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RESEARCH ARTICLE

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Seismic constraints on magma chambers at Hekla and Torfajökull volcanoes, Iceland

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Abstract Hekla and Torfajökull are active volcanoes at a rift–transform junction in south Iceland. Despite their location next to each other they are physically and geologically very different. Hekla is an elongate stratovolcano, built mainly of basaltic andesite. Torfajökull is a prominent rhyolitic centre with a 12-km-diameter caldera and extensive geothermal activity. The scope of this study is to examine the propagation of body waves of local earthquakes across the Hekla–Torfajökull area and look for volumes of anomalous S-wave attenuation, which can be evidence of magma chambers. So far the magma chamber under Hekla has been modelled with various geophysical means, and its depth has been estimated to be 5–9 km. A data set of 118 local earthquakes, providing 663 seismic rays scanning Hekla and Torfajökull, was used in this study. The major part, 650 seismograms, did not show evidence for S-wave attenuation under these volcanoes. Only six seismograms had clear signs of S-wave attenuation and seven seismograms were uncertain cases. The data set samples Hekla well at depths of 8–14 km, and south part of it also at 4–8 km and 14–16 km. Western Torfajökull is sampled well at depths of 4–14 km, eastern and southern Torfajökull at 6–12 km. Conclusions cannot be drawn regarding the existence of magma beyond these depth ranges. Also, magma volumes of smaller dimensions than about 800 m cannot be detected with this method. If a considerable molten volume exists under Hekla, it must be located either above 4 km or below 14 km. The former possibility seems unlikely, because Hekla lacks geothermal activity and persistent seismicity, usually taken as expressions of a shallow magma chamber. An aseismic volume with a diameter of 4 km at the depth of 8 km in the west part of Torfajökull has been inferred in earlier studies and interpreted as evidence for a cooling magma chamber. Our results indicate that this volume cannot be molten to a great extent because S-waves travelling through it are not attenuated. Intense geothermal activity and low-frequency earthquakes are possibly signs of magma in the south part of Torfajökull, but a magma chamber was not detected there in the areas sampled by this study.

Keywords Hekla · Iceland · Magma chamber · S-wave attenuation · Torfajökull

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Introduction

The eruptive activity in Iceland is expressed by volcanic systems that are comprised of a central volcano—the location of the highest productivity—and a fissure swarm transecting it (e.g. Jakobsson 1979, Björnsson and Einarsson 1990). Hekla and Torfajökull are active central volcanoes at the junction of the South Iceland seismic zone and Eastern volcanic zone in south Iceland (Fig. 1; Einarsson and Saemundsson 1987). Hekla is an elongate ENE–WSW-oriented summit crossed by a fissure swarm with the same orientation. Hekla is one of the most active volcanoes in Iceland and has erupted at least 18 times in the last 1,100 years. The latest eruption occurred in February–March 2000 (Einarsson 2000; Stefánsson et al. 2000) and produced an estimated lava volume of 0.17 km^3 (Ólafsdóttir et al. 2002). The volcano Torfajökull has a caldera, 12 km in diameter, and an extensive high-temperature geothermal field (Saemundsson 1972, 1982). Fissure swarms of Torfajökull extend to NE and SW of the volcano. Last large eruptions have taken place in the

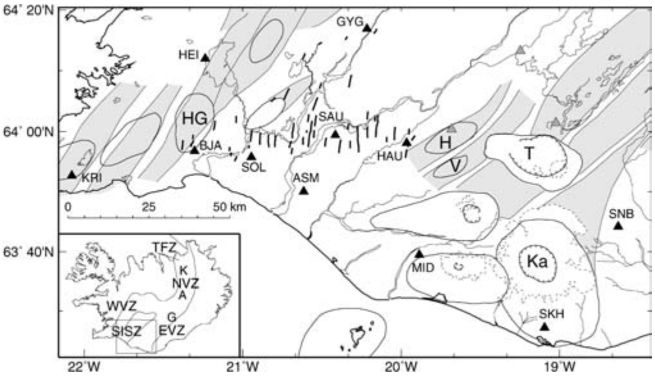

Fig. 1 Index map of the Hekla–Torfajökull area. *Black triangles* are the digital South Iceland Lowland (SIL) seismograph stations and *grey triangles* the analogue stations. *Thick black lines* are the faults of the South Iceland seismic zone. The central volcanoes are outlined, their fissure swarms are *shaded grey* (Einarsson and Saemundsson 1987) and the calderas are *hatched* (Jóhannesson et al. 1990). Central volcanoes referred to in the text are *H* Hekla; *T* Torfajökull; *V* Vatnafjöll; *Ka* Katla; *HG* Hengill–Grensdalur; *G* Grímsvötn; *A* Askja; *K* Krafla. *Dashed lines* mark the glaciers. The *smaller index map* shows the locations of the areas related to the plate boundary: Western (*WVZ*), Eastern (*EWZ*) and Northern (*NVZ*) volcanic zones, South Iceland seismic zone (*SISZ*) and Tjörnes fracture zone (*TFZ*). All the figures are made with the Generic Mapping Tools program (Wessel and Smith 1998)

Torfajökull system in about A.D. 150, 900 and 1480 (Larsen 1984). Geologically, Hekla and Torfajökull are unlike most Icelandic rift-zone volcanoes, as typified by Krafla (Björnsson et al. 1977). Moreover, although they are located side by side, they are very dissimilar to each other. The volcanic products of Hekla are basaltic andesites in contrast with the rift zone tholeiites (Sigvaldason 1974). Torfajökull is a rhyolitic complex, the largest silicic centre in Iceland (McGarvie 1984).

In this paper, we study the propagation of body waves of local earthquakes across the Hekla–Torfajökull area and look for volumes of anomalous attenuation. Magma chambers are likely to attenuate seismic waves, particularly S-waves. Lack of anomalous attenuation, therefore, may be taken as evidence for the absence of large batches of magma.

Various geophysical methods have been used for outlining the magma chamber under Hekla. Kjartansson and Grönvold (1983) concluded from geodetic measurements that the top of the magma reservoir of Hekla is roughly at 8 km depth. Eysteinsson and Hermance (1985) found a 'pocket' of low-resistivity material at approximately 8 km depth beneath Hekla in a magnetotelluric survey. They suggest it to be a manifestation of a local magma chamber. Sigmundsson et al. (1992) made GPS measurements of the deflation of 1991 Hekla eruption and determined a pressure source at 9 km depth (+6, −7 km). Linde et al. (1993) obtained a source depth of 6.5 km (the top at about 4–5 km) from inversion of strainmeter data of that eruption. Tryggvason (1994) defined a magma pressure source at 5–6 km depth from tiltmeter data.

In their magnetotelluric survey, Eysteinsson and Hermance (1985) found a low-resistivity area under Torfajökull, similarly to that beneath Hekla. Guðmundsson (1988) discusses the existence of the Torfajökull magma chamber, and suggests the top of the chamber is at about 3 km depth, based on the measurements of Eysteinsson and Hermance, and the existence of high-temperature geothermal activity. In our study of Torfajökull seismicity (Soosalu and Einarsson 1997), we detected indications of magma in the form of a spherical volume with centre at 8 km depth and diameter of 4 km that is void of earthquakes and surrounded by hypocentres. We interpreted this volume to be a cooling magma chamber. At Torfajökull, we have also observed numerous low-frequency earthquakes that have been recorded since 1985 (Brandsdóttir and Einarsson 1992). They are mainly clustered in the southern part of the caldera and may indicate an active magma volume there.

Magma chambers of other Icelandic volcanoes

Earlier studies have found evidence for magma chambers beneath other active volcanoes in Iceland, including Krafla, Grímsvötn, Hengill–Grensdalur, Katla and Askja (Fig. 1). At Krafla volcano in north Iceland, Einarsson (1978) used records of local earthquakes and observations of high shear wave attenuation for outlining a shallow crustal magma reservoir of size of ~2×7 km, with an upper boundary at 3 km depth and a lower boundary not deeper than at 7 km depth. Björnsson et al. (1979), using geodetic and gravimetric data, modelled the Krafla magma source with a spherical volume at a depth of 3 km. Tryggvason (1986) interpreted his ground deformation measurements to be expressions of multiple magma reservoirs under Krafla caldera. According to him, the magma source of Krafla would consist of a shallow chamber with a centre at 2.6 km and three other interconnected reservoirs below it down to about 30 km depth. Brandsdóttir et al. (1997) studied the crustal structure of Krafla using seismic reflection and refraction data, and found a broad high-velocity dome rising from the lower crust. They identified the Krafla magma chamber as a body of low seismic velocities on top of the dome. According to their study, the chamber is a 0.75–1.8-km-thick lens with a top at 3 km depth. Its width in a N–S direction is 2–3 km and in a E–W direction it is 8–10 km. They obtained no evidence of partially molten material at mid-crustal depths below the chamber.

At the subglacial Grímsvötn volcano, Björnsson et al. (1982) suggested a shallow magma chamber to be the source of the heat flux of the intense geothermal activity. Einarsson and Brandsdóttir (1984) studied the earthquake activity related to a minor eruption in 1983. They concluded that the precursory events were caused by brittle failure of the crust above and around an inflating magma chamber, and the swarm related to the onset of the eruption by failure of the chamber walls and migration of magma. These events were located slightly SE of the main caldera. Guðmundsson and Milsom (1997) modelled structures under Grímsvötn using gravity and magnetic data. According to their study, a possible interpretation of the data is a magma chamber at about 2 km depth, 0.8 km thick and 4 km across. GPS measurements at Grímsvötn (Sturkell et al. 2003a) revealed inflation before an eruption in 1998, eruption-related subsidence, and uplift after the eruption. The source depth was modelled to be located beneath the centre of the Grímsvötn caldera, at least to 1.6-km depth. Alfaro et al. (2003) studied regional and teleseismic P-wave delays through the Grímsvötn volcano and find evidence for a low-velocity body, probably caused by partial melt, at depths below the 4-km-thick brittle region where earthquakes occur.

Foulger and Toomey (1989) made a tomographic study at the area of Hengill–Grensdalur volcanic complex and found a small, low-wave-speed body beneath the Hengill volcano that was tentatively interpreted as a magma body. However, after revisiting the area and combining the earlier data with an improved survey (Miller et al. 1998), no such body was detected and the earlier discovery was considered an artefact of the method. Thus, they did not find bodies that might contain melt in the upper crust beneath the Hengill volcanic system. However, we find qualitative evidence in seismic records of local earthquakes that clearly indicate S-wave shadow effects in the Hengill area. No systematic study has been made of this yet.

At Katla, south of Torfajökull, Guðmundsson et al. (1994) carried out a two-dimensional seismic undershooting. They found a shallow magma chamber, about 5 km across and with a bottom at 3 km depth, the top was not well resolved. This body is expressed by P-wave delays and S-wave shadowing. The chamber is underlain by rocks of average or high velocity. In the summer of 1999 either a minor subglacial eruption or a shallow intrusion, accompanied by seismic activity, occurred at Katla. GPS measurements at Katla in 1993–2000 show an outward displacement from the centre of the caldera, which is consistent with magma accumulation at shallow depth (Sturkell et al. 2003b).

Tryggvason (1989) and Sturkell and Sigmundsson (2000) have measured ground deformation at the Askja volcano in north Iceland since 1966, beginning 5 years after its latest eruption. Their observations indicate a shallow magma chamber, modelled with a point source at about 3 km depth near the centre of the main caldera of Askja.

Data and methods

It can be theoretically expected that if a seismic ray between a hypocentre and a seismic station passes molten or partially molten material it will change the appearance of the seismogram with the S-wave visibly attenuated or even missing. We used records of 118 local earthquakes observed by the digital SIL seismograph network (Fig. 1; Stefánsson et al. 1993) for scanning possible locations of melt below the Hekla and Torfajökull volcanoes. These earthquakes occurred in 1991–1999 in the area around Hekla and Torfajökull (63°42′–64°18′N, 18°30′–20°12′W). We added observations of three local analogue stations (Fig. 1) to improve the location accuracy and made relocations with the program HYPOINVERSE (Klein 1978; Soosalu and Einarsson 1997). We studied 663 seismic rays between SIL stations and the hypocentres that had paths travelling under Hekla, and its fissure swarm, Torfajökull or Katla (Fig. 2). Some rays go under both Torfajökull and Hekla/Hekla fissure swarm, some under both Torfajökull and Katla.

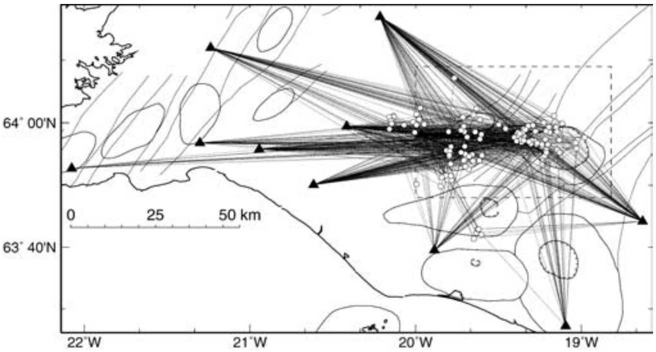

Fig. 2 Map showing all the seismic rays used in this study, from earthquakes in the Hekla–Torfajökull area to digital SIL seismograph stations. *White dots* show the epicentres of the events. *Dashed rectangle* marks the location of the study area used in later maps

The seismic rays were traced with the program SEIS83 (Cervený and Pšencik 1983) using a one-dimensional crustal model (see Soosalu and Einarsson 1997), which is based on refraction profiles in the region and consists of layers with constant velocity gradient. This approximate model was assumed to be precise enough for this purpose. The relative locations given by SEIS83 were transformed to geographical coordinates with simple calculations.

The station SKH is located on the southern part of Katla volcano (Fig. 1), so some rays observed by SKH pierce both Torfajökull and Katla. Thus, eventual S-wave attenuation may have occurred under Katla, as well as under Torfajökull. The magma chamber under the Katla caldera, detected by Guðmundsson et al. (1994), is quite near the surface in the uppermost 3 km. All our rays that travel under Katla caldera area pass it at depths of 9–13 km. Thus, at least the shallow magma body does not affect our results. Stations BJA and HEI are located near the Hengill volcanic system (Fig. 1). The local structures of Hengill are not likely to affect the appearance of the rays observed by these stations as BJA is located south of the central volcano itself. The rays to HEI travel only under the fissure swarm and not the Hengill volcano.

According to the appearance of the S-waves we classified the rays into three categories:

1. Normal: S-arrival in the record is clear, indistinguishable from S-waves in other areas.

2. Uncertain: it is difficult to judge if S is normal or if something has affected it.
3. Abnormal: S-wave is clearly attenuated, or even absent.

Observations

By far, most of the seismic records with ray paths travelling under Hekla and Torfajökull look normal. An S-wave onset is easily picked and it fits to travel-time calculations. In total, 650 records of 663, with rays passing Hekla, Torfajökull and/or Katla, were classified normal. Thirty-nine of these were records of Hekla low-frequency earthquakes, lacking high frequencies, but having a distinct S-wave (described in Soosalu and Einarsson 1997). The rays of normal looking records are presented in depth slices in Fig. 3a–l. They cover Hekla rather well at the depth level of 8–14 km, and the southern part of it at 4–8 and 14–16 km. The western part of the Torfajökull volcano is well-covered at 4–14 km, and the eastern and southern parts of Torfajökull are quite well covered at 6–12 km. Figure 4 shows an example of a Torfajökull earthquake that has normal looking records at every observing station. The P and S data look clear and their arrival times are consistent with the crustal model.

The seismic rays of records that were classified as uncertain or abnormal were very few. They were recorded from eight earthquakes, all in Torfajökull, occurring at 3.3–12.7 km depth. Five of these events had only one record with uncertain or abnormal appearance, two had two, and one had four such records. The rays are shown in depth slices in Fig. 5a–j and similarly as the normal rays in Fig. 3a–l. They travel under Torfajökull at 6–12 km and under Hekla at 10–16 km. Seven records, observed by stations BJA, HEI, KRI and MID, were such that it was hard to judge if a real S-wave exists or not. The seismograms are shown in Fig. 6. Only six records were definitely abnormal looking, and they were observed only by stations BJA and HEI. The seismograms are presented in Fig. 7.

Our results must be evaluated with the following reservations in mind:

1. The absence of the S-waves may be a source effect. A source effect is implausible if only records of part of the observing stations miss the S-wave.
2. In exceptional cases, an S-wave can be absent if the station is located in the direction of one of the nodal points of the focal sphere.
3. A magma chamber may cause a lens effect. If the magma volume is small it may only bend the seismic rays without affecting their appearance. Thus, the seismic rays look normal even though molten material exists. This effect limits our resolution to bodies that are larger than the characteristic wavelength of the waves used, i.e. about 800 m.

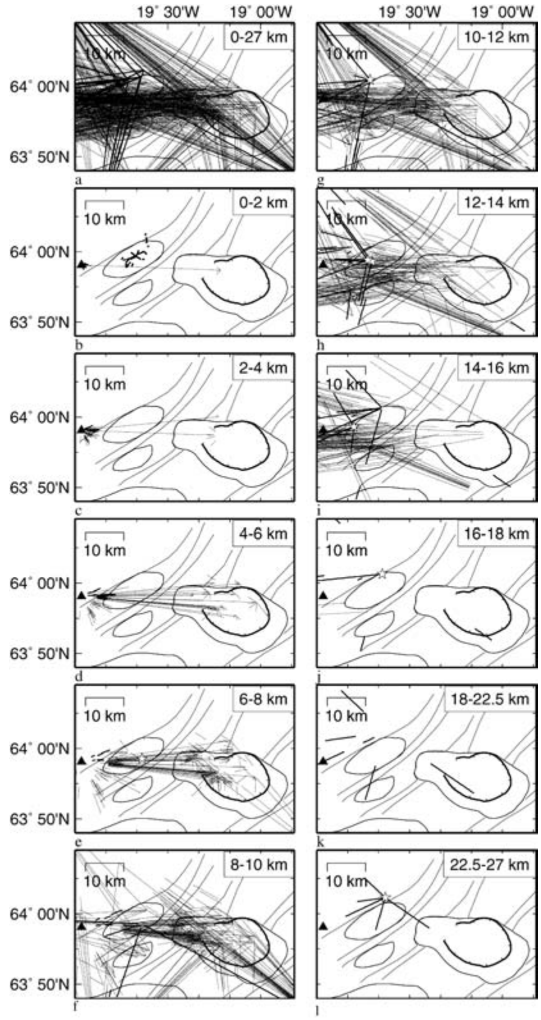

Fig. 3 Depth slices of the Hekla–Torfajökull area with seismic rays of the 'normal' seismograms. **a** Entire rays are plotted; **b–j** 2-km-thick depth slices, **k**, **l** the two deepest 4.5-km-thick slices. The depth interval is marked in the *upper right corner of each figure*. The surface map is shown in every figure for reference. *Dashed lines* show the rays of the seismograms with normal appearance. The *solid lines* mark rays from low-frequency Hekla earthquakes, the seismograms of which are normal as well. Hypocentres of the low-frequency Hekla earthquakes are plotted with *white stars* at their actual depths. **b** The fissures of the 1970, 1980–1981 and 1991 Hekla eruptions. The *black triangle* is the closest digital seismograph station, HAU

Hekla

Almost all seismic records that have rays travelling under Hekla or its fissure swarm look normal and, thus, they give no hint of molten material under this volcano. If Hekla has a prominent magma chamber it has to be either shallower than about 4 km or deeper than about 14 km. The seismic ray coverage of Hekla area is poor in the uppermost 4–5 km, below that it is quite good down to 15 km. Only a few rays cross Hekla at greater depth. It is worth noticing that many seismic rays passing Hekla have already travelled under Torfajökull, as a large number of

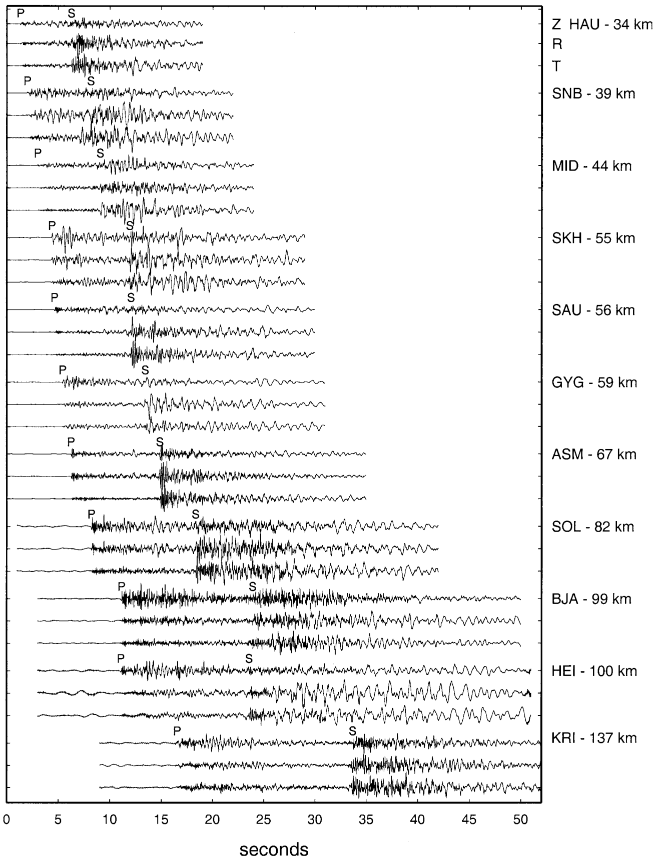

Fig. 4 A Torfajökull earthquake at 5.4 km depth (location is shown with a *black dot* in Fig. 9), local magnitude 2.4. All the SIL stations that observed the event (see Fig. 1, distances mentioned) have recorded both P and S clearly. No filtering is done to the data, and amplitude unit is arbitrary. Z is the vertical component, R radial and T transverse. The relative onset time difference between observing stations is the same as in reality

earthquakes used in this study occurred within the Torfajökull volcano.

The rays of the few uncertain or abnormal records under Hekla or its fissure swarm travel at 10–16 km, but at those depths there are rays of normal seismograms as well. As these are records of Torfajökull earthquakes, it is possible that the S-waves have already suffered attenuation there.

In our earlier study of the background seismicity in the Hekla–Torfajökull area (Soosalu and Einarsson 1997), we observed earthquakes beneath Hekla at 8–14 km depths. These events had peculiar low frequency appearance, but still had clear S-wave arrivals at every seismic station. These features point to brittle failure, possibly with a low stress drop. The S-waves speak against molten material at these locations. One Hekla earthquake, of those containing only low frequencies, occurred at the considerable depth of 26 km. Despite its deep location, clear and fitting S-waves were observed at all the SIL stations recording it.

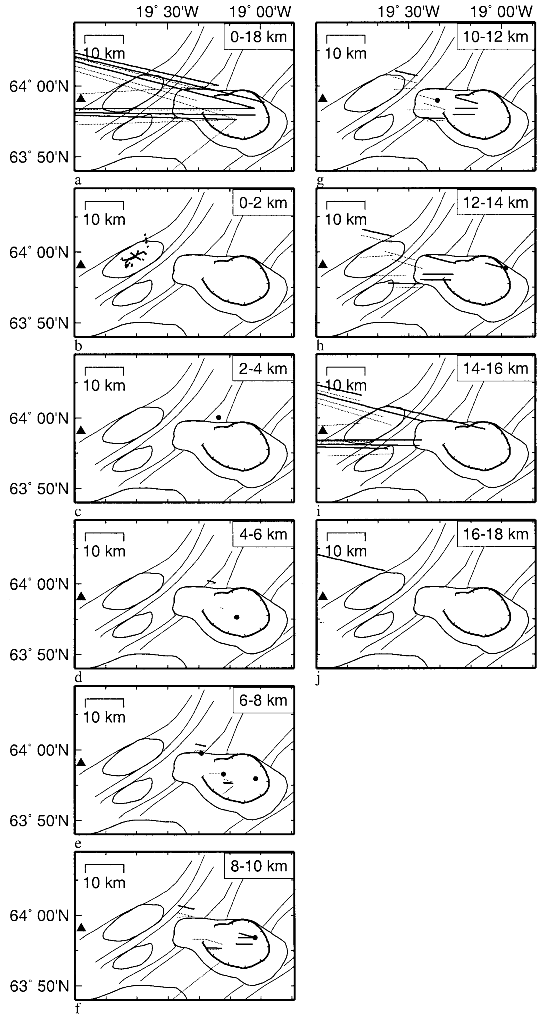

Fig. 5a–j Depth slices of the seismic rays in the Hekla–Torfajökull area having 'uncertain' and 'abnormal' looking seismograms. **a** Entire rays are plotted; **b–j** 2-km-thick depth slices, down to 18 km. The depth interval is marked in the *upper right corner of each figure*. The surface map is shown in every figure for reference. *Dashed lines* show the rays of the seismograms with uncertain appearance, *solid line rays* of the seismograms with abnormal appearance. Hypocentres of the corresponding earthquakes are plotted with *black dots* at their actual depths. **b** The fissures of the 1970, 1980–1981 and 1991 Hekla eruptions. The *black triangle* is the closest digital seismograph station, HAU

Torfajökull

Similar to Hekla, the majority of seismic records with rays travelling in the Torfajökull area look normal. Thus, we do not have evidence for considerable molten volumes in areas from which we have data. We have a good seismic ray coverage from about 5 km depth down to some 14 km in the western part of the volcano and to 10–12 km in the eastern part; at greater depth the rays get fewer. Waves travelling under Torfajökull in these areas look normal. Due to the unfavourable hypocentral distribution we have only a couple of rays in the uppermost 5 km at Torfajökull.

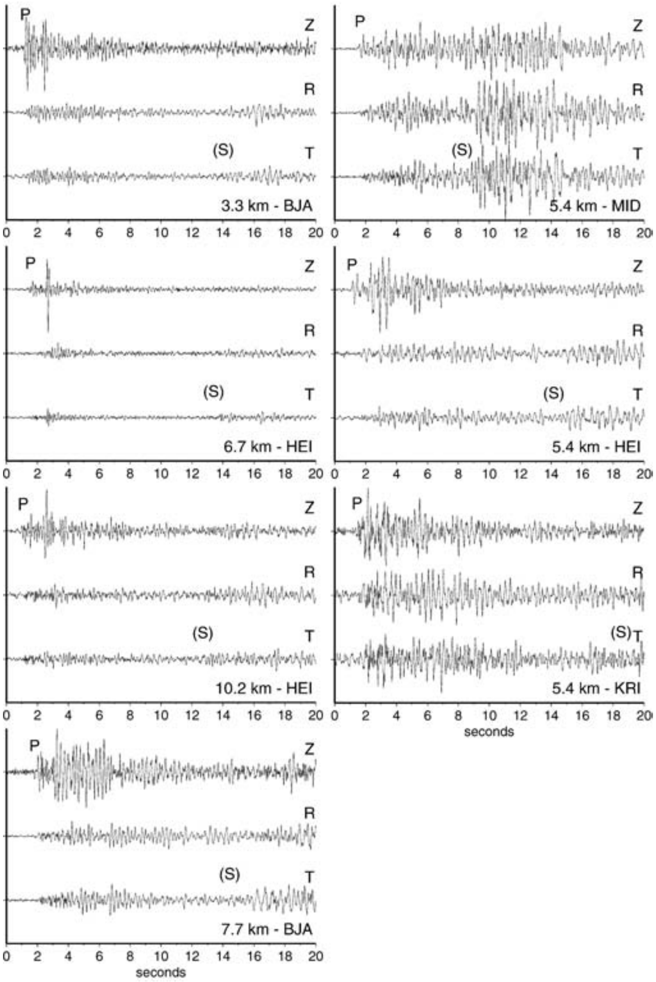

Fig. 6 The seven seismograph records that were uncertain looking, band-pass filtered at 1.0–20 Hz. The names of the stations and the depths of the events are mentioned. P marks the actual picked P-arrival, (S) the calculated arrival time of an S-wave. Amplitudes are arbitrary and not to scale with each other. Z is vertical, R radial and T the transverse component

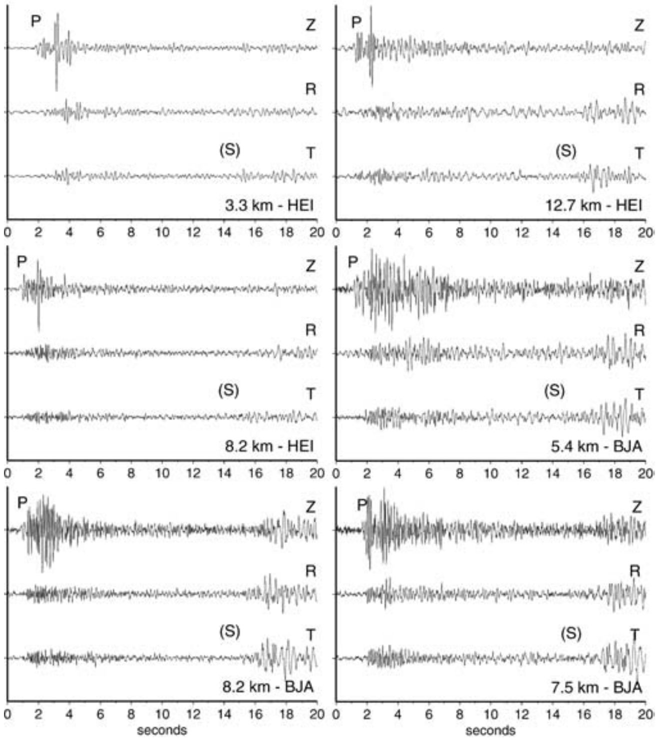

Fig. 7 The six seismograms that had abnormal appearance, band-pass filtered at 1.0–20 Hz. The names of the stations and the depths of the events are mentioned. P marks the actual picked P-arrival, (S) the calculated arrival time of an S-wave. Amplitudes are arbitrary and not to scale with each other. Z is vertical, R radial and T the transverse component

The few uncertain or abnormal looking records have rays that have travelled under Torfajökull, mainly at about 6–14 km. Their ray paths do not differ much from those of the normal records. Also, the majority of seismograms that have rays piercing the 4-km-diameter cooling magma chamber at 8 km depth in the west part of the volcano (Soosalu and Einarsson 1997) look normal. Thus, the S-waves are not attenuated there. This suggests that the magma chamber with its aseismic volume consists of hot, but not molten material, at least not molten to a great extent.

Torfajökull is a persistent source of low-frequency earthquakes. They are small in magnitude and often occur in swarms. They are difficult to locate, as their P-arrivals are typically very small and both P- and S-waves are emergent (Fig. 8). However, we have obtained locations for a set of Torfajökull low-frequency earthquakes (Fig. 9). Most of them originate at 6–10 km depth beneath the south part of the caldera. The peculiar nature

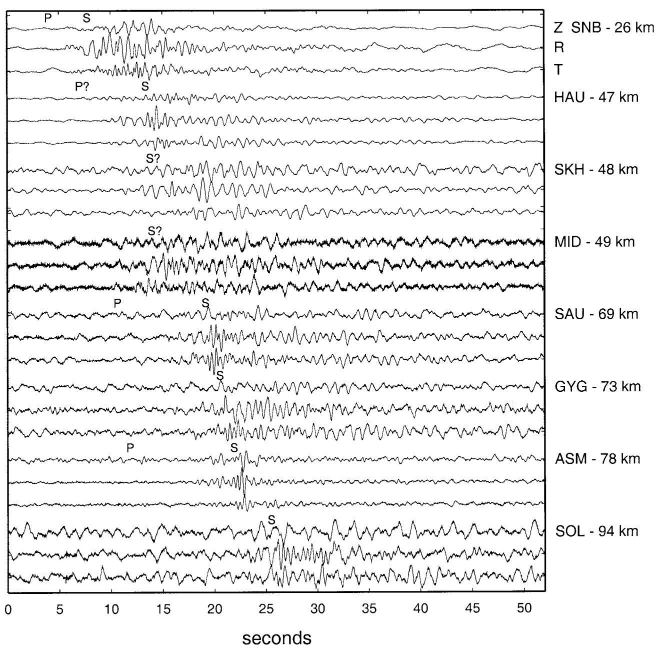

Fig. 8 An example of a Torfajökull low-frequency earthquake at 9.8 km depth. Its location is shown with a *black star* in Fig. 9 and it had a local magnitude of 1.1. The data of HAU, SKH and MID are band-pass filtered at 1.5–20 Hz; because of high noise, the other records are unfiltered. The amplitude unit is arbitrary. Z is vertical, R radial and T the transverse component. The relative onset time difference between observing stations is the same as in reality

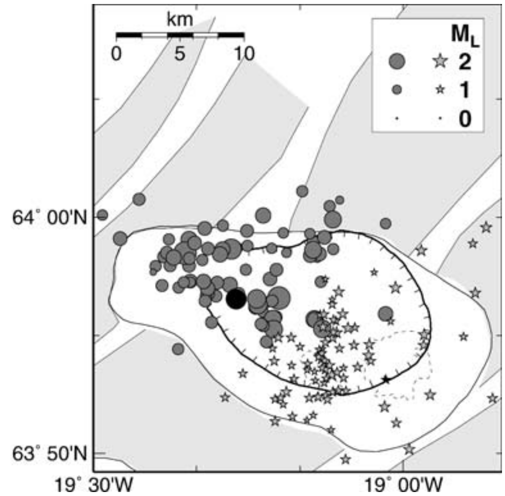

Fig. 9 Low-frequency earthquakes at Torfajökull in July 1994–April 2000 (*grey stars*). Plotted are 85 events that fulfil the location criteria: root mean square time error ≤ 0.2 s, horizontal error ≤ 1.0 km, vertical error ≤ 2.0 km, largest gap between observing stations $\leq 180°$, minimum number of observed P-waves =3 and minimum number of S-waves =4. For comparison, high frequency earthquakes (*dark grey dots*, observed in July 1991–October 1995, rms ≤ 0.2 s, horizontal error ≤ 1.0 km, vertical error ≤ 2.0 km, gap $\leq 180°$) are shown surrounding the cooling magma chamber that has a centre at 8 km depth at 63°58′N, 19°14′W (Soosalu and Einarsson 1997)

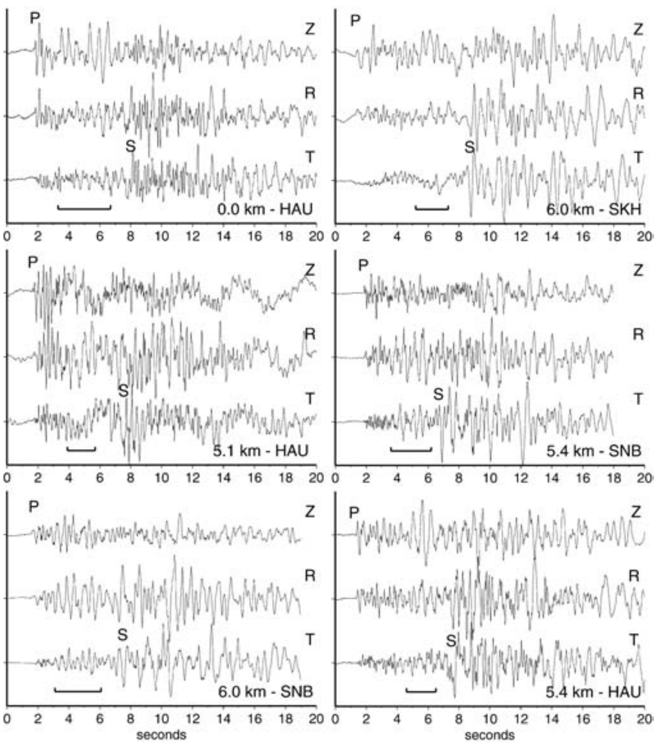

Fig. 10 Examples of seismic records with a low-frequency wave package between the P- and S-waves: the data are unfiltered. The names of the stations and the depths of the events are mentioned. P- and S-arrivals are shown, and the low-frequency package marked with a *horizontal bar*. Amplitudes are arbitrary and not to scale with each other. Z is vertical, R radial and T the transverse component

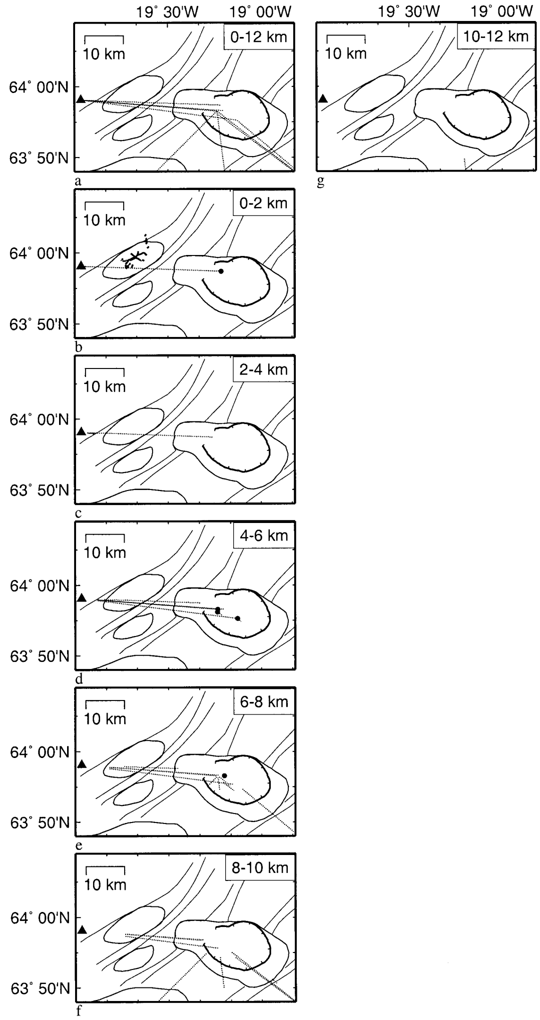

Fig. 11 Depth slices showing the seismic rays of the records with low-frequency wave package between the P- and S-waves. The corresponding earthquakes are shown at their actual depth

of these earthquakes may possibly be related to active magma in the south part of Torfajökull. However, our seismic ray data do not give indications of molten magma there (compare Fig. 9 with Figs. 3 and 5).

There were also other interesting waveforms in the appearance of the seismograms. Nine records, observed by stations HAU, SNB, SKH and MID, showed a clear low-frequency signal between P and S. Examples are shown in Fig. 10. All these records are classified as normal. These seismograms are products of five earthquakes at Torfajökull, four of them near the SW edge of our 'cooling magma chamber' and one beneath the middle of the caldera (Fig. 11). Four have focal depths of 5–8 km depth and one less than 1 km. Low-frequency phases were not observed at all the seismic stations for these events. Therefore, it appears that the low-frequency phase is a path effect. We suggest that, without further argumentation, these rays have travelled somewhere in the contact area of a solid and molten area where the low-frequency signal is produced.

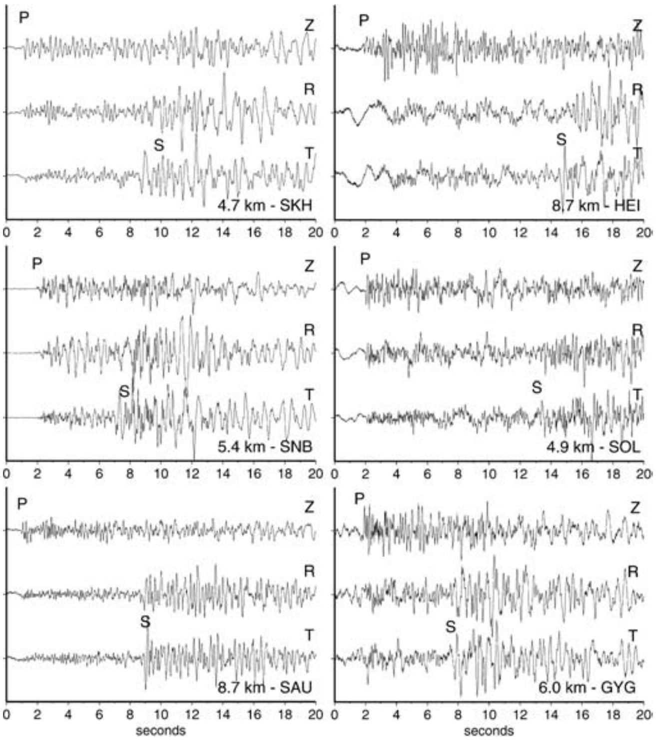

Fig. 12 Examples of seismic records with a long P-wave train, the data are not filtered (see also the record of the station MID in Fig. 6). The names of the stations and the depths of the events are mentioned. P- and S-arrivals are shown. Amplitudes are arbitrary and not to scale with each other. Z is vertical, R radial and T the transverse component

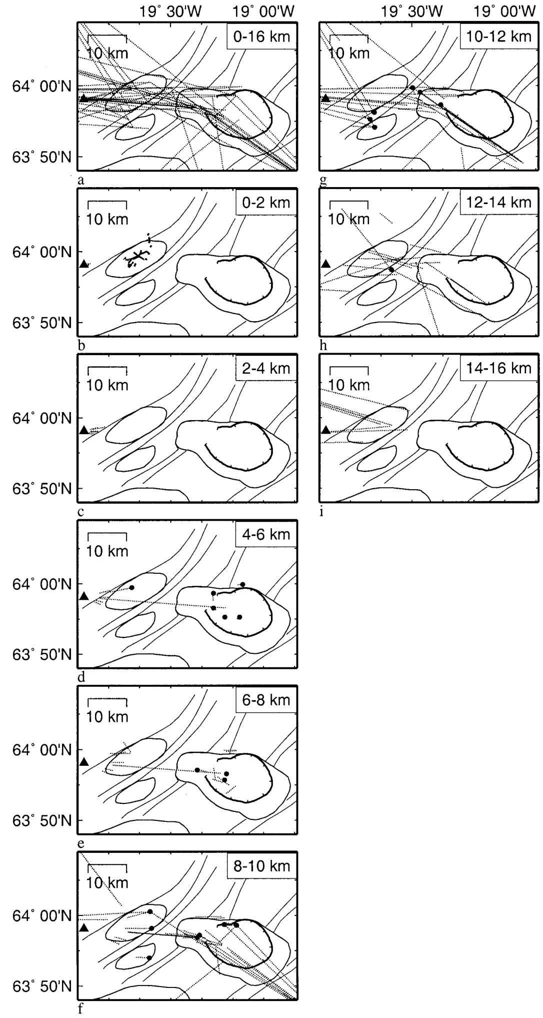

Fig. 13 Depth slices showing the seismic rays of the records with a long P-wave train. The corresponding earthquakes are shown at their actual depth

Some of the seismograms have very long-lasting P-wavetrains (examples in Fig. 12). We defined the P-wavetrain 'long' if the amplitude of the Z-component record did not diminish clearly before the arrival time of the S-wave. We found 50 such seismograms. Thirty-one of the seismic rays had travelled under Torfajökull, and 36 under Hekla or its fissure swarm (Fig. 13). Seismograms with a long P-wavetrain were observed at every station except the distant KRI. Forty-nine of the rays had normal appearance and one of uncertain character was recorded at station MID (see Fig. 6). The events that produced seismograms with a long P-wavetrain were always observed to have shorter P-wavetrains at least at some of the stations. These earthquakes were located either around the 'cooling magma chamber' of Torfajökull (at 5–12 km depth) or in an earthquake lineament (at 6–13 km depth) crossing the middle parts of Hekla and the central volcano Vatnafjöll south of it. This is the eastern one of the two N–S earthquake lineaments observed in the Hekla-Vatnafjöll area (Soosalu and Einarsson 1997). The long P-wavetrain consists of later P-phases than the direct P and is apparently caused by heterogeneities at the boundary between the cooling chamber and surrounding rock. Heterogeneities in the structures of the volcanoes of Hekla and Vatnafjöll might cause secondary P-phases there.

Discussion

Neither Hekla nor Torfajökull is a typical rift zone volcano, but they are anomalous in the Icelandic plate-tectonic setting. They are located in a tectonically unique area, at a junction between a transform South Iceland seismic zone, a rift segment and a flank segment of the eastern volcanic zone. North of Torfajökull, the volcanism is of rifting type. Torfajökull itself is located at the site where the rifting propagates towards the south-west into the flank zone (Óskarsson et al. 1982). Hekla stands at the junction of the transform and the volcanic zone. This is reflected in the seismicity of Hekla because, in non-eruptive times, Hekla earthquakes have a similar distribution as that in the eastern end of the South Iceland seismic zone and are not related to the volcano itself (Soosalu and Einarsson 1997).

It is to be expected that a shallow magma chamber shows its presence through high-temperature geothermal activity and persistent small-scale seismicity. This is the case with the Icelandic volcanoes that are known to have shallow magma chambers, such as Krafla and Grímsvötn. On the contrary, Hekla has practically no geothermal activity, and only few earthquakes occur there in non-eruptive periods. Thus, an existence of a shallow (in uppermost 5 km) magma chamber at Hekla does not seem likely. Indeed, various geophysical measurements suggest that the depth to the Hekla chamber is in the depth range of 5–9 km. Our observations, however, point to an even deeper source, as we see no evidence for a considerable molten volume in the uppermost 14 km.

All the other estimates for the depth of the Hekla magma chamber, 5–9 km, are quite consistent with each other, although our observations do not support this. Provided that the Hekla chamber actually is located at this depth range, it has to be too small for us to detect. The amount of material produced in eruptions can give constraints in estimating the size of a magma chamber. It is commonly believed that only a fraction of the contents of a reservoir is drained in an eruption until underpressure inside the chamber leads to cessation of the eruption. Bower and Woods (1998) have made a theoretical study on explosive eruptions (andesitic to rhyolitic magma) and estimate the amount of erupted material to be in maximum ~10% of the total contents for a shallow chamber and only ~0.1–1.0% for a deep chamber.

Based on observations at Krafla, a basaltic volcano, we can get an estimate on relationship between the erupted material and the size of its magma reservoir. Brandsdóttir et al. (1997) give limitations to the size of the chamber: 0.75–1.8×2–3×8–10 km, which gives volume constraints of 12–54 km^3. Tryggvason (1980) made geodetic measurements at Krafla in 1975–1979, during a volcanic episode, and estimated the volume of material flowing out from the Krafla magma chamber in this period to be 0.481 km^3, 0.479 km^3 of which was travelling as intrusions to north and south and only 0.002 km^3 of which was erupted on the surface as basaltic lava. At later phases of the episode, in 1980–1984, the activity was more focused on lava production, and in total the Krafla lavas are estimated to be 0.25 km^3 (Saemundsson 1991). Still, the amount of intrusive and eruptive products leaving the magma chamber is just a fraction of the total volume estimates of the chamber.

The recent Hekla eruptions, in 1970, 1980–1981, 1991 and 2000 have produced lava and tephra in the range of 0.2–0.3 km^3 (Grönvold et al. 1983; Guðmundsson et al. 1992; Ólafsdóttir et al. 2002). With a 10%-drainage assumption this would lead to a magma chamber with a size of 2–3 km^3, i.e. it should be large enough to be detected with our method. However, we cannot control the validity of this estimate.

It is possible that the Hekla magma chamber is a network of interconnected patches of magma rather than a simple and more voluminous structure. However, the geochemical observations made on Hekla lavas show that the composition of products during the course of an eruption is rather uniform (Grönvold et al. 1983; Karl Grönvold, personal communication, 2003) and, thus, does not support a complicated magma-chamber structure.

The Hekla eruptions start very suddenly. Precursory seismicity develops only 25–80 min before an eruption (Einarsson and Björnsson 1976; Grönvold et al. 1983; Guðmundsson et al. 1992; Einarsson 2000; Stefánsson et al. 2000; Soosalu and Einarsson 2002). Paradoxical as it may sound, the lack of seismicity preceding eruptions can be taken as a sign of a deeper magma source. The stress change related to inflating magma chamber is distributed in a wider area and would occur aseismically until a dyke starts propagating.

The quick onset of an eruption fed from great depth is problematic. We have evidence against a voluminous magma chamber above 14 km depth. The strain signals showed 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 has to travel 14 km or more during half an hour, it requires at least a velocity of 7.8 m/s for the ascending magma. Small eruption-related earthquakes were observed to start about 80 min before the 2000 Hekla eruption, but they appear rather to be related to general adjustment of the stress field than directly to the tip of a propagating dyke (Soosalu, Einarsson and Þorbjarnardóttir, in preparation).

For comparison, earthquakes related to propagating dykes during the eruptive episode of Krafla in 1975–1984 showed various propagating velocities, mostly between 0.5 and 1.2 m/s (Einarsson 1991). Another comparable example is the Heimaey eruption in 1973 with 30 h of precursory seismicity. Earthquakes during the eruption indicated that the magma came from depth of more than 15–25 km (Einarsson and Björnsson 1979). If the precursory earthquakes can be taken as a sign of a commencing dyke, we get a propagation rate of 0.2 m/s (25 km in 30 h). Rubin (1995) points out that the transport rates depend upon the magma viscosity, and gives propagation velocities of 0.01–10 m/s for mantle-derived dykes, based on computations and laboratory experiments. The 1975–1984 eruptions at Krafla were basaltic (e.g. Brandsdóttir et al. 1997), the 1991 Hekla eruptives were basaltic andesite (Guðmundsson et al. 1992), and thus more viscous than the material at Krafla. The eruptive material in the beginning of the 1973 Heimaey eruption was alkali basaltic andesite (Jakobsson 1979), and a contact with seawater may have cooled it and made it more viscous in the uppermost couple of kilometres. One would expect Hekla magma to have lower velocities than the Krafla magma and perhaps velocities similar to the Heimaey magma.

A high propagation speed of the magma of Hekla does not sound very realistic. An explanation given by Sacks and Linde (2001; Selwyn Sacks, personal communication, 2001) for a rapid start of a Hekla eruption is that the gas phase is released from the magma inside the reservoir, and accumulated in the upper part of the reservoir, forcing 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, spouting out first the gases from the upper part of the chamber. Because the gas phase is the first one to be erupted, the eruption can easily commence more rapidly than an eruption starting with flowing magma. Linde et al. (1993) have modelled the magma chamber to be between 4 and 9 km in depth, considerably shallower than the 14 km discussed here. The gas release explanation is consistent with the observation that the Hekla eruptions begin with an explosive phase emitting gases and tephra, and subsequently calm down to erupt flowing lava (Grönvold et al. 1983; Guðmundsson et al. 1992).

Torfajökull is an extraordinary volcano for a spreading plate border region with its extensive rhyolitic volcanism. This likely explains why our magma chamber observations differ from observations made at other Icelandic volcanoes, mainly basaltic in composition. Similar to the Icelandic volcanoes with shallow magma chambers it has considerable geothermal activity, and continuous seismic activity. A shallow magma chamber was not found at Torfajökull with our seismic methods. Instead, in the west part of its caldera, we have found a cooling, but mostly solidified magma volume, much deeper (8 km) and much larger (diameter of 4 km) than observed elsewhere in Iceland. Apparently, the volume has been even larger when it started to cool. The shallow magma chambers found at Icelandic volcanoes so far are rather small structures compared to sizes of calderas and the volcano massifs themselves, approximately with dimensions of 1–2 km.

A most likely candidate for an active magma chamber at Torfajökull is below the south part of the caldera, where high-temperature geothermal activity and frequent small low-frequency earthquakes are focused. Though we did not find traces of magma with our method in the volumes that our seismic rays traversed; this area is an attractive subject for further magma chamber studies.

Conclusions

No prominent volumes of molten magma could be found at Hekla or Torfajökull in the areas we could cover with the seismic ray method. Only a tiny fraction of the seismic records with ray paths travelling under Hekla and/or Torfajökull showed attenuated S-waves. Our observations suggest that if Hekla has a substantial magma chamber it has to be located either in the uppermost 4–5 km, which is not supported by other geophysical measurements, or below about 14-km depth.

Torfajökull with its large caldera and fairly recent eruptive activity is a promising candidate for having a considerable magma reservoir. We did not find evidence for large volumes of molten material anywhere beneath Torfajökull. However, with these data, we cannot exclude the existence of a shallow magma chamber in the south or east part of Torfajökull. We have detected a large aseismic volume in the west part of the caldera and interpreted it to be a cooling magma chamber. Small low-frequency earthquakes occur persistently in the south part of the caldera and may reflect active magma there. These events appear to be deeper than 6 km, but the depth resolution for them is not very good.

Some seismic records showed a low-frequency wave package between the P- and S-waves. These were observed from a handful of earthquakes around the cooling magma chamber of Torfajökull, and may have been produced in connection with patches of molten magma.

A few seismograms had an anomalously long P-wavetrain. After the direct P multiple secondary P-arrivals are recorded by the seismograph station. These observations indicate scattering around heterogeneities, either at their origin around the Torfajökull cooling magma chamber and middle-east Hekla and Vatnafjöll, or along the ray path, e.g. in the south part of the Torfajökull caldera.

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