of short period, permanent stations was greatly increased in the early seventies and has since varied between 20 and 40 stations. In addition, dense, multielement networks have been in operation in different areas over short periods of time. These studies have yielded accurate hypocentral depths and fault plane solutions for a large number of small events.

3. Ocean-bottom seismographs have been deployed in a few areas on the Mid-Atlantic Ridge, notably near 1°N, 23°N, 37°N, and 45°N. For practical reasons, these studies are limited in space and time and their results must be interpreted with these limitations in mind.

FROM THE EQUATOR TO THE AZORES

This part of the Mid-Atlantic Ridge is cut by an unusually large number of fracture zones with large offsets (Vogt, this volume, Ch. 12). South of 25°N, almost all the fracture zones offset the ridge crest to the left, but to the north all offsets are to the right. This shapes the ridge system into an arc-like structure, concave towards east, reflecting the original shape of the continents at the time of break-up. The present plate boundaries between the North American and African Plates and the South American and African Plates are faithfully traced by the seismicity, which shows a narrow zone of brittle deformation. In most places the zone is 20 km wide or less, which is just about the resolution of the teleseismic locations. The boundary between the North and South American Plates, on the other hand, is not delineated by a well-defined seismic zone. The intraplate seismicity west of the Mid-Atlantic Ridge at latitudes 10–20°N appears to be slightly higher than normal, indicating plate deformation in a broad zone separating the North and South American Plates.

All major transform faults can be identified on the seismicity map by one or more of their seismic characteristics, that is east-west alignment of epicenters, offsets in the ridge crest seismic zone, and fault plane solutions indicating strike-slip faulting in the transform sense. All these characteristics have been found in the Vema, 15°20′, Kane, and the Doldrums fracture zones. Other major fracture zones, such as the Atlantis F. Z., are only identifiable on the seismicity map as an offset in the ridge crest seismicity. Most fracture zones have a clear E-W seismicity belt, up to 610 km in length. Examples of right-lateral and left-lateral transform faulting are shown in Figure 2.

All fault plane solutions obtained for earthquakes at or near the ridge crest in this sector show normal faulting. Although the orientation of the fault planes and stress axes cannot be determined with confidence, the stress conditions in the axial regions appear to be non-uniform and may change considerably over short distances and with time.

Two seismic peculiarities found near the Oceanographer and Hayes fracture zones are worth mentioning. A prominent cluster of epicenters is found west of the plate boundary, where the Oceanographer Fracture Zone joins the ridge axis (Fig. 2). This area has shown persistent activity over the last 20 years with several events of magnitude 5 or larger. The events occur as far as

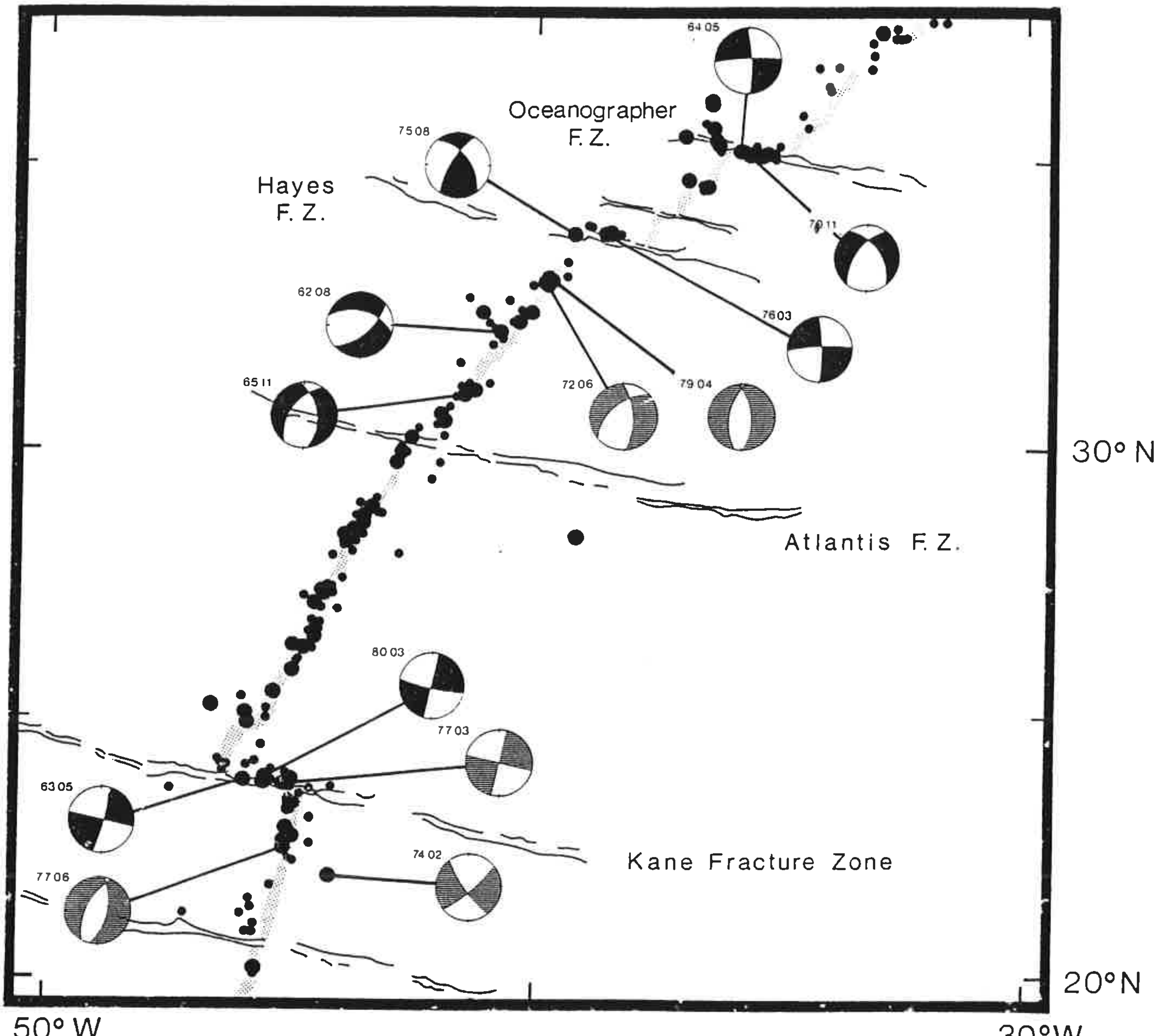

Figure 2. Epicenters and fault plane solutions along the Mid-Atlantic Ridge between 20 and 38°N, showing transform faulting in the Kane, Hayes, and Oceanographer fracture zones, and normal faulting along the ridge crest. Epicenters are from the PDE lists of the U.S. Geological Survey for the period 1963–1981, only epicenters determined with 10 or more stations are included. Fault plane solutions are shown schematically on lower hemisphere stereographic projection of the focal sphere, compressional quadrants are black. Numbers denote year and month of the event. Solutions are from Sykes (1967), Udias and others (1976), and Einarsson (1985).

100 km off the plate boundary, which is much more than the expected uncertainty in the locations. No unusual topographic features are known in this area, and the nature of this seismic activity is unclear. The other case is a fault plane solution of one event near the western end of the Hayes transform fault that shows a significant but unusual component of thrust faulting. Large scale vertical movements are indicated by the presence of transverse ridges in some of the major fracture zones (Bonatti, 1978; Bonatti and Chermak, 1981; Bonatti and others, 1983) and seem to be an integral part of the tectonic regime of a transform fault. Occasional occurrence of thrust faulting events in fracture zones should therefore not be too surprising.

This sector of the ridge system has been the object of several microearthquake surveys using both ocean bottom and free floating instruments. Duschenes and others (1983), studying the hypocentral resolution of microearthquake surveys carried out at sea, concluded that networks consisting of sonobuoys alone (Reid and MacDonald, 1973, Spindel and others, 1974) are only of marginal use in work of this kind and warned against overinterpretation of these data.

Earthquakes were recorded by ocean bottom instruments in the FAMOUS area (Francis and others, 1977). These originated both in the median valley and in the adjacent fracture zone, but since only two instruments were operational, too much significance should not be attached to the locations. Rowlett (1981) studied seismicity at the eastern ridge junction of the Oceanographer Fracture Zone, using two ocean bottom seismographs for about five days. Hypocenters could, of course, not be determined with any accuracy, but making some assumptions it was concluded, that the recorded seismicity was distributed across the corner between the ridge axis and the transform fault. Similar results were obtained by Rowlett and Forsyth (1984) for the western ridge intersection of the Vema Fracture Zone and by Francis and others (1978) for the eastern ridge intersection of the St. Paul Fracture Zone. In the latter study hypocentral depths could be resolved. Earthquakes were found to originate at 0–8 km depths.

To date, the most extensive ocean bottom seismic experiment on the Mid-Atlantic Ridge was conducted at the ridge axis near 23°N, slightly south of the Kane Fracture Zone (Toomey and others, 1985). Ten instruments were in operation for a period of three weeks. Earthquakes were located beneath the central valley and the eastern valley wall, with hypocentral depths between 5 and 8 km. Composite fault plane solutions show normal faulting. These results indicate that this section is undergoing active extension and has cooled to temperatures within the brittle field of behavior to a depth of at least 7–8 km, even in the center of the median valley.

THE AZORES-GIBRALTAR ZONE

In the Atlantic, the Eurasian and African Plates are separated by a seismic zone that extends westward from the Straits of Gibraltar to form a triple junction with the Mid-Atlantic Ridge within the Azores archipelago. The nature and evolution of this plate boundary and the triple junction has been the subject of considerable debate. See for example Krause and Watkins (1970), McKenzie (1972), Laughton and Whitmarsh (1974), Udias and others (1976), Hirn and others (1980), Searle (1980), Udias (1982), and Moreira (1982).

The zone can be divided into three sections according to its seismic characteristics. The westernmost section extends from the triple junction, through the Azores, to the eastern end of the archipelago. On the seismicity map the zone appears relatively narrow, comparable to the zone on the Mid-Atlantic Ridge. Fault plane solutions show that strike-slip is the principal mode of faulting, one nodal plane with an easterly strike and the other striking northerly. It appears likely that the northerly striking nodal planes of the fault plane solutions are the fault planes, which are therefore not parallel to the main seismic zone or the trend of the archipelago (Hirn and others, 1980). The earthquakes here seem to occur in response to a stress field set up and maintained by relative movement of the two plates, but the seismic zone has not developed into a steady state feature. Several authors have concluded that the plate boundary in this region has a complex history (e.g. McKenzie, 1972; Laughton and Whitmarsh, 1974). The situation here may be somewhat similar to the South Iceland Seismic Zone where faults active in individual earthquakes strike transversely to the overall trend of the zone (Einarsson and Eiríkson, 1982).

The middle section of the Azores-Gibraltar zone has been seismically quiet for several decades. It is either locked or temporarily inactive.

East of 18°W the plate boundary is no longer defined by a narrow seismic belt. The high, but diffuse seismicity in this region shows that plate deformation occurs within a 400–500 km wide zone that extends into the Gulf of Cadiz and is connected to the seismic belt of Morocco and Algiers. Earthquakes in this zone may reach magnitude 8 (as did the shocks of February 1969 and May 1975) or even more (as must have been the case in 1755 when Lisbon was destroyed). Focal mechanisms are characterized by thrust faulting; occasionally strike-slip mechanisms are seen. A common feature is the maximum compressional axis trending N to NW, reflecting convergence between the Eurasian and African Plates.

FROM THE AZORES TO ICELAND

This section of the Mid-Atlantic plate boundary consists of two gently arcuate ridges, offset near 52°N by the Charlie-Gibbs Fracture Zone, a major transform fault. All other fracture zones, including the Kurchatov Fracture Zone near 40.5°N, have too small an offset to be resolved on the seismicity map. The ridge axis immediately to the north of the Azores is relatively straight and plate separation occurs at a right angle to the plate boundary. Seismic activity is fairly uniform both in space and time and earthquakes larger than magnitude 5 are rare (Einarsson, 1979). Two fault plane solutions show normal faulting at the ridge axis. Using three instruments, an ocean bottom seismograph study was conducted in the axial region near 45°N (Lilwall and others, 1977, 1978). All the recorded activity, including a swarm, was located in a narrow, elongate zone under the median valley.

The ridge section between 48 and 51°N appears to have some peculiar features. The seismic zone has a general NNW trend. Oblique spreading must therefore occur along this plate boundary, if the spreading direction is assumed to be parallel to the Charlie-Gibbs Fracture Zone immediately to the north. The structure of this part of the 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) to result from asthenospheric flow southward from the Iceland hot spot. One must then assume that the intersection of the transverse ridge with the plate boundary is the locus of unusually high production of eruptive material that slowly migrates southward along the plate boundary.

Focal mechanism studies in this region show some unexpected results. Fault plane solutions for two events at or near the ridge axis, near 49.5°N and 51°N, have a significant component of thrust faulting. The axis of maximum compression is horizontal, trending NE nearly perpendicular to the basement ridges. Several possible explanations for these unusual fault plane solutions were given by Einarsson (1979), including magmatic activity within a central volcano complex. Forcible intrusion of viscous magma at shallow depth can cause thrust faulting in the adjacent region, and the deflation of a magma chamber will cause reverse faulting in the chamber roof. Explanations of this kind are favoured here, especially since the only other known examples of reverse faulting at a divergent plate boundary are found under the Bárdarbunga central volcano in the eastern rift zone of Iceland (see later in this paper). It is therefore suggested that the transverse ridges described by Johnson and Vogt (1973) are the traces of central volcano complexes.

The Charlie-Gibbs Fracture Zone, between 52 and 53°N, offsets the ridge crest about 350 km to the left. The structure of this zone is described in considerable detail in the literature, most recently by Searle (1981). The structure is dominated by two parallel troughs, 45 km apart, that are separated by an E-W ridge. In the western part of the transform section the northern trough is more pronounced, but in the eastern part the southern trough is better developed. Vogt and Avery (1974) and later Searle (1981) concluded from topography, sonographs, and magnetic and seismic data that the two troughs were joined by a spreading center cutting across the transverse ridge at 31°45′W. This is seen clearly in the seismicity map in the back of this volume. During the last two decades, at least, practically all the recorded seismicity has been limited to the northern trough west of the central spreading axis. Transform faulting is shown by five fault plane solutions. The southern trough has been seismically inactive, except possibly at the eastern junction with the ridge axis. One fault plane solution indicates oblique faulting at this junction.

Kanamori and Stewart (1976) studied the source parameters of earthquakes and the mode of strain release along the Charlie-Gibbs Fracture Zone. They found that earthquakes of M_s between 6 and 7 occurred with an average repeat time of 13 years, the latest in 1967 ($M_s = 6.5$) and 1974 ($M_s = 6.9$). Seismic moment (M_0) was found to be anomalously large for the respective M_s, and similarly M_s was anomalously large if compared to m_b. These disparities were explained by large fault length (60–70 km) and anomalously low dislocation particle velocity (20 cm/s). Okal and Stewart (1982), who studied this phenomenon for a number of fracture zones, concluded that "slow" earthquakes tended to occur in fracture zones in the neighborhood of hot spot volcanism.

The seismic zone north of the Charlie-Gibbs Fracture Zone follows the crest of the Reykjanes Ridge to Iceland. A systematic variation along the plate boundary is seen in several of the characteristics of the ridge. Near its southern end the ridge trends N-S, has a rough topography, a well developed rift valley and high rate of seismicity. These features are rather typical for the Mid-Atlantic Ridge, and here the spreading occurs at a right angle to the plate boundary. Near 56°N the ridge bends to a N35°E direction. All available fault plane solutions in this area show normal faulting (Einarsson, 1979; Tréhu and others, 1981). The spreading north of this point appears to be oblique to the plate boundary. Near 58½°N another change occurs. The topography becomes smooth and the central valley gives way to a central horst (Fleischer, 1974). The seismicity, which is relatively high to the south, is low north of 58½°N. This seismicity pattern has persisted for the past half century at least (Francis, 1973). The northern part of the Reykjanes Ridge thus seems to have some of the characteristics of a fast spreading ridge, in spite of its low spreading rate of 1 cm/yr (Talwani and others, 1971). There seems to be a general consensus in the literature to ascribe this to the proximity to the Iceland hot spot (see e.g. Vogt, 1978).

ICELAND

The plate boundary in Iceland is displaced to the east by two major fracture zones, the South Iceland Seismic Zone in the south and the Tjörnes Fracture Zone in the north. Both zones have rather complex structures and lack the clear topographic expression typical of oceanic fracture zones. They are defined primarily by their high seismicity, earthquake focal mechanisms, and configuration with respect to the spreading axes (Ward, 1971; Tryggvason, 1973). The largest earthquakes in Iceland occur within these zones and may exceed magnitude 7 (M_s). The divergent plate boundary between the fracture zones is expressed by volcanic rift zones, with one branch in northern Iceland and two parallel branches in southern Iceland. Earthquakes occur along the rift zones, but they rarely exceed magnitude 5 (m_b). This activity is highly clustered both in time and in space, and a large part of it appears to be related to central volcanoes.

Teleseismic locations of earthquakes of the period 1963–1981 are shown in Figure 3. This map shows many of the characteristics of Icelandic seismicity, even though some of the

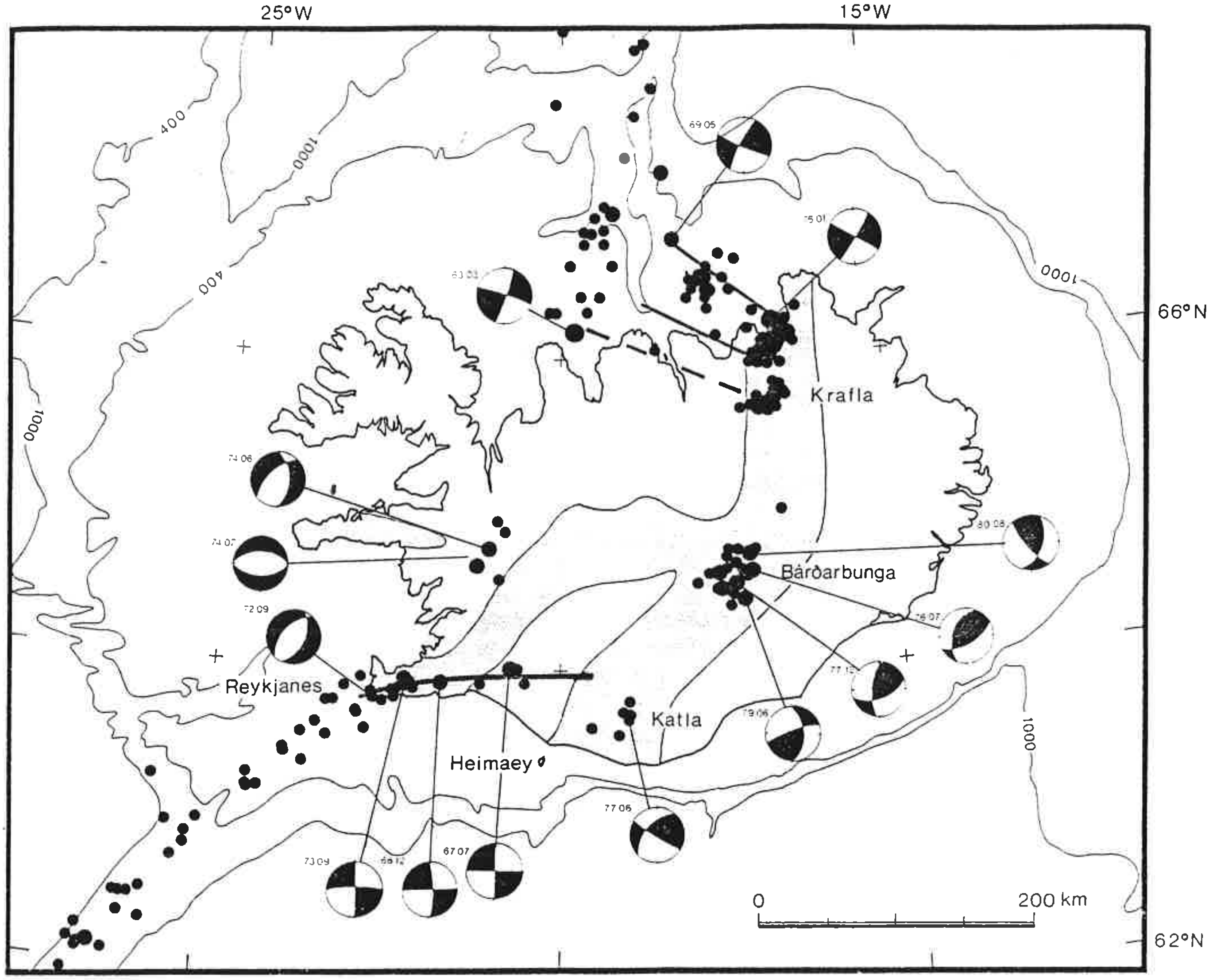

Figure 3. Epicenters and single event focal mechanism solutions in the Iceland area. Epicenters are taken from the PDE lists of the U.S. Geological Survey for the period 1963–1981, only epicenters determined with 10 or more stations are included. Larger dots are events of $m_b = 5$ and larger. Focal mechanism solutions are shown schematically on lower hemisphere stereographic projection of the focal sphere, compressional quadrants are black. For further references on the solutions see Einarsson (1979, 1985). The volcanic zones are stippled red and red lines show the seismic belts in the fracture zones.

locations may be in error by as much as 40 km. Concentration of activity is seen in the Tjörnes Fracture Zone near the coast of northern Iceland, in southwestern Iceland on the Reykjanes Peninsula, and in the South Iceland Seismic Zone. The focal mechanisms indicate strike-slip faulting. If the easterly striking nodal planes are taken as fault planes, the sense of motion is right-lateral in northern Iceland and left-lateral in southwestern Iceland, which is consistent with a transform fault interpretation of these zones. Outside of the fracture zones, clusters of activity are seen in the Borgarfjördur area in western Iceland, in the volcanic zone in Central Iceland, and near the volcanoes Katla in southern Iceland and Krafla in northern Iceland. More detailed information reveals that each of these zones has distinct seismological characteristics, which will be described below.

Historical seismicity of the three last centuries is shown in Figure 4. Except for the Borgarfjördur earthquakes of 1974, all major events are within the two transform zones. In southern Iceland the events are concentrated in a narrow, E-W belt, but in northern Iceland the estimated epicenters are distributed in a broader zone. The same pattern is seen in the instrumentally determined epicenters shown in Figure 3. It is clear that the large-scale seismotectonic pattern of the two zones is different, even though there are similarities in some of the details.

The Reykjanes Peninsula is an area of high seismicity and

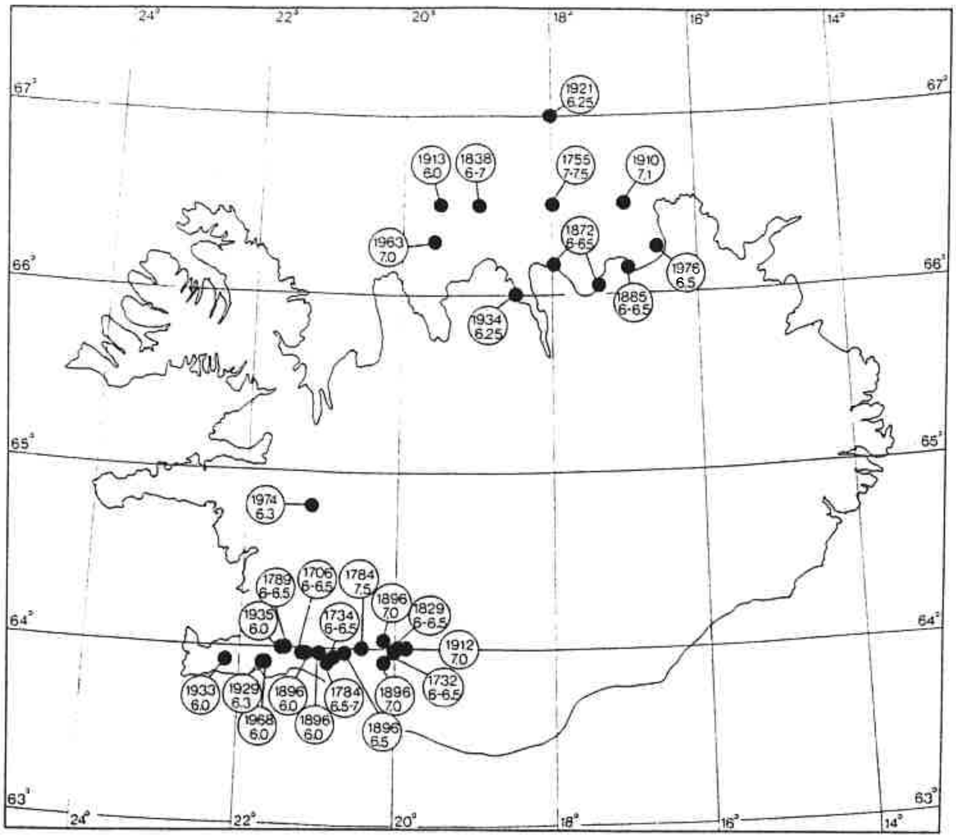

Figure 4. Estimated epicenters of historical, destructive earthquakes in Iceland after 1700. Year of occurrence and magnitude are given in the circles. Magnitudes of events prior to 1900 are estimated from the destruction areas. From Björnsson and Einarsson (1981).

recent volcanism that forms a transition between the Reykjanes Ridge to the west, and the western volcanic zone and the South Iceland seismic zone to the east (Fig. 3). The Mid-Atlantic plate boundary as defined by the seismicity enters Iceland near the tip of Reykjanes and then runs along the peninsula in an easterly direction. Detailed studies by Klein and others (1973, 1977) show that the seismic zone is less than 2 km wide in most places (Fig. 5). The earthquakes are mostly at a depth of 1–5 km and are not located on a single fault. The seismicity seems to be caused by deformation of the brittle crust above a deeper seated, aseismic deformation zone. Small scale structures can be resolved in the seismicity within the zone. Several seismic lineations or faults can be identified, striking obliquely or even transversely to the main zone.

Focal mechanisms have been determined for a large number of small earthquakes, using data from dense, local networks, and for two earthquakes larger than m_b 5 using teleseismic data. The minimum compressive stress is consistently oriented in a horizontal, NW direction. The maximum compressive stress rotates between the vertical direction, causing normal faulting on NE-striking faults, and the horizontal NE direction, causing strike-slip faulting on N or E striking faults. Thus the stress regime is characterized by the NW-trending minimum stress. The other principal stresses are probably nearly equal and may change directions according to local, or time dependent, conditions. Dykes open up against the minimum stress and strike NE, as shown by the eruptive fissures observed on the surface.

The mode of strain release changes systematically along the peninsula. Near the tip of Reykjanes, earthquakes occur in swarms; that is, in sequences where no single event is much larger than the others. Normal faulting is the most common faulting mechanism. Toward the east earthquakes tend to occur more in mainshock-aftershock sequences and strike-slip faulting becomes more prominent. The plate boundary in the eastern part of the peninsula has not been mapped in detail because of the low activity there in recent years.

The South Iceland Seismic Zone bridges the gap between the two branches of the volcanic rift zones in South Iceland. Few instrumentally determined epicenters are available, but the epicenters of historical events determined from the destruction areas (Fig. 4) define a narrow, E-W trending zone. In spite of this E-W orientation there is no indication on the surface of a major E-W fault in this area. However, many of the earthquakes have been associated with surface faulting; but the faults or fault systems have a N-S orientation (Einarsson and Eiríksson, 1982). The earthquakes appear to be caused by right-

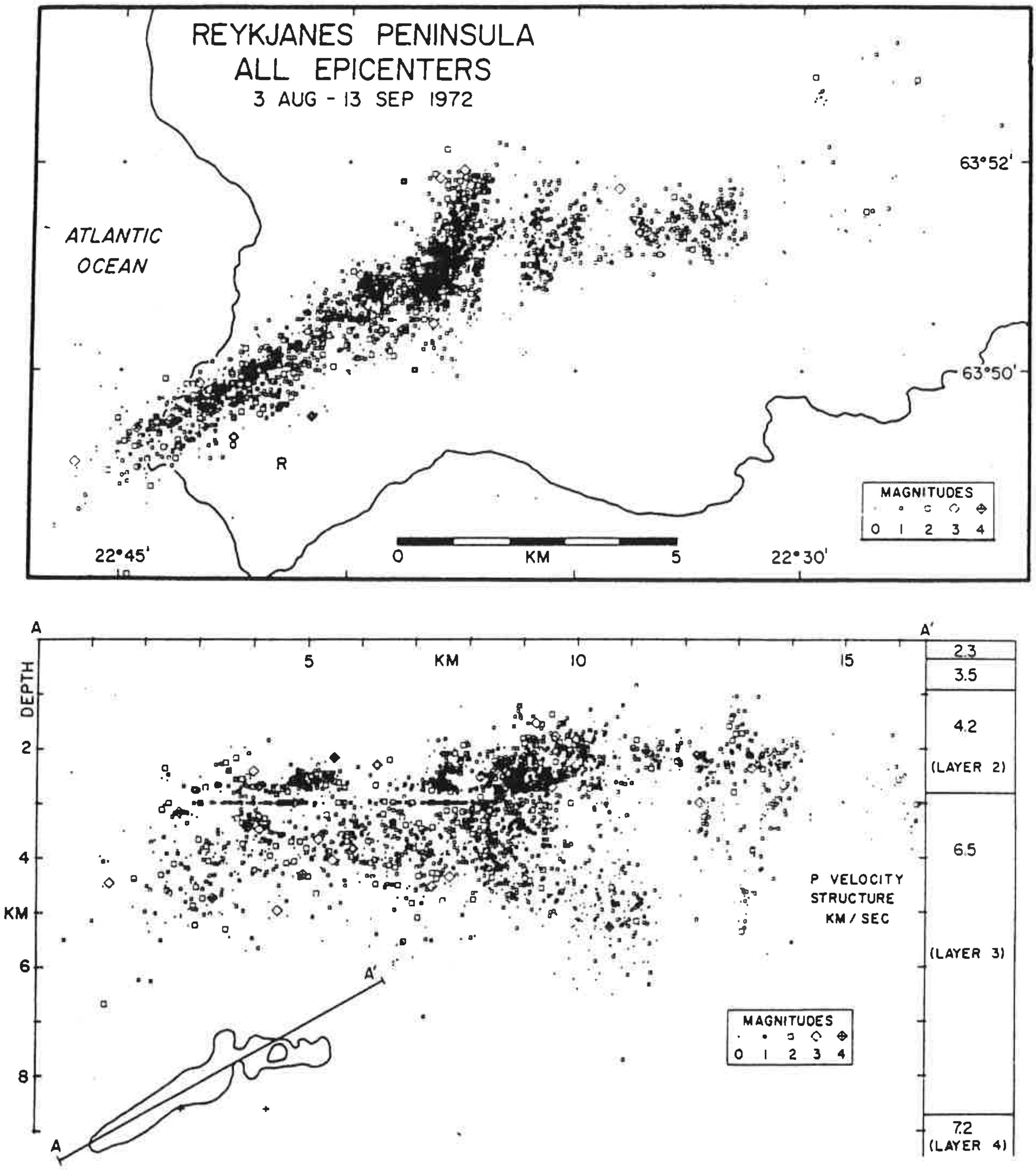

Figure 5. Epicenters and a hypocentral cross section of the earthquake swarm of Aug.-Sept. 1972, on the Reykjanes Peninsula in Iceland, from Klein and others (1977). Horizontal and vertical error of most hypocenters is less than 0.5 km, velocity model used in the locations is shown at the right. The position of the cross section with respect to the epicentral zone is shown in the inset.

lateral strike-slip on northerly striking faults, arranged side by side within the E-W zone. This is supported by the N-S elongate destruction areas (Einarsson and others, 1981). The reason for this unusual configuration is poorly understood, but a tentative interpretation (Einarsson and Eiríksson, 1982) is that the transform zone is migrating sideways in response to the southward propagation of the eastern rift zone.

The strain in the South Iceland Seismic Zone is released in sequences of large earthquakes with a recurrence time of 50–120 years. The sequence often starts with a magnitude 7–7½ shock in the eastern part of the zone followed by slightly smaller shocks in the western part.

The Tjörnes Fracture Zone also has a complicated structure, but in a different way. It is a broad zone of faulting and seismicity that connects the southern end of the submarine Kolbeinsey Ridge to the volcanic rift zone in North Iceland. The seismicity is too diffuse to be associated with a single fault or a simple plate boundary. Instead, the transform motion appears to be taken up by a series of parallel NW-striking faults or seismic zones (Einarsson, 1976; Einarsson and Björnsson, 1979). Two such zones have been identified; the Grímsey zone marking the northern boundary of the Tjörnes Fracture Zone off shore and the Húsavík zone, which can be traced on land from the rift zone through the town of Húsavík, and then offshore as a bathymetric feature (Saemundsson, 1974) and a strong gravity anomaly (Pálmason, 1974). A third zone has been tentatively identified near the southern boundary of the Tjörnes F. Z. Three fault plane solutions confirm transform faulting.

Earthquakes in the volcanic zones are generally smaller than in the fracture zones. Volcanic eruptions are usually accompanied by earthquakes, but between eruptions most parts of the volcanic zones are seismically quiet. A few areas of persistent seismic activity are found, the most prominent ones in Central Iceland and near the subglacial volcano Katla in South Iceland (Fig. 3).

The seismic area in Central Iceland, near the center of the Iceland hot spot, is largely covered by the ice cap Vatnajökull so its tectonic structure is poorly known. Recent studies of ERTS images of this area (Thorarinsson and others, 1973) seem to indicate that the structure is dominated by a group of central volcanoes, and it is tempting to relate the earthquakes to volcanic processes. An unusual sequence of earthquakes occurred in this area during the period 1974 to 1980. Seven earthquakes of m_b 5 and larger occurred in this period, but events of that size were unknown there before. Fault plane solutions of four of these events (Fig. 3) all show that the sequence is associated with reverse faulting in the caldera region of the subglacial Bárdarbunga volcano. It is difficult to see how this type of faulting could be related to plate movements. Brittle failure of the crust above a deflating magma chamber, however, could produce reverse faulting earthquakes, and this is considered to be the most likely explanation for the Bárdarbunga events.

The subglacial Katla volcano is located near the southern end of the eastern volcanic zone, south of its junction with the South Iceland Seismic Zone. The structure of this part of the zone is characterized by several central volcanoes; rifting structures are less significant. Historic eruptions of Katla have been preceded by felt earthquakes and, because of the potential danger of future eruptions, the seismicity at Katla is monitored by a relatively dense seismograph network. The epicenters located so far delineate two active areas 15 km apart. One poorly constrained fault plane solution indicates strike-slip with a significant component of reverse faulting. As in the case of Bárdarbunga, a deflating magma chamber may offer an explanation for this type of faulting. The seismic activity in the Mýrdalsjökull area shows a pronounced annual cycle. The probability of an earthquake occurring within a given time interval is several times higher in the second half of the year than in the first half. This annual cycle was first noted by Tryggvason (1973) for the years 1952–1958 and has been confirmed by later data.

The Heimaey eruption in 1973 was preceded by an intensive swarm of small earthquakes that started 30 hours before the eruption. Earthquakes also accompanied the eruption, but seismicity declined as lava production diminished. No shock reached local magnitude 4. The earthquakes during the eruption originated at the depth of 15–25 km and occupied a spherical volume centered under Heimaey. It seems likely that the erupted magma was either stored or formed within this volume.

The depth of the Heimaey earthquakes is much larger than observed elsewhere in Iceland. In this area the upper boundary of the anomalous mantle underlying Iceland is at a depth of 12–15 km. Earthquakes at the depth of 15–25 km may be taken to imply brittle failure in the Icelandic, anomalous mantle where creep or ductile behaviour is normally assumed. In this volcanic region 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.

A major rifting episode has been in progress since 1975 in the volcanic rift zone in northeastern Iceland. The activity has been confined to the Krafla central volcano and its associated fissure swarm (Björnsson and others, 1977, 1979) and provides a demonstration of a process that seems to play an important role in Icelandic tectonics. The activity is characterized by repeated cycles of relatively slow inflation and rapid deflation of the volcano. Magma apparently accumulates at a constant rate under the volcano during the inflation periods and during the deflation events the magma escapes from the reservoir area. Each cycle of activity is accompanied by a characteristic pattern of seismic activity as described by Brandsdóttir and Einarsson (1979) and Einarsson and Brandsdóttir (1980). Continuous volcanic tremor starts in the caldera region at about the same time as the deflation. Small earthquakes also occur in the caldera, but the epicentral area is soon extended along the Krafla fissure swarm to the north or to the south, as shown by the example in Figure 6, the rifting event of July 1978. The rate of propagation of the seismic activity is highest during the first few hours, typically 0.5 m/sec., but the speed decreases as the deflation rate decreases and the epicentral zone is extended. The earthquake activity culminates after the maximum in tremor and deflation rate is reached. The largest earthquakes are located within a well defined, but each time different, section of the fissure swarm. Local magnitude only rarely exceeds 4.5. The depth of hypocenters is in the range 0–9 km. Extensive fault movements, both normal faulting and fissuring, occur in the area of maximum earthquake activity. The propagating seismic activity suggests that the magma escaping from the Krafla reservoir is injected laterally into the fissure swarm to form a dyke. The dykes may be as long as 40–60 km.

The first and the most violent deflation event started on Dec.

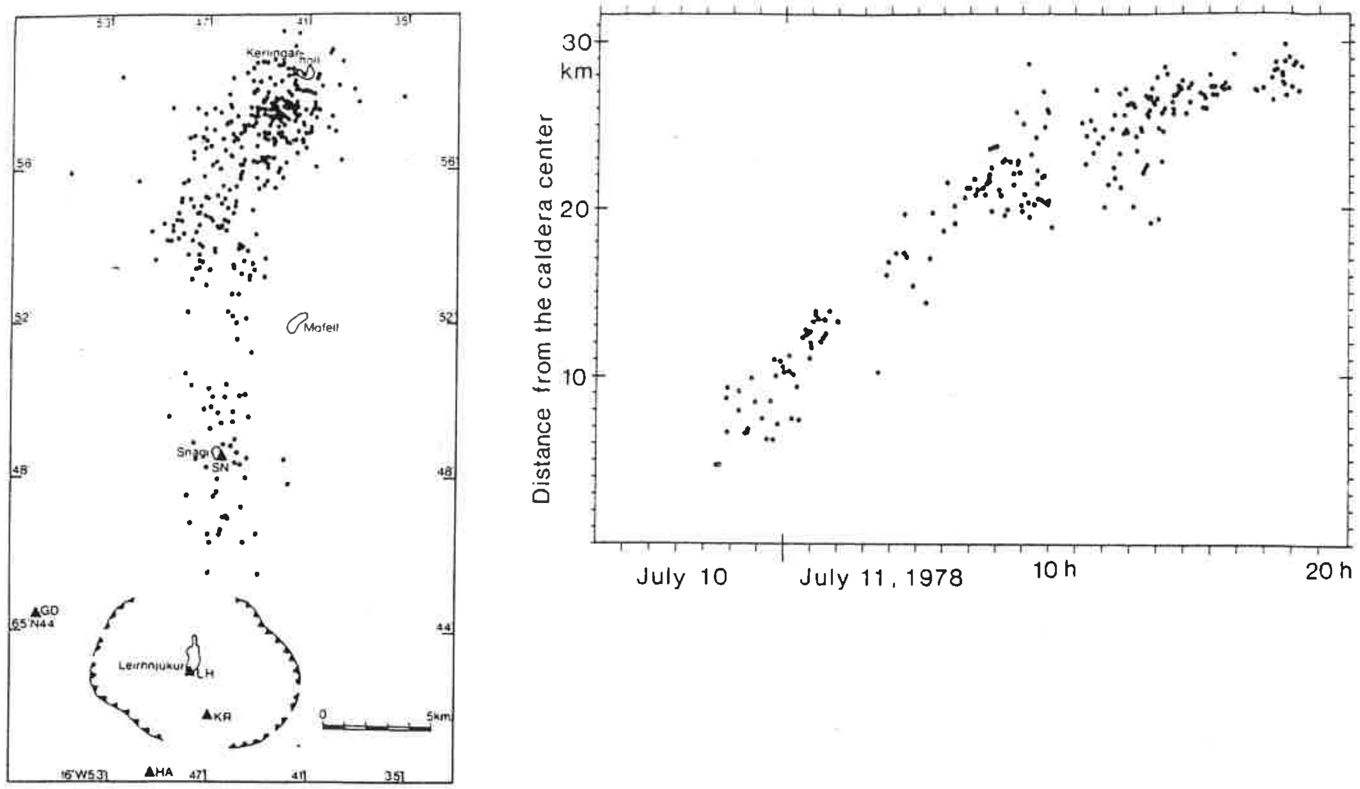

Figure 6. Epicentral map and migration of epicenters during the intrusion of July 1978 in the Krafla fissure swarm in North Iceland. The origin in the time-distance plot corresponds to the center of the caldera and the time when deflation of the caldera region and volcanic tremor began. After Einarsson and Brandsdóttir (1980).

20, 1975. The deflation of the caldera exceeded 2 m and the accompanying earthquake swarm lasted about eight weeks. Most of the epicenters that appear in the northern part of the volcanic zone in Figure 3 belong to this swarm. The largest earthquakes were confined to two separate areas. One area was within the caldera where earthquakes were apparently associated with faulting above the deflating magma reservoir. Depth of most hypocenters was 0–4 km. The largest earthquakes reached magnitude 5. The other epicentral area was near the junction between the Krafla fissure swarm and the Grímsey seismic lineament. The largest earthquake was of m_b 6 and the focal mechanism shows right-lateral strike-slip along the Grímsey zone. This earthquake sequence demonstrates well the relationship between rifting along the diverging plate boundary and transform faulting in the fracture zone. The present Krafla events are assumed to be the result of interaction between magma pressure under the Krafla volcano and rifting of the plate boundary. The rifting is triggered by increasing magma pressure in the reservoir when a fluid-filled extensional crack propagates horizontally along the Krafla fault swarm. The driving force of this process is the tectonic stress at the plate boundary, but the mode of strain release is modified by the presence of fluid.

Prior to July 1980 the deflation events were associated with mostly rifting and subsurface transport of magma. Eruptions to the surface were only minor. But character of the deflation events changed in July 1980. In most events since then a large proportion of the mobilized magma has reached the surface and rifting has played a secondary role.

Earthquakes are rare outside of the volcanic zones and the transform zones in southern and northern Iceland. However, intraplate earthquakes are known in the Iceland region, for example, near the insular shelf margin east of Iceland and in Borgarfjördur in western Iceland where a significant earthquake sequence was recorded in 1974. This sequence was studied with portable instruments, giving detailed locations and fault plane solutions (Einarsson and others, 1977). The depth of hypocenters was 0–8 km and the area was shown to be undergoing horizontal extension.

THE PLATE BOUNDARY IN THE ARCTIC

North of Iceland the plate boundary follows the Kolbeinsey Ridge to the Jan Mayen Fracture Zone. This part of the plate boundary is anomalous in several ways, because of its relatively high elevation, low seismicity, and its asymmetric position with respect to the adjacent continents. All these features may be related in some way to the existence of the Iceland hot spot to the south. The topographic anomaly is part of a much larger regional bulge centered on Iceland. The low seismicity is comparable to that of the northern part of the Reykjanes Ridge and thus displays a certain symmetry with respect to Iceland. The off-center position of the Kolbeinsey Ridge is the result of one or more ridge jumps to the west that may be related to drifting of the plate boundary off the Iceland mantle plume. The axis of the Kolbeinsey Ridge is offset by two minor transform faults, the Spar Fracture Zone near 69°N and another small fracture zone, probably located near 71°N, immediately south of the Eggvin Bank. Neither zone is clearly expressed on the seismicity map. The plate boundary in the Eggvin Bank region is poorly defined and may be complicated by the topographic high suggested by Saemundsson (this volume) to be the trace of a hot spot in this area.

The Jan Mayen Fracture Zone is the most significant fracture zone in the Arctic region, displacing the ridge axis 210 km to the right. The transform section of the fracture zone is a pronounced but highly asymmetric trough, because of the high topography connected with the Eggvin Bank and the Jan Mayen continental sliver to the south. The volcanic island of Jan Mayen protrudes into the fracture zone about 55 km west of the eastern ridge intersection. The overall trend of the transform section is 110°, and all fault plane solutions are consistent with left-lateral strike-slip along a plane with strike varying between 120° in the east to 102° in the west.

During the last two decades there has been a marked difference in the strain release pattern between the western and the eastern parts of the Jan Mayen Fracture Zone. Few, but large earthquakes have occurred in the western part. In the eastern part, on the other hand, the seismicity is characterized by relatively frequent, but smaller, earthquakes. Much of this activity is concentrated where Jan Mayen protrudes into the fracture zone. Soernes and Fjeldskaar (1980) concluded that the activity had increased in this area, notably after the eruption of Beerenberg on Jan Mayen in 1970. Earthquake swarms are found to be associated with Beerenberg, but most of the larger earthquakes seem to be related to slip along the transform fault.

In most respects the Mohns Ridge is a typical mid-oceanic ridge, centrally located in the Greenland-Norwegian basin and uninterrupted by fracture zones. The spreading is slightly oblique, especially at its western end, where it joins with the Jan Mayen Fracture Zone at an angle of 120°. The seismic zone is well defined, narrow and continuous along the rifted crest. The seismicity is higher than that of the Kolbeinsey Ridge and the northern Reykjanes Ridge, but comparable to other parts of the Mid-Atlantic Ridge. Fault plane solutions by Conant (1972) and Savostin and Karasik (1981) show normal faulting at the ridge axis.

The pattern of seismicity changes abruptly from the Mohns to the Knipovich Ridge (Vogt, this volume, Ch. 12). Spreading along the Knipovich Ridge is highly oblique and the seismicity is diffuse. This diffuse distribution of epicenters is not an artifact of the location accuracy as one can see by comparison with the tightly concentrated seismicity of the Mohns Ridge immediately to the south. Vogt and others (1982) note that the change in the seismicity pattern coincides with a change in the ridge crest topography. The rift valley widens where the seismicity drops off. The scattered epicenters imply that deformation takes place within a wide zone, a feature possibly inherited from the time when this part of the plate boundary changed from being primarily of strike-slip character to being a divergent boundary. Subsequent jumping of the rift axis may also contribute to the complexity.

The intraplate seismicity is considerably higher east than west of the Knipovich Ridge. This asymmetry was noted by Vogt and others (1982) in connection with asymmetry of other features of the Arctic ridge system such as topography, gravity, and spreading rate. A good part of the intraplate events east of the plate boundary appears to be related to a seismicity anomaly in Svalbard.

The axis of the Knipovich Ridge approaches the continental margin west of Svalbard and is there displaced to the left by two 100 km long transform faults connected by a short ridge axis. The southern fracture zone has been relatively quiet in the last two decades, whereas several earthquakes larger than $m_b = 5$ have occurred in the northern one. Fault plane solutions of five events show transform faulting (Horsfield and Maton, 1970; Conant, 1972; Savostin and Karasik, 1981).

The Svalbard Archipelago stands out on most seismicity maps as an area of high seismicity. Mitchell and others (1979) and Bungum and others (1982) operated temporary local networks in 1976, 1977, and 1979, and found that most of the activity was concentrated in two zones, although scattered activity was also recorded in other parts of the archipelago. Teleseismic maps show activity on the continental shelf west of Svalbard, possibly connecting it to the Knipovich plate boundary. The concentrated earthquake zones are elongate in a WNW-ESE direction and fault plane solutions show left-lateral strike-slip. This seismicity was attributed by Bungum and others (1982) to an interaction between the present tectonic stress field and older zones of weakness. Savostin and Karasik (1981), on the other hand, felt that this seismicity implied the existence of a separate plate, the Spitsbergen microplate, even though its eastern boundary could not be clearly delineated.

Northwest of Svalbard, near 83°N, the plate boundary turns to a NE trend and follows the Nansen (Gakkel) Ridge across the Arctic Basin. The Nansen Ridge is uninterrupted by large offset transform faults over a distance of more than 2000 km. The seismicity is moderately high and all available fault plane solutions show normal faulting at the ridge crest (Savostin and Karasik, 1981; Conant, 1972; Sykes, 1967). The seismicity of this part of the plate boundary is elaborated on further by Fujita (1985).

EARTHQUAKE SEQUENCES IN TIME AND SPACE

Earthquake sequences are often classified into three principal types (Mogi, 1963): mainshock-aftershock sequences, foreshock-mainshock-aftershock sequences, and earthquake swarms, that is, sequences without one event that is distinctly larger than the others. Examples of different types of sequences are shown in

Figure 7. The sequence of September 1969 in the 15°20′ Fracture Zone is a typical mainshock-aftershock sequence, with four recorded aftershocks following a strike-slip earthquake ($m_b = 5.7$) within a few hours. The sequence of October 1974 in the Charlie-Gibbs Fracture Zone is similar except that a foreshock preceded the mainshock by nine minutes. The Reykjanes Ridge sequence of March 1967 is a typical swarm that terminated a very active period in this part of the plate boundary. Furthermore, sequences are found that are difficult to classify according to this scheme, for example, the complex sequence of December 1975 to February 1976 in northern Iceland, involving deflation of the Krafla volcano, rifting and intrusion in the rift zone, and transform faulting. It may be described as two earthquake swarms and a mainshock-aftershock sequence superimposed on one another. Sequences that are best described as a series of mainshocks also exist, for example, in the Charlie-Gibbs Fracture Zone in July 1965 (Sykes, 1970) and in the South Iceland Seismic Zone in 1896 (Einarsson and others, 1981). In these sequences each mainshock is outside the source area of the previous one.

Soon after the installation of networks of long period seismographs it became apparent that earthquake swarms were common along the world rift system. Sykes (1970) studied the spatial distribution of swarms and found that most of them originated in the crestal area of the ridges. Fracture zone seismicity, on the other hand, is characterized by mainshock-aftershock sequences, although swarms occasionally occur there. Clustering of events, both in time and space, is thus found to be more common along the rift zones than along fracture zones (Francis and Porter, 1971).

In addition to clustering of epicenters within distances of ten kilometers, several examples can be found of apparent temporal correlation of seismicity over distances of hundreds of kilometers along the plate boundary. These include active periods in the vicinity of the Vema Fracture Zone in 1979, the Charlie-Gibbs Fracture Zone in 1965–1967 (Einarsson, 1979), and the South Iceland Seismic Zone (Einarsson and others, 1981). These examples demonstrate temporal correlation of activity over large areas, and yet the source volume of each earthquake is much too small for one event to directly influence the occurrence of the others. It seems likely that the events are triggered by a regional strain pulse that affects a large part of the plate boundary, possibly related to large scale plate motions.

Several cases of migration of seismicity along the plate boundary have been documented, ranging over distances of kilometers to thousands of kilometers, and at rates from 1 to 100 km per day.

Limond and Recq (1981) observed an apparent migration of seismic events along the western boundary of the Eurasian Plate, from Jan Mayen southward to the Azores and then east toward Gibraltar. By correlating the seismicity of different subsections of this plate boundary, a migration velocity of 3 to 10 km per day was found. The significance of the correlation is critically dependent on the occurrence of three earthquake sequences; a large swarm in March 1967 on the Reykjanes Ridge near 56°N, a

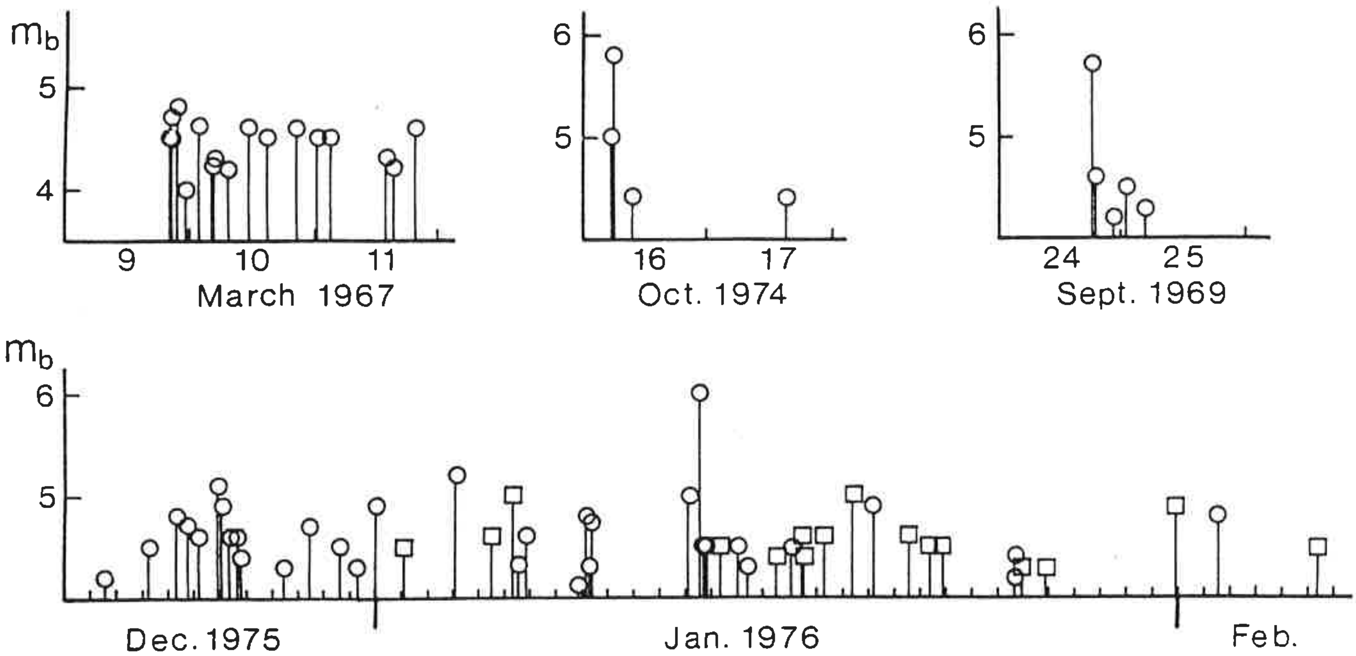

Figure 7. Magnitude as a function of time for four earthquake sequences, a swarm on the Reykjanes Ridge near 56°N, a foreshock-mainshock-aftershock sequence in the Charlie-Gibbs Fracture Zone, a mainshock-aftershock sequence in the 15°20′ Fracture Zone, and a mixed sequence in northern Iceland associated with deflation of the Krafla Volcano (squares), and rifting and transform faulting near the junction of the Krafla fissure swarm and the Grimsey seismic zone (circles). Data from the PDE listings of the U.S. Geological Survey. Note the different time scales.

swarm north of the Azores near 41°N in January 1968, and a large earthquake on February 28, 1969 in the Goringe Bank area. Removing these three peaks from the time series destroys the correlation.

Migration of epicenters was already mentioned in conjunction with lateral injection of magma from the Krafla Volcano in northern Iceland (see Fig. 6), but a direct connection between rifting of a divergent plate boundary and strike-slip motion along an adjacent transform fault was also found during the Krafla rifting episode. The first magmatic event of the Krafla Volcano that occurred between December 1975 and February 1976, began with an eruption and deflation of the caldera region, followed by migration of epicenters along the rift zone towards the transform fault where strike-slip faulting subsequently occurred (Björnsson and others, 1977). An 80 km long section of the plate boundary was involved during this two month event. In a typical Krafla rifting event the hypocenters probably mark the tip of a propagating, fluid-filled crack. The rate of propagation is governed by the rate at which magma can be fed to the crack tip, which again is a function of the magma viscosity, dike width, dike length, and the pressure gradient (Einarsson and Brandsdóttir, 1980). Migration speeds in the range of 0.5–4 km per hour have been observed.

Migration of hypocenters is well documented for the Reykjanes Peninsula earthquake swarm of September 1972 in Iceland (Klein and others, 1977). The activity started with small earthquakes the central part of the epicentral zone, and then spread laterally away from this nucleation point at a rate of 1–2 km per day. Individual subswarms also showed similar behaviour, usually spreading bilaterally from the center of each subswarm. Klein and others (1977) suggested that the migration was governed by a propagating dislocation in a viscoelastic fault zone or by a fluid diffusion process.

Major earthquake sequences in the South Iceland Seismic Zone follow a pattern of migration, beginning with a large (M_S about 7) event in the eastern part of the zone followed by smaller shocks farther west (Einarsson and others, 1981). Even though the first event is the largest one of the sequence, the later events are not aftershocks. They occur outside the source area of the first event but are probably triggered by the change in the regional strain field associated with it. The time delay between events may be due to viscoelastic response of the crust or coupling between the elastic lithosphere and the viscous asthenosphere.

THE FREQUENCY-MAGNITUDE DISTRIBUTION AND B-VALUE

Earthquakes in general are found to follow a relation of the form: $\log N = A - bM$ where N is the number of events of magnitude larger or equal to M. A and b are constants, describing the overall level of seismicity and the relative importance of large versus small earthquakes. Thus a low b-value means that large earthquakes are relatively frequent, whereas high b-values are found where the seismicity is characterized by small events. The value of b depends on the type of magnitude scale used, but has been found to be remarkably similar between different seismic areas. A significant difference was, however, demonstrated between the b-value of earthquakes along the fracture zones and the axial areas of the Mid-Atlantic Ridge (Francis, 1968 a, b). These two populations of events had b-values of 0.99 and 1.72 respectively, if the m_b-scale was used, and 0.65 and 1.33 if surface wave magnitudes (Ms) were used. Francis used earthquakes of the period 1963 to 1967 in his study. A larger data set, for the period 1963–1981, reveals an even greater difference between ridge and fracture zone earthquakes. Magnitude distribution plots are shown in Figure 8 for events along the ridge axis 40–51.9°N and 53.1–60°N (dots), and for four major fracture zones, the Charlie-Gibbs, Kane, 15°20′, and Vema zones (crosses). The magnitude detection limit is around $m_b = 4.5$ and is not significantly variable between areas. The plots are reasonably linear and the difference in the slopes is evident. Maximum likelihood estimates of the respective b-values give 2.3 ± 0.3 and 1.2 ± 0.3 (95% confidence limits).

The b-values of ridge axes and fracture zone earthquakes in some way reflect the difference between these two tectonic regimes, although the exact physical mechanism responsible for the different b-values is unclear. The b-value in fracture zones is close to that of most other seismic zones, so it seems to be the high b-value at ridge crests that needs explaining. In laboratory tests on

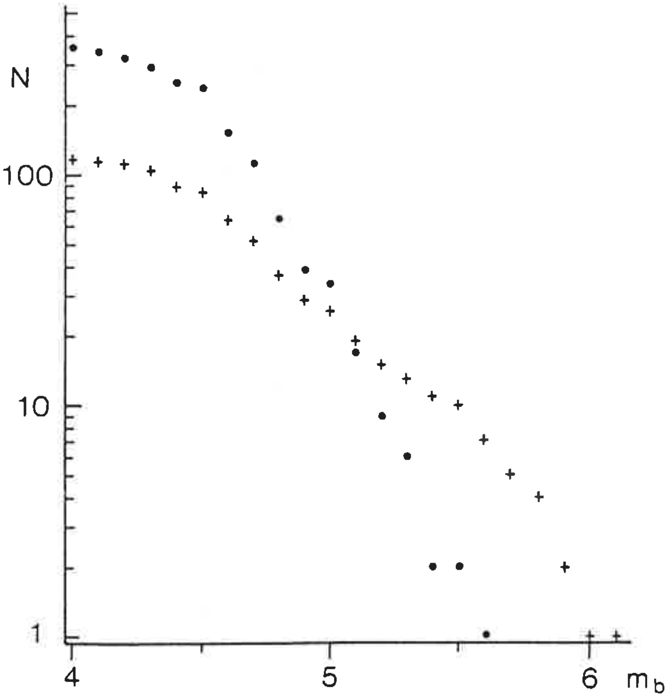

Figure 8. Cumulative number as a function of magnitude (m_b) for earthquakes along ridge axes (dots) and fracture zones (crosses). Data are taken from the PDE lists of the U.S. Geological Survey for the period 1963–1981.

microcracking, high b-values have been found to be associated with heterogeneous material and concentrated sources of stress (Mogi, 1963) or fracturing at low stress (Scholz, 1968). Cataclastic deformation has also been found to be accompanied by microcracking with high b-values (Scholz, 1968). Although a direct correspondence between microfracturing in laboratory specimens and large scale fracturing of the crust during earthquakes has not been experimentally verified, it seems plausible that some of these physical factors may contribute to the high b-values observed at the crest of the Mid-Atlantic Ridge. The crust in the axial zone is known to be extensively fractured by normal faulting and fissuring. Intrusions further add to the heterogeneity of the crust. Uniform loading of this heterogeneous zone by the relative motion of the adjacent plates leads to a heterogeneous stress field, which is further complicated by the local stress field around new magmatic intrusions, both of mechanical and thermal origin. Hydrothermal circulation has an effect by lowering the shear strength of the rock and by changing the thermal conditions. When this zone is loaded towards failure, the areas of high stress and/or low strength will fail first, even at relatively low average stress. Conditions for a large earthquake, high, uniform stress over a wide area, are not likely to develop. Thus the scale of the inhomogeneities is directly related to the source dimensions of the earthquakes. Inhomogeneities with linear dimensions of 1 to 10 km may be expected to significantly affect the observed b-value of earthquakes in the magnitude range 4 to 6, if one assumes that the relationships between magnitude, seismic moment, and source dimensions given by Kanamori (1977) apply for ridge crest earthquakes.

For further discussion of high b-values at the axis of mid-oceanic ridge the reader is referred to Francis (1968), Sykes (1970) and Francis and Porter (1971).

DEPTH OF HYPOCENTERS AND THE QUESTION OF MAGMA CHAMBERS

The source depth of earthquakes gives important information on the physical state of the crust and constraints for models of crustal growth. Thus the deepest hypocenters in an area mark the transition from brittle to totally ductile behavior of crustal material. The depth of this transition depends on temperature, creep rate, and water content (Meissner and Strehlau, 1982).

Brittle failure does not take place within magma bodies, but earthquakes may occur in regions surrounding and even beneath magma pockets and intrusions in response to pressure changes in the magma. The depth distribution of hypocenters therefore constrains the possible locations of magma chambers.

Routine teleseismic locations of Mid-Atlantic Ridge earthquakes give little information about their source depth, except that they all fall into the category of shallow events, that is, depth less than 70 km. Reliable determination of source depth along the oceanic part of the ridge system is therefore restricted to a few teleseismic studies of individual earthquakes and to ocean bottom surveys where three or more instruments have been deployed temporarily. In Iceland, on the other hand, dense temporary networks have been operated in several areas, and in some parts of the active zones the permanent seismograph network gives reliable depth determinations on a routine basis.

Teleseismic studies of earthquake sources have been made for several Mid-Atlantic events, yielding focal depth as a resulting parameter (Weidner and Aki, 1973; Duschenes and Solomon, 1977; Hart, 1978; Tréhu and others, 1981). Ridge crest earthquakes are generally found to be shallower than 5 km, and fracture zone events possibly slightly deeper. Although the few events studied by teleseismic data represent only a small fraction of the total number of events recorded from the Mid-Atlantic Ridge system, the results deserve serious attention. These events are among the largest and most significant earthquakes in this area during the period of instrumental observation.

Results of studies where local seismograph arrays have been used to determine focal depths are compared in Figure 9. Most of the results are from Iceland, where 5–23 stations have been used. Only two ocean bottom investigations are included, from the ridge crest near 45°N where three instruments were used, and near the eastern end of the St. Paul's transform fault where four instruments were deployed.

Three of the examples in Figure 9 are from the Krafla area in the axial rift zone in northern Iceland. These are only representative samples of seismicity during the magmatic and tectonic activity that began in this region in 1975. The first example shows the depth distribution of hypocenters within the caldera of the Krafla Volcano during one inflation period of the volcano between the deflation and rifting events of January 20 and April 27, 1977. Earthquakes in the caldera region are correlated with the level of inflation, and may be regarded as the response of the magma chamber roof to the increasing magma pressure. Seismicity is highest between the depth of 1 and 4 km, above and surrounding the top of the magma chamber or chambers, as determined from surface deformation (Björnsson and others, 1979) and attenuation of S-waves (Einarsson, 1978). Below is an earthquake-free zone at 5–6 km depth, but earthquakes again occur at about 7 km depth. S-waves also propagate without abnormal attenuation at 7 km depth under the volcano, presumably marking the bottom of the magma chamber.

In contrast to the inflation periods, characterized by relatively continuous seismicity extending over weeks or months, the rifting events are accompanied by intense earthquake swarms lasting hours or days. Epicenters migrate along the rift zone, out of the caldera and away from the magma chamber, either to the north or to the south (Brandsdóttir and Einarsson, 1979; Einarsson and Brandsdóttir, 1980). Depth distribution of two such swarms is shown in Figure 9. It is remarkable that the depth distribution is significantly different during these two intrusion events, and yet the epicentral zones are nearly identical. The event of September 1977 was associated with extensive faulting and rifting at the surface within a 16 km long segment of the rift zone, extending from the northern caldera rim southward across the caldera and along the southern rift zone. A small lava eruption

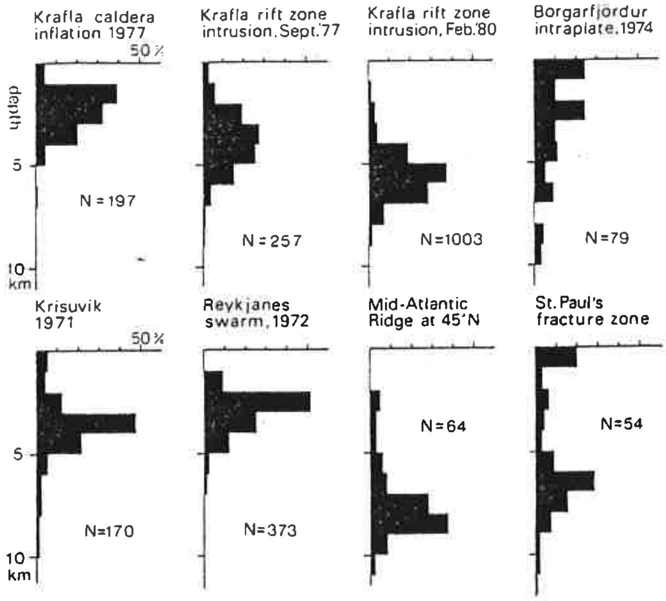

Figure 9. Histograms for depth distribution of hypocenters in Iceland (Krafla, Borgarfjördur, Krísuvík, Reykjanes) and on the submarine part of the ridge system (45°N, St. Paul's Fracture Zone). N is the total number of events used in the plot.

occurred near the northern caldera rim and a small amount of basaltic pumice was erupted through a drill hole in the southern rift. The hypocentral distribution suggests a dike intrusion mainly at the depth of 3–4 km (Brandsdóttir and Einarsson, 1979) in the southern rift. The February 1980 event was in many respects different. Only minor rifting and faulting was observed at the surface, no eruption occurred, and the maximum in the depth distribution was at 5–7 km. This suggests that the dike of February 1980 was injected below the September 1977 dike. The lesson to be learned here is that the frequency of earthquakes versus depth during a short recording period is not only dependent on material properties and physical state of the crust, it also depends on the previous course of events in the area. It is significant, however, that earthquakes occur down to the depth of 9 km, even under the most active part of the rift zone.

The depth distribution of the intraplate earthquakes in Borgarfjördur in 1974 (Einarsson and others, 1977) is shown in Figure 9 for comparison. These earthquakes were associated with normal faulting of a 3–7 m.y. old crust within the North American Plate. The hypocenters reach a depth of 9–10 km, not significantly deeper than the rift zone earthquakes.

The results of two studies on the Reykjanes Peninsula are shown in Figure 9, one in Krísuvík area in the central part of the peninsula (Klein and others, 1973), and the other near the tip of Reykjanes (Klein and others, 1977). Dense, multi-element arrays were used in both studies and most of the hypocenters have vertical errors considerably less than 1 km. In Krísuvík most hypocenters are between 2 and 5 km depth, but earthquakes also occur close to the surface and as deep as 9 km. Most of the Reykjanes earthquakes belong to a large swarm of more than 17000 recorded events ($M \leq 4.4$) that occurred during a six week recording period in 1972. Earthquakes occur at 1–7 km depth, but the distribution is sharply peaked at 2–3 km depth. The peak is not as sharp if the distribution of energy release is considered. The energy distribution is essentially flat between 2 and 5 km depth (Klein and others, 1977).

The activity recorded by Lilwall and others (1977, 1978) at the ridge crest near 45°N is located between 2 and 11 km depth, with a peak near 8 km. The data of Francis and others (1978) from the eastern junction of the St. Paul's Fracture Zone with the ridge axis show a similar peak at 6–8 km depth, but in addition there is a peak at 0–1 km depth. Duschenes and others (1983) suggest that the shallow peak may be artifacts of the location technique used. The experience from Iceland warns against assigning too much significance to the peaks in the distribution, especially since both data sets were obtained during short recording periods. It is of considerable interest to note that earthquakes occur down to 11–12 km depth under the rift valley of the Mid-Atlantic Ridge. The results of Toomey and others (1985) show brittle behavior down to at least 7–8 km depth beneath the central valley at 23°N.

Several models of crustal generation by sea-floor spreading assume a magma chamber at small but usually unspecified depth beneath and extending along the spreading center (e.g. Cann, 1974; Sleep, 1975, 1978; Rosendahl, 1976; Hall and Robinson, 1979). Evidence for and against the existence of such a magma chamber was recently reviewed by Sanford and Einarsson (1982) and Lewis (1983), who concluded that in the case of the Mid-Atlantic Ridge no evidence could be found for an extensive magma chamber at crustal levels (see also Macdonald, this volume). Moreover, depth of hypocenters and the efficient propagation of crustal S-waves across the crestal zone of the ridge seem to preclude anything but small, isolated pockets of magma in the crust. Evidence for such small magma chambers is found under the Krafla Volcano (Einarsson, 1978) and possibly on the Reykjanes Peninsula in Iceland (Klein and others, 1977). Evidence is also found for a zone of melting in the mantle under the Mid-Atlantic Ridge. Molnar and Oliver (1969), for example, found a narrow zone of inefficient propagation of the Sn phase centered on the ridge crest. Sn is a shear wave that normally propagates over long distances in the lithosphere at subcrustal levels. This zone of partial melting appears to be particularly well developed in the mantle under Iceland, where it is characterised by low density, low P-velocity, high P- to S-velocity ratio, high S-wave attenuation and a layer of low electrical resistivity near the crust-mantle boundary (Sanford and Einarsson, 1982).

CONCLUSIONS

1. The eastern boundary of the North American Plate, which follows the Mid-Atlantic Ridge system and its continuation into the Arctic, is marked by a continuous belt of seismicity, that is less than 30 km wide in most places.

2. Earthquakes in fracture zones occur almost exclusively

within its transform section, connecting the ridge crests on either side.

3. A distinct difference in seismicity pattern is found between the ridge crests and fracture zones. Mainshock-aftershock sequences and normal b-values are found along the fracture zones, whereas the seismicity along ridge crests is characterized by earthquake swarms and high b-values. The difference is probably related to structural heterogeneity in the zone of volcanism and crustal accretion along the ridge crest.

4. Almost all available fault plane solutions in fracture zones show strike-slip faulting in the transform sense, sometimes with a small component of reverse faulting.

5. A large majority of fault plane solutions along ridge crests indicates normal faulting with tension axes perpendicular to the trend of the ridge.

6. Hypocenters reach depths of 8 km under at least some parts of the Mid-Atlantic Ridge, and seem to rule out the existence of a large, continuous magma chamber at crustal levels. If shallow magma chambers play an important role in the process of crustal genesis at the Mid-Atlantic Ridge, they must be intermittent features and of limited extent, possibly similar to the magma chamber at 3-7 km depth under the Krafla Volcano in the axial rift zone in northeastern Iceland.

7. Focal mechanism solutions with large components of reverse faulting have been found at the ridge axis in two areas, near 50°N and in the Bárdarbunga central volcano near the center of the Iceland hot spot. Several mechanisms may be found to explain reverse faulting at a divergent plate boundary, but the one favoured here is brittle failure above a deflating, localized magma chamber.

8. Several cases of earthquake migration or correlation over considerable distances are found. In the case of the Krafla Volcano epicentral migration along the rift occurs during deflation of the caldera region, indicating lateral injection of magma. In other cases the correlation of seismic activity occurs over too great distances for each earthquake to directly influence the occurrence of the other. In these instances regional strain pulses associated with the large scale plate motions must be invoked to explain the correlations.

REFERENCES

Barazangi, M., and Dorman, J.
1969 : World seismicity of E.S.S.A. Coast and Geodetic Survey epicenter data for 1961–1967: Bulletin of the Seismological Society of America, v. 59, p. 369–380.

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

Björnsson, A., Johnsen, G., Sigurdsson, S., Thorbergsson, G. and Tryggvason, E.
1979 : Rifting of the plate boundary in north Iceland 1975–1978: Journal of Geophysical Research, v. 84, p. 3029–3038.

Björnsson, S. and Einarsson, P.
1981 : Jardskjálftar (Earthquakes); in Náttúra Íslands (in Icelandic), ed. H. Thórarinsdóttir, S. Steinthórsson: Almenna Bókafélagid, Reykjavík, p. 121–155.

Bonatti, E.
1978 : Vertical tectonism in oceanic fracture zones: Earth and Planetary Science Letters, v. 37, p. 369–379.

Bonatti, E., and Chermak, A.
1981 : Formerly emerging crustal blocks in the equatorial Atlantic: Tectonophysics, v. 72, p. 165–180.

Bonatti, E., Sartori, R., and Boersma, A.
1983 : Vertical crustal movements at the Vema Fracture Zone in the Atlantic: Evidence from dredged limestones: Tectonophysics, v. 91, p. 213–232.

Brandsdóttir, B., and Einarsson, P.
1979 : Seismic activity associated with the September 1977 deflation of the Krafla central volcano in NE Iceland: Journal of Volcanology and Geothermal Research, v. 6, p. 197–212.

Bungum, H., Mitchell, B. J., and Kristoffersen, Y.
1982 : Concentrated earthquake zones in Svalbard: Tectonophysics, v. 82, p. 175–188.

Cann, J. R.
1974 : A model for oceanic crustal structure developed: Geophysical Journal of the Royal Astronomical Society, v. 39, p. 169–187.

Conant, D. A.
1972 : Six new focal mechanism solutions for the Arctic and a center of rotation for plate movements; [M.A. Thesis] Columbia University, New York.

Duschenes, J., Lilwall, R. C., and Francis, T.J.G.
1983 : The hypocentral resolution of microearthquake surveys carried out at sea: Geophysical Journal of the Royal Astronomical Society, v. 72, p. 435–451.

Duschenes, J. D., and Solomon, S. C.
1977 : Shear wave travel time residuals from oceanic earthquakes and the evolution of oceanic lithosphere: Journal of Geophysical Research, v. 82, p. 1985–2000.

Einarsson, P.
1976 : Relative location of earthquakes within the Tjörnes Fracture Zone: Societas Scientiarum Islandica, Greinar V, p. 45–60.
1978 : S-wave shadows in the Krafla caldera in NE-Iceland, evidence for a magma chamber in the crust: Bulletin Volcanologique, v. 41, p. 1–9.
1979 : Seismicity and earthquake focal mechanisms along the mid-Atlantic plate boundary between Iceland and the Azores: Tectonophysics, v. 55, p. 127–153.
1985 : Compilation of earthquake fault plane solutions in the North Atlantic and Arctic Oceans: Manuscript, 27 pp.

Einarsson, P. and Björnsson, S.
1979 : Earthquakes in Iceland: Jökull, v. 29, p. 37–43.

Einarsson, P., Björnsson, S., Foulger, G., Stefánsson, R., and Skaftadóttir, T.
1981 : Seismicity pattern in the South Iceland seismic zone: in Earthquake Prediction—An International Review, Maurice Ewing Series 4, American Geophysical Union, p. 141–151.

Einarsson, P. and Brandsdóttir, B.
1980 : Seismological evidence for lateral magma intrusion during the July 1978 deflation of the Krafla volcano in NE-Iceland: Journal of Geophysics, v. 47, p. 160–165.

Einarsson, P. and Eiríksson, J.
1982 : Earthquake fractures in the districts Land and Rangárvellir in the South Iceland Seismic Zone: Jökull, v. 32, p. 113–120.

Einarsson, P., Klein, F. W., and Björnsson, S.
1977 : The Borgarfjördur earthquakes of 1974 in West Iceland: Bulletin of the Seismological Society of America, v. 67, p. 187–208.

Einarsson, T.
1968 : Submarine ridges as an effect of stress fields: Journal of Geophysical Research, v. 73, p. 7561–7575.

Fleischer, U.
1974 : The Reykjanes Ridge—A summary of geophysical data; in Geodynamics of Iceland and the North Atlantic Area, ed. L. Kristjánsson: Dordrecht, Netherlands, Boston, U.S.A., D. Reidel Publishing Company, p. 17–31.

Francis, T.J.G.
1968a: The detailed seismicity of mid-ocean ridges: Earth and Planetary Science Letters, v. 4, p. 39–46.
1968b: Seismicity of mid-oceanic ridges and its relation to properties of the upper mantle and crust: Nature, v. 220, p. 899–901.
1973 : The seismicity of the Reykjanes Ridge: Earth and Planetary Science Letters, v. 18, p. 119–124.

Francis, T.J.G., and Porter, I. T.
1971 : A statistical Study of mid-Atlantic Ridge earthquakes: Geophysical Journal of the Royal Astronomical Society, v. 24, p. 31–50.

Francis, T.J.G., Porter, I. T., and Lilwall, R. C.
1978 : Microearthquakes near the eastern end of St. Paul's Fracture Zone: Geophysical Journal of the Royal Astronomical Society, v. 53, p. 201–217.

Francis, T.J.G., Porter, I. T., and McGrath, J. R.
1977 : Ocean-bottom seismograph observations on the Mid-Atlantic Ridge near 37°N: Geological Society of America Bulletin, v. 88, p. 664–677.

Fujita, K., Hasegawa, H., Wetmiller, R., and Forsyth, D.
1985 : Seismicity and focal mechanisms in the Arctic region: in press

Gutenberg, B., and C. F. Richter
1941 : Seismicity of the Earth: Geological Society of America, Special Papers No. 34, 125 pp.
1949 : Seismicity of the Earth and Associated Phenomena: Princeton University Press, 273 p.

Hall, J. M., and Robinson, P. T.
1979 : Deep crustal drilling in the North Atlantic Ocean: Science, v. 204, p. 573–586.

Hart, R. S.
1978 : Body wave observations from the September, 1969, North Atlantic Ridge earthquake (Abstract): EOS, Transactions of the American Geophysical Union, v. 59, p. 326.

Hirn, A., Haessler, A., Hoang Trong, P., Wittlinger, G., and Mendes Victor, L. A.
1980 : Aftershock sequence of the January 1st, 1980, earthquake and present-day tectonics in the Azores: Geophysical Research Letters, v. 7, p. 501–504.

Horsfield, W. T., and Maton, P. I.
1970 : Transform faulting along the De Geer line: Nature, v. 226, p. 256–257.

Husebye, E. S., Gjoystdal, H., Bungum, H., and Eldholm, O.
1975 : The seismicity of the Norwegian and Greenland Seas and adjacent continental shelf areas: Tectonophysics, v. 26, p. 55–70.

Isacks, B., Oliver, J., and Sykes, L. R.
1968 : Seismicity and the new global tectonics: Journal of Geophysical Research, v. 73, p. 5855–5899.

Johnson, G. L., and Vogt, P. R.
1973 : Mid-Atlantic ridge from 47° to 51°N: Geological Society of America Bulletin, v. 84, p. 3443–3462.

Kanamori, H.
1977 : The energy release in great earthquakes: Journal of Geophysical Research, v. 82, p. 2981–2987.

Kanamori, H., and Stewart, G. S.
1976 : Mode of strain release along the Gibbs fracture zone, mid-Atlantic ridge: Physics of Earth and Planetary Interiors, v. 11, p. 312–332.

Klein, F. W., Einarsson, P., and Wyss, M.
1973 : Microearthquakes on the mid-Atlantic plate boundary on the Reykjanes Peninsula in Iceland: Journal of Geophysical Research, v. 78, p. 5084–5099.
1977 : The Reykjanes Peninsula, Iceland, earthquake swarm of September 1972 and its tectonic significance: Journal of Geophysical Research, v. 82, p. 865–888.

Krause, D. C., and Watkins, N. D.
1970 : North Atlantic crustal genesis in the vicinity of the Azores: Geophysical Journal of the Royal Astronomical Society, v. 19, p. 261–283.

Laughton, A. S., and Whitmarsh, R. B.
1974 : The Azores-Gibraltar plate boundary; in Geodynamics of Iceland and the North Atlantic Area, ed. L. Kristjánsson: Dordrecht, Netherlands, D., Reidel Publishing Company, p. 63–81.

Lewis, B.T.R.
1983 : The process of formation of oceanic crust: Science, v. 220, p. 151–157.

Lilwall, R. C.
1982 : Seismicity of the oceanic rifts; in Continental and Oceanic Rifts, ed. G. Pálmason: Geodynamics Series, v. 8, American Geophysical Union, p. 63–80.

Lilwall, R. C., Francis, T.J.G., and Porter, I. T.
1977 : Ocean-bottom seismograph observations on the Mid-Atlantic Ridge near 45°N.: Geophysical Journal of the Royal Astronomical Society, v. 51, p. 357–370.
1978 : Ocean-bottom seismograph observations on the Mid-Atlantic Ridge near 45°N—further results: Geophysical Journal of the Royal Astronomical Society, v. 55, p. 255–262.

Limond, W. Q., and Recq, M.
1981 : Possible earthquake migration along the Western European plate boundary: Journal of Geophysical Research, v. 86, p. 11623–11630.

McKenzie, D.
1972 : Active tectonics of the Mediterranean region: Geophysical Journal of the Royal Astronomical Society, v. 30, p. 109–185.

Meissner, R., and Strehlau, J.
1982 : Limits of stresses in continental crust and their relation to the depth-frequency distribution of shallow earthquakes: Tectonics, v. 1, p. 73–89.

Mitchell, B. J., Zollweg, J. E., Kohsmann, J. J., Cheng, C.-C., and Hang, E. J.
1979 : Intraplate earthquakes in the Svalbard archipelago: Journal of Geophysical Research, v. 84, p. 5620–5626.

Mogi, K.
1963 : Some discussion on aftershocks, foreshocks and earthquake swarms—The fracture of a semi-infinite body caused by an inner stress origin and its relation to the earthquake phenomena, 3: Bulletin of the Earthquake Research Institute, Tokyo University, v. 41, p. 615–658.

Molnar, P., and Oliver, J.
1969 : Lateral variations of attenuation in the upper mantle and discontinuities in the lithosphere: Journal of Geophysical Research, v. 74, p. 2648–2682.

Moreira, V.
1982 : Seismotectonics of mainland Portugal and its adjacent region in the Atlantic (in Portuguese, with English abstract), Lisbon, 29 pp.

Okal, E. A., and Stewart, L. M.
1982 : Slow earthquakes along oceanic fracture zones: Evidence for asthenospheric flow away from hotspots? Earth and Planetary Science Letters, v. 57, p. 75–87.

Pálmason, G.
1974 : The insular margins of Iceland; in The geology of continental margins, ed. C. A. Burk and C. L. Drake: Berlin, Springer Verlag, p. 375–379.

Reid, I., and Macdonald, K.
1973 : Microearthquake study of the Mid-Atlantic Ridge near 37°N, using sonobuoys: Nature, v. 246, p. 88–90.

Rosendahl, B. R.
1976 : Evolution of oceanic crust, 2. Constraints, implications, and inferences: Journal of Geophysical Research, v. 81, p. 5305–5314.

Rowlett, H.
1981 : Seismicity at intersections of spreading centers and transform faults: Journal of Geophysical Research, v. 86, p. 3815–3820.

Rowlett, H., and Forsyth, D. W.
1984 : Recent faulting and microearthquakes at the intersection of the Vema Fracture Zone and the Mid-Atlantic Ridge: Journal of Geophysical Re-

search, v. 89, p. 6079–6094.
Saemundsson, K.
1974 : Evolution of the axial rifting zone in Northern Iceland and the Tjörnes fracture zone: Geological Society of America Bulletin, v. 85, p. 495–504.
Sanford, A. R., and Einarsson, P.
1982 : Magma chambers in rifts; in Continental and oceanic rifts, ed. G. Pálmason: Geodynamics Series, v. 8, American Geophysical Union, p. 147–168.
Savostin, L. A., and Karasik, A. M.
1981 : Recent plate tectonics of the Arctic Basin and of northeastern Asia: Tectonophysics, v. 74, p. 111–145.
Scholz, C. H.
1968 : The frequency-magnitude relation of micro-fracturing in rock and its relation to earthquakes: Bulletin of the Seismological Society of America, v. 58, p. 399–415.
Searle, R.
1980 : Tectonic pattern of the Azores spreading centre and triple junction: Earth and Planetary Science Letters, v. 51, p. 415–434.
1981 : The active part of Charlie-Gibbs Fracture Zone: A study using sonar and other geophysical techniques: Journal of Geophysical Research, v. 86, p. 243–262.
Sleep, N. H.
1975 : Formation of oceanic crust: Some thermal constraints: Journal of Geophysical Research, v. 80, p. 4037–4042.
1978 : Thermal structure and kinematics of mid-oceanic ridge axis, some implications to basaltic volcanism: Geophysical Research Letters, v. 5, p. 426–428.
Soernes, A., and Fjeldskaar, W.
1980 : The local seismicity in the Jan Mayen area: Norsk Polarinstitutt, Skrifter Nr. 172, p. 21–32.
Spindel, R. C., Davis, S. B., Macdonald, K. C., Porter, R. P., and Philips, J. D.
1974 : Microearthquake survey of median valley of the Mid-Atlantic Ridge at 36°30′N: Nature, v. 248, p. 577–579.
Stover, C. W.
1973 : Seismicity and tectonics of the East Pacific Ocean: Journal of Geophysical Research, v. 78, p. 5209–5220.
Sykes, L. R.
1965 : The seismicity of the Arctic: Bulletin of the Seismological Society of America, v. 55, p. 519–536.
1967 : Mechanism of earthquakes and nature of faulting on the mid-ocean ridges: Journal of Geophysical Research, v. 72, p. 2131–2153.
1970 : Earthquake swarms and sea-floor spreading: Journal of Geophysical Research, v. 75, p. 6598–6611.
Talwani, M., Windisch, C. C., and Langseth, M. G.
1971 : Reykjanes Ridge crest: A detailed geophysical study: Journal of Geophysical Research, v. 76, p. 473–517.
Tams, E.
1922 : Die seismischen Verhältnisse des Europäischen Nordmeeres: Centralblatt für Mineralogie, Geologie und Palaeontologie, Jahrgang 1922, p. 385–397.
1927a: Die seismischen Verhältnisse des offenen Atlantischen Ozeans: Zeitschrift für Geophysik, v. 3, p. 361–363.
1927b: Die seismischen Verhältnisse des offenen Atlantischen Ozeans: Gerlands Beiträge zur Geophysik, v. 18, p. 319–353.
Thorarinsson, S., Saemundsson, K., and Williams, R. S.
1973 : ERTS—1 Image of Vatnajökull: Analysis of glaciological, structural and volcanic features: Jökull, v. 23, p. 7–17.
Toomey, D. R., Solomon, S. C., Purdy, G. M., and Murray, M. H.
1985 : Microearthquakes beneath the median valley of the Mid-Atlantic Ridge near 23°N: Hypocenters and focal mechanisms: Journal of Geophysical Research, in press.
Tréhu, A. M., Nábělek, J. L., and Solomon, S. C.
1981 : Source characterization of two Reykjanes Ridge earthquakes: Surface waves and moment tensors; P waveforms and non-orthogonal nodal planes: Journal of Geophysical Research, v. 86, p. 1701–1724.
Tryggvason, E.
1973 : Seismicity, earthquake swarms and plate boundaries in the Iceland region: Bulletin of the Seismological Society of America, v. 63, p. 1327–1348.
Udias, A.
1982 : Seismicity and seismotectonic stress field in the Alpine-Mediterranean region: Alpine Mediterranean Geodynamics. American Geophysical Union, Geodynamics Series, v. 7, p. 75–82.
Udias, A., Lopez Arroyo, A., and Mezcua, J.
1976 : Seismotectonics of the Azores-Alboran region: Tectonophysics, v. 31, p. 259–289.
Van Bemmelen, R. W.
1964 : The evolution of the Atlantic mega-undation: Tectonophysics, v. 1, p. 385–430.
Vogt, P. R.
1978 : Long wavelength gravity anomalies and intraplate seismicity: Earth and Planetary Science Letters, v. 37, p. 465–475.
Vogt, P. R., and Avery, O. E.
1974 : Detailed magnetic surveys in the northeast Atlantic and Labrador Sea: Journal of Geophysical Research, v. 79, p. 363–389.
Vogt, P. R., Kovacs, L. C., Bernero, C., and Srivastava, S. P.
1982 : Asymmetric geophysical signatures in the Greenland-Norwegian and Southern Labrador Seas and the Eurasian Basin: Tectonophysics, v. 89, p. 95–160.
Ward, P. L.
1971 : New interpretation of the geology of Iceland: Geological Society of America Bulletin, v. 82, p. 2991–3012.
Weidner, D. J., and Aki, K.
1973 : Focal depth and mechanism of mid-ocean ridge earthquakes: Journal of Geophysical Research, v. 78, p. 1818.
Whitmarsh, R. B., and Lilwall, R. C.
1983 : Ocean-bottom seismographs; in Structure and Development of the Greenland-Scotland Ridge, ed. M.H.P. Bott, S. Saxov, M. Talwani and J. Thiede. Plenum Press, New York and London, p. 257–286.
Wilson, J. T.
1965 : A new class of faults and their bearing on continental drift: Nature, v. 207, p. 343–347.

MANUSCRIPT ACCEPTED BY THE SOCIETY FEBRUARY 25, 1985

ACKNOWLEDGMENTS:

The author received help from various institutions and individuals during the preparation of this paper. Financial contribution was obtained from the Icelandic Science Fund. Extensive use was made of the WWSSN film chips library at Lamont-Doherty Geological Observatory, the help and hospitality of Drs. R. Bilham and D. Simpson is gratefully acknowledged. Sigfús Johnsen, Sigurdur E. Pálsson, and Bryndís Brandsdóttir helped with computing and plotting, Kristín Pálsdóttir and Ágústa Thorláksdóttir typed the manuscript.

Printed in U.S.A.