Earthquakes and present-day tectonism in Iceland

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ABSTRACT


The mid-Atlantic plate boundary in Iceland is expressed by a series of seismic and volcanic zones. The structure of the plate boundary is strongly influenced by the Iceland hot spot. The relative motion of the Mid-Atlantic Ridge with respect to the hot spot leads to ridge jumps, propagating rifts and other complexities. Most large earthquakes in Iceland occur within two transform zones that connect the presently active Northern and Eastern Volcanic Zones to the ridges offshore. In the south the South Iceland Seismic Zone is marked by a 10–15 km wide, E-trending epicentral belt. The large earthquakes occur by faulting on N–S striking right-lateral faults. The left-lateral transform motion along the zone thus appears to be taken up by slip on numerous parallel faults by counterclockwise rotation of the blocks between them (bookshelf tectonics). It is argued that the South Iceland Seismic Zone is a transient feature, migrating sideways in response to propagation of the Eastern Volcanic Zone. In Northern Iceland the transform motion is taken up along the Tjörnes Fracture Zone. At least three parallel, NW-trending seismic belts have been identified within the zone. The seismicity of the volcanic zones is characterized by spatial clustering of epicenters. Most clusters coincide with central volcanoes. Rifting structures such as fissure swarms and normal faults are mostly aseismic except during episodes of rifting and magmatism such as the present events in Krafla. Earthquake recording has been used very successfully at Krafla to monitor the level of inflation and deflation of the volcano, and to trace the path and speed of lateral magmatic intrusions into the associated fissure swarm. Seismic activity at the Bárðarbunga volcano in Central Iceland correlates in time with the Krafla events, and it seems as if inflation of Krafla is followed by deflation of Bárðarbunga. It is postulated that the pressure drop in the partially molten mantle beneath Krafla is transmitted to neighboring volcanoes, leading to magma withdrawal from their shallow reservoirs. Bursts of seismicity of the Katla volcano in South Iceland in 1967 and 1977 may similarly be the result of magma withdrawal in response to the 1964–1967 Surtsey and 1973 Heimaey eruptions. Annual periodicity seen in the Katla seismicity is explained as the result of the triggering effect of pore pressure in the crust beneath the glacier covering Katla. Several volcanoes exhibit persistent, low-magnitude seismicity. In the Hengill volcano in Southwest Iceland, many events involve a non-double-couple mechanism. The seismicity is interpreted as the result of extensional failure and heat extraction from a cooling magma chamber.

Two classes of intraplate earthquakes have been identified in Iceland. One includes events in the lithospheric block between the transform zones. These events are related to crustal extension above the hot spot. The other class includes events on the insular shelf off the east and southeast coasts which are possibly caused by a differential cooling rate in the crust across the shelf edge.

Introduction

Most of the seismicity of Iceland is in some way related to the mid-Atlantic plate boundary that crosses the country (Fig. 1). The relatively simple tectonic picture of a mid-oceanic plate boundary with spreading centers and transform faults, however, does not readily apply to Icelandic tectonics. In Iceland, the plate boundary is superimposed on a large hot spot with a presumed deep root in the mantle. The excessive hot spot volcanism has built up the Icelandic lava pile and is responsible for Iceland’s existence. The volcanism is fed by a partial melting layer that underlies most of the island (Schmeling, 1985; Gebrände et al., 1980). The plate boundary slowly drifts westwards with respect to the hot spot, resulting in successive eastwards jumps of the spreading zones. It is postulated that a new rift zone starts propagating from the center of the hot
spot when the plate boundary has migrated some critical distance off it. At times, two or more parallel zones have been active simultaneously (Jóhannesson, 1980; Helgason, 1985). At least one ridge jump has occurred in North Iceland, and in South Iceland a second ridge jump seems to be presently in progress. The Western Volcanic Zone (WVZ in Fig. 1) is being replaced by the Eastern Volcanic Zone (EVZ), which seems to be propagating southwestwards, away from the center of the hot spot in east Central Iceland (Pálmason, 1981; Einarsson and Eiríksson, 1982; Óskarsson et al., 1985; Meyer et al., 1985). A propagation speed of 3.5–5 cm/yr has been estimated from the apparent rotation of the Western Volcanic Zone (Einarsson, 1988b).

In response to this unstable situation, complex fracture zones have developed in the north (the Tjörnes Fracture Zone, TFZ) and south (the South Iceland Seismic Zone, SISZ) which connect the eastward displaced rift zones in Iceland to the spreading centers of the Mid-Atlantic Ridge to the north and south of the island. Differential movements are taken up by these fracture zones. Because they are relatively young and the stress field that induces them unstable, complex series of

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Fig. 1. Epicenters and focal mechanisms of earthquakes in the Iceland area. Epicenters are taken from the earthquake lists (PDE) of the U.S. Geological Survey for the period 1963–1987; only epicenters determined using ten or more stations are included. Larger dots are events of magnitude \( m_b \) \( \geq 5 \). Focal mechanisms are shown schematically on lower hemisphere stereographic projections of the focal sphere; compressional quadrants are black. For further references to the solutions see Einarsson (1987, 1988a), Bjarnason and Einarsson (1989) and Foulger (1988a, b). The volcanic zones (WVZ, EVZ, and NVZ) are stippled, and seismic belts in the South Iceland Seismic Zone (SISZ) and the Tjörnes Fracture Zone (TFZ) are shown with heavy lines.
faults are formed that are not oriented parallel to the spreading direction (Einarsson, 1976; Einarsson et al., 1981; Einarsson and Eiríksson, 1982). In comparison with most large submarine fracture zones, they exhibit subdued surface morphology, which may indicate that the individual fractures are only transient.

The structure of the volcanic zones is characterized by structural units called volcanic systems (Saemundsson, 1974, 1978; Jakobsson, 1979). These volcanic systems are arranged subparallel or in an en echelon fashion within the volcanic zones (Fig. 2). Each volcanic system consists of a central volcano and a fissure swarm that transects it. Volcanic production is highest in the central volcano of each system, but fissure eruptions may occur out to considerable distance in the fissure swarm area. A geothermal field, silicic volcanism and a caldera structure are frequently associated with the central volcano.

As an addition to this complex picture, the volcanism and seismicity are not confined to the continuous plate boundary. Intraplate volcanic zones, or flank zones (Jakobsson, 1972; Saemundsson, 1978) exist (e.g. on Snæfellsnes in West Iceland and south of the ridge-transform intersection near 64°N in South Iceland). These zones contain active volcanoes, but the overall production rate is relatively low and rifting structures are insignificant. Intraplate earthquakes also occur, mainly in West and North Iceland and along the insular shelf edge off the east coast (Einarsson, 1989).

This paper is intended as overview of the seismicity of Iceland and a review of current ideas on how it relates to other tectonic and volcanic

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**Fig. 2.** Epicenters of earthquakes in Iceland, 1982–1985. Shown are all earthquakes of magnitude ≥ 2 that could be located by local seismic data with errors smaller than 5 km. Almost all epicenters have errors smaller than 2 km. Active volcanic systems, with their central volcanoes, fissure swarms, and calderas, are also shown (Einarsson and Saemundsson, 1987).
phenomena. This work is based mostly on the analysis of new seismic data from a local, short-period seismograph network. The network has expanded greatly since 1974, and since then between 30 and 40 stations have been in operation (Einarsson and Björnsson, 1987). Station spacing is 50–100 km in most of the active zones, but in some areas 10–20 km is more typical. Epicentral locations are computed with the HYPOINVERSE program (Klein, 1978). The formal standard errors of almost all the epicentral locations in Figs. 2 and 4–8 are smaller than 2 km.

Reykjanes Peninsula

The Reykjanes Peninsula is a zone of high seismicity and 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. The plate boundary can be defined as a narrow seismic zone that enters Iceland near the tip of Reykjanes (Fig. 3) and runs along the peninsula. Detailed studies by Klein et al. (1973, 1977) show that the seismic zone is less than 2 km wide in most places. The earthquakes mostly occur at a depth of 1–5 km and are not located on any one particular fault.

The seismicity seems to be caused by deformation of a brittle crust above a deeper seated, aseismic deformation zone. Small-scale structures and seismic lineations can be resolved within the zone, 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 (Klein et al., 1973, 1977), and for two larger events using telesismic data (Ward, 1971; Einarsson, 1979). The minimum compressive stress in this area is consistently horizontal, oriented in a northwesterly direction. The maximum stress rotates between the vertical direction, causing normal faulting on NE-striking faults, and the horizontal, northeasterly, direction, resulting in 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 direction according to local, or even time-dependent conditions. Dikes are formed when fissures open up against the minimum stress. They strike northeast, as shown by the eruptive fissures at the surface.

The mode of strain release changes systematically along the peninsula. Near the tip of Reyk-

Fig. 3. Seismic zone and fissure swarms on the Reykjanes Peninsula. The shaded areas are epicentral areas of the largest earthquake swarms during the period 1971–1975 (data from S. Björnsson, in Halldórsson et al., 1984). Fissures and faults are from the geological map of Sæmundsson and Einarsson (1980).
janes, earthquakes occur in swarms, i.e. 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 half of the peninsula has not been mapped in detail because of the low rate of activity there in recent years.

The Hengill triple junction

The Reykjanes Peninsula plate boundary splits up near 21.5°W into the Western Volcanic Zone and the South Iceland Seismic Zone. This triple junction is an area of high, persistent seismicity, and shows up on the seismicity map (Figs. 2 and 4) as a dense cluster of events. The seismicity of the area has been studied by Foulger (1988 a, b). The most prominent structural feature is the Hengill central volcano with its fissure swarm. The seismicity extends over a much wider area, however, including the extinct Grensdalur central volcano, 5 km east of Hengill. Hypocenters were found to be in the depth range 1–7 km. Focal mechanisms were determined for numerous small events. About half of them showed results similar to those found on the Reykjanes Peninsula, i.e. normal and strike-slip faulting in response to a horizontal, NW-oriented minimum compressive stress. The other half showed evidence of a non-double-couple component in the mechanism. Some of the data could be explained by pure extensional failure. These events were mostly located in areas of geothermal activity, and the mechanisms were

Fig. 4. Epicenters in the South Iceland Seismic Zone, 1974–1986 and May 1987. Thin lines show faults and fissures of the volcanic zones (from Saemundsson and Einarsson, 1980; Jóhannesson et al., 1982), heavy lines are strike-slip faults of the seismic zone (from Einarsson and Saemundsson, 1987). The dense cluster on the left-hand side of the map is in the Hengill geothermal area. The circular structures in the lower right corner are the calderas of the Tindfjöll and Eyjafjöll volcanoes.
interpreted as the result of active heat extraction from the hot crust. Circulating groundwater causes cooling and contraction of the rock, thus reducing the confining stress and leading to extensional failure. Thermal and regional tectonic stress is released and the water gains access to new hot rock through cracks propagating into the hot rock body in the root of the volcanoes (Björnsson et al., 1982).

South Iceland Seismic Zone

Plate divergence in the southern part of Iceland is accommodated by two subparallel rift zones, the Western and the Eastern Volcanic Zones. The gap between them is bridged in the south by a transform zone, the South Iceland Seismic Zone, which takes up the transform motion between the Reykjanes Ridge and the Eastern Volcanic Zone (Fig. 1). No comparable seismic zone presently links the northern end of the Western Volcanic Zone to the plate boundary in Central Iceland. Partly for this reason one can conclude that rifting is dying out in the Western Zone, and is being taken over by the Eastern Zone.

The South Iceland Seismic Zone crosses the populated lowland in South Iceland from west to east, and extends into the Eastern Volcanic Zone, which has a width of about 60 km at this latitude. This junction is marked by the Hekla, Vatnajökull and Torfajökull volcanic systems (Fig. 2). Throughout the history of Iceland, the seismic zone has produced numerous destructive earthquakes, some of which have been well documented. The zone was identified as a transform fault zone by Ward (1971), primarily on the basis of its geometric relationship to the ridge system and one poorly constrained fault plane solution that showed strike-slip faulting on an E–W or N–S striking plane. The spatial and temporal pattern of the seismicity was summarized by Einarsson et al. (1981). The zone is delineated by areas of destruction created by historical earthquakes, surface ruptures, and instrumentally determined epicenters. It is oriented E–W and is 10–15 km wide. The destruction areas of individual earthquakes and surface faulting show, however, that each event is associated with faulting on N–S striking planes, perpendicular to the main zone. Destruction areas are elongate in the N–S direction, and detailed mapping of surface fractures (Einarsson and Eiríksson, 1982) reveals N–S trending arrays of en-echelon tension fractures indicating right-lateral faulting. The overall left-lateral transform motion along the zone thus appears to be accommodated by right-lateral faulting on many parallel, transverse faults and counterclockwise rotation of the blocks between them. Einarsson and Eiríksson (1982) speculated that this was an indication of the transient nature of the plate boundary in South Iceland, and suggested that the seismic zone was migrating southwards in response to propagation of the Eastern Volcanic Zone. (It is noteworthy here that a mode of crustal movement similar the left/right-lateral combination just mentioned has been described in Southern California by Nicholson et al. (1986).

Major earthquake sequences, which affect most of the zone, tend to last from a few days to about 3 years. These sequences occur at average intervals of 80–100 years. Each sequence typically begins with a magnitude 7 event in the eastern part of the zone, followed by smaller events further west. The last sequences occurred in 1732–1934, 1784 and 1896, so the next sequence is expected soon (Einarsson, 1985). In addition to these major sequences, large localized events may occur near the eastern and western ends of the zone (such as the magnitude 7 earthquake west of Hekla in 1912).

The Vatnajökull earthquake in 1987 ($M_e = 5.8$) was the largest event in the zone since 1912. It occurred in the eastern part of the zone, at the junction with the Eastern Volcanic Zone. Locations of foreshocks and aftershocks show that the earthquake was associated with slip on a N-striking fault, and inversion of teleseismic body waves shows right-lateral strike-slip on a nearly vertical fault (Bjarnason and Einarsson, 1989). The rupture was initiated at a depth of 11 km, near the base of the crust, and propagated upwards. However, it did not extend to the surface. The depth range of the entire earthquake sequence was 6–14 km. The strike-slip mechanism gives the event a definite resemblance to earthquakes in the seismic zone to the west. Volcanic processes such as injection or withdrawal of magma, or thermal contrac-
tion as envisaged by Pálmason (1981), lead to dip-slip earthquakes and are considered unlikely as causes of the Vatnafjöll earthquake. Transform faulting is thus shown to extend well into the rift zone, possibly as far as the Torfajökull central volcano.

The long-term forecast of a major earthquake sequence within the next few years has led to a modest effort in earthquake prediction research in the South Iceland Seismic Zone (Einarsson et al., 1981). The area was designated by the Council of Europe as one of five test areas for earthquake prediction in Europe. Radon has been monitored at a few sites since 1977, and premonitory radon anomalies have been identified in association with a few small earthquakes (Hauksson and Goddard, 1981; Hauksson, 1981). Volumetric strainmeters have been in operation in boreholes at seven sites since 1979. Gradual strain buildup as well as strain events resembling slowquakes have been recorded (Stefánsson et al., 1983; Stefánsson and Halldórsson, 1988). A geodetic network measuring $15 \times 65$ km was installed in 1984 mainly in order to detect coseismic and possibly preseismic strain (Thorbergsson, 1985; Erlingsson and Einarsson, 1985, 1990). This network was augmented by a GPS survey in 1986 (Foulger et al., 1987, 1989). Several short-period analog seismographs have been in operation in the area for over a decade, and a new digital network is presently being designed by a group of Nordic seismologists (see Stefánsson et al., 1986).

Fig. 5. Epicenters of earthquakes in North Iceland (1981–1986) and structural elements of the Northern Volcanic Zone and Tjörnes Fracture Zone. Lines offshore show normal faults and flexures, from McMaster et al. (1977); fissure swarms of the Northern Volcanic Zone are from Saemundsson (1978). The contours off the mouth of Eyjafjörður are the 10 mGal (outer) and 0 mGal (inner) gravity isolines of the strong gravity low associated with the Hú Savik faults (Pálmason, 1974). NW–SE trending topographic scarps east of Eyjafjörður possibly mark the eastwards continuation of the Dalvík seismic zone.
Tjörnes Fracture Zone

The other major fracture zone of Iceland, the Tjörnes Fracture Zone, is located near the north coast. It is a broad zone of seismicity, transform faulting and crustal extension that connects the southern end of the submarine Kolbeinsey Ridge to the volcanic zone in North Iceland. The seismicity is too diffuse to be associated with one 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 can be readily identified in Figs. 2 and 5. The northernmost one, the Grimsey zone, which is located entirely offshore, has in recent years been the scene of frequent earthquake swarms, especially near its western end. These swarms may last days or weeks, and their maximum magnitude exceeds 5. Two fault plane solutions show right-lateral strike-slip faulting along the zone (Einarsson, 1987). In spite of the clear NW–SE alignment of epicenters the structure of the Grimsey zone is dominated by N–S structural elements. A series of graben-like troughs with this trend have been identified (Fig. 5) (McMaster et al., 1977). These structural elements are most likely related to crustal dilation, whereas the larger earthquakes seem to be associated with horizontal shear and transform motion. In some respects this situation resembles that on the Reykjanes Peninsula, where the fissure swarms trend obliquely to the seismic zone. The seismic zone on the Reykjanes Peninsula also lacks a clear surface expression.

The second zone is about 40 km south of the first one, and is well defined by the seismicity near its western end (Figs. 2 and 5). Here it is also marked by a pronounced negative gravity anomaly (Fig. 5) (Pálsson, 1974). The eastern end has been seismically quiet in recent years, but the fault zone can nevertheless be traced on the ocean bottom to the coast at the town of Húsavík, continuing on land into the volcanic zone. The Húsavík Fault Zone is expressed in the surface geology as a zone of NW–SE transform faulting (Saemundsson, 1974; Young et al., 1985). A major earthquake sequence occurred in this zone in 1872, with extensive damage and surface ruptures in Húsavík (Björnsson and Einarsson, 1981). A third zone is indicated about 30 km south of the Húsavík faults. Earthquakes in recent years have been primarily in its western part, but a damaging earthquake did occur in the town of Dalvík in 1934. Attempts to find a surface expression of the western part of this zone have so far failed. East of Dalvík, on the other hand, the continuation of the zone is marked by a major valley. It is likely that the M 7 earthquake of 1963 in the mouth of Skagafjörður (Fig. 1 and 5) occurred in the western end of the Dalvík zone, offshore. Fault plane solutions of that event show right-lateral strike-slip along the zone (Stefánsson, 1966; Sykes, 1967).

Northern Volcanic Zone

The Northern Volcanic Zone is about 50 km wide and is composed of five volcanic systems arranged in echelon along the plate boundary. Three of the central volcanoes have developed calderas, Krafla, Askja and Kverkfjöll. Most of the zone displays quite low seismic activity (only three prominent clusters show up on the map in Fig. 2) and yet a major rifting episode has been in progress within this zone since 1975. The activity has been confined to the Krafla central volcano and its associated fissure swarm (Fig. 5). An 80 km long segment of the plate boundary has been activated and the accumulated rifting has reached a maximum of about 9 m.

Like most of the central volcanoes within the rift zones of Iceland the Krafla central volcano is a rather indistinct structure. The weak, divergent crust does not support high volcanic edifices and subsidence plays a major role in the development of these volcanoes. Therefore, rather than being prominent mountains, they are areas of high rates of volcanism, acidic volcanism, and geothermal activity. Some of them have caldera structures and all have an associated fissure swarm running through them. The Krafla central volcano has produced mostly basaltic material in the form of interglacial or postglacial lavas, or subglacial formations. A 10 × 7 km caldera is marked by fault scarps and arcuate structures (Saemundsson, 1982). On the basis of S-wave shadows, a crustal
magma chamber is inferred to underlie the central part of the caldera. This chamber is thought to derive its magma from a subcrustal, partially molten layer which is apparent from seismic and magnetotelluric measurements.

The current activity at Krafla has been studied with various methods, including field observations and seismological, geodetic, gravimetric, geochemical and petrological methods (e.g. Einarssson, 1978; Einarssson and Brandsdóttir, 1980; Wendt et al., 1985; Sigurðsson, 1980; Torge and Kanni- giesser, 1985; Tryggvason, 1980, 1983, 1986; Johnsson et al., 1980; Grönvold and Mäkipää, 1978; Grönvold, 1988; Björnsson et al., 1977). The following conceptual model of events is derived from these observations.

The activity of Krafla is primarily influenced by two processes, inflow of magma into the crustal chamber and rifting of the plate boundary. The regulating parameters are the magma pressure in the chamber and tectonic stress at the plate boundary. Tectonic stress presumably builds up slowly with time due to plate movements. If magma begins flowing into the magma chamber, it can trigger the release of tectonic stress and thus initiate rifting. Fissures propagate away from the magma chamber near the center of the volcano, and magma flows laterally along the fissure swarm to the north and south of the volcano for tens of kilometers and forms a dike. Propagation velocity may reach 0.5–1.2 m/s shortly after the initiation of intrusion, but then it gradually slows down as the dike lengthens. The pressure in the magma chamber drops and the volcano deflates during the emplacement of the dike. This process can last anything from hours to a few weeks, depending on the length of the dike, and the deflation can amount to centimeters or a few meters in the center of the caldera. The fissure swarm is strongly affected during the emplacement of the dike. Extensive fissuring and normal faulting occurs above the dike, the central part of the swarm subsides and the flanks are uplifted. Displacements, both horizontal and vertical, are of the order of tens of centimeters to meters. If magma inflow continues after a deflation event, the volcano reinflates at a typical rate of 1–10 mm/day for weeks or months. Many such inflation-deflation cycles may occur during a rifting episode that may last several years to over a decade. In the beginning of an episode the tectonic stress is high and most of the magma is emplaced in dikes in the crust. As the tectonic stress is released, the crust supports higher magma pressures, and an increasing proportion of the mobilized magma reaches the surface in eruptions.

Each phase of the activity is accompanied by its own characteristic type of seismic activity:

**Inflation earthquakes:** During inflation periods, earthquakes occur within the caldera, in the magma chamber roof. This activity starts when the stress in the roof exceeds the previous highest stress level, and is after that well correlated with the rate of inflation. The largest earthquakes reach magnitude 4. The activity stops immediately when inflation stops or deflation begins. Figure 6 shows the epicenters of earthquakes during a few inflation periods in 1982–1983. They are located at shallow levels in the crust (1–3 km), above the crustal magma chambers as shown by the S-wave shadow zones. Earthquakes have also been located at depths of 7–8 km, i.e. below the crustal chambers.

**Deflation earthquakes:** When the magma chamber deflates the deviatoric stress in the chamber roof is first relaxed and earthquakes cease. If the pressure in the magma chamber falls below a certain level, the roof begins to fail and collapses. Thus earthquake activity increases again in response to falling pressure. The deviatoric stress in the roof during this process is the opposite to that produced during inflation. Inflation leads to extension and normal faulting in the roof. Deflation leads to contraction and reverse faulting. The deflation earthquakes occupy the same hypocentral volume as the inflation earthquakes. The amount of deflation necessary to produce deflation earthquakes seems to be about 1 m in Krafla and earthquakes as large as magnitude 5 have been recorded.

**Rifting earthquakes:** Intense swarms of earthquakes accompany rifting and dike intrusions in the fissure swarm of Krafla. The earthquakes begin near the caldera shortly after deflation sets in, and then they propagate along the fissure swarm away from the caldera, presumably marking the propagating edge of the dike. Both high- and
low-frequency events occur, the latter appearing to be associated with surface faulting. The earthquakes only seldom exceed magnitude 4, in spite of extensive surface faulting.

*Intrusion tremors:* A continuous tremor accompanies deflation and dike intrusion. The tremor begins as soon as deflation starts and its amplitude is roughly correlated with the rate of deflation. The amplitude is fairly uneven and spasmodic, and the frequency spectrum appears broad. The tremor is mixed with rifting earthquakes.

*Eruption tremors:* When a dike reaches the surface and an eruption begins, a new type of
tremor appears on the seismograms. The eruption tremor has an even amplitude and a narrow frequency spectrum, with a predominant frequency below 3 Hz. Its amplitude appears to be correlated with the vigor of the eruption, as seen in, for example, the height of the lava fountains. The eruptions themselves are usually not accompanied by earthquakes.

During inflation periods almost all the changes measured occur within the caldera or in its immediate surroundings. The inflation bulge is centered on the center of the caldera, as shown by elevation changes, tilt measurements, distance measurements and gravimetry. These data fit well to a Mogi-type model of an inflating magma chamber at a depth of 2.6–3.0 km. The inflation is also reflected in the gradual widening of fissures within the caldera. Sometimes, changes in geothermal activity and composition of fumarolic gases are detected during inflation, particularly if the previous deflation event was accompanied by intrusion into the magma chamber roof within the caldera.

During deflation events, measurable changes occur both within the caldera and outside it. The onset of deflation is usually abrupt. Tiltmeters

Fig. 7. Epicenters in Central Iceland, 1975–1985. Volcanic systems are from Einarsson and Saemundsson (1987) with slight modifications based on new radio echosoundings of H. Björnsson (1988). The volcanic centers marked with B, G, K, T and A are Bárðarbunga, Grímsvötn, Kverkfjöll, Thórdarhyrna and Askja. The Vatnajökull ice sheet covers a large part of this area.
show a reversal in tilt within minutes or tens of minutes, and an intrusion tremor appears simultaneously on the seismograms. The inflation bulge subsides and rifting earthquakes appear outside the caldera. The deflation rate reaches a maximum within hours and then slowly decreases as the dike lengthens. If an eruption occurs, it begins 1–7 hours after the onset of deflation. So far, eruptions have lasted from a few hours to 14 days. The magma is highly fluid, basaltic in composition, and is erupted through 1–9 km long fissures. In several cases, large quantities of lava have flowed from the surface down into fissures, inducing secondary rifting that sometimes amounts to meters. Sometimes, very small phreatic eruptions lasting a few hours occur. Small steam fields have also been formed in the fissure swarm days or weeks after an intrusion.

Two main types of magma have so far been erupted during the Krafla events (Grönvold, 1984, 1988). The lavas are spatially separated, the more evolved type erupted in the caldera region, and the more primitive type north of the caldera. This may indicate that the magma chamber is divided and fed by different magma sources, or that the magma chamber is vertically stratified, with the more primitive and dense magma sitting below the more evolved one.

The Krafla activity is not over yet. Intermittent inflation has been continuing between 1985 and 1989 following a substantial eruption and deflation in September 1984.

Alternative models of the Krafla activity have been proposed by Pálmason (1981) and Björnsson (1985).

Three clusters of epicenters appear near the Askja central volcano (Fig. 7). One of them is located beneath the southeastern part of the volcano. The activity here is persistent, and every year a few events occur that are large enough to be located. The nature of this activity is unclear. Vertical crustal movements have been measured within the caldera, and interpreted as the result of inflation and deflation of a magma chamber beneath the central part of the volcano (Tryggvason, 1987). If our experience from Krafla can be applied here, the seismicity does not seem to be related to this process. The epicenters are spatially more correlated with the geothermal activity, so perhaps the Hengill volcano provides a closer analogy. If so, the earthquakes may be related to thermal cracking and heat extraction in the crust.

The second epicentral cluster is located east of Askja. Most of these events belong to two swarms that occurred in the summers of 1982 and 1983. The epicenters delineate a 20 km long, NE-SW trending zone. This zone possibly extends beneath the southeastern part of the Askja volcano and thus includes the epicentral cluster mentioned above. This zone does not correspond to any known geological structure at the surface, and its trend is oblique to most faults in the area.

The third cluster is northwest of Askja. It also does not correspond to any known structures. This area is flat and covered with recent lava flows.

**Volcanoes in Central Iceland**

It has been recognized for some time that seismic activity is generally high in the northwestern Vatnajökull area (Tryggvason, 1973; Brandsdóttir, 1984), the presumed central area of the Iceland hot spot (Sigvaldason et al., 1974). The area is largely covered by the Vatnajökull ice sheet, and its tectonic structure is poorly understood. It seems to be dominated by a group of central volcanoes, some of which have calderas (Thórarinsson et al., 1973; Björnsson, 1988). When the relatively dense seismograph network was installed in Iceland in 1973–1977, accuracy in epicentral determination improved and it became clear that almost all the seismic activity was associated with the different central volcanoes in the area, i.e. Bárðarbunga, Grímsvötn and Kverkfjöll (Fig. 7). By far the most seismically active was Bárðarbunga, but about 15 km south of it an east-west belt of seismicity is also quite active. This belt coincides with a row of subglacial geothermal areas (Björnsson, 1983). Together, the seismicity and the geothermal activity indicate that the belt represents a row of central volcanoes.

An unusual sequence of events began in June 1974 with an earthquake of \( m_b = 5 \). Since then, eight earthquakes of \( m_b \geq 5 \) have occurred at Bárðarbunga. Prior to 1974, earthquakes of this magnitude are not known to have occurred in this
area, since the beginning of instrumental observation in the 1920's (Tryggvason, 1973; Brandsdóttir, 1984). The present earthquake sequence is the most significant seismic occurrence in this area for at least half a century.

Five of the Bárdarbunga earthquakes were large enough to allow a fault plane solution to be obtained from P-wave first motions (Einarssson, 1986, 1987, 1988a). A common feature of all the solutions is a large component of reverse faulting.

To find reverse faulting at Bárdarbunga was certainly an unexpected result. Bárdarbunga is located on a divergent plate boundary near the center of the Iceland hot spot (see, for example, Sigvaldason et al., 1974). Focal mechanisms of earthquakes in other divergent parts of the mid-Atlantic plate boundary show almost exclusively normal faulting (Einarssson, 1987; Huang et al., 1986). Normal faulting at ridge crests is generally thought to be related to the formation of the rift valley by the down-dropping of a “keystone” at the boundary of the diverging plates. In this light it is clear that the Bárdarbunga earthquakes require an explanation that does not depend directly on movements of the major plates.

The first observation that comes to our attention is that earthquakes at Bárdarbunga are tightly clustered in the caldera region of the volcano. The depths of the hypocenters are not well constrained in this part of Iceland, but all data are consistent with a shallow source. If the caldera can be taken as evidence of the existence of a crustal magma chamber beneath Bárdarbunga, the earthquakes originate around, but mostly above that chamber. The earthquakes are therefore likely to be related to brittle failure of the crust in response to stress changes caused by pressure fluctuations in the magma chamber. It can be shown that pressure decrease in the magma chamber can lead to horizontal compression above the chamber and therefore reverse faulting in the chamber roof. It thus appears that the pressure in a crustal magma chamber beneath Bárdarbunga began decreasing in or shortly before 1974.

There is a striking correlation in time between the seismic activity at Bárdarbunga and the magmatic activity at Krafla central volcano in the rift zone 110 km to the north. Earthquakes at Bárdarbunga began when magma flow is inferred to have started into the Krafla crustal magma chamber. The activity culminated and faded away simultaneously in both volcanoes. In both volcanoes events of this extent and magnitude are unusual on a time scale of 50–250 years. The events are therefore likely to be causally related. The magma pressure in the shallow chamber of Bárdarbunga decreases when magma flows into the shallow Krafla chamber.

To explain this causal connection the following model of magma flow is proposed:

As stated earlier, a layer of partially molten material beneath the 10–20 km thick crust in Iceland is inferred from seismic and magnetotelluric data. This layer acts as a conductor of pressure variations between different volcanic systems, even if their magmas do not mix. Excessive magma accumulated in this layer beneath Krafla and was driven by buoyancy up into the crustal chamber. The pressure in the magma layer was locally reduced, and this pressure release pulse was transmitted to neighboring volcanic systems. Some of them, Bárdarbunga in particular, responded by magma flow out of the crustal chamber, down into the magma layer.

The subglacial Grímsvötn Volcano stands out on the epicentral map with a dense cluster of epicenters (Fig. 7). Most of this activity is related to an eruption in May, 1983 (Einarsson and Brandsdóttir, 1984). The eruption was preceded by increased seismic activity for 3–5 months, which was probably caused by inflation of a shallow magma chamber beneath the southeastern rim of the Grímsvötn caldera. An intense earthquake swarm immediately preceded the outbreak of the eruption, presumably when the chamber walls ruptured and magma flowed to the surface. After the eruption broke out, the earthquake activity was replaced by a volcanic tremor that continued off and on for 5 days while the eruptive activity lasted. The volcano was seismically quiet for several months after the eruption due to its deflated state. However, apparently magma was accumulating in its shallow chamber because seismicity increased again in the spring of 1984. This activity terminated in August 1984 with a short burst of volcanic tremor. It is inferred from
these observations that a small, subglacial eruption took place and reduced the pressure in the shallow magma chamber. No other evidence of this eruption has been found, and apparently the magma chamber has not been recharged since.

Epicentral clusters are seen in Kverkfjöll and Thórdarhyrna (K and T in Fig. 7), demonstrating activity in these volcanic centers.

Volcanoes in South Iceland

The Eastern Volcanic Zone southwest of the Vatnajökull volcanoes is devoid of earthquakes. The structure here is characterized by fissure swarms emanating from the Vatnajökull volcanoes to the northeast and a group of central volcanoes in South Iceland to the southwest. Similar to the

Fig. 8. Epicenters in the southern part of the Eastern Volcanic Zone, 1979–1985. The Vatnajökull earthquake sequence of May 1987 is also included. It forms the cluster south of Hekla. Volcanic fissures and faults are from the geological map by Jóhannesson et al. (1982). The Katla volcano and large parts of the Torfajökull and Eyjafjöll volcanoes are covered by glaciers.
Northern Volcanic Zone, the fissure swarms here have been largely aseismic in recent decades. The last known major magmatic event in this zone was in the 1860's (Thorarinsson and Sigvaldason, 1972).

South of its junction with the South Iceland Seismic Zone near 64°N the structure of the Eastern Volcanic Zone changes. The structure is dominated by a group of central volcanoes (Torpajökull, Hekla, Vatnajökull, Tindfjöll, Eyjafjöll, Katla and Vestmannaeyjar), and rifting structures are insignificant. Some of the volcanoes (Katla, Torfajökull and Vestmannaeyjar) display persistent seismic activity, whereas others (e.g. Hekla) are seismically active only during eruptions.

Historic eruptions of the subglacial volcano Katla have been preceded by felt earthquakes. The most recent eruptions were in 1918 and a possible small, subglacial eruption in 1955. In spite of the lack of eruptions in the last few decades, Katla is the most seismically active volcano of the area. The epicenters clearly define two separate clusters, one beneath the southeastern part of the Mýrdalsjökull glacier, where the Katla volcano has traditionally been supposed to be, and the other under the southwestern part of the glacier, defining a second volcanic center (Fig. 8). One poorly constrained fault plane solution in the southeastern cluster indicates strike-slip with a significant component of reverse faulting (Fig. 1). As in the case of Bárðarbunga, a deflating magma chamber may offer an explanation for this type of faulting. A pronounced annual cycle in the Katla seismicity was demonstrated by Tryggvason (1973). Brandsdóttir and Einarsson (1989) show that the periodicity is primarily in the activity of the southwestern cluster. Almost all seismic events there occur during the latter half of the year. To explain this it is postulated that due to glacier melting the pore pressure in the crust beneath the glacier is higher in the fall than in the spring. The pore pressure then triggers slip on faults in the stressed crust above the magma chamber. The stress is supplied by changes in the magma pressure, and the pore pressure acts as a trigger by lowering the friction on faults.

The Torfajökull central volcano is moderately seismically active. It is characterized by extensive rhyolitic volcanism, a large caldera or a circular structure, and vigorous geothermal activity. No eruptions have occurred in the last few hundred years. The most recent eruptions and most of the geothermal activity are concentrated in the western part of the caldera. This is also where most of the present earthquake activity is located (Fig. 8). This spatial correlation and the persistent nature of the seismicity suggests that the earthquakes may be related to heat mining and thermal cracking, as was suggested for the Hengill seismicity.

Small, low-frequency earthquakes were discovered near Torfajökull after a seismograph was set up in the area in 1985. Four prolonged swarms of these events were recorded in 1986 (Brandsdóttir and Einarsson, 1989). This activity could not be accurately located, but is thought to reflect minor magmatic movements in the roots of the volcano.

The Vestmannaeyjar volcanic system went through a period of unrest, with the Surtsey eruption in 1964–1967 and the Heimaey eruption in 1973. The area is still seismically active, but the activity diminishes year by year. Most of the events in recent years were located near Surtsey.

The Heimaey eruption was preceded by an intense swarm of small earthquakes that started to develop 30 hours before the eruption. Earthquakes also accompanied the eruption, but seismicity declined as the magma emission rate diminished. The earthquakes during the eruption originated at a 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.

It is considered possible that the Surtsey and Heimaey eruptions led to pressure drop in the partially molten mantle below and subsequent deflation of Katla. Pulses of seismicity occurred at Katla in 1967 and 1977, four years after the beginning of the eruptions. This is a much longer response time than in the case of Krafla and Bárðarbunga. A longer response is not unreasonable in this flank zone, where a greater depth to the magma layer and a lower degree of partial melting is expected.
Western Volcanic Zone

The Western Volcanic Zone is presumably a dying rift zone giving way to the Eastern Volcanic Zone which becomes the main spreading center in South Iceland. This hypothesis is supported by the observations that historical fissure eruptions have been limited to the Eastern Zone and that transform faulting occurs near 64° N and not in Central Iceland. The Hreppar crustal block between the two rift zones therefore seems to be moving with the North American plate. In spite of the clear decline in volcanic productivity of the Western Volcanic Zone in the last few thousand years, there has been considerable rifting (Saemundsson, 1986), leading to graben subsidence which is most clearly seen in the Thingvellir graben in the southern part of the zone.

The idea of a dying Western Zone does not find much support in the seismicity of recent years, however (Fig. 2). The zone is moderately seismically active, and, at least on the time scale of the last decades, much more active than the Eastern Rift Zone. The seismicity is characterized by swarms. The clusters on the epicentral map of Fig. 2 are all due to single swarms. The largest occurred in June–July 1985, and had a maximum magnitude of $m_b = 4.3$.

Intraplate earthquakes

Only a small proportion of Icelandic earthquakes can be termed truly intraplate, i.e. not directly related to the plate boundary or the volcanic zones. Two classes of intraplate earthquakes can be identified (Einarsson, 1989). In the

![Fig. 9. Intraplate earthquakes (1930–1987), plate boundaries and volcanic flank zones in the Iceland area (from Einarsson, 1989). Epicenters of swarms and single events are shown with dots. Bathymetric contours are drawn every 400 m.](image-url)
first are earthquakes occurring west of the main rift zone through Iceland, within the tongue of lithosphere that is bounded by the two transform zones in North and South Iceland. This seismicity is associated with internal deformation of the North American plate. Best known are the Borgarfjördur events of 1974 in central West Iceland (Einarsson et al., 1977), which comprise a prolonged series of earthquakes with a maximum magnitude of 6. The epicentral area was at least 25 km long and the earthquakes were clearly not associated with displacement on one particular fault. Most of the hypocenters were at a depth of 0–8 km, but events did occur down to the 10 km depth. Thus, fracturing extended down through most of the crust. Fault plane solutions show normal faulting, indicating horizontal extension. This is also consistent with surface faulting observed during the earthquakes.

The areal extent of the lithospheric tongue may be seen in the observation that the transform zones bounding it in the north and south are not parallel, but diverge towards the west. This phenomenon is most likely related to the excess production of the Iceland hot spot, although the exact mechanism is unclear. The extension could be achieved by underplating and intrusion of magma into the crust from the partially molten layer below.

The second major class of Icelandic intraplate earthquakes consists of events originating on the insular shelf off Eastern and Southeastern Iceland (Fig. 9). Most of them have epicenters very close to the shelf edge, which is thus shown to be a seismogenic structure. The shelf edge in this area probably represents a structural discontinuity formed after about 36 Ma, when the main discharge area of the Iceland hot spot shifted from the Iceland–Faeroe Ridge to produce the present Icelandic basement. The discontinuity is expressed in surface topography as well as in crustal and mantle structure. There is probably also an age jump of 15–25 Ma and a resulting thermal discontinuity across the shelf edge. The present seismicity is thus interpreted as the result of differential lithospheric response to loading or a different cooling rate on either side of the shelf edge.

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