Reviewed research article

Seismic characteristics of the Hekla volcano and its surroundings, Iceland

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Abstract — *The volcano Hekla is located in south Iceland at the junction of a transform segment, the South Iceland seismic zone, and a ridge segment, the Eastern volcanic zone, of the mid-Atlantic plate boundary. Hekla is one of the most active volcanoes in Iceland, with at least 18 eruptions during the last 1100 years. In recent decades it has had relatively small eruptions, approximately once in a decade, most recently in 1991 and 2000. During non-eruptive periods Hekla is virtually aseismic, and does not give long-term or intermediate-term precursory warnings before its eruptions. Eruption-related seismicity starts 25–80 minutes before its onset. Hundreds of small volcano-tectonic earthquakes (magnitude < 3), related to magma intrusion, occur during the first hours when the eruption is violent and explosive. This seismicity soon diminishes, along with the eruptive activity. Subsequent eruptive activity consists mainly of lava effusion and occasional gas bursts associated with very few earthquakes. Volcanic tremor, continuous low-frequency vibration of the ground, starts simultaneously with the eruption, and continues throughout it. It is most vigorous during the explosive onset, and decreases along with the eruptive activity. The few earthquakes at Hekla and its immediate vicinity during non-eruptive times are small (magnitude < 2), and apparently not related to the Hekla volcano itself. They follow a distribution similar to the events of the South Iceland seismic zone. They form two north-south lineaments analogous to the seismic zone faults, and occur mainly at depths of 8–12 km. Thus seismically the Hekla area has a dual nature: on one hand the seismicity is ruled by the tectonics of the South Iceland seismic zone, and on the other hand by the internal processes of the volcano. Transform zone tectonics dominate during the non-eruptive periods. The volcano-related seismicity of Hekla is almost exclusively associated with eruptions. Volcanic tremor has never been recorded during non-eruptive periods.*

INTRODUCTION

The two most recent eruptions of Hekla occurred in January 17–March 11, 1991 and February 26–March 8, 2000. These events produced a rather similar amount of eruptives of basaltic-andesitic composition, 0.15 km^3 (Guðmundsson *et al*. 1992) and ~0.2 km^3 (Höskuldsson *et al*. submitted), respectively. Both eruptions had a short-lived initial Plinian phase and were most vigorous in the first hours. Effusion of lava began at the same time as the explosive activity, or shortly after. Initially, large segments of fissures were active, but the eruptions became more localised during the later phases. In both cases, lava production was largest during the first days of activity. Seismically, the eruptions were very similar. No long-term precursory seismicity was detected. The onset of each eruption was accompanied by an initial swarm of hundreds of small volcano-tectonic earthquakes (M_L < 3) which increased in magnitude towards the onset of the eruption, few earthquakes in the later phases and continuous low-frequency volcanic tremor with dominant frequencies at 0.7–0.9 Hz (Soosalu and Einarsson 2002; Soosalu *et al*. 2003, 2005).

The 1991 and 2000 Hekla eruptions are the first for which digital seismic data exist, facilitating quantitative research on patterns of the eruption-related seismicity. Eruptions in 1970 and 1980–1981 were similar in size and behaviour (Einarsson and Björnsson 1976; Grönvold *et al.* 1983; Brandsdóttir and Einarsson 1992). Qualitative research based on analogue seismograms conducted on these eruptions point to very similar seismic characteristics. In order to make a comparison of seismicity during eruptive and non-eruptive times we conducted a study of the background seismicity at Hekla and its vicinity during a non-eruptive period in 1991-1996 (Soosalu and Einarsson 1997). Together with Hekla, the study area (63°42'–64°18'N and 18°30'–20°12'W) covered the neighbouring volcanoes, Torfajökull to the east and Vatnafjöll to the south, and the eastern end of the South Iceland seismic zone, a transform zone of the mid-Atlantic plate boundary (Figure 1).

In this paper we draw general conclusions about the seismic nature of the Hekla volcano and its immediate surroundings, based on digital and analogue data from June 1990 until mid-August 2005. We define the characteristics related to the relatively small Hekla eruptions that have been occurring during recent decades. We also describe the nature of seismicity at Hekla during non-eruptive times, which appears to be unrelated to Hekla as a volcano but rather follows the pattern of the transform zone to the west.

GEOLOGICAL SETTING OF THE HEKLA AREA

The central volcano Hekla in south Iceland is a ridge elongated in the ENE-WSW direction, formed by repeated eruptions and reaching an altitude of 1500 m above sea level. Recent eruptions have had a tendency to take place along a radial fracture system, as well as along the main Hekla fissure that splits the volcano lengthwise. Hekla's fissure swarm extends NE and SW of the summit. The central volcanoes next to Hekla, also discussed in this study, are Torfajökull in the east and Vatnafjöll in the south (Figure 2).

Hekla is located in a tectonically complex area, at the junction between the transform-like South Iceland seismic zone and the Eastern volcanic zone, the east branch of the chain of active volcanic systems crossing the middle of Iceland (Figure 1). North of Hekla and Torfajökull, the volcanic zone is characterized by rifting activity, whereas to the south it has the nature of a non-rifting flank zone. At the location of Torfajökull, rifting is propagating to the southwest (Óskarsson *et al.* 1982).

Hekla is not a typical rift zone volcano, due to its tectonic setting and peculiar petrology. The products of the Hekla volcanic system range from basalts through basaltic andesites to dacites and rhyolites (Jakobsson 1979). The more acidic products are issued from the volcanic edifice, while the basaltic products come from the fissure swarm. Petrologically, Hekla is more akin to the group of volcanoes in the volcanic flank zone to the south-east.

Hekla is one of the most active volcanoes in Iceland and has erupted at least 18 times since Iceland was colonized in the ninth century (Guðmundsson *et al.* 1992). Since the major eruption of 1104 AD until the 1947–1948 eruption, the activity was characterized by relatively large eruptions about twice a century (Þórarinsson 1967). Within the last decades Hekla has changed its eruptive pattern. Smaller eruptions with volumes of about 0.2–0.3 km^3 have occurred about every ten years, in 1970, 1980–1981, 1991, and 2000.

Torfajökull is a major rhyolitic complex with a 12-km-diameter caldera (Sæmundsson 1972, 1982), an outstanding high-temperature geothermal field (McGarvie 1984), and fissure swarms stretching both NE and SW of the central volcano. The latest eruption in the Torfajökull area occurred at the end of the fifteenth century (Larsen 1984). The Vatnafjöll central volcano, south of Hekla, does not have any caldera or geothermal areas. The fissure swarm of Vatnafjöll is elongated in the NE-SW direction, parallel to the fissure swarm of Hekla. No eruptions are known to have occurred in Vatnafjöll during the last 1100 years (Bjarnason and Einarsson 1991).

The South Iceland seismic zone is a 70–80 km long and 10–15 km wide zone in the South Iceland lowland. It acts as an E-W transform, but is characterized by abundant seismicity on N-S right-lateral

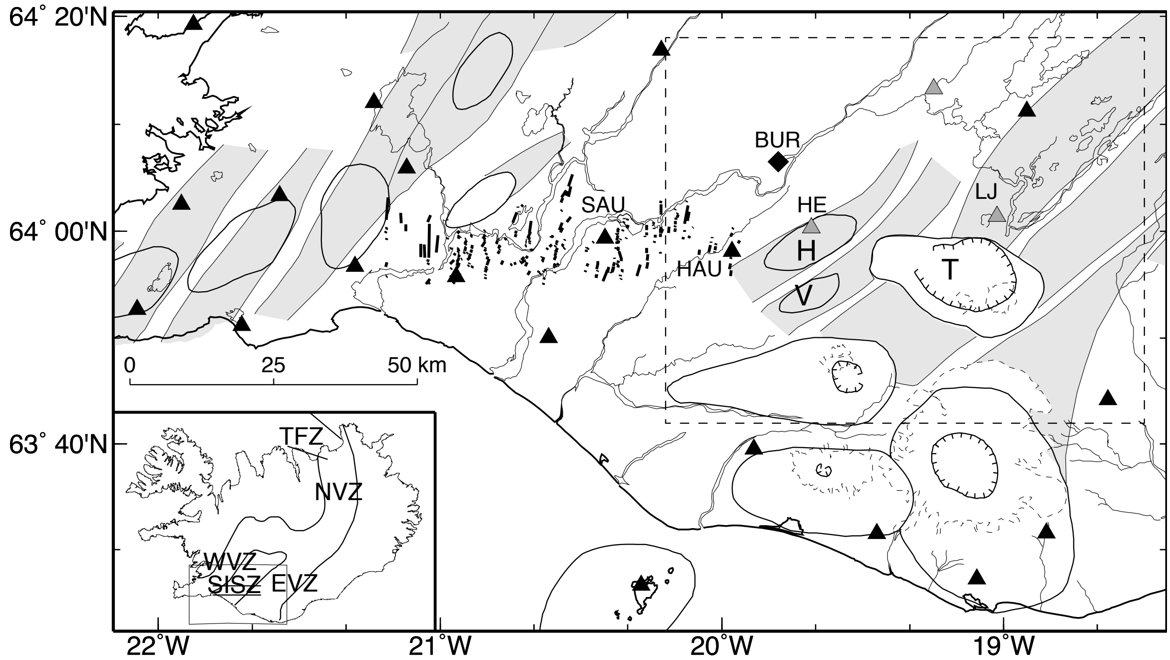

Figure 1. Index map of the Hekla area and the seismograph network. Black triangles are the digital SIL seismograph stations and grey triangles the analogue stations, those mentioned in the text are named. The black diamond represents the strain station BUR. The central volcanoes are outlined, their fissure swarms are shaded grey (Einarsson and Sæmundsson 1987) and the calderas are hatched (Jóhannesson *et al*. 1990). Named central volcanoes are: H = Hekla, V = Vatnafjöll, T = Torfajökull. Thick black lines are the faults of the South Iceland seismic zone. Short dashed lines mark the glaciers. The box with dashed outline shows the study area. Smaller index map shows the locations of the Western (WVZ), Eastern (EVZ) and Northern (NVZ) volcanic zones, the South Iceland seismic zone (SISZ), and the Tjörnes fracture zone (TFZ). All the figures are drawn using the Generic Mapping Tools program (Wessel and Smith 1998). – *Yfirlitskort af Suðurlandi og skjálftamælanetinu. Svartir þríhyrningar tákna mæla í stafræna landsnetinu, gráir þríhyrningar eru hliðrænir mælar, sem skrá með penna á pappír. Svartur tígull táknar þenslumælinn við Búrfell, BUR. Megineldstöðvar, sprungusveimar og öskjur eru sýndar. Megineldstöðvarnar Hekla, Vatnafjöll og Torfajökull eru merktar með H, V og T. Þykkar, svartar línur eru sniðgengi á skjálftabelti Suðurlands. Jöklar eru sýndir með strikalínum og strikalínukassi afmarkar rannsóknarsvæðið, sem sýnt er nánar á 2. mynd.*

strike-slip faults, and events up to magnitude 7.1 have occurred (Einarsson 1991; Stefánsson *et al*. 1993). The earthquakes have a tendency to deepen towards east: in the westernmost part of the zone they are typically located in the uppermost 6 km and in east at 6–12 km depth. The east end of the seismic zone, about 15 km in width, is included in this study of seismicity in the Hekla area.

THE SEISMIC DATASET

This study is based principally on seismic data gathered by the digital seismic stations of the SIL (South Iceland Lowland) network (Figure 1), maintained by the Icelandic Meteorological Office. The SIL network has gathered data since 1990 (Stefánsson *et al*. 1993). Originally the network consisted of eight digi-

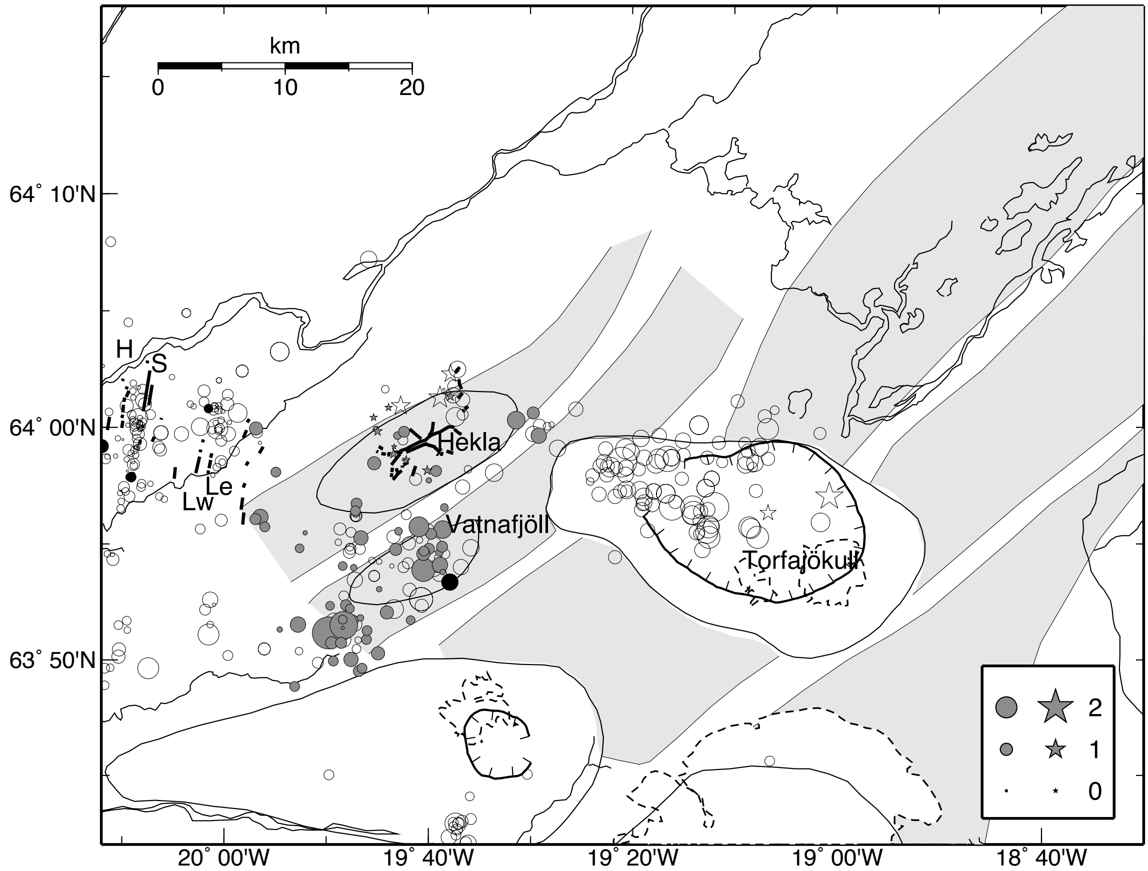

Figure 2. The well-located earthquakes in the Hekla area at non-eruptive times in the period June 1990 – mid-August 2005. Seismicity in the whole study area has been examined until end-October 1995 and in the smaller Hekla-Vatnafjöll area (63°48'–64°05'N, 19°25'–19°56'W) until mid-August 2005. Black symbols are earthquakes that occurred before the 1991 Hekla eruption, open symbols events that occurred between the 1991 and 2000 eruptions, and grey symbols after the 2000 eruption. Dots denote ordinary high-frequency earthquakes and stars low-frequency earthquakes. The inset shows the sizes of events of corresponding local magnitudes (M_L). Fissures of the 1970, 1980–1981, 1991 and 2000 Hekla eruptions are shown. H-Hellar fault, S-Skarðsfjall fault, Lw-western Leirubakki fault, Le-eastern Leirubakki fault. – *Upptök jarðskjálfta á Heklusvæðinu á tímabilum þegar ekki er gosvirkni í eldstöðinni. Sýndir eru einungis vel staðsettir skjálftar. Á svæðinu umhverfis Heklu og Vatnafjöll (63°48'–64°05'N og 19°25'–19°56'V) eru sýndir skjálftar á tímabilinu frá júní 1990 og fram í miðjan ágúst 2005. Utan þessa svæðis eru einungis sýndir skjálftar fram í október 1995. Svört tákn sýna skjálfta fyrir gosið 1991, ófyllt tákn sýna skjálfta milli gosanna 1991 og 2000, og grá tákn sýna skjálfta eftir gosið 2000. Venjulegir hátíðniskjálftar eru sýndir með hringlaga táknum, stjörnur sýna lágtíðniskjálfta. Stærð táknanna fer eftir stærð skjálftanna samkvæmt M_L-kvarða. Gossprungur sem voru virkar í Heklugosunum 1970, 1980–1981, 1991 og 2000 eru sýndar. Einnig eru sýnd sniðgengi við Hella (H), Skarðsfjall (S), og Leirubakka (Lw og Le).*

tal three-component stations in the south Iceland lowland area, and more recently it has been expanded to other parts of Iceland (Böðvarsson *et al.* 1999). Because the study area is situated at the edge of the original SIL network, three permanent analogue vertical-component stations in the vicinity of Hekla were used in addition to get good station coverage for the whole area (Figure 1). The events were relocated with the location program HYPOINVERSE (Klein 1978), using a crustal model consisting of layers with constant velocity gradients (Gebrande *et al.* 1980). The model is an average model with time delay corrections used for the seismograph stations to improve the location accuracy. The location procedure is described in detail in Soosalu and Einarsson (1997).

The dataset used spans the period from June 1990 to mid-August 2005 and covers the two latest Hekla eruptions. Over one-thousand small earthquakes (M_L < 3) were detected at Hekla and its surroundings during this period. Of these events, we plot in the subsequent maps only the well-located ones, using the location criteria: root mean square travel-time residual (rms) $\leq$ 0.2 s, horizontal error (erh) $\leq$ 1.0 km, vertical error (erz) $\leq$ 2.0 km, and largest gap between observing stations $\leq 180°$.

SEISMICITY IN THE HEKLA AREA BETWEEN ERUPTIONS

During non-eruptive times Hekla is characterised by scarce seismicity (Einarsson 1991; Soosalu and Einarsson 1997, 2002; Soosalu *et al.* 2005). The earthquakes are sporadic and small in size, typically $M_L \leq 1$. As an illustration of the seismicity at Hekla and its vicinity, earthquakes in the area during non-eruptive times, June 1990–January 1991, March 1991–February 2000 and March 2000–mid-August 2005 are plotted in Figure 2. Their hypocentral distribution is shown in Figure 3. Inter-eruption Hekla seismicity is not volcano-related, but reflects the transform tectonics of the South Iceland seismic zone together with the earthquake activity at the Vatnafjöll volcano south of Hekla. The earthquakes at Hekla and Vatnafjöll are loosely clustered in two N-S lineaments, similar to the faults of the South Iceland seismic zone. One of the lineaments cuts the SW parts of the volcanoes and the other the middle-NE parts of them. The earthquakes occur typically at 8–13 km depth, similar to earthquakes at the east end of the South Iceland seismic zone. Three events in the NE part of Hekla were located at considerable depths of 16, 18 and 26 km. For all these events the nearest station was within the range of 17 to 18 km.

During non-eruptive periods, the surroundings of Hekla are far more seismically active than the volcano itself. Most intensive seismicity occurs at the Torfajökull volcano; e.g. during the period 1990–1995 about 200 events, up to local magnitude 2.8 were located in the Torfajökull area (Soosalu and Einarsson 1997). Torfajökull is characterised by dual seismicity: high-frequency volcano-tectonic earthquakes occur in the western part of its caldera and low-frequency earthquakes in the south (Soosalu and Einarsson 2003).

Seismicity in the easternmost section of the South Iceland seismic zone, next to Hekla, is clustered in two elongate N-S lineaments, roughly at longitudes 20°W and 20°10'W, both during eruptive and non-eruptive periods for Hekla (Soosalu and Einarsson 1997, 2002). Small earthquakes, up to local magnitude 2, are common in this area, and over 190 were observed in the period 1990–1995. They occurred typically in the depth range of 6–12 km. The events were located between latitudes 64°08'N and 63°49'N, but were most abundant around latitude 64°N, where surface faults can be seen.

From the start of the study period, June 1990 to half an hour before the start of the 1991 eruption, on January 17, no earthquakes were located at Hekla or its immediate vicinity (Soosalu and Einarsson 2002). Thus, no long-term or intermediate term precursors were observed prior to the eruption. Earthquakes in this period were confined to the Torfajökull volcano and South Iceland seismic zone.

Minor seismicity was observed within the Hekla-Vatnafjöll area following the eruption in January 17–March 11, 1991 until June 1, 1991 when an unusual swarm of earthquakes occurred beneath the northern flank of Hekla, most at less than 3 km depth. The SIL network detected about thirty events with magni-

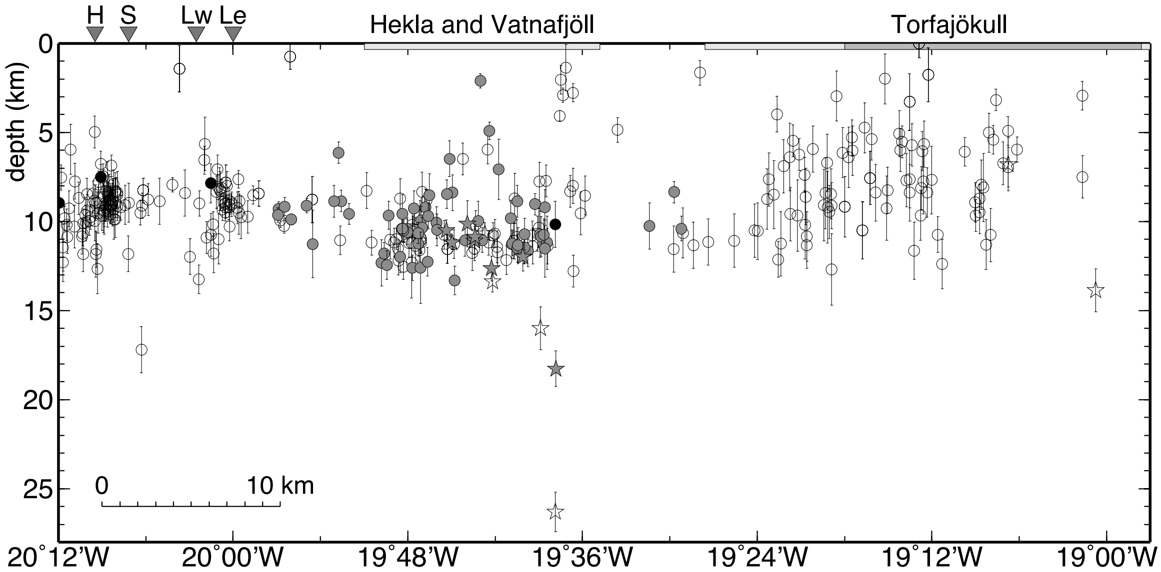

Figure 3. Vertical cross section of the study area from South Iceland seismic zone to east edge of the Torfajökull caldera (between latitudes 63°48.6'–64°05'N), seen from south and without vertical exaggeration. Plotted are all the well-located earthquakes at these latitudes. Black symbols are earthquakes that occurred before the 1991 Hekla eruption, open symbols events that occurred between the 1991 and 2000 eruptions, and grey symbols after the 2000 eruption. Dots denote ordinary high-frequency earthquakes and stars low-frequency earthquakes. Inverted triangles show the surface locations of the South Iceland seismic zone faults: H-Hellar fault, S-Skarðsfjall fault, Lw-western Leirubakki fault, Le-eastern Leirubakki fault. The locations of the central volcanoes are shown with grey bars, the location of the Torfajökull caldera with darker grey shade. – *Lóðrétt, A-V snið í gegnum skjálftabeltið á Suðurlandi, Heklu og Torfajökulseldstöðina. Horft er úr suðri. Lóðréttur og láréttur kvarði er hinn sami. Teiknuð eru öll skjálftaupptök milli 63°48.6'og 64°05'N, sama gagnasafn og sömu tákn og á 2. mynd. Þríhyrningar á yfirborðinu sýna staðsetningar misgengja við Hella (H), Skarðsfjall (S) og Leirubakka (Lw og Le). Gráar rendur við yfirborðið sýna staðsetningu megineldstöðvanna, Heklu, Vatnafjalla og Torfajökuls. Torfajökulsaskjan er sýnd með dekkri rönd.*

tudes of 0.6–1.7 during about ten hours, and the analogue station HE at Hekla recorded over seventy additional small events. The analogue stations HE and LJ also recorded three volcanic-looking events within the swarm, similar in appearance to hybrid events (Chouet 1996), with a high-frequency onset followed by a low-frequency coda. No volcanic tremor was detected during this earthquake swarm. After June 1991, Hekla was seismically quiet for more than a year. Later, sporadic small earthquakes started to occur, and their local magnitudes were typically under 1.

A distinct feature of Hekla seismicity during non-eruptive times is the low-frequency character of the events (Figure 4a). The earthquakes occurring at Hekla proper have a peculiar appearance: they consist of only low frequencies (main frequency content being below 5 Hz) but have clear S-waves, similar to tectonic earthquakes. In contrast to this, the earthquakes related to the eruptions are ordinary-looking high-frequency events (Figure 4b). High-frequency earthquakes may continue to occur for a few months after eruptive activity, but the few inter-eruption Hekla earthquakes have a low-frequency appearance. This phenomenon has been observed after both the 1991

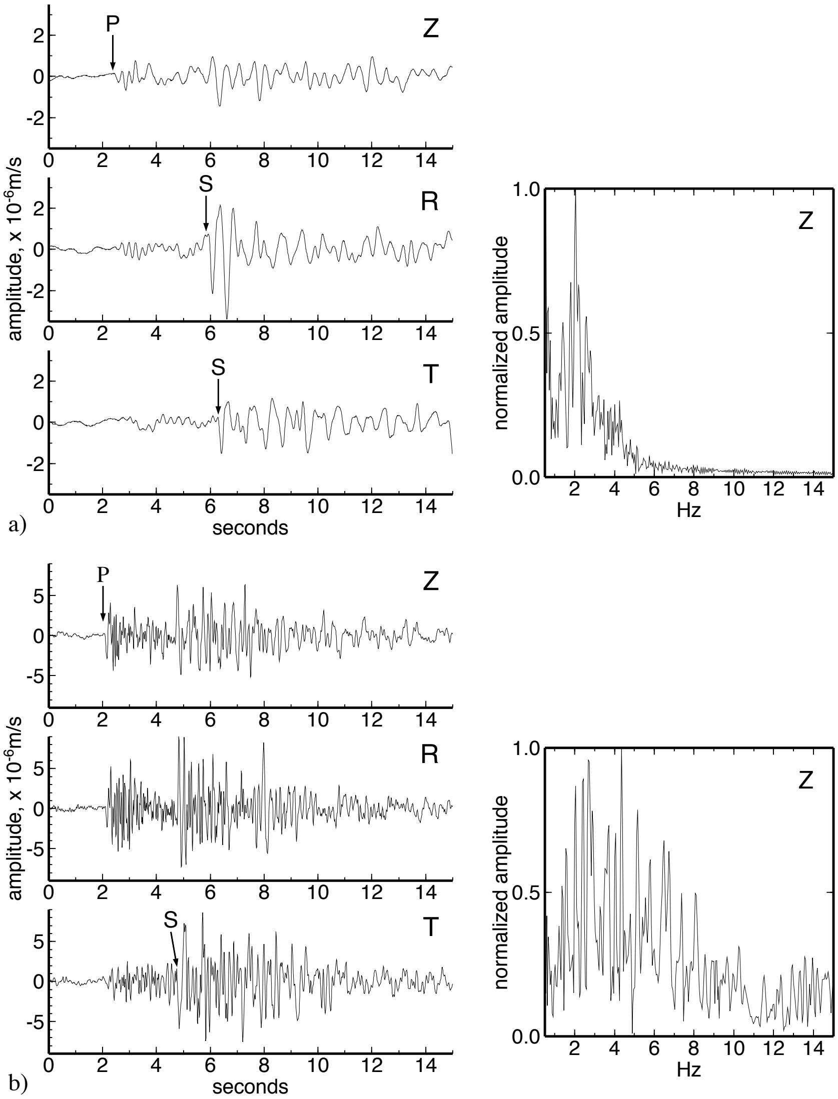

Figure 4. An example of a) a low-frequency Hekla earthquake with its spectrum (March 27, 1999, M_L 1.2, at 16 km depth). S-wave splitting can be discerned in this sample, a feature not further studied in this paper. b) For comparison a high-frequency Hekla earthquake is shown (February 26, 2000, 17:45, 34 minutes before the onset of the eruption, M_L 1.2, at 6 km depth). Both records are from station HAU, Z–vertical, R–radial, T–transverse component. The data are 0.5 Hz high-pass filtered because of microseismic noise. The spectra (vertical component) of both events are taken over the first 10 seconds. The plots are clipped in the low end at 0.5 Hz (modified from Soosalu *et al.* 2005). – *Sýnishorn skjálftalínurita frá mælistöðinni í Haukadal. a) Lágtíðniskjálfti með upptök undir Heklu, ásamt tíðnirófi (27. mars, 1999, stærð M_L = 1.2). b) Hátíðniskjálfti með upptök undir Heklu til samanburðar (26. febrúar, 2000, kl. 17:45, 34 mínútum áður en gos hófst, stærð M_L = 1.2). Sýndir eru þrír þættir hreyfingarinnar, Z lóðréttur, R stefna frá upptökum, T stefna þvert á upptök. Gögnin eru háhleypisíuð við 0.5 Hz til að losna við jarðóróa. Tíðniróf eru af lóðrétta þættinum og ná yfir fyrstu 10 sekúndur skjálftaritsins. Tíðnir fyrir neðan 0.5 Hz eru ekki sýndar.*

(Soosalu and Einarsson 2002) and the 2000 eruptions (Soosalu *et al.* 2005). After the 1991 eruption, the last low-frequency event was detected in March 1999. High-frequency earthquakes started to occur at Hekla proper at the end of the decade: three of them were observed in February 1998 and one in July 1999. After then, no earthquakes of any sort were detected at Hekla itself before the 2000 eruption, which started on February 26.

Since the 2000 eruption, Hekla seismicity has been modest. Until mid-August 2005 nineteen low-frequency Hekla events with M_L magnitudes between 0.3 and 1.6 have been detected. A clearly high-frequency Hekla earthquake (M_L 0.9) was observed at 5 km depth in September 2004, and another one (M_L 0.4) at 6 km depth in March 2005.

SEISMIC BEHAVIOUR OF HEKLA DURING ITS ERUPTIONS

Hekla is a notorious volcano because of the short warning time before its eruptions. Seismicity related to eruptions starts gradually and reaches the detection threshold very shortly before the onset; approximately 25 minutes in the eruptions in 1970 (Einarsson and Björnsson 1976) and in 1980 (Grönvold *et al.* 1983), half an hour in 1991 (Guðmundsson *et al.* 1992; Soosalu and Einarsson 2002) and 79 minutes in 2000 (Einarsson 2000; Soosalu *et al.* 2005).

Based on knowledge gained from seismicity and strain observations in 1991 (Linde *et al.* 1993), the 2000 Hekla eruption was successfully predicted. In 1991 the initial swarm of small earthquakes was accompanied by a compressive strain signal observed at a strain station 15 km from Hekla (BUR in Figure 1), interpreted as a result of a feeder dyke propagating at depth towards the surface. The same pattern was observed in 2000, which led to a successful short-term prediction for the eruption, some 50 minutes before its onset (Ágústsson *et al.* 2000; Stefánsson *et al.* 2000).

Two expressions of seismicity are observed during the Hekla eruptions: volcano-tectonic earthquakes (Figure 5a) and low-frequency volcanic tremor (Figure 5b). The majority of the earthquakes occur around the onset of the eruption. During later phases of the eruption, only occasional earthquakes are observed. The tremor starts simultaneously with the eruptive activity, continues throughout it and fades away together with it. No tremor has ever been observed at Hekla during non-eruptive periods. The tremor is most vigorous during the first hours of eruptions which are characterized by explosive activity, and declines later.

The initial earthquake swarm

The initial earthquake swarms in 1991 and 2000 were very similar (Figure 6). The very first observed events were tiny, of about magnitude 0. The size of the events grew towards the onset of the eruption and culminated around it. In 1991 the detection threshold was higher than in 2000, as the analogue station, HE, on the flank of the volcano was broken and the closest digital station, HAU, was down. Both of these stations operated well in 2000. The seismicity in 1991 quite likely started in a similar manner as in 2000 but reached the detection threshold later, first 30 minutes before the onset of the eruption, compared to the 79 minutes of 2000. Both in 1991 and 2000 a strain signal of contraction, indicating propagation of an intrusion from depth, was observed at the close strain station about half an hour before the eruption reached the surface (Linde *et al.* 1993; Ágústsson *et al.* 2000). The earthquake swarm started some 50 minutes before the strain station observed any deformation. It is possible that small strain changes occurred but went undetected because of the large distance (15 km) to the nearest strain station (Alan Linde, *pers. comm.* 2004).

The initial earthquake swarm soon became very intense, and events were observed with intervals of less than one minute. The events were relatively small; in 1991 the maximum magnitude was M_L 2.5 and in 2000 M_L 2.1. In 1991 the total seismic energy of the initial earthquakes corresponded to a single event of M_L 3.4 and in 2000 to an event of M_L 3.2.

In total, 380 earthquakes were detected during the initial swarm in 1991, some 60 of which occurred before the onset of eruptive activity. In total, 340 events were observed in 2000, 208 of which preceded the start of the eruption. Figure 7 shows the epicentral map of well-located earthquakes on February 26, 2000. In the first few hours the earthquakes were rather even in magnitude, ≥ 2 in 1991 and ≤ 2 in

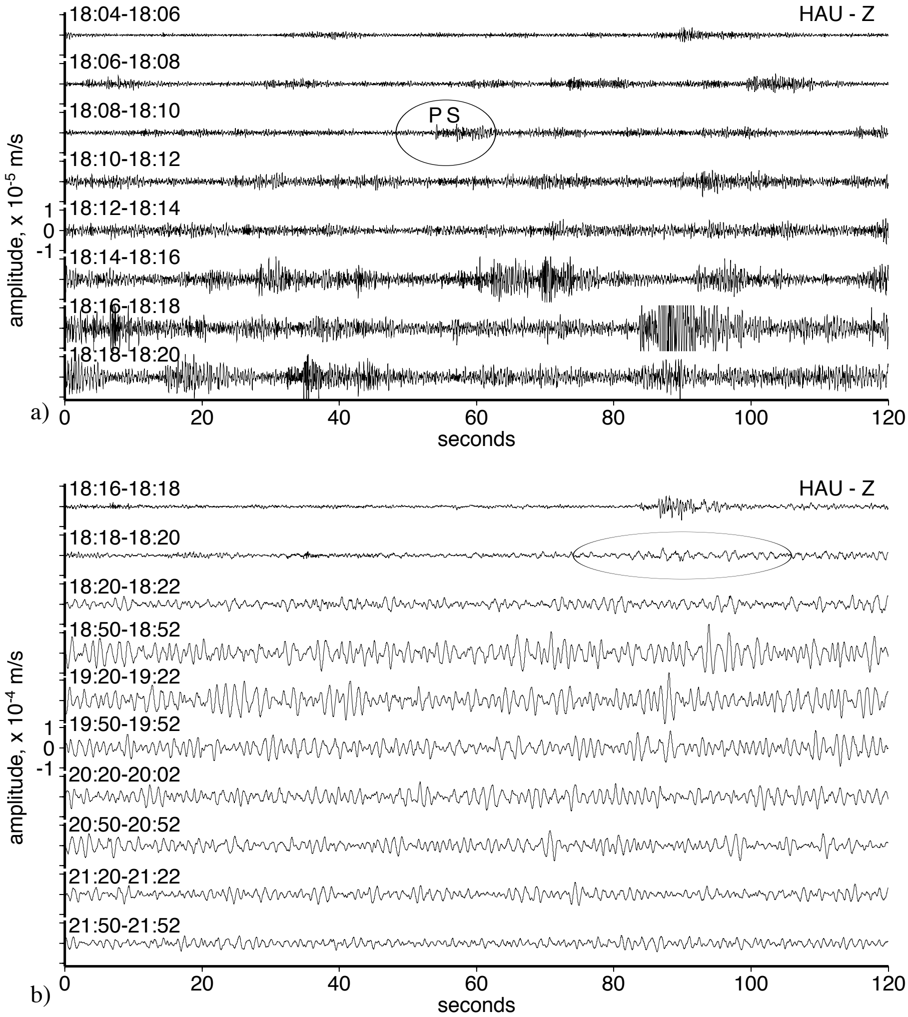

Figure 5. a) Vertical component seismograms at the station HAU until the onset of the 2000 Hekla eruption at 18:19 GMT. At first sporadic small earthquakes occur (one event is circled as an example, P and S arrivals are shown). Towards the start of the eruption they become larger and more frequent, and finally merge into a continuous-looking signal. The data are high-pass filtered at 2 Hz because of microseismic noise. The M_L 2.1 earthquake at 18:17 is clipped because of illustration purposes. b) Vertical component seismogram samples at HAU at the onset of the 2000 eruption. They are unfiltered and their amplitude is at a 10-times smaller scale than in Figure 5a. The first two traces at 18:16–18:20 overlap with those in Figure 5a. The low-frequency tremor amplitude increases quickly, stays high for 1.5–2 hours, and then clearly starts to decline. The circle indicates the time when the eruption started according to the eyewitnesses. – *a) Skjálftarit Haukadalsmælisins (lóðréttur þáttur) 26. febrúar 2000, frá kl. 17:56 og þangað til gos hefst kl. 18:19. Skjálftar eru strjálir til að byrja með, einn þeirra er merktur með hring, P- og S-bylgjur eru greinilegar. Skjálftarnir stækka og verða tíðari þega nær dregur gosbyrjun. Að lokum renna þeir saman í samfellda hreyfingu. Stærsti skjálftinn, M_L = 2.1, varð kl. 18:17. Gögnin eru háhleypisíuð við 2 Hz. b) Skjálftarit Haukadalsmælisins af lóðrétta þætti hreyfingarinnar við byrjun gossins. Sjónarvottar staðfesta að gosið byrjaði kl. 18:19 en þá vex útslag óróans hratt. Útslagið er mikið í 1,5–2 klukkustundir en síðan dregur úr því. Takið eftir að mælikvarðinn er 1/10 af því sem er í skjálftaritinu í a). Engum síum er beitt.*

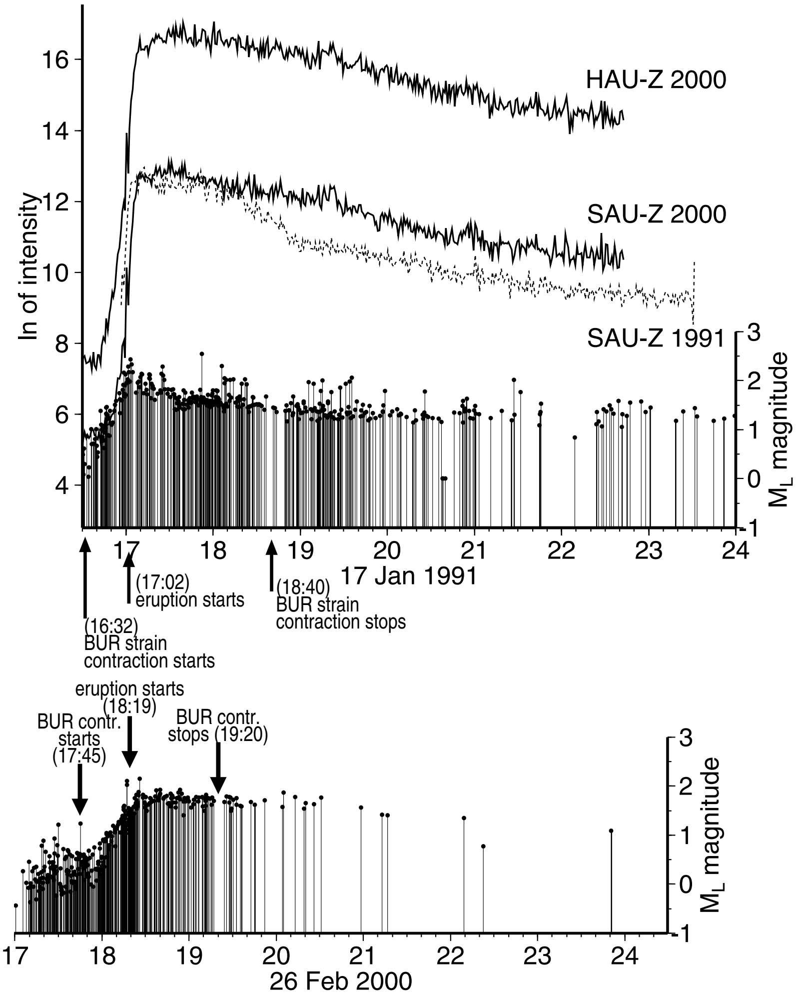

Figure 6. Temporal distribution of the earthquakes around the onset of the 1991 and 2000 eruptions with their M_L magnitudes (scale to the right), the onset times are aligned at the same line. The magnitude scale is on the right. Changes in strain are also marked. Tremor intensity curves, for SAU in 1991 (dashed line), and HAU and SAU in 2000 (bold lines) are shown, as well (scale to the left). Only the vertical components (Z) are shown. The horizontal components had similar, slightly higher values. High peak in the very end of the SAU 1991 curve is an artefact. The frequency band used is 0.5–3.0 Hz, to damp the effect of the low-frequency microseism of oceanic origin and the high-frequency earthquakes. The intensity is calculated as averaged energy over 60-second intervals, and its natural logarithm is plotted. – *Skjálftavirkni í tengslum við gosbyrjanir 1991 og 2000. Stærð skjálfta (M_L) er sýnd sem fall af tíma fyrir bæði gosin. Tímakvarðinn er sá sami fyrir bæði gosin og upphafstímar gosanna eru látnir standast á. Styrkur óróans er einnig sýndur sem fall af tíma eins og hann kemur fram á mælum í Saurbæ (SAU) og Haukadal (HAU). Einungis lóðrétti þátturinn er sýndur. Óróagögnin eru bandhleypisíuð á bilinu 0.5–3.0 Hz til að útiloka truflanir frá úthafsöldum og hátíðniskjálftum. Styrkurinn er reiknaður sem 60 sekúndna meðaltal af orku.*

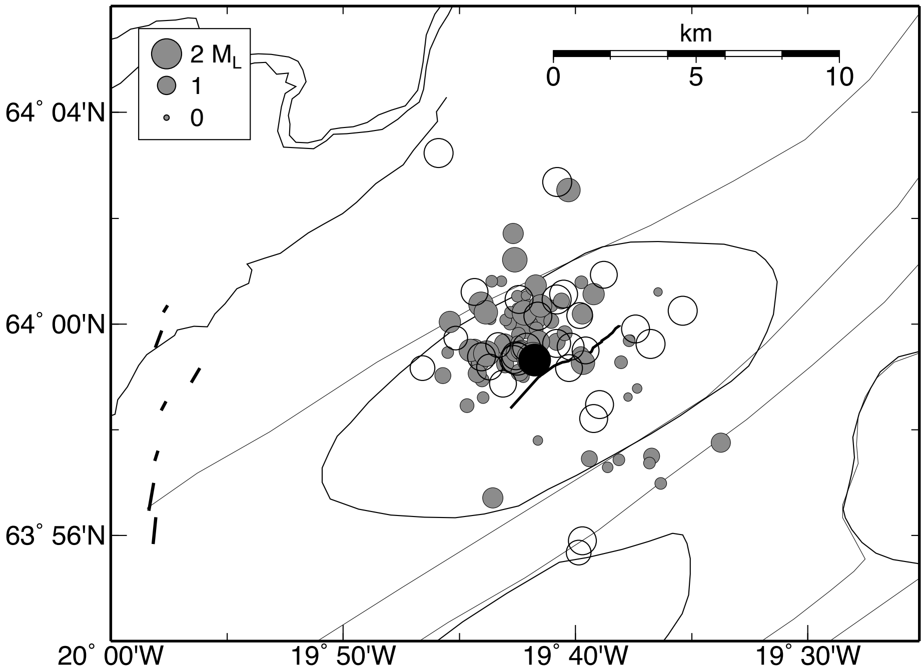

Figure 7. Earthquakes during the onset phase of the Hekla 2000 eruption, open circles show the earthquakes before the start of the eruption at 18:19 GMT and grey dots the events after. The black dot is a magnitude-2.1 earthquake at 3 km depth at 18:17. The magnitude scale is given in the inset. The fissure of the 2000 eruption is shown with a thick line. Black bars at about 20°W are the easternmost faults of the South Iceland seismic zone. – *Upptök jarðskjálfta undir Heklu sem urðu í tengslum við eldsuppkomuna 2000. Hringir tákna skjálfta sem urðu áður en gosið kom upp klukkan 18:19, gráir deplar skjálfta eftir að gosið kom upp. Svarti depillinn sýnir upptök stærsta skjálftans (M_L = 2.1), en hann varð á 3 km dýpi klukkan 18:17, rétt áður en gosið sást. Stærð táknanna sýnir stærð skjálftanna. Gossprungan sem var virk 2000 er sýnd með feitri línu. Einnig eru sýndar sprungur sem voru virkar í jarðskjálftanum 1912, austast á skjálftabelti Suðurlands.*

2000, showing a slight decline with time. In Figure 6 small events ($M_L < 1$) are not seen after the start of the eruptions, due to the masking effect of the high-amplitude volcanic tremor. After some hours the earthquakes become sporadic and soon stop occurring altogether. Neither in 1991 nor in 2000 were earthquakes observed on the second day of the eruption, and few events were observed during the later phases in general.

The location accuracy of events on January 17, 1991 was not good because of insufficient station coverage. The events of February 26, 2000 were far better observed and the depths of many earthquakes could be constrained rather well (Soosalu *et al*. 2005). All the first events, from 17:00 GMT on, were very small and those that could be located were shallow, at 0–4 km depth. After 17:36 the main activity jumped to 4–9 km depth. After the start of the eruption at 18:19 earthquakes were observed at all depths from the surface down to 14 km, but mainly at 2–12 km depth.

All the Hekla earthquakes forming the initial swarms, as far as was possible to discern, can be classified as normal, high-frequency earthquakes with clear S-phases (see Figure 4b) and are indistinguishable from earthquakes caused by brittle failure. Actually, the higher-frequency content was used for visual picking of the events from the continuous low-frequency volcanic tremor. Separate, low-frequency volcanic earthquakes were either not recorded or they were hidden in the tremor.

The volcanic tremor

Volcanic tremor at Hekla has only been observed during eruptions. Tremor starts simultaneously with the eruptive activity, thus its appearance in seismograms can be taken as the seismic expression of the onset of the eruption. The vigour of the tremor ceases together with the eruptive activity.

On January 17, 1991, continuous low-frequency volcanic tremor started at 17:02 GMT, marking the onset of eruptive activity, thirty minutes after the first observed earthquake (Soosalu *et al*. 2003). On February 26, 2000, the tremor appeared at 18:19, 79 minutes after the detection of the first earthquake (Soosalu *et al*. 2005). The simultaneous beginning of the eruption was verified by an eyewitness account by radio at the same moment (see Figure 5b).

In 1991 the station SAU, 35 km west of Hekla, was the closest digital station to record continuous tremor data at the onset of the eruption. In 2000, in addition to data from SAU, data were also available from the station HAU, 15 km west of Hekla, and from a few other more distant stations. The amplitude of the volcanic tremor at Hekla rises rapidly, within minutes, and becomes the dominant feature in the seismograms, effectively masking the earthquakes. At the onset of the 1991 eruption, the maximum reduced displacement (e.g. McNutt 1994) calculated from the records of the station SAU was about 8 cm^2 (Soosalu *et al*. 2003).

Tremor remains most vigorous during the first hour of the eruption, and subsequently declines. The amplitude of the tremor oscillates. Episodes of high-amplitude tremor, lasting from a few seconds to about ten seconds, are separated by moments of lower amplitude. To follow the development of the vigour of the tremor in a simple manner, we calculated its intensity using the procedure presented by Þorbjarnardóttir *et al*. (1997). The overall behaviour of the tremor during the first hours of the eruption is shown in a graph expressing its intensity at SAU in 1991 and both SAU and HAU in 2000, together with the observed earthquakes (Figure 6). In 1991, intensity began to decline sharply about one hour after the onset. This was not observed in 2000.

The spectrum of the tremor at the onset of both of the eruptions is remarkably similar. Generally it is observed that volcanic tremor has a peaked spectrum, typically with one dominant and a few subdominant frequency peaks (e.g. Aki *et al*. 1977; Chouet 1992; Seidl *et al*. 1981; Ferrick *et al*. 1982). The Hekla tremor also had this pattern. The characteristic frequency band of the tremor is 0.5–1.5 Hz and most of the time one single outstanding peak existed within the frequency band 0.7–0.9 Hz (Figure 8a-b). Occasionally there were two or three approximately equal high peaks. A handful of subdominant peaks sporadically appeared within the 0.5–1.5 Hz band. The location of the maximum peak was markedly constant throughout the first hours, for which we have continuous data. The amplitudes of the peaks decrease with time, but the spectral range remains the same. Although the general pattern is similar at various stations, some local spectral differences due to path effects were observed in 2000, with data from several stations.

The source of volcanic tremor is often inferred to be shallow, thus the tremor consists mainly of surface waves (e.g. McNutt 1986; Gordeev *et al.* 1990; Gordeev 1992; Goldstein and Chouet 1994; Ripepe *et al.* 1996). Our particle motion analysis of the data from the station HAU, both in 1991 and 2000, shows evidence for surface waves, particularly Rayleigh waves, and a shallow source of the tremor (Soosalu *et al*. 2003, 2005). The Hekla tremor attenuates faster with distance than the earthquakes, and is indisputably visible at remote digital seismograph stations only at the beginning of the eruption. This also indicates a shallow origin for the tremor, shallower than the earthquakes during the first hours of the eruption. The tremor is closely related to degassing, as it first appears when the conduit is open and is most intense in the beginning, when the eruption has a violent, explosive phase. When eruption vigour subsides, the tremor intensity also declines. Occasional tremor-bursts may reflect gas-bursts.

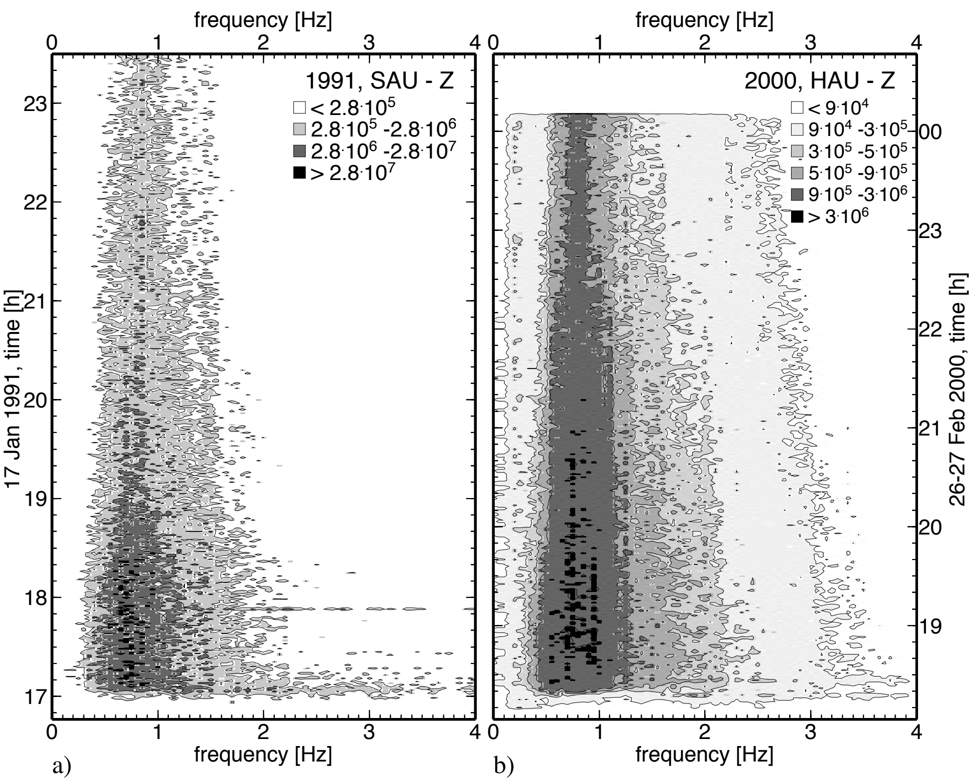

Figure 8. a) A spectrogram of the vertical component data at the station SAU for the first hours of the 1991 eruption. b) a spectrogram of the HAU vertical component data in the beginning of the 2000 eruption. The 2000 dataset ends at 00:10 on February 27. The graphs are adjusted thus that the onsets of the eruptions are aligned. The spectra are fast Fourier transform power spectra. The data consist of detrended and demeaned 60-second-long non-overlapping time windows with a 0.5 s cosine taper. No filtering is done. Amplitude scales of the tremor peaks are arbitrary and not comparable. The abscissa is the frequency and it is cut at 4 Hz because peaks above it are of negligible size. The ordinate is the time. – *Tíðnisnið fyrir skjálftarit af gosbyrjunum árin 1991 og 2000. a) Tíðnisnið frá Saurbæ fyrir gosbyrjun 1991, lóðréttur þáttur. b) Tíðnisnið frá Haukadal fyrir gosbyrjun 2000, lóðréttur þáttur. Tíðnisniðin sýna hvernig tíðniróf breytist með tíma. Tímakvarðarnir á lóðrétta ásnum eru stilltir af þannig að upphafstímar gosanna standast á. Háu tíðnirnar, sem eru til hægri á sniðunum, sýna jarðskjálftavirkni. Hún er fyrst og fremst tengd gosbyrjuninni og deyr síðan hratt út með tímanum. Gosórói hefur áberandi lága tíðni, þ.e. sést á vinstri hluta sniðanna. Hann er mestur fyrst og minnkar síðan hægt með tímanum meðan gosið varir.*

The seismicity during the eruptions after the onset day

Continuous digital records of Hekla tremor exist only for the first hours of both the 1991 and the 2000 eruptions. However, there are samples of tremor throughout these eruptions, and in both cases a general declining trend with occasional fluctuations is observed (Soosalu *et al.* 2003, 2005). The disappearance of the tremor from seismic records of the analogue station

HE at the flank of the volcano is taken as the end of the eruption.

After the initial swarm of earthquakes on January 17, 1991, earthquake activity at Hekla and the Hekla-Vatnafjöll area was modest during the eruption (Soosalu and Einarsson 2002). On January 19–February 21 fourteen events (M_L 1.1–2.6) were observed in the Hekla-Vatnafjöll area, nine of them at Hekla proper. At the end of the eruption, on March 11, the analogue station HE, on the flank of the volcano, recorded a swarm of about thirty small events before noon. They likely represent conduit collapse after the volcanic activity had ceased. The depths of the Hekla events after the onset day, about 8–12 km, were similar to typical events in the eastern part of the South Iceland seismic zone. All the earthquakes recorded by the SIL network during the eruption at Hekla were high-frequency tectonic events with distinct S-phases, none of them looked like low-frequency volcanic earthquakes (Chouet 1996). During later phases of the 2000 eruption, only one earthquake, with a size of M_L 0.8 was detected on March 1. It does not have a well-constrained location, due to large gaps between the stations.

DISCUSSION

During non-eruptive times the few earthquakes which occurred at Hekla do not have an apparent correlation to Hekla as a volcano. Instead, the seismicity in the area around Hekla and the Vatnafjöll volcano to the south have the same characteristics. The earthquakes cluster loosely along two N-S lineaments and occur mainly at 8–13 km depth, similar to the distribution of seismicity at the eastern end of the South Iceland seismic zone. A magnitude 5.9 (Mw) earthquake occurred in the SW part of Vatnafjöll in 1987. Its fault plane solution showed right-lateral strike-slip faulting on a N-S striking fault, i.e. characteristics of South Iceland seismic zone earthquakes (Bjarnason and Einarsson 1991). It was thus discovered that "bookshelf faulting", the seismicity pattern of the seismic zone, continues to the east as far as western Vatnafjöll, some 10 km further east than the surface expression of the seismic zone. A portion of the earthquakes in our data set occurred on the same lineament as the Vatnafjöll earthquake with its fore- and aftershocks. Another, fuzzier, N-S lineation can be discerned further east, through the central parts of Hekla and Vatnafjöll. Fault plane solutions for five events in this area are primarily of the strike-slip type (Soosalu and Einarsson 1997). Thus, South Iceland seismic zone tectonics extend well into the volcanic zone, according to our observations, all the way to longitude 19°40'W.

Depth estimates for Hekla earthquakes before the onset of the eruptions (Kristín Vogfjörð and Sigurður Th. Rögnvaldsson, unpubl. data; Soosalu *et al.* 2005) point to a shallow origin for the first earthquakes. Although it is likely that the initial earthquakes are related to stress changes caused by the intruding magma reaching the surface, it is clear that they are not forming a propagating front close to the tip of the intrusion.

We suggest that the lack of seismicity preceding Hekla eruptions is evidence for a deep magma source. The stress change related to a deep-seated, inflating magma chamber is distributed over a wider area and occurs aseismically until a dyke starts propagating. We have studied seismic rays between SIL stations and local earthquakes to look for signs of volumes of magma (Soosalu and Einarsson 2004). We did not find evidence for a substantial magma chamber at Hekla in the volume we could cover, i.e. the depth range of 4–14 km. This is in contrast with former geophysical studies which place a magma chamber under Hekla at 5–9 km depth (Kjartansson and Grönvold 1983; Eysteinsson and Hermance 1985; Sigmundsson *et al.* 1992; Linde *et al.* 1993; Tryggvason 1994). New interpretation of strain data by Sturkell *et al.* (2005a) suggest a Hekla magma chamber at 11 km depth, with a radius of 2 km, in line with our suggestion. Because of scarce data we could not examine well the eventual existence of a molten volume in the uppermost 4 km under Hekla. However, Hekla lacks the typical expression of a shallow magma chamber, such as persistent microearthquake activity and geothermal systems, and it is thus considered unlikely.

Our method was restricted to volumes with dimensions larger than about 800 m (see Soosalu and Einarsson 2004). If the Hekla magma chamber actually is located somewhere at 5–9 km, it must be too small for us to detect. The amount of erupted material can provide constraints for a size estimate of a magma chamber. It is generally assumed that only a fraction of the contents of a reservoir is drained during an eruption, until the pressure-drop inside the chamber leads to cessation of the eruption. In a theoretical study on explosive eruptions (andesitic to rhyolitic magma), Bower and Woods (1998) estimate the maximum amount of erupted material to be $\sim$10% of the total contents for a shallow chamber and only $\sim$0.1–1.0% for a deep chamber. The Hekla eruptions in 1970, 1980–1981, 1991 and 2000 produced lava and tephra of about 0.2 km^3 (Grönvold *et al.* 1983; Guðmundsson *et al.* 1992; Höskuldsson *et al.* submitted). With an assumption of 10%-drainage this suggests a magma chamber of 2–3 km^3, which should be large enough to be detected by our method.

The Hekla magma chamber may be a network of interconnected patches of molten material, rather than a simple voluminous structure. However, geochemical analysis of Hekla lavas shows that the composition of products during the course of an eruption is quite uniform (Grönvold *et al.* 1983; Karl Grönvold 2003, *pers. comm.*), thus not supporting a complicated magma chamber structure.

The quick onset of an eruption fed from great depth sounds problematic and rather unrealistic. Strain signals show that the dyke started propagating half an hour before the onset of the Hekla eruptions in 1991 and 2000 (Linde *et al.* 1993; Ágústsson *et al.* 2000). If the magma travels 14 km or more during half an hour, it requires at least a velocity of 7.8 m/s for the ascending magma. Sacks and Linde (2001; Selwyn Sacks 2001, pers.comm.) suggest that the rapid start of a Hekla eruption is the result of degassing. The gas phase is released from the magma inside the reservoir, and accumulates in the upper part of the reservoir, which forces the level of the liquid magma to sink. The pressure in the magma chamber increases due to the ascent of gas bubbles until an eruption starts, first extruding the gases from the upper part of the chamber. Because the gas phase erupts first, the eruption can easily commence more rapidly than an eruption starting with a lava flow. The gas release explanation is in harmony with the observation that the Hekla eruptions begin with an explosive phase emitting gases and tephra, and subsequently calm down to lava effusion (Grönvold *et al.* 1983; Guðmundsson *et al.* 1992; Höskuldsson *et al.* submitted).

Volcanic tremor during the two eruptions of 1991 and 2000 was very similar. It started simultaneously with the eruption and had a stable frequency-band during the first hours, although the eruptive activity and the amplitude of the tremor varied. The characteristic spectral band was about 0.5–1.5 Hz and the maximum peaks were around 0.7–0.9 Hz. This is at the lower end of the frequencies generally observed at active volcanoes in the world, mainly 0.1–8 Hz (Konstantinou and Schlindwein 2002).

A number of possible sources for volcanic tremor have been proposed in the literature. Some models explain the tremor as the result of resonant effects produced by the geometry of volcanic conduits. Turbulent motion in the vapour-gas-magma mixture makes the volcanic pipes oscillate (e.g. Seidl *et al.* 1981; Ferrick *et al.* 1982), and the frequency content of the tremor may vary with the length of the conduit. The characteristic low frequencies of Hekla tremor could indicate that the magma channel of Hekla is very large, i.e. the conduit would extend to a considerable depth and the magma chamber be at a deep level. Although the degassing-related origin of the tremor is shallow, the resulting vibration can occur in the long channel and produce the characteristic low frequencies. Other models suggest that volcanic tremor is produced by vibrations of tensile, fluid-filled, jerkily or suddenly opening cracks (Aki *et al.* 1977; Chouet 1981, 1985). In these models the excess pressure and degassing in the fluid generates the trembling. According to Chouet (1992) volcanic tremor is the response of the tremor-generating system to sustained bubble oscillations in the fluid. Julian (1994) explains the cause of the volcanic tremor to be nonlinear excitation by fluid flow, analogous to the excitation mechanism of musical wind instruments.

Volcanic tremor often begins prior to the actual surface outbreak of an eruption and may extend beyond the duration of surface activity (e.g. Chouet 1981; Montalto *et al.* 1995). This was not the case at Hekla, where the tremor started at the same time as the eruption and also terminated simultaneously with the eruptive activity. Apparently the tremor-producing mechanism at Hekla could not start before the magma conduit was opened, indicating that the tremor is closely related to degassing of magma. Schick (1988, 1992) states that strong tremor is not necessarily accompanied by strong lava emission, but strong degassing of a volcano does coincide with strong tremor. In the light of our observations on the latest Hekla eruptions, the Hekla tremor reflects the vigour of the eruption rather than the amount of produced lava.

The sudden swarm of at least one hundred shallow earthquakes at Hekla, which occurred in June 1991, is an unusual phenomenon. It may have been a failed attempt to revive the eruptive activity after its cessation on 11 March. Two former cases of resumed activity at Hekla are known: the August 1980 eruption continued after several months of quiescence in April 1981 (Grönvold *et al.* 1983), and the major eruption in 1766-1768 died down for six months in between (Þórarinsson 1967). The June 1991 earthquake swarm was not accompanied by a strain signal indicating the start of an intrusion, as was observed in January 1991 (Kristján Ágústsson 2000, *pers. comm.*). Interestingly, the 1980–1981 eruption was similar in this sense. In the initial phase of the eruption on August 17, 1980 an intrusion-related strain signal was observed, but not when the eruption continued on April 9, 1981 (Ragnar Stefánsson 2003, *pers. comm.*). An additional observation supporting an attempt to resume the eruption is the three volcanic-looking earthquakes that occurred during the swarm.

Following both the 1991 and 2000 eruptions, Hekla earthquakes have been few, and with an unusual low-frequency appearance, but with clear S-wave arrivals. This points to brittle failure in the crust rather than to a volcanic origin. Apparently, the reason for the low-frequency appearance is that the crust is still hot and weak after the eruption, and breaks under low stress-drop. Low-frequency volcanic earthquakes typical for many volcanoes in the world (e.g. Chouet 1996) have almost never been observed at Hekla. The only known exception so far occurred during the June 1991 swarm. High-frequency earthquakes at Hekla proper are observed almost exclusively during its eruptions. High-frequency Hekla events are generated during times of high strain, i.e. during an eruption or an attempt to resume an eruption.

The frequency content of the few inter-eruption earthquakes at Hekla can potentially be useful for long-term anticipation of eruptions by giving hint of a strain build-up. After 1991, the few Hekla earthquakes which occurred had a low-frequency appearance, until small high-frequency events were detected in February 1998 and July 1999. The re-appearance of high-frequency events may indicate that stress is starting to build up at Hekla and a new eruption is in preparation. The signal is vague, though, because the events are small and few. Our post-2000 eruption dataset demonstrates that Hekla events have again had a low-frequency appearance, until September 2004 when a clearly high-frequency earthquake occurred in the central part of Hekla. Another high-frequency event was observed at Hekla in March 2005. These events may be the first seismic indications that stress is building up again at Hekla. In addition, current tilt observations suggest increasing magma pressure under Hekla (Sturkell *et al.* 2005b).

Seismicity at the east end of the South Iceland seismic zone is of interest because of its similarity to inter-eruption seismicity in the Hekla-Vatnafjöll area. We have studied the area east of 20°12'W and observed that the earthquakes mainly occur along two N-S lineaments (Soosalu and Einarsson 1997, 2002). The seismicity is highest in the area of mapped surface faults (approximately 10 km in length), but in total, the epicentral lineaments are considerably longer, about 20–30 km. Our observations are in harmony with the boundary element calculations of Hackman *et al.* (1990) which imply that the South Iceland seismic zone faults have to be longer than observed on the surface, or the zone cannot accommodate the required transform deformation. Nearly all the hypocentres are concentrated at 6–12 km depth, with a peak at 8–10 km. This is consistent with the general pattern of earthquake depths within the seismic zone; hypocentres deepen towards the east (Stefánsson *et al.* 1993).

We interpreted earlier (Soosalu and Einarsson 1997) that the earthquake lineaments of the seismic zone are associated with the Hellar fault (in the west) and the Leirubakki fault (in the east) which are visible at the surface (Einarsson and Eiríksson 1982). However, the hypocentres are displaced about 1 km east from the faults at the surface, and thus the faults should be dipping approximately 80°. Recent field mapping has revealed two formerly unknown faults, Skarðsfjall fault east of the Hellar fault (Einarsson *et al*. 2002; Figure 2) and a fault east of the Leirubakki fault (Einarsson *et al*. 2003). The faults in the Leirubakki area are currently named as the western (formerly known) and the eastern (newly found) Leirubakki fault (Figure 2). The recently mapped faults are located above our earthquake lineaments and are more likely the origins for them.

Although the volcano Torfajökull is next to Hekla, its seismic behaviour is completely different. It is far more active, as small earthquakes occur persistently. Two sorts of events are identified: high-frequency events in the western parts of its caldera, hypothesised to be related to a cooling, but solidified magma chamber (Soosalu and Einarsson 1997) and low-frequency events in the south, which apparently are related to active magma (Soosalu and Einarsson 2003). Torfajökull seismicity is volcano-related and not affected by the transform tectonics of the South Iceland seismic zone.

CONCLUSIONS

The seismicity at Hekla and its immediate surroundings is quite unique and has a dual nature. In non-eruptive periods there is little seismicity and the few earthquakes that do occur are not related to the volcano itself. Instead, they have the same characteristics as the seismic activity in the South Iceland seismic zone located to the west of Hekla. The ultimate eastern terminus of transform tectonics of the South Iceland seismic zone therefore lies between the volcanoes Hekla and Torfajökull.

The Hekla eruptions in recent decades have been quite similar in size and general behaviour. Volcano-related seismicity occurs at Hekla in the form of an initial earthquake swarm, continuous low-frequency tremor and eventual sporadic small earthquakes during later phases of the eruption. The eruption-related seismicity starts only tens of minutes earlier with a swarm of hundreds of small earthquakes which increase in size towards the onset of the eruption. The sizes of earthquakes, $M_L < 3$, culminate around the very start of the eruption. They turn subsequently to a slight decline in size, continue for a few hours and then stop altogether. Only few earthquakes occur during the later phases. Rather little seismic energy is released during opening of eruptive conduits at Hekla, corresponding to a single event of M_L 3.4 in 1991 and M_L 3.2 in 2000, respectively. All the detected events during eruptions have been high-frequency, volcano-tectonic earthquakes.

Low-frequency volcanic tremor begins simultaneously with the onset of the eruption, when a conduit is open, and is closely related to degassing. Within minutes it becomes the dominant element in the seismic records. It is most violent during the first hours, continues throughout the eruption and fades away together with it. The characteristic frequency band of Hekla tremor is 0.5–1.5 Hz, with one or a few dominant peaks at 0.7–0.9 Hz. Large attenuation of tremor with distance compared to the eruption earthquakes indicates that the tremor has a shallower origin than the earthquakes. Particle motion observations point to a large amount of surface waves in the tremor signals.

With close seismograph stations the initial earthquake swarm related to the onset of an eruption can be detected soon after it begins, and with combined use of seismicity and strain observations it is possible to foresee Hekla eruptions on a short-time scale, approximately within an hour. The spectral low-frequency character of the inter-eruption seismicity at Hekla proper and its change to high-frequency seismicity when the strain is building up may provide a tool for long-term forecasting of Hekla eruptions, but its validity needs testing with future observations.

Acknowledgements

The colleagues at the Science Institute, University of Iceland, at the Icelandic Meteorological Office and at the Institute of Seismology, University of Helsinki are thanked for help and advice. The Icelandic Meteorological Office provided the digital SIL data. The National Power Company of Iceland funds the analogue seismograph stations. H. Soosalu has had financial support from NorFA, Suomalainen Konkordialiitto (Finnish Concordia Foundation), the Magnus Ehrnrooth Foundation of the Finnish Society of Sciences and Letters, the Finnish Cultural Foundation and the Vilho, Yrjö and Kalle Väisälä Foundation of the Finnish Academy of Science and Letters. Critical comments by Susanna Falsaperla, an anonymous reviewer and the editor, Bryndís Brandsdóttir, improved the paper.

ÁGRIP

Skjálftavirkni Heklu og nánasta umhverfis hennar

Eldgosin í Heklu 1991 og 2000 urðu innan tiltölulega þétts nets jarðskjálftamæla og sköpuðu þannig ný tækifæri til rannsókna á eðli og innviðum eldstöðvarinnar. Hekla er meðal virkustu eldstöðva Íslands og er staðsett á flekaskilum þar sem mætast skjálftabelti Suðurlands og eystra gosbeltið. Eldstöðin er þó ekki dæmigerð, hvorki fyrir íslenskar eldstöðvar né eldstöðvar á gliðnunarsvæðum. Gosvirkni Heklu hefur verið með nokkuð reglubundnum hætti á sögulegum tíma, með 1–2 gos á öld. Síðan 1970 hefur þó þetta munstur breyst og síðustu gos hafa verið á um 10 ára fresti. Gosin 1970, 1980–1981, 1991 og 2000 hafa verið hvert öðru lík, bæði hvað varðar goshætti, og magn og gerð gosefna. Skjálftavirkni í Heklu er um margt nokkuð sérstæð. Milli gosa er skjálftavirknin lítil, skjálftar eru fáir og smáir. Ekki hafa verið borin kennsl á neina langtímaaukningu sem rekja megi til vaxandi kvikuþrýstings í rótum eldstöðvarinnar fyrir gos. Skammtímaforboði fyrir gos er þó greinilegur í skjálftavirkninni. Ákafar hrinur smáskjálfta hafa mælst á undan öllum síðustu gosum Heklu og byrja þær 25–80 mínútum áður en gos kemur upp. Skjálftar mælast síðan fyrstu klukkustundir gossins meðan mestur gangur er í því en síðan dregur úr bæði gosvirkni og skjálftum. Auk jarðskjálfta mælist eldvirkniórói, þ.e. stöðugur lágtíðnititringur, meðan á gosum stendur. Óróinn byrjar nánast um leið og gosið og nær hámarki á fyrstu klukkustundinni. Útslag hans er í góðu samræmi við ákafann í gosinu og hann deyr út um leið og gosvirkni lýkur. Órói hefur aldrei mælst nema þegar gos er uppi. Þeir fáu og smáu skjálftar sem mælast í og við Heklu milli gosa virðast standa í litlu sambandi við eldstöðina. Þeir virðast hins vegar sverja sig í ætt við skjálftabelti Suðurlands. Þeir verða flestir á 8–12 km dýpi og raða sér á línur með N-S stefnu svipað og gerist á skjálftabeltinu vestan Heklu.

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