Along-Strike Seismic Structure of the Northern Volcanic Zone, Iceland

William Menke (1,2), Michael West (1,2), Bryndís Brandsdóttir (3) and David Sparks (1)

(1)Lamont-Doherty Earth Observatory, Palisades NY USA
(2)also Department of Earth and Environmental Sciences, Columbia University
(3)Science Insitute, University of Iceland, Reykjavík, Iceland
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SUMMARY

Seismic data from the B96 array in northern Iceland are used to constrain the compressional and shear velocity structure of the Northern Volcanic Zone (NVZ) (the mid-Atlantic plate boundary in northern Iceland) and its western rift flank. Traveltimes from P waves crossing the NVZ from a fan shot in eastern Iceland demonstrate that the structure of the mid-crust is dominated by the Krafla high-velocity dome. Neighboring central volcanoes have less prominent anomalies, suggesting that they have a more subsidiary role in the formation of the lower crust. Traveltimes from P and S waves from shots and microearthquakes north and south of the array are used to constrain the along-strike velocity structure of the western rift flank. Two dome structures are detected at depths down to 7 km. They may be the fossil roots of extinct central volcanoes. P, S, PmP and SmS wave traveltimes indicate that the compressional to shear wave velocity ratio in the crust is 1.75-1.76, with no significant variation detected between the mid and lower crust. The crust is 25-31 km thick, with the southward thickening occuring in an abrupt step. The relatively high Pn and Sn wave apparent velocities of 8.00 +/- 0.1 km/s and 4.31 +/- 0.04 km/s, respectively, from an earthquake in southern Iceland, are consistent with a "mantle lid", that is, a layer of sub-solidus mantle separating the Moho from a deeper, partial melt zone.

INTRODUCTION

The Northern Volcanic Zone (NVZ) is the present-day expression of the divergent mid-Atlantic plate boundary in northern Iceland (Figs. 1 and 2). It is composed of an en echelon series of volcanic systems, each containing a central volcano and an associated fissure swarm (Sæmundsson 1974, Einarsson and Sæmundsson 1987). The northern part of the NVZ, about which this study is concerned, consists of three such volcanic systems: Þeistareykir, Krafla and Fremri-Námur (from north to south), with 20-40 km spacing. Most of the volcanic rocks are tholeiitic basalts (Jakobsson 1972). Minor amounts of acidic rocks, such as dacites and rhyolites, are associated with crustal remelting at the central volcanoes (O'Nions and Grönvold 1973, Sigmarsson et al. 1991).

The NVZ began to develop 6-7 Ma ago, when rifting and volcanism jumped eastward from the older Húnaflói-Skagi Volcanic Zone (HSVZ) (Sæmundsson 1974). The shift in rifting activity from the HSVZ to the NVZ seems to have been abrupt as the NVZ has widened by 100-120 km during the last 6-7 Ma (Sæmundsson 1995), implying an average spreading rate similar to the present full-rate of 18 mm/year (DeMets et al., 1994). The Tjörnes Fracture Zone (TFZ) was, however, activated at least 2 m.y. prior to the rift jump (Young et al. 1985) during which pro-rifting central volcanoes, were active in the region between two volcanic zones (Jancin et al. 1985).

The B96 seismic array that we discuss in this study is located on the western flank of the NVZ (Fig. 2). It strikes north-south, parallel to the rift axis. The surface outcrop of the unconformity between rocks erupted at the HSVZ and the NVZ is about 10-15 km west of the array. The unconformity has a shallow (10-25 deg) dip to the southeast (Jancin et al. 1985, Sæmundsson 1995), but it is probably interrupted at depth by several southeast- or eastward dipping normal faults (Fig. 3). The B96 array sits on basalts that were erupted from the NVZ at 0-3 Ma

(Jóhannesson and Sæmundsson 1989). However, pro-rifting NVZ rocks (Jancin et al. 1985) and possibly HSVZ-derived rocks in the deep crust probably extend a few km east of the array (Fig. 3).

The 1994 Faroes-Iceland Ridge Experiment (FIRE) (White et al. 1996, Staples et al. 1997, Brandsdóttir et al. 1997) includes an east-west transect of the NVZ through the Krafla central volcano to eastern Iceland. The center of the Krafla caldera is located approximately 40 km east of the B96 array (Fig. 3). A dramatic change in crustal thickness is observed along the transect, from 25 km in the western (0-7 Ma) part, to 19 km beneath Krafla (0-1 Ma), to 35 km in the Tertiary eastern fjords. Staples et al. (1997) argue that the differences in crustal thickness represent differences in overall melt productivity during three periods of spreading. The major change in topography that occurs near Bárðardalur (from HSVZ-derived and pro-rifting NVZ highlands to the west to NVZ-derived lowlands to the east, Fig. 2) is associated with this change in crustal thickness. Hence we would place the maximum eastward extent of subsurface pro-rifting NVZ and possibly HSVZ-derived rocks on the western rift flank at few kilometers to the east of the B96 array.

The central volcanoes of the NVZ are major eruptive centers that, based primarily on geological mapping, appear to be at least 0.3-0.5 m.y. old (Sæmundsson, personal communication 1995). Little is known, however, about their stability over longer periods of time, their relative productivity, or how that productivity varies with time. Krafla, arguably the best studied of them, is known to possess a shallow (3-4 km) magma chamber that is underlain by a dome with anomalously high seismic velocity (the velocity contrast is about 7% at 10 km depth). Brandsdóttir et al. (1997) invoke the theory of lower crustal solid-state flow developed by

Pálmason (1973), Henstock et al. (1991) and Phipps Morgan & Chen (1993) and interpret this dome as due to olivine-rich cumulates, that form at the base of the magma chamber an which are then subsequentally advected to downward and outward during the spreading process. Citing the dome's 33 km width, Brandsdóttir et al. (1997) argue that Krafla may be at least 1.8 m.y. old.

The purpose of this study is to examine whether the Krafla dome is a unique feature, or whether similar domes are present elsewhere along strike of the NVZ (e.g. beneath the Þeistareykir and Fremri-Námur central volcanoes). The array also provides an opportunity to explore the seismic structure of the pro-rifting NVZ and possibly HSVZ-derived material of the western rift flank.

DATA ACQUISITION

The north-south striking B96 array is about 80 km long, and mostly located along the eastern side of Bárðardalur and Kaldakinn valleys (Fig. 2). It consists of 25 three-component seismometers, spaced at approximately 3 km. All stations used GPS-timed, Refraction Technology 72A-07 digitizers. The southern station used a Streckeisen STS-2 broadband geophone, the northern used a Guralp CMG40-T broadband geophone, and the other stations used Mark Product L22-D short period (2 Hz) geophones. Data was recorded continuously at 50 samples/s. The array was operational during April-May, 1996. The data are available from the Incorporated Research Institutions for Seismology as PASSCAL dataset XE-96.

Three chemical explosions were recorded: Shot N (100 kg), in a bay just north of the array; Shot S (100 kg), in a lake just south of the array; and shot E (150 kg), 120 km away on the east coast (Fig. 2). All shots used dynamite detonated in 30 m of water. Numerous microearthquakes were also recorded. Earthquake hypocenters were determined using arrival time data from the SIL array (Icelandic Meterological Office) and the U. Iceland array.

SHALLOW COMPRESSIONAL VELOCITY PROFILE

The shallow structure beneath the array (along the strike of the rift flank) was investigated using traveltimes of P waves from the N and S shots. P waves in this distance range (0-80 km) sample the earth down to a maximum depth of about 8 km.

The P waves arrive with an apparent velocity of about 6.66 +/- 0.05 km/s in the 20-80 km distance range. This relatively high crustal velocity, typical of solid basalt (Flóvenz et al. 1985), indicates that low-porosity rocks are within a few kilometers of the surface. Traveltime flucuations about this trend have an amplitude of 0.1-0.2s and a wavelength of 20-40 km, and are strongly anti-correlated between the N and S shots (Fig. 4). The anticorrelation is due to a parallax effect, and indicates that the corresponding heterogeneity must be several km deep. (Near-surface heterogenities, such as sediment cover, would be positively correlated). A 2D velocity model was constructed to match the observed travel times using the RTMOD forward program (Caress et al. 1992) and trial-and-error fitting (Fig. 5).

The crustal model has two notable features: 1) The uppermost 1 km at the northern 20 km of the model has significantly lower velocities than rocks at corresponding depths to the south (the surface velocity is 2.4 km/s in the north and 3.7 km/s in the south). This low-velocity material is likely Holocene sediment accumulation near the coast. 2) The structure of the shallow crust is dominated by two domes, each 30-40 km wide and spaced 30 km apart, that have detectable velocity contrast down to a depth of about 7 km. The compressional velocity contrast of these dome anomalies is very large. For instance, at 6 km depth, the compressional velocity varies from 6.8 km/s in the middle of the domes to 6.4 km/s between the domes (i.e. about 6% contrast).

The domes have a similar width (25-30 km) and velocity contrast (6%) to the dome beneath Krafla, which is about 30-40 km wide in its north-south dimension and has a velocity contrast of 7% (Brandsdóttir et al. 1997). Furthermore, the 30 km spacing between the two domes is comparable to the along-strike spacing of NVZ central volcanoes. We suggest that the domes represent the fossil roots of extinct central volcanoes.

We assume that these volcanos were active during the onset of rifting, 6-7 M years ago, since the domes sit within in the oldest NVZ-derived rocks near the border faults of the NVZ's western margin. They would be structurally analogous to 7-3 Ma volcanoes situated just west of the currently active Western Volcanic Zone (WVZ), (Jóhannesson 1997). NVZ-derived, 6-5 Ma acidic rocks, associated with an (informally named) Náttfaravík central volcano occur about 10 km west of the northern end of the B96 array (Jancin et al. 1985). The northern of the two domes might possibly be associated with this volcano. Other acidic rocks outcrop 10-15 km northwest of the southern dome (Jóhannesson and Sæmundsson 1989), and might be associated with the southern dome.

FAN SHOT THROUGH THE NORTHERN VOLCANIC ZONE

Shot E, off the coast east of Iceland, samples a swath across northeastern Iceland. P-waves turn in the mid-crust (10-11 km) beneath the NVZ (Fig. 5 and 6), and cross the Þeistareykir, Krafla and northern half of the Fremri-Námur central volcanos (Fig. 7). We use traveltimes of these P waves to examine along-strike structure of the NVZ.

Traveltime anomalies can arise from heterogeneities anywhere along the path, but are most sensitive to structures near the turning point (beneath the NVZ). Structure near the source is probably only a minor contributor to the traveltime anomalies, since seismic waves depart within a narrow cone of angles, sampling more-or-less the same material. The traveltime perturbation due to upper-crustal (<5 km) heterogeneities beneath the array is more significant (0.1-0.2 s), but may be accounted for by using the previously determined shallow velocity profile. After this correction, traveltime anomalies ranging between -0.15 and 0.4 s are observed (Fig. 7).

We cannot correct for the effect of deeper structure associated with the two domes (Fig. 5) because we have little information about their eastward extent. However, since the area east of the array has been extensively rifted it is likely that the domes have been truncated within at most 15 km of the array. We performed a sensitivity test to determine the effect of different widths. We ray traced through a series of hypothetical dome structures constructed from a "dome-top" velocity profile taken from under the 30 km mark on the array (Fig. 8) and a "background" velocity taken from the 0 km mark. The test indicates that even a 15 km wide anomalous feature would not perturb the velocity by more than 0.3 seconds, half of the observed anomaly. Thus the overall shape of the fan shot is likely attributable to deep features within the NVZ.

A single fan shot cannot, of course, image the structure of the lower crust. Indeed, if the crust had a complicated spatial pattern of velocity anomalies, the traveltime pattern of a single fan shot might be impossible to interpret. Fortunately, the observed pattern is very simple, with the most significant feature being the early arrival along raypaths reaching the central parts of the array. This feature is almost certainly due to the Krafla dome (Brandsdóttir et al. (1997): the maximum traveltime advance occurs along raypaths that pass through the northwestern part of Krafla caldera; and its amplitude consistent with a ~40 km wide dome with a velocity contrast of 7%. A simple model of the high-velocity region indicates that it is 40-50 km in the north-south dimension (Figs. 9 and 10), with the ampitude of the anomaly generally decreasing both northward and southward from a maximum at the northern edge of Krafla caldera. The data do not constrain the east-west width of the feature, since width trades off with the velocity contrast. However, a velocity contrast of more than about 7% (and hence a width narrower than about 30 km) can be ruled out because it leads to unacceptably large fluctuation in P wave amplitude, which were observed to be approximately constant across the array.

Brandsdóttir et al. (1997) interpret the high mid-crustal velocities of the dome to be due to olivine-enriched basalts that formed at the base of shallow (<10 km) magma chambers and were subsequently advected downward. The strength of the observed mid-crustal velocity anomaly is therefore a proxy for the amount of cumulates that were produced in the recent past. Krafla central volcano has clearly been more productive, in this sense, than the neighboring Þeistareykir and Fremri-Námur central volcanoes (i.e. they have no distinct domes). We speculate that much of the mid crust beneath these two less productive central volcanoes may have been produced near Krafla, and subsequently advected along the axis of the NVZ.

The fan shot traveltimes also have a short-wavelength feature centered just south of the Krafla caldera (Fig. 7). Arrival there are delayed by 0.2-0.35 s with respect to the fastest paths crossing the dome, but advanced by 0.1-0.3 s with respect to those near Fremri-Námur. The velocity model has a corresponding narrow feature (Fig. 9). We cannot unambiguously attribute this feature to processes occuring at Krafla, since it is sufficiently small that it might be due to heterogeneties occuring elsewhere along the raypaths. If it is due to structure at Krafla, we speculate that this feature may be caused by fine-scale complexity in the cumulate production and advection process.

LOWER CRUSTAL STRUCTURE AND MOHO DEPTHS FROM TRAVELTIMES OF REGIONAL MICROEARTHQUAKES

Record sections of magnitude 2.5-4.0 microearthquakes occuring 50-100 km north and south of the ends of the array contain many identifiable seismic phases, including P, PP, PPP, PmP, S, SS and SmS (Fig. 11). The P and S waves sample down to the mid-crust (15 km depth) near the northern and southern ends of the array, and have apparent velocities of about 6.74 +/- 0.10 km/s and 3.82 +/- 0.05 km/s respectively. The compressional to shear wave velocity ratio of 1.765 +/- 0.036 is similar to mid-crustal values reported elsewhere in Iceland (e.g. Menke et al. 1994a, Menke et al. 1996, Staples et al. 1997). The PmP and SmS Moho-reflected phases sample down to the lowermost crust. They have a compressional to shear wave velocity ratio of 1.755 +/- 0.05, statistically indistinguishable from the mid-crustal value. No evidence for lower-crustal heating (or partial melting) is observed.

Traveltimes from microearthquakes occuring 50-100 km north and south of the ends of the array were used to extend the compressional velocity model to 200 km width and 32 km depth (Fig. 12 and 13). (The shear wave structure was not explicitly modelled, given the near-constant compressional to shear wave velocity ratio). The shallow structure could be mapped only crudely, because of the lack of stations outside the 80 km array. However, PP and PPP traveltimes provide some constraints on the shallow velocity structure. These multiple P phases indicated that the region between Vatnajökull and the southern end of the array has significantly lower near-surface (<2 km) velocities than the array region. This difference is probably due to thick NVZ-derived lava flows that occur south of the array, and is consistent with the overall pattern of near-surface compressional velocity in Iceland noted by Flóvenz et al. (1985).

The PmP and SmS arrivals tightly constrain Moho depth. Traveltimes from all but the southernmost Vatnajökull event (1110434) are consistent with a flat Moho at 25 km depth. Traveltimes from event 1110434, which samples the most southerly region of Moho, requires a depth of 31 km. We therefore model the Moho with an abrupt southward deepening at the position of approximately the southern end of the array.

We have assembled these Moho depth determinations together will all previously published data for Northern Iceland (Fig. 14). The northernmost B96 measurements and the westernmost FIRE measurements (Staples et al. 1997) agree closely. They sample Moho within 20 kilometers of one another and both give a depth of 25 km. The stepping down (to 31 km in the southeast) of the crust that occurs at the southern end of the B96 array is generally in agreement with the very thick (35 km) value reported for the NVZ just north of Vatnajökull (square symbol in Fig. 14). Yet the abruptness of the transition indicates a behavior more complicated than a simple thickening of the crust towards a region of greater magmatic productivity in the central part of the hotspot (as was suggested by Bjarnason et al. 1993). The step may possibly represent some kind of deep geological boundary. Staples et al. (1997) propose that the abrupt rise in Moho (to 19 km) observed near Krafla is due to the magmatic productivity in the northern NVZ being less than at the older HSVZ. This explanation will not work for the B86 step, since the Moho deepens - not shallows - toward the center of the southern NVZ. An important objective for future research will be to establish how short wavelength (10 km), high amplitude (5-15 km) Moho topography is created and how it is sustained over time periods of 3-6 million years.

EVIDENCE FOR MANTLE LID FROM MANTLE PHASES

Pn and Sn mantle-refracted waves were observed from earthquakes in the Mýrdalsjökull region of south Iceland (e.g. event 1171351 M=3.8, Fig. 15) and just east of the Kolbeinsey Ridge (e.g. event 1080943 M=4.0), north of Iceland (KB in Fig. 1). The path from Kolbeinsey Ridge crosses the NVZ while the path from Mýrdalsjökull crosses central Iceland just west of Vatnajökull (dashed lines in Fig. 2). Pn and Sn apparent velocities from the Kolbeinsey Ridge are 7.66 +/- 0.04 km/s and 3.96 +/- 0.03 km/s, respectively, while those from Mýrdalsjökull are about 4% higher, 8.00 +/- 0.1 km/s and 4.31 +/- 0.04 km/s, respectively. The Mýrdalsjökull Pn data are consistent with the value of 8.01 +/- 0.15 km/s reported by Menke et al. (1996) for another central Iceland path (Vatnajökull to southwest Iceland). The compressional to shear wave apparent velocity ratio is 1.93 +/- 0.01 for the Kolbeinsey event and 1.85 +/- 0.02 for the Mýrdalsjökull event.

The Mýrdalsjökull data are particularly interesting because they sample the upper mantle above the Iceland "plume anomaly" that has been studied by Wolfe et al. (1997). These authors use teleseismic S wave tomography to identify a low-velocity (by 4%) shear wave velocity anomaly in the 100-400 km depth range beneath central Iceland (shallower depths are unresolved in their inversion). At the observed ranges of 200-250 km, the Mýrdalsjökull Pn and Sn waves turn in the uppermost mantle, only a few (10-15) kilometers beneath the Moho. The rays upon which we base the apparent velocity measurements cross a flat portion of Moho beneath the array, and thus are not influenced by Moho dip. We thus take the apparent velocity measurements as a proxy for shallow mantle velocities. The measurement may be influenced by upper mantle velocity anisotropy. However, the azimuth of N20E is close to 45 degrees from the N30W shear wave high-velocity direction observed by Menke et al. (1994b) and Bjarnason et al. (1996). The velocities thus represent intermediate values.

Some insight into the thermal state of the uppermost mantle sampled by the Mýrdalsjökull Pn and Sn waves can be obtained by comparing the observed shear wave velocity (4.31 km/s) and velocity ratio (1.85) to values measured for the mantle beneath mid-ocean ridges. Nishimura and Forsyth (1989) estimate a velocity of 3.98 km/s and a ratio of 2.04 at 45 km depth for 0-4 Ma mantle beneath the East Pacific Rise (EPR). Forsyth (1992) notes that the low shear wave velocity for the EPR probably requires melt, since it is well below the 4.18 km/s value that is thought to characterize peridotite at a temperature just below its solidus. The very uppermost mantle beneath central Iceland thus seems to be considerably cooler than at a comparable depth beneath the EPR, and appears not to contain melt. The central Iceland values are more similar to those occuring in the subsolidus mantle "lid" of the 4-20 Ma EPR lithosphere, which has a shear velocity (at 45 km depth) of 4.37 and a ratio of 1.86 (Nishimura and Forsyth 1989). Thus Iceland, though it sits over a hot mantle plume, appears to have a layer of subsolidus mantle (or lid) separating the crust from the melting region. This result is consistent with mantle melting models (Menke and Sparks 1995), which place the top of melting about 25 km below Moho.

The observed Kolbeinsey Ridge shear wave velocity of 3.96 km/s is in close agreement with the 3.9 km/s value measured previously by Evans and Sacks (1980) from surface wave data, and probably requires partial melt. The top of melting appears to be at shallower depths north of Iceland than near the center of the Iceland mantle plume, within the NVZ.

RESULTS

1. Mid-crustal (10-11 km) rocks beneath the axis of the NVZ have anomalously high velocities

(by as much as 7%), with the anomaly being concentrated beneath the northern part of Krafla central volcano. This feature is associated with the Krafla dome, previously interpreted by Brandsdóttir et al. (1997) as caused by shallow magma chamber-derived olivine cumulates, that are subsequently advected into the lower crust. The high-velocity anomaly is weaker beneath the neighboring Þeistareykir and Fremri-Námur central volcanoes. Krafla may therefore be a more significant source of lower crustal rocks than its neighbors.

2. Two dome structures, similar in size and velocity contrast to the Krafla dome, occur on the western flank of the NVZ. These domes may be the fossil roots of extinct central volcanoes that were active prior to NVZ rifting, or at its onset, at 6-7 M years ago.

3. The crust along the western rift flank of the NVZ is 25-31 km thick, with the southward thickening occuring as an abrupt step. The compressional to shear wave velocity ratio is about 1.76 throughout the crust, including the region just above the Moho. Abrupt changes in Moho depth in northern Iceland have also been observed on the FIRE profile (Staples et al. 1997).

4. Pn velocities of 8.0 +/- 0.1 km, and Pn/Sn apparent velocity ratio of 1.85 +/- 0.02, for a path that crosses central Iceland, imply that the uppermost mantle is sub-solidus. This result is consistent with Menke and Sparks (1996) prediction of a mantle "lid" above the zone of partial melt.

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Figure Captions

Fig. 1. Map of Iceland, showing the three major brances of the Neovolcanic Zones (WVZ, EVZ and NVZ), locations of seismometers of B96 array (triangles), central volcanoes (circular outlines), fissure swarms (line segments), glaciers (shaded regions), and shot (squares) and earthquake (stars) locations. Letters indicate place names, KB (Kolbeinsey ridge), RR (Reykjanes ridge), TFZ (Tjörnes Fracture Zone), WF (western fjords), EF (eastern fjords), H (Húnaflói bay), S (Skagi peninsula); HSVZ (the extinct Húnaflói-Skagi Volcanic Zone).

Fig. 2. Map of eastern Iceland. The B96 array (triangles) is situated in the Bárðardalur (BD) and Kaldakinn (KK) valleys, immediately to the west of the Þeistareykir (Þ), Krafla (KR) and Fremri-Námur (FN) central volcanoes. Other major volcanoes within the NVZ and EVZ are Askja (A), the subglacial hotspot systems of Kverkfjöll (KV), Bárðarbunga (BB) and Grímsvötn

(G), Mýrdalsjökull (MJ), and the extinct Náttfaravík (Na) central volcano. Shotsites (N, S, and E) and earthquakes (stars) are also shown.

Fig. 3. Cross-section of the crust across the western flank of NVZ. Known (solid) and hypothetical (dashed) geological boundaries are shown. See text for further discussion.

Fig. 4. Traveltimes, reduced to 6.5 km/s, of the N (triangles) and S (crosses) shots. Traveltimes were picked "by eye" from high quality seismograms. Solid curves are predicted traveltimes for the velocity model in Figure 5.

Fig. 5. 2-D compressional velocity model, along a north-south profile extending from Íshólsvatn (south) to Skjálfandi (north), based on traveltime data from these shots.

Fig. 6. Predicted ray path from E shot to a typical station (labeled stn) of the B96 array. The ray passes beneath the center of the NZV at 10.6 km depth.

Fig. 7. Fan shot corrected traveltime anomalies (bar graph) and ray paths (dotted lines) portrayed on a map of northeastern Iceland. See text for further discussion.

Fig. 8. Effect of dome width is estimated by calculating traveltime anomaly associated with domes of varying widths. (left) Ray from a surface source at 120 km arrives at a station at the origin after sampling a dome with vertical sides. (right). Traveltime anomaly as a function of dome half width, for two shapes of dome: dome with sloping sides (triangles), dome with vertical sides (circles).

Fig. 9. Compressional velocity anomalies along the ray surface of the fan shot (shaded, with contours), with superimposed geographical features. This surface is 10-11 km deep beneath the

NVZ.

Fig. 10. Fan shot corrected traveltime anomalies (circles). Solid curves are predicted traveltimes for the velocity model in Fig. 9.

Fig. 11. (Bottom) Low-pass filtered, vertical component record section, reduced to 6.5 km/s, of Vatnajökull earthquake 1270028 observed on the B96 array. Note clear P, PmP, PP, and PPP phases. (Top) Unfiltered transverse-horizontal component record section, reduced to 3.8 km/s. Note clear S and SmS phases. Amplitudes are section-normalized.

Fig. 12. 2-D model of compressional velocity along a north-south profile extending from Vatnajökull to northern Iceland, based on traveltimes of P, PmP, PP and PPP phases. The Moho declines from a depth of about 25 km in the north to 31 km in the south.

Fig. 13. Traveltime data (triangles) for five regional earthquakes, and corresponding prediction based on velocity model in Fig. 12. Epicenters are shown in Figs. 1 and 2. The small (0.2 s) static anomaly in event 1260156, is most likely caused by a small error in the hypocentral location.

Fig. 14. Map of crustal thickness measurements (symbols) in northwestern Iceland, and speculative 5 km contours (curves). Fissure swarms (line), central volcanoes (circular outlines), glaciers (shaded), and selected crustal thickness measurements in km (symbols). Symbol type refers to data sources: (square) Our reinterpretation of Gebrande et al.'s [1980] Shot D; (star) Menke et al.'s [1996] reinterpretation of Gebrande et al.'s [1980] Shot E; (triangles) Staples et al. [1997]; and (diamonds) B96 results. Note concerning (square): Shot D has a second arrival that we, following Bjarnason et al. (1993), identify as PmP. It has a traveltime, reduced to 7.0 km/s, of 4.0 s at 100 km and 3.1 s at 150 km, which corresponds to a 35 km deep Moho, assuming a

crustal velocity structure similar to our regional model (Fig. 12).

Fig. 15. (Bottom) Low-pass filtered, vertical component record section, reduced to 7.5 km/s, of Mýrdalsjökull earthquake 1171351 observed on the B96 array. The central part of the array was not in operation at the time of this event. Note clear P and Pn phases. (Top) Low-pass filtered transverse-horizontal component record section, reduced to 4.0 km/s. Note clear S and SmS phases. Amplitudes are section-normalized.