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Faroes-Iceland Ridge Experiment
2. Crustal structure of the Krafla central volcano

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Abstract

The seismic velocity structure of the Krafla central volcano is characterized by large variations in compressional velocity. A 40 km wide high-velocity dome extends from the lower crust (11–14 km depth) beneath the volcano narrowing upward. A magma chamber sits at its top near 3 km depth. It is defined by both 0.2–0.3 s compressional wave delays and shear wave shadowing to be 2–3 km N-S, 8–10 km E-W, and 0.7–1.8 km thick. The near-surface structure (uppermost 2.5 km) of the Krafla caldera is approximately flat-lying, with only minor lateral heterogeneities. The crust beneath the magma chamber has low shear wave attenuation and anomalously high compressional and shear wave velocities. Shear waves, reflected from a 19 km deep Moho, are clearly visible for some paths through the crustal volume below the magma chamber, even though the more shallow diving S waves are severely attenuated. The midcrust beneath the shallow magma chamber cannot contain partial melt or even be at near-solidus temperatures. The Krafla central volcano plays a major role in crustal genesis along the plate boundary. The high-velocity dome, in our view, represents crust generated in and around the magma chamber, which has subsequently been advected to greater depths.

Introduction

Iceland has been built by hot spot volcanism at a slow spreading ridge. The volcanism is caused by interaction between the mantle plume and the rifting at the divergent Mid-Atlantic plate boundary. Repeated eastward jumps of the rift axes have occured over the past 25 m.y. [*White et al.,* 1995], apparently so that the rift axes remain centered over the hot spot, resulting in the present expression of the neovolcanic zones in Iceland (Figure 1). The rift zone in northern Iceland (the Northern Volcanic Zone (NVZ)) became active following an eastward jump 6-7 m.y. ago [*Sæmundsson,* 1979].

Volcanism in Iceland is confined to volcanic systems which, in many ways, correspond to the spreading segments of the Mid-Atlantic Ridge. Each volcanic system consists of a central volcano, where the productivity is highest, and a transecting fissure swarm. Geological, K/Ar dating, and paleomagnetic studies indicate that each central volcano has a life span of 0.3–1.0 m.y., after which the center of activity shifts to a new location along the plate boundary (K. Sæmundsson, personal communication, 1995). The central volcanoes are circular in outline and 20–40 km in diameter. They are spaced 20–70 km apart along the plate boundary (Figure 1). Many have a caldera, acidic volcanism and a high-temperature geothermal system associated with them [*Sæmundsson,* 1978; *Einarsson and Sæmundsson,* 1987].

Northern Volcanic Zone

The NVZ is composed of seven NNE-SSW elongated volcanic systems arranged en echelon along the plate boundary. The plate boundary is about 50 km wide in northern Iceland but broadens to approximately 80 km in central Iceland, where the three southernmost volcanic systems of the NVZ overlap (Figure 1). Fissure swarms, characterized by rifting structures such as crater rows, normal faults, and open fissures transect the central volcanoes at an azimuth perpendicular to the regional spreading direction.

Variations in petrology and topography of the NVZ represent the different evolutionary stages of individual volcanic systems. Central volcanoes with developed calderas represent the most mature volcanic systems [*Sæmundsson,* 1982]. The petrogenesis of the NVZ is reflected in the chemical composition of its tholeiitic rock suites. The rift zone volcanoes are fundamentally basaltic in composition, with relatively small volumes of more evolved rocks [*Sæmundsson,* 1978; K. Grönvold, personal communication, 1995]. Petrochemical and isotopic data indicate that the rhyolites are formed by remelting of hydrated crust [*O'Nions and Grönvold,* 1973; *Sigmarsson et al.,* 1991; *Jónasson,* 1994] and that crustal assimilation plays a significant role in the petrogenesis of the most differentiated rocks as well as for the rift zone tholeiites [*Óskarsson et al.,* 1982; *Nicholson et al.,* 1991]. In order to explain these processes, we must evoke crustal magma chambers large enough to remelt the adjacent crust and produce silicic magmas. It is thus evident that crustal magma chambers must play an important role in the crustal genesis of Iceland. Based on the mapped distribution of silicic rocks [*Jóhannesson and Sæmundsson,* 1989], one can easily point out the central volcanoes which must have, at some point in their history, hosted crustal magma chambers. The Krafla central volcano is one of them.

Volcanic activity in the NVZ during the last three centuries has been mainly in the Krafla, Askja, Bárdarbunga, and Grímsvötn volcanic systems. During this time, three rifting episodes occurred in the Askja system (1874-1876, 1921-1933, and 1961-1962) and two in the Krafla volcanic system (1724-1729 and 1974-1985). The eruptive activity during these episodes indicates that magma accumulates in small, shallow, crustal magma chambers and is intermittently injected away from the magma chamber into dike systems parallel to the ridge [*Björnsson et al.,* 1977; *Sigurdsson and Sparks,* 1978; *Sæmundsson,* 1991; *Brandsdóttir,* 1993]. The recent rifting activity in the Krafla volcanic system offered a unique opportunity to study crustal rifting and magmatism associated with a shallow crustal magma chamber situated on a rift axis [*Björnsson et al.,* 1977; *Einarsson,* 1978, 1991a,b; *Brandsdóttir and Einarsson,* 1979; *Einarsson and Brandsdóttir,* 1980; *Tryggvason,* 1980, 1984, 1986; *Grönvold,* 1988; *Björnsson,* 1985; *Brandsdóttir and Menke,* 1992; *Arnott and Foulger,* 1994a,b].

Krafla Volcanic System

The Krafla volcanic system is made up of the Krafla central volcano and an approximately 100 km long, transecting fissure swarm. The central volcano is a major eruptive center which, based on geological mapping, K/Ar dating, and paleomagnetic studies, is less than 500,000 years old (K. Sæmundsson, personal communication, 1995). The Krafla central volcano is approximately 21 km long by 17 km wide and encloses a 10 km by 7 km caldera that was formed 100,000 years ago during the last interglacial period. It has since been filled with younger hyaloclastites and basaltic lavas [*Björnsson et al.,* 1977; *Sæmundsson,* 1978, 1982, 1991]. The central volcano has a broad, relatively flat terrain (ranging from 300 to 500 m in elevation) made up of alternating basaltic lava and hyaloclastite sequences [*Ármannsson et al.,* 1987], but individual hyaloclastic table mountains and rhyolitic ridges rise up to 900 m elevation.

Two high-temperature geothermal areas occur within the Krafla volcanic system. One is located 5 km south of the Krafla caldera but the other, the NW-SE aligned, 15 km^2, Krafla-Leirhnúkur geothermal field, is located inside the Krafla caldera. The eastern part of the Krafla-Leirhnúkur geothermal field is presently utilized by the Krafla power plant, which started operation in 1978.

The 1974-1985 rifting activity in the Krafla volcanic system consisted of long periods of inflation, during which magma accumulated at a shallow depth within the caldera region, and short deflation periods when the magma was intruded laterally into the associated fissure swarm or was erupted at the surface [*Tryggvason,* 1984; *Björnsson,* 1985; *Einarsson,* 1991b]. The inflation/deflation cycles were regulated by increasing/decreasing pressure in the Krafla magma chamber and by tectonic stress at the diverging plate boundary. The inflation periods were characterized by intense seismicity in the crust above the Krafla magma chamber, whereas migrating earthquake activity accompanied the lateral dike intrusions and rifting in the Krafla fissure swarm [*Brandsdóttir and Einarsson,* 1979; *Einarsson and Brandsdóttir,* 1980]. During 1974-1985, twenty inflation/deflation cycles took place in the Krafla volcanic system, activating most of the 100-km-long fissure swarm [*Björnsson,* 1985]. Nine of these deflations were accompanied by basaltic fissure eruptions. The last deflation/eruption event took place in September 1984 after which the magma chamber inflated normally during the succeeding 3 months. The inflation halted in early 1985. Brief periods of inflation occurred in 1986-1989, during which increased seismic activity in the uppermost 3 km around the center of inflation was generated by the increasing differential stress in the crust above the magma chamber. Currently, the seismic activity (less than one microearthquake per day) is mostly confined to the geothermal regions.

Einarsson [1978] used local earthquakes recorded during an inflation period in order to delineate two regions of high shear wave attenuation. He inferred that there is a shallow crustal magma reservoir with an upper boundary at a depth of approximately 3 km, near the center of inflation in the caldera. This magma chamber is smaller than the caldera, about 2 km by 7 km, with the long axis oriented E-W and divided near its top. *Einarsson's* [1978] shear wave data did not precisely define the lower boundary of the magma chamber but constrained it to be no deeper than 7 km.

Brandsdóttir and Menke [1992] performed waveform studies on earthquakes which originated in the Krafla caldera during a brief inflation period in July 1988 in order to study the thickness of the magma chamber. They identified reflections from beneath the magma chamber which indicated that it was less than 1 km thick in the central northern part of the caldera, but they could not constrain its lateral dimensions. Geodetic measurements support the existence of a shallow magma chamber at a depth of 3 km within the caldera and have been used to argue for the existence of multiple magma reservoirs at depth [*Tryggvason,* 1986].

Arnott and Foulger [1994a] used local earthquakes and artificial sources recorded at a local array of seismic stations to construct a coarse three-dimensional velocity model for the shallow (<3 km) structure of the Krafla central volcano. Their model is strongly laterally heterogenous, with high-velocity bodies, which they interpret as shallow crystalline intrusives, located around the caldera rim. However, the refraction data we present below do not detect these high-velocity bodies. Instead, they provide strong evidence that the shallow structure of the caldera, or at least the parts of it crossed by our lines, is relatively flat lying.

This paper uses data from three different seismic surveys at the Krafla central volcano: The 1991 Tjörnes Fracture Zone (TFZ) experiment [*Sturkell et al.,* 1993], the 1993 Krafla reflection/refraction survey, and the 1994 Krafla undershooting experiment, which was a part of the Faroes-Iceland Ridge Experiment (FIRE) [*White et al.,* 1996; *Staples et al.,* this issue]. It also uses some data recorded by a seismic array operated in 1989 around the Askja central volcano.

The undershooting experiment, which is the focus of this paper, was aimed at mapping the seismic structure of the Krafla central volcano and its immediate vicinity in order to further our understanding of the processes of melt extraction from a mantle plume and its emplacement within the crust at a slow spreading ridge. The experiment targeted the lower part of the Krafla magma chamber and the crust-mantle boundary beneath the neovolcanic zone. Our data also constrain the thickness and volume of the Krafla magma chamber and the temperature structure of the Icelandic crust.

Data

The 1994 Krafla undershooting experiment deployed 38 seismic recorders with three configurations of instruments: 18 Reftek model 72A-06 recorders with Mark Products model L22D geophones; 16 Scintrex model PRS-4 recorders with L22D geophones; and four Scintrex model PRS-4 recorders with Sprengnether (S-6000) 2-Hz geophones. The same PRS4/-L22D instruments were used in the 1991, 1993, and 1994 Krafla surveys and the 1989 Askja array.

The Reftek instruments recorded continuously at 50 samples/s and were timed by Global Positioning System (GPS) clocks. The PRS4 instruments recorded only timed windows at 100 samples/s; their quartz crystal clocks were synchronized frequently using the 60 kHz radio time signal from Rugby, England. All geophones had three components, with a flat response to ground velocity above their 2 Hz corner frequency. Further information about the general performance of L22-D geophones is given by *Menke et al.* [1991].

Seismic signals were generated by detonating explosive charges on the bottom of lakes and on the seafloor around the coast of northern and eastern Iceland (Table 1), using portable clocks which were calibrated frequently against the Rugby time code. The seismometers were installed on solid hyaloclastite or lava outcrops and covered with piles of sand and gravel for reliable ground coupling. The travel times were hand picked from a computer screen, with an accuracy of ±0.01–0.12 s for first arrivals and ±0.02–0.15 s for later arrivals.

Layout of the 1989, 1991, 1993, and 1994 Surveys

Two refraction profiles were shot across the caldera (Figure 1). They were oriented E-W and N-S and crossed at the 1724 explosion crater Víti. The profiles were measured twice, at high resolution within the caldera during the 1993 Krafla reflection survey and at a somewhat coarser resolution across the entire NVZ during the 1994 FIRE.

The 1994 survey extended 20 km N-S and 55 km E-W, with a station spacing of 500 m within the caldera and 1–4 km elsewhere. The E-W profile was surveyed from six shot points, two to the west of the caldera (Eyjafjördur and Skjálfandi), three to the east (Sænautavatn, Lögurinn, and Reydarfjördur), and one within it (Víti) [*White et al.,* 1996; *Staples et al.,* this issue].

The 1993 survey had a station spacing of approximately 200 m. It was conducted during late winter, on frozen ground. The unreversed N-S profile extended 4 km north from a shot point at Víti. The E-W profile was reversed and extended 9 km west from a shot point in Víti to a shot point at Gæsadalur at its western end. It included a third shot point at Leirhnúkur.

We also use data from a sideline of the 1991 Tjörnes Fracture Zone study [*Sturkell et al.,* 1993] which consisted of a single NNE-SSW line with nine stations, located at an average interval of 2.5 km along the western part of the Krafla fissure swarm. A single shot in Öxarfjördur, about 65 km north of the profile, is used here.

One earthquake in Öxarfjördur (70 km north of Krafla) that was recorded in 1989 by an array of seven seismometers near the Askja central volcano (75 km south of Krafla) is also of interest because the propagation path crosses Krafla.

General Data Reduction and Analysis

Primary data reduction, carried out during the fieldwork, consisted of computer filing individual shots and earthquakes along with all relevant station, source, and timing information.

All data were transcribed from their native format into the Lamont-Doherty Earth Observatory's Ad Hoc (AH) format. Timing errors and geophone misorientations were corrected at this stage, and the waveform data were rotated into vertical, radial, and transverse components. At each processing stage, the data were plotted as record sections and examined carefully for evidence of errors.

Arrival times were picked by eye from computer displays of the waveform data for all observed seismic phases. Band-pass filtering was used occasionally to improve the signal-to-noise ratio of the waveform data. However, the picks were always checked against the original seismograms to guard against timing errors introduced by the filtering process.

Upper Structure of the Krafla Caldera From Seismic Refraction

The 1993 and 1994 reflection/refraction surveys were the first high-resolution refraction experiments to be conducted within the Krafla caldera. The close (200 m) station spacing and near-shot (100 m) station locations provide new data on the shallowest (<400 m) structures which older experiments could not resolve. The 1993 profiles generally constrain the compressional wave velocity down to a depth of about 2 km. The 10-km-long ray paths turn at that depth, within the roof of the Krafla magma chamber.

The E-W profile crosses both of *Einarsson's* [1978] shear wave attenuation zones, as well as the inflation/-deflation center, which was active during the Krafla rifting episode [*Tryggvason,* 1980]. The N-S profile traverses the eastern shear wave attenuation zone. Both profiles overlap the Krafla geothermal well fields [*Ármannsson et al.,* 1987] and extend well beyond the caldera rims.

Ármannsson et al. [1987] describe the subsurface lithology and tectonics of the Krafla-Leirhnúkur geothermal field, based on 24 borehole sections. The borehole cross sections provide fundamental information about the lithology of the uppermost 2200 m of the crust along a 2-km-long section of the E-W profile just east of Víti and along a 3-km-long section of the N-S profile south of Víti. The subsurface formations can be divided into two major sections which most likely represent the overall crustal structure of the Krafla caldera: an upper, extrusive section and a lower, intrusive section. This lithological information is useful in interpreting the seismic velocities.

The observed travel times along the reversed 1993 E-W profile (Figure 2a) from the Gæsadalur and Víti shot sites are almost identical in the 2.5–8.0 km range. The overall structure is flat-lying, with no systematic dip of structures. Only subdued lateral heterogeneities are present: a small local deviation from the average (calculated) travel time curve appears 2.5 km east of the Gæsadalur shot point (circles in Figure 2a) where the P wave from Víti is advanced 0.1 s relative to the Gæsadalur profile. This difference is most likely due to slightly higher-velocity material in the uppermost 1 km within the caldera than at comparable depths outside it. Small (0.1 s) systematic travel time differences are also observed between the N-S and E-W profiles. For instance, waves on ray paths parallel to the Krafla fissure swarm south of Víti (circles in Figure 2b) are about 12% faster than those west of Víti, along ray paths perpendicular to the fissure swarm. This difference might possibly be due to seismic anisotropy caused by the shallow cracks and fissures, since it is similar in magnitude to anisotropy identified in the shallow crust of the Western Volcanic Zone (WVZ) [*Menke et al.,* 1994].

Extremely low apparent P wave velocities of about 1.1 km/s are observed through the porous, recent lava flows adjacent to the shot sites. Elsewhere in Iceland, near-surface compressional velocities range from 1.6 to 2.0 km/s in the neovolcanic zones and increase progressively with age to 3.0–4.7 km/s in the Tertiary regions [*Flóvenz and Gunnarsson,* 1991; *Staples et al.,* this issue].

Velocity Models of the Uppermost Crust Within the Krafla Caldera

Two P wave shadow zones are evident on the E-W profile, within the Krafla caldera, at ranges of 3 km and 5 km (Figures 3a, 3b and 4a). The 3-km shadow zone is visible from all the 1993 shots, but the 5-km shadow zone is seen only on two. It is not visible from the central shot point (Lh, on Figure 1), which has poor signal-to-noise ratios at these ranges. Travel time analyses, supplemented by full-waveform synthetic seismograms (Figures 3 and 4), indicate that these shadow zones are caused by two thin, approximately flat-lying, low-velocity layers that occur at depths of about 1300 m and 1700 m below the surface of the caldera (Figure 2c). The vertical travel time of 0.08–0.10 s through these low-velocity layers is tightly constrained by the travel time offset. Their thickness trades off to some degree with the velocity. Our best fit gives a thickness of 240 m for the shallower layer, with a compressional velocity of 3.0 km/s. Calculated thicknesses greater than 375 m (using a P velocity of 4.2 km/s) result in unacceptably little overlap in the travel time branches. Hence a thickness of 375 m should be considered as an upper bound. The lower layer has similar characteristics. The compressional velocity ranges from 5.9 to 6.3 km/s just above and below the low-velocity layers. The borehole data indicate that the low-velocity zones are within the extrusive layers. However, they do not seem to correspond to any specific lithologies *Ármannsson et al.* [1987] have identified within the extrusive sequence. The deeper 6.3 km/s material corresponds to the intrusives (which are mostly gabbros).

The 3-km shadow zone is absent on the N-S profile, north of Víti (Figures 2b and 5c), suggesting that the low-velocity layers are not continuous across the entire caldera. Although no lower seated low-velocity layers are detected elsewhere along the N-S profile, the station spacing along that profile is probably too large to completely rule them out.

The low-velocity zones at 1300 and 1700 m depth do not exhibit strong shear wave attenuation associated with them and therefore cannot represent layers of magma. Indeed, a comparison of the observed shear wave (Figure 4a) with full waveform synthetic seismograms indicate that the shear wave amplitudes are more consistent with a shear wave quality factor of 300 within the layers (typical of solid rocks), than with a quality factor of 5 (typical of magma). The travel time of the S wave tracks closely that of the P wave (Figure 3b), with V_p/V_s=1.76, which is also diagnostic of a solid material. Thus the very low-velocity is probably due to high porosities. *Wyllie et al.'s* [1958] formula, which is a good fit to measurements from Tertiary rocks at Reydarfjördur [*Christensen and Wilkens,* 1982], can be used to estimate porosity if the compressional velocity of the rock matrix is known. Using a matrix velocity appropriate for basaltic intrusives (6.25 km/s) and a fluid velocity appropriate for liquid water (1.5 km/s), we infer a porosity of 15–34% within the low-velocity layer. However, *Stefánsson* [1981] argues that the steam fraction in the lower zone of the Krafla geothermal field is 10–20%, an effect which would lower our porosity estimates to 12–27%. The estimated porosity in the upper zone of the Krafla geothermal field, between 200 and 1100 m depth, is 15% [*Stefánsson,* 1981]. The main aquifers of the Leirbotnar geothermal well field, SSW of Víti, follow the hyaloclastite/basalt boundaries at 400 m and 800 m depth as well as the upper boundary of a granophyre intrusion at 1900–2100 m depth, whereas the main aquifers of the Sudurhlídar system SE of Víti are associated with horizontal acidic intrusions at 900–1200 m depth, [*Ármannsson et al.,* 1987]. Hence it is plausible that the low-velocity layers we observe at 1300 m and 1700 m depth are associated with similar high-porosity, steam- and water-charged geothermal aquifers. Their apparent absence in the northern part of the caldera may reflect regional variations in the aquifers, as contrasted to variations in lithology.

Comparison to Arnott and Foulger's Model of Krafla

Our model of the compressional velocity structure in the uppermost 2.5 km of the Krafla caldera differs in two fundamental ways from the model proposed by *Arnott and Foulger* [1994a]. Their model has higher near-surface velocities and is more laterally heterogeneous than ours. Both of these features of their model imply features in the corresponding travel time curves that are not observed in our data (Figure 5). Their model predicts travel times that are as much as 0.3 s less than we observe, especially for the shorter ranges (<3 km). This difference arises because *Arnott and Foulger* [1994a] constrain the upper 0.5 km of their model to have relatively high near-surface velocities (>3 km/s), much higher than we observe. Their model also predicts large differences (0.1–0.3 s) in the Víti and Gæsadalur travel time curves, out to a distance of about 10 km. This difference arises because of a large-scale horizontal gradient in velocities in their model, increasing toward the west (i.e., Víti is updip, whereas Gæsadalur is downdip). The smaller velocity heterogeneities also cause 0.1–0.2 s undulations in the travel time curves with wavelengths of a few kilometers, which are not evident in our travel time data. Thus we conclude that *Arnott and Foulger* [1994a] have overestimated the amount of lateral heterogenity present in the upper 2 km of the Krafla caldera, or at least the parts of it crossed by our lines. The structure of the uppermost 2 km of the caldera is relatively flat-lying, consistent with the borehole data [*Ármannson et al.,* 1987].

Structure of the Northern Volcanic Zone and the Krafla Central Volcano as Determined Through Seismic Undershooting

Some seismic work was conducted in the NVZ prior to our experiment [*Pálmason,* 1963; *Gebrande et al,* 1980; *Zverev et al.,* 1980]. Although some of these profiles lie on the flanks of central volcanoes, none of the older profiles crosses the center of a volcanic system.

In the discussion below we rely mostly on the 1994, N-S and E-W profiles, which cross the central part of the Krafla caldera and intersect near the Víti crater. We use sources (shots and earthquakes) which ranged in distance from 50 km to 250 km and have P and S waves with computed ray paths that dive deep (>10 km) into the crust, cutting across the upper caldera at steep (< 45°) angles of incidence. By mapping the lower crustal structure and identifying crustal reflections across the Krafla central volcano we can examine in detail a section of the Mid-Atlantic plate boundary in northeastern Iceland.

P and PmP Waves

Clear P waves are observed from all FIRE shots, which are to the north, northwest, south, east, and west of the Krafla caldera. The P wave apparent velocity increases from about 6.5 to 7.0 km/s as the waves cross the flank of the caldera (meaning a 10–20 km wide band surrounding the caldera rim) (Figure 6). Stations located on the caldera rims have the largest travel time anomaly, an advance of 0.1–0.3 s. This feature is due to a a high-velocity dome beneath the caldera (discussed further below). A pronounced decrease in apparent velocity, to 6.2 km/s, and a significant reduction in amplitude (by a factor of 5–10) occur as the rays cross the caldera (Figure 6). Secondary arrivals, interpreted as due to a shallow magma chamber (discussed further below), are observed from all azimuths. The caldera rim anomaly is especially clear at a station 65 km from the Öxarfjördur shotsite on the N-S line just before the onset of low P wave amplitudes, which are confined to the southern part of the caldera (Figure 6a). The secondary arrival appears on this profile at a range of 66.6 km.

The region of amplitude reduction exhibits a parallax effect. Its position on the N-S profile (Figures 6a and 6b) from a shot in the north differs by about 2 km from its position from a shot in the south; and its position on the E-W profile (Figures 6c and 6d) from a shot in the west differs by about 1.5 km from its position from a shot in the east. This parallax implies that the source of the amplitude reduction must be at several kilometers depth and cannot be due to the shallower geothermal fields. We discuss more precise depth estimates below.

We observe wide-angle reflections from all the shot sites outside the Krafla region (Figure 7, see also *Staples et al.* [this issue]). We interpret the reflecting interface as the Moho and the reflections as PmP because *White et al.* [1996] observe clear Pn phases, with apparent velocities of 7.6–7.8 km/s, from the continuation of this reflector in eastern Iceland. We also observe Pn-like apparent velocities of 8.1±0.2 km/s from an Öxarfjördur earthquake observed on the 1989 seismic array near the Askja central volcano (Figure 8). Waves traveling along this path have calculated P and PmP ray paths that bottom beneath the Krafla caldera at about 10 and 20 km depth, respectively.

A clear PmP phase is also observed on the the 1991 N-S profile, which recorded a shot in Öxarfjördur (Figure 9). Travel times along the neovolcanic zone are consistent with the cross-axes crustal model of *Staples et al.* [this issue], which has a Moho that rises from 25 km depth at the western edge of the NVZ, to 19 km beneath Krafla, and then descends to 35 km depth in eastern Iceland. The P wave in this profile, which is southwest of the center of the caldera, has no significant amplitude or travel time anomalies. This confirms our assumption that those anomalies are caused by structures near the center of the caldera and not along more distant parts of the propagation path.

S and SmS Waves

After generating record sections from each shot site we were able to identify shear phases from most of the detonations. Although hardly any S energy was observed on the vertical component record section from individual shot sites, clear S phases occur on the radial-horizontal component record section from those same shot sites. This indicates that the shear wave energy is generated by P to S_V conversion at the shot site. Still, we were not able to identify shear phases from the Sænautavatn shot, recorded on the E-W profile, nor from the Askja and Öxarfjördur shots, recorded on the N-S profile. We attribute the lack of S energy from these shots to shot site conditions because we observe S waves along their ray paths from other sites. For instance, although we do not observe shear waves from the 1994 Öxarfjördur shot, we see S waves propagating along the neovolcanic zone from earthquakes in the same region, such as the one included in this study. We also record S waves from the Langjökull earthquake at some stations on the N-S profile. Why S waves were not generated in lake Sænautavatn and the Askja caldera lake remains unexplained, although in the latter case strong attenuation from the Askja magma chamber [*Brandsdóttir et al.,* 1992] may be a factor.

The S wave travel times for the Eyjafjördur, Skjálfandi (Figure 10), and Lögurinn shot sites and the Öxarfjördur earthquake are all consistent with V_p/V_s in the 1.76–1.79 range, at least for distances out to those where the waves are attenuated under the Krafla caldera. Ratios in the range of 1.9–2.0 are ruled out for these distances, because they predict S arrivals several seconds later than is observed.

The E-W profile crosses both of the shear wave attenuation zones defined within the Krafla caldera by *Einarsson* [1978], and the N-S profile crosses the eastern attenuation zone. Again, we see large variations in shear wave velocities and energy within the caldera, coinciding with the observed P wave variations (Figure 11). The shear wave energy is fairly constant at the flank stations, whereas drastic changes occur between individual caldera stations.

We observe clear SmS phases within the neovolcanic zone from the Skjálfandi (Figure 10), Eyjafjördur, and Reydarfjördur shots along the E-W profile and from the earthquake in Öxarfjördur, along the N-S profile (Figure 12a). Our estimates of V_p/V_s ratios for SmS are variable. V_p/V_s=1.76 and V_p/V_s=1.79 are good estimates for Skjálfandi and Reydarfjördur, respectively. On the other hand, we find V_p/V_s=1.88 for arrivals from the Öxarfjördur shot. The two events have quite different SmS reflection points, Skjálfandi's being west of the Northern Volcanic Zone and Öxarfjördur's being within it. Thus the difference in ratios might plausibly be related to higher temperatures in the lower crust beneath the Northern Volcanic Zone. However, this interpretation should be considered speculative until confirmed with further observations.

Crustal Structure Across the Krafla Central Volcano

We have created two-dimensional N-S (Figure 13a) and E-W (Figure 13b) velocity models across the Krafla central volcano, using the *Caress et al.* [1992] RTMOD forward modeling program and an iterative, trial and error approach to fitting travel times. The compressional velocity is represented by continuous linear splines on a triangular mesh. We began with a smooth, two-dimensional structure based on a combination of our travel times, the previously published crustal models of *Pálmason* [1963] and *Flóvenz and Gunnarsson* [1991] and an unpublished north Iceland average crustal model, NOR, developed by the Science Institute of the University of Iceland. We iteratively perturbed this starting model until we were able to fit the small travel time anomalies. The observed P arrivals are fitted to better than 0.1 s except in the vicinity of the Krafla magma chamber where we have misfits of 0.1–0.2 s caused by insufficient station coverage within the Krafla caldera (Figure 13c). The ray coverage, which is derived from four shot points on the E-W profile and two shot points and one earthquake on the N-S profile, enables us to resolve the crustal structure of the uppermost 10–15 km of the Krafla central volcano and to determine the crustal thickness at different midpoints from individual sources.

Our final profiles fit the overall travel times to within an rms error of 0.05 s. The model has three main features:

1. A broad, high-velocity dome, or "chimney" characterizes the substructure of the volcano. Under the central volcano, the 6.5 km/s isovelocity surface (*Pálmason's* [1963] layer 3) rises up to 2.6 km depth in a narrowing chimney. The base of the chimney, defined by the 6.7 km/s isovelocity surface, is about 33 km wide E-W and 37 km wide N-S but narrows toward the surface. The shape of this chimney is constrained by the regional P wave travel time data, which have a 15–20 km wide, 0.1–0.3 s advance centered over the caldera (Figures 6, 10, and 14). An advance of 0.15 s in P wave travel time is detectable 10–30 km east of Víti, for the western shots (Figure 14). At these ranges, waves from Eyjafjördur and Skjálfandi shot points have computed ray paths that pass through the chimney at 8-10 km depth. The velocity anomaly associated with the chimney extends at least to these depths.

2. The Krafla magma chamber sits at the top of the high-velocity dome with its top at a depth of 3 km. It is a 2–6 km wide, 0.75–1.8 km thick lens of low velocities, with a nominal velocity of 3 km/s. We discuss the specific features of the data that constrain these dimensions below.

3. The thickness of the crust decreases toward the Krafla central volcano where it has a minimum thickness of 19 km [see also *Staples et al.,* this issue].

Two-Dimensional Finite Difference Seismograms of Krafla

Our compressional velocity model of Krafla (Figure 13) is based on travel times alone. In this section we test whether the model correcly predicts the observed pattern of amplitudes.

We compute synthetic seismograms using a 2-D elastic finite difference code. The P velocity is defined in a 20 km wide by 10 km deep box centered around Krafla, with a grid spacing of 17.4 m. The S velocity is computed from the P velocity according to V_p=1.76 V_s, except in the magma chamber where it is assumed to be 1 km/s. The density is computed by $\rho = 1.85 + 0.165\ V_p$. The upper surface of this box is a free surface, while the sides and bottom are energy absorbing. A line of receivers is placed along the top, and a plane compressional wave is created near the bottom, with a 4-Hz Ricker wavelet time function. The elastic wave equation is then solved with an implicit finite difference technique, with a time step of

1 ms.

The first case we examine is for a model containing a high-velocity dome but no magma chamber. The 0.2-s advance in first arrival travel times is clearly visible, both for normal (Figure 15a) and oblique (Figure 15b) incidence of the plane wave. In the oblique case, low amplitudes are observed on the side of the model away from the incident direction. This effect, also observed in the data, is caused by defocusing of the waves by the chimney structure. In this model, no secondary arivals are evident.

We then added a 1-km-thick magma chamber to the model. A prominant secondary arrival, delayed 0.2–0.3 s from the first arrival, and corresponding to P waves that pass through the magma chamber is visible in both the normal (Figure 15c) and oblique (Figure 15d) incidence cases. Amplitudes in the defocused region also decrease.

The case of oblique incidence with a magma chamber (Figure 15d) compares favorably with the data (Figures 6 and 12b). The overall pattern of amplitudes of both the first and second arrival are matched well.

In the oblique incidence cases, the edge of the region of high amplitudes associated with the chimney and the center of the region of secondary arrivals are both shifted by about 2 km to the right of the actual positions of the edge of the chimney and the center of the magma chamber. This parallax would decrease if the magma chamber were to be placed shallower in the crust. Thus any observed parallax can be used to constrain the depth of the top of the chimney and magma chamber.

Dimensions of the Magma Chamber

The size of the region of intense shear wave attenuation can be used as a measure of the lateral dimension of the magma chamber. We use the ratio of amplitudes of P and S waves to quantify this attenuation (Figure 16), and find that the zone of high attenuation (or "shadow") is 2–3 km wide in the N-S direction and 8–10 km wide in the E-W direction. These dimensions agree with the oblong shape determined by *Einarsson* [1978].

Some parallax is visible in the N-S data, with the center of the shadow at 2.5 km south of the Víti crater for incidence from the north and about 1.5 km north of Víti for incidence from the southwest. Thus the center of the 2-km-wide (measured north-south) magma chamber is located about 0.5 km south of Víti. The 2-km displacement of the center of the shadow from the center of the magma chamber is consistent with the 2-km shift of features determined by the finite difference modeling, giving a depth of about 3 km for the top of the magma chamber.

The thickness of the magma chamber can be estimated by the observed 0.2–0.3 s delay experienced by P waves that cross the structure (Figure 12b). Its thickness is 0.75–1.8 km, assuming that the magma velocity is 3 km/s and the background velocity is 5–6 km/s. Its volume (assuming a tabular shape) is therefore 12–54 km^3.

As mentioned above, the edges of the region of P wave attenuation, which is due to defocusing due to the chimney, also exhibit a parallax of about 2 km, indicating that the top of this feature is also at a depth of 3 km.

Krafla Structure Summary

1. The near-surface (0–2 km) structure of the Krafla caldera is approximately flat-lying, with only subdued lateral heterogeneities. Two thin (240–375 m) layers of low compressional velocity occur west of the Víti crater at depths of 1300 and 1700 m. They are interpreted as regions of high porosity related to geothermal aquifers.

2. The top of the Krafla magma chamber lies at a depth of 3 km. The magma chamber has a width of 2–3 km in the N-S direction and 8–10 km in the E-W direction and a thickness of 0.75–1.8 km. A zone of high S wave attenuation is associated with the magma chamber. However, this zone does not extend down to Moho, as SmS phases, which cross the central volcano at lower crustal depths, are detected. No low-velocity zones are detected in the midcrust elsewhere within the neovolcanic zone, contrary to predictions based on the interpretation of high midcrustal electrical conductivities [*Beblo and Björnsson,* 1980] which would place partial melt at these depths.

3. A broad, high-velocity dome (chimney), rising from the lower crust toward the magma chamber, characterizes the substructure of the Krafla central volcano. The general structure of the chimney supports the crustal accretion models proposed by *Menke and Sparks* [1995], in which crust is created at shallow depths in the volcanic zone and then subsides, by solid-state creep, to greater depths.

4. The shear velocities in the crust, as deduced from S and SmS travel times, are consistent with V_p/V_s=1.76–1.79, with two exceptions: shear waves

do not propagate through the shallow magma chamber beneath Krafla and *SmS* travel times for paths along the NVZ, north of Krafla, are consistent with V_p/V_s=1.88, which may imply higher temperatures in the lower crust there.

5. The upper crust (V_p <6.5 km/s) is thinner in the NVZ than in the older, adjoining regions. This effect has been noted previously by *Pálmason* [1963] for the NVZ and by *Bjarnason et al.* [1993] for the WVZ. *Pálmason* [1971] found the shallowest depth to the isovelocity surface 6.3 km/s associated with volcanic centers and suggested that the depth contours of this surface may be used to map volcanic centers. Our study supports this hypothesis.

Discussion and Conclusions

There has been little consensus on the nature of *Pálmason's* [1963] layer 3 (P3) in Iceland. *Flóvenz* [1980] equated layer P3 to layer 3 of the oceanic crust. Others associated P3 with a metamorphic facies of basaltic rocks [*Pálmason,* 1971] or with low-porosity basalts with high contents of epidote [*Christensen and Wilkins,* 1982; *Flóvenz and Gunnarsson,* 1991], rather than a heterogeneous mixture of basaltic intrusives and extrusives [*Bödvarsson and Walker,* 1964] or a sheeted dike complex [*Walker,* 1975].

Walker's [1975] interpretation of the nature of layer P3 was inferred from Tertiary central volcanoes in eastern and southeastern Iceland. His model of an intense regional swarm of intrusive sheets is consistent with its oceanic counterpart, which has been interpreted to consist of dikes (layer 2C, V_p=6.1 km/s) and intrusive gabbros (layer 3A, V_p=6.8 km/s) [*Harrison and Bonatti,* 1981; *White et al.,* 1992]. Seismic velocities near 6.8 km/s (in oceanic layers 2B-3 range), have been drilled at Ocean Drilling Program (ODP) hole 504b at a depth of 2060 m below seafloor [*Alt et al.,* 1993]. Although the Iceland Research Drilling Project (IRDP) hole in Reydarfjördur did not penetrate layer P3, it penetrated a number of dikes with a mean compressional velocity of 6.02±0.26 km/s, whereas the lava flows had a mean compressional velocity of 5.67±0.65km/s [*Christensen and Wilkins,* 1982]. The dike intensity in the IRDPhole varied between 31 and 44% of total rock volume [*Robinson et al.,* 1982]. Our dense refraction profile and the lithological cross sections of the Krafla geothermal fields, where gabbro is found at 2 km depth, confirm that *Bödvarsson and Walker's* [1964] interpretation of the composition of layer P3 was correct. A *P* velocity of 6.25 km/s has to be associated with basaltic intrusives (dikes) at shallow depths within a central volcano. However, outside the central volcanoes, the *P* velocity of 6.25 km/s can be ascribed to all the above mentioned compositional variations.

Except for the Krafla magma chamber itself, the crust in this part of the NVZ contains very little melt. No low-velocity anomalies, which might be indicative of partial melt, are identified in the midcrust. Instead, this region is characterized by faster than normal compressional and shear wave velocities. This result casts further doubt on the old model of the structure of Iceland, which hypothesized pervasive partial melt beneath the volcanic zones at 10–15 km depth.

Reexamination of data from the South Iceland Seismic Tomography (SIST) profile [*Bjarnason et al.,* 1993], which crosses the (WVZ) about 10 km north of the Hengill central volcano in southwest Iceland shows that the *S* wave travel time is well-predicted by V_p/V_s=1.79 out to ranges of at least 120 km, and no zone of attenuation is detected [*Menke et al.,* 1996]. This result also supports the idea that zones of attenuation are concentrated beneath central volcanoes and do not extend significantly along the plate boundary. Reviewing older data, we cannot find any seismic measurements that support the thin, hot crust model. In Iceland, as elsewhere, compressional velocites of 7.0–7.4 km/s should be attributed to the crust.

Our identification of Moho-reflected shear waves (*SmS*) from many shot sites in northeast Iceland is further support for *Menke et al.'s* [1996] contention that *Gebrande et al.* [1980] misidentified *SmS* as *S*. Their high V_p/V_s ratios are thus artifacts of comparing the travel times of two phases, *P* and *SmS*, with significantly different ray paths. Our study finds normal V_p/V_s ratios for both diving (*P* and *S*) and reflected (*PmP* and *SmS*) phases.

Our model of the crustal structure supports the theory that the Krafla central volcano plays a major role in crustal genesis along the Mid-Atlantic plate boundary in northeastern Iceland. The central volcano and its related fissure swarm define a volcanic segment similar to those identified on the Mid-Atlantic Ridge south of Iceland [*Appelgate and Shor,* 1994]. We interpret the high-velocity chimney to represent crust, generated in and around the magma chamber itself, which has subsequently been advected to greater depths such as has been proposed by *Menke and Sparks* [1995]. This is similar to the model proposed by *Henstock et al.* [1993] for generating the entire normal oceanic crustal section from high-level sill-like magma chambers. Two important classes of questions arise from this scenario:

First, the way in which our snapshot of the volcano evolves over geologic time needs explanation. Before the Mid-Atlantic plate boundary jumped eastward, a broad depression formed at the present position of the Northern Volcanic Zone, at 6–10 Ma [*Sæmundsson,* 1995]. Immediately after the jump, volcanic activity seems to have begun in the southern part of the NVZ, near Vatnajökull, and migrated north, until a mature rift zone was formed. How Krafla fits into this scenario is unclear. K. Sæmundsson (personal communication, 1995) gives 0.5 Ma as an upper bound for its age, so approximately 5 Myr of geological history are unaccounted for. Certainly the 33 km E-W width of the base of the high-velocity chimney would seem to imply that this region had been a major volcanic center for far longer (1.8 Ma, based on a 9 mm/yr [*DeMets et al.,* 1994] half-spreading rate, if we assume that the volcanic activity started with a single dike in the center of the present caldera). The future of the Krafla central volcano is also worthy of speculation. Is its structure stable? Or does the presence of the high-velocity chimney interfere with upward melt transport, so that the formation of a new central volcano, at a different location along the plate boundary, will be favored in the near future?

Second, we have detected no feature in the seismic data that is related to melt transport through the lower crust. Nevertheless, such transport necessarily must occur. The Krafla magma chamber seems unconnected to the mantle, the zone of high shear wave attenuation, which we take as a proxy for temperature, seems restricted to the midcrust and above. Admittedly, our resolution at these depths is poor, and there may be structures that we have missed. On the other hand, the lack of such melt transport structures is consistent with the observation that melt transport takes place episodically. The recurrance time of Krafla rifting episodes is of the order of 200–1000 years and is a manifestation of intermittent melt injection from the mantle. What crustal conduit does melt take during these injection events?

To conclude, our model of crustal structure of the Northern Volcanic Zone in general, and the Krafla central volcano in particular, reenforces the model of Iceland as a slow spreading oceanic ridge. Iceland falls at the extreme end of ridges, to be sure. Its crustal thickness is 3–5 times that of "normal" oceanic crust; it has permanent magma chambers not found on the Mid-Atlantic Ridge itself. On the other hand, the overall pattern of crustal structure and the processes that create it closely follow the oceanic norm.

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Figure 1. (Opposite left) Map of the Krafla central volcano showing the 1991, 1993, and 1994 seismic stations (small circles), the Gæsadalur (G), Leirhnúkur (Lh) and Víti (V) shot sites (large circles) and geologic features. (bottom right). Interpretation of geologic features. (top right) Map of Iceland showing the seismic stations of the 1994 FIRE line (triangles), shot sites (stars), and earthquakes (squares). Langjökull earthquake (LANG); Eyjafjördur (E); Skjálfandi (SK); Öxarfjördur 1991 and 1994 shot sites and earthquake (ÖX); Víti, Krafla; Askja; Sænautavatn (S); Lögurinn (L); and Reydarfjördur (R). Central volcanoes (darker shaded patches) and associated fissure swarms (shaded lines) in the neovolcanic zones of Iceland are also shown. Bárdarbunga (Bb), Grímsvötn (G). Basemap after *Einarsson and Sæmundsson* [1987].

Figure 2. Travel times and 1-D velocity model for shallow (<3 km) structure across the Krafla central volcano. (a) Reversed 1993 E-W profile with shots at Gæsadalur (circles) and Víti (crosses). (b) 1993 profile of stations north (crosses) and south (circles) of the Víti shot point. Note the similarity of travel times for all shots, indicating that only subdued lateral heterogeneities are present. Two shadow zones are visible in the E-W data. (c) One-dimensional velocity model based on the reversed E-W profile. Low-velocity zones correspond to shadow zones in the data.

Figure 3. Vertical component, reduced record sections of a shot in Víti observed on the 1993 E-W line. (a) Trace-normalized plot showing first arrival picks (grey lines). Note the two shadow zones at 3 and 5 km range due to low-velocity zones. (b) Same as Figure 3a, except plot has been section-normalized, with a gain function that is quadratic in distance. (c) Synthetic seismograms computed using an integral transform method and the 1-D velocity model in Figure 2c. The plot is section-normalized, with the same gain function as in Figure 3b. Phases corresponding to those in Figures 3a and 3b are shown as grey lines. Note that the shadow zones are clearly visible in the synthetics, and the variation of amplitude with distance is similar to the data in Figure 3b.

Table 1. Shot Locations Used in This Study

Date	Day	Time, UT	Latitude	Longitude	Water Depth, m	Charge, kg	Location
Aug. 10, 1991	222	1930:00.108	66°15.294'N	16°35.435'W	159	101.8	E35, Öxarfjördur.*
Aug. 13, 1991	225	0030:00.200	65°43.067'N	16°45.462'W	40	100	Víti, Krafla*
April 9, 1993	099	2330:00.000	65°43.067'N	16°45.462'W	12	10	Víti, Krafla
April 10, 1993	100	1600:00.015	65°43.067'N	16°45.462'W	5	5	Víti, Krafla
April 10, 1993	100	1700:00.017	65°42.94'N	16°48.03'W	12	10	Leirhnúkur*
April 10, 1993	100	1800:00.019	65°44.262'N	16°55.992'W	5	20	Gæsadalur*
April 11, 1993	101	1100:00.005	65°43.067'N	16°45.462'W	8	15	Víti, Krafla
April 11, 1993	101	1200:00.006	65°42.94'N	16°48.03'W	12	15	Leirhnúkur*
April 11, 1993	101	1300:00.008	65°44.262'N	16°55.992'W	5	15	Gæsadalur*
April 11, 1993	101	1600:00.014	65°44.262'N	16°55.992'W	5	10	Gæsadalur*
April 11, 1993	101	1700:00.016	65°42.94'N	16°48.03'W	12	15	Leirhnúkur*
April 11, 1993	101	1900:00.019	65°43.067'N	16°45.462'W	40	30	Víti, Krafla*
July 27, 1994	208	1101:00.00	65°43.067'N	16°45.462'W	40	50	Víti, Krafla*
July 28, 1994	209	1101:00.00	65°43.067'N	16°45.462'W	40	7.2	Víti, Krafla*
July 29, 1994	210	0801:00.01	66°16.774'N	16°29.740'W	50	100	Öxarfjördur*
July 30, 1994	211	1601:00.01	65°2.597'N	16°46.155'W	50	100	Askja
Aug. 2, 1994	214	2301:00.00	65°0.993'N	14°10.602'W	59	200	Reydarfjördur*
Aug. 3, 1994	215	1401:00.01	65°5.987'N	14°45.260'W	50	150	Lögurinn*
Aug. 3, 1994	215	2201:00.01	65°16.668'N	15°31.410'W	23	100	Sænautavatn*
Aug. 4, 1994	216	1401:00.00	65°43.067'N	16°45.462'W	40	93	Víti, Krafla*
Aug. 5, 1994	217	2201:00.00	66°01.595'N	17°38.543'W	54	200	Skjálfandi*
Aug. 6, 1994	218	2001:04.00	65°56.783'N	18°15.373'W	64	200	Eyjafjördur*

* Shot detonated on the bottom.

Figure 4. Radial-horizontal component, reduced record sections of a shot in Gæsadalur observed on the 1993 E-W line. (a) Section-normalized plot, where a gain function that is quadratic in distance has been applied. Note the clear S wave. Two travel time curves are shown (grey lines). The lower one is computed from the P wave travel times using the assumption that V_p/V_s=1.78. The top curve is for V_p/V_s=2.0. The bottom curve is a better fit. (b) Synthetic seismograms computed using an integral transform method and the 1-D model described in the text (with V_p/V_s=1.78). The same normalization and gain function is used as in Figure 4a. Phases corresponding to those in Figure 4a are shown (grey curves). The travel time and variation in amplitude with range is similar to what is observed in Figure 4a. (c) Same as Figure 4b, except for an Earth model that assumes melt in the low-velocity zones (V_s=1 km/s, Q_s=5). S wave amplitudes decline more rapidly with distance than is observed in the data.

Figure 5. P wave travel times for N-S and E-W profiles crossing the Krafla central volcano. Data (squares), predictions of our model (solid lines), and predictions from *Arnott and Foulger's* [1994a] model (dotted lines) are shown. The dotted curve was calculated by ray tracing through Arnott and Foulger's published model. Corrections for station elevation above mean sea level (the top of Arnott and Foulger's model) were made using formula provided by G. Foulger (personal communication, 1995). The locations of the Gæsadalur (G), Leirhnúkur (L), and Víti (V) shot sites are marked. (a) Central part of E-W profile. (b) Complete E-W profile. (c) N-S profile.

Figure 6. Vertical component record sections, reduced to 6.5 km/s, showing P wave undershooting the Krafla central volcano from the (a) north, (b) south, (c) west, and (d) east. The position of the Víti crater (V) and caldera rims (R) are indicated. Amplitudes have been section-normalized. The four sections are similar, indicating that the heterogeneity beneath Krafla has a rough cylindrical symmetry. See text for further discussion.

Figure 7. Vertical component record section, reduced to 6.5 km/s and band-passed 1–4 Hz, from the Skjálfandi (SKJA), Víti, and Reydarfjördur (REYD) shot sites. Amplitudes have been trace-normalized. The P and PmP phases are clearly visible.

Figure 8. Vertical component record section, reduced to 7 km/s, showing the P wave of a small earthquake in Öxarfjördur observed near the Askja central volcano. The timescale is arbitrary. Amplitudes have been trace-normalized. The waves cross the Krafla region, which is about halfway between Öxarfjördur and Askja, at a depth of 20–30 km. Note the high apparent velocity of 8.1±0.1 km/s (grey line), which suggests that the P wave has turned in the mantle.

Figure 9. Vertical component record section, reduced to 6.5 km/s, of a shot in Öxarfjördur observed on the 1991 N-S line. Amplitudes have been section-normalized. The northern end of this line is in the western part of the Krafla caldera, with the rest of the line extending south of it. Clear P and PmP phases are observed. No P wave delay or attenuation is observed, because the line does not cross the Krafla magma chamber.

Figure 10. P, PmP, S, and SmS waves from the Skjálfandi shot observed on the 1994 E-W line. (a) Radial-horizontal component record section, reduced to 4 km/s. (b) vertical component record section, reduced to 6.5 km/s. Amplitudes are section-normalized. The position of the Víti crater (V) and caldera rims (R) are indicated. The P and PmP wave picks (shaded curves in Figure 10b), after being scaled by a factor of 1.76 (grey curves in Figure 10a), are a good estimate of the S and SmS arrival times.

Figure 11. Radial-horizontal component record sections, reduced to 4 km/s, of two shots observed on 1994 E-W profile. Amplitudes are section-normalized. (a) Eyjafjördur shot to the west of Krafla. (b) Lögurinn shot to east of Krafla. The position of the Víti crater (V) and caldera rims (R) are indicated. Note that in both record sections the S wave is strongly attenuated as it crosses the center of the caldera.

Figure 12. P and S waves from an earthquake in Öxarfjördur undershooting the Krafla volcano. Positions of Víti crater (V) and caldera rims (R) are indicated. (a) Section-normalized record section of the radial component, reduced to 4 km/s. Note the S wave, which is shadowed as it crosses the magma chamber lying beneath central part of the caldera, and the SmS phase. (b) Section-normalized record section of the vertical component, reduced to 6.5 km/s. Note the phase labeled P_2, which has traveled through the magma chamber, and the PmP phase. (c) Finite difference synthetic seismograms, shown as a section-normalized record section reduced to 6.5 km/s, calculated using the 2-D model of Krafla discussed in the text. Compare with Figure 12b.

Figure 13a. Crustal structure of the Krafla central volcano. A two-dimensional, compressional velocity model derived from travel time data from the E-W profile, structure across the plate boundary. The shaded region on top of the high-velocity dome denotes the magma chamber.

Figure 13b. Structure along the plate boundary. Only the part of the N-S model beneath the central volcano itself is well-resolved as the length of our N-S profile is only 22 km (Figure 1).

Figure 13c. Observed (circles) and calculated (grey curves) travel times corresponding to the model.

Figure 14. P wave travel time anomalies, for the part of the 1994 E-W line that crosses the Krafla caldera from the western shot sites, Eyjafjördur (EYJA) and Skjálfandi (SKJA), relative to an average travel time curve constructed from the shot sites east of Krafla. The positions of the Víti crater (V) and caldera rims (R) are shown. The maximum advance in travel times of 0.2–0.3 s, due to the high-velocity chimney, occurs within the caldera. A smaller advance of 0.1–0.2 s extends to 30–35 km east of Víti. See text for further discussion.

Figure 15. Synthetic seismograms of P waves undershooting the Krafla central volcano, computed using a two-dimensional, elastic finite difference method. (a) Normal incidence, model has chimney only. (b) Oblique incidence from left, chimney only. (c) Normal incidence, model has both chimney and magma chamber. (d) Oblique incidence from left, both chimney and magma chamber. The location of the top of the chimney (grey bar) and magma chamber (solid bar) are shown. The broad 0.2-s advance in travel times caused by the chimney, the diminished amplitudes on the positive distances caused by defocusing of rays by the chimney, and the localized 0.2–0.3 s delay caused by the magma chamber are all clearly visible.

Figure 16. Amplitude ratios of P and S waves undershooting Krafla from the (a) north, (b) southwest, (c) east, and (d) west. The positions of the Víti crater (V) and caldera rims (R) are shown. Peaks in this ratio occur when the S wave experiences strong attenuation, believed to be due to the magma chamber.