TABLE 4: CRUSTAL MODELS

1973-model		1993-model		EW-1995		GP-Icel. 1963		KATLA-94	
depth (km)	P-vel. (km/s)	depth (km)	P-vel. (km/s)	depth (km)	P-vel. (km/s)	depth (km)	P-vel. (km/s)	depth (km)	P-vel. (km/s)
0.0	2.4	0.0	1.1	0.00	1.20	0.0	2.8 (0)	0.00	3.30
0.5	3.8	0.1	2.0	0.26	2.67	0.7	4.2 (1)	0.50	4.00
2.5	5.2	0.2	3.0	0.60	4.10	2.0	5.1 (2)	1.50	5.00
5.0	6.5	1.0	4.5	0.80	4.84	3.65	6.3 (3)	4.50	6.00
		2.0	5.5	1.00	5.40				
		5.0	6.5	1.24	5.93				
				1.28	3.07				
				1.44	2.98				
				1.48	6.16				
				1.60	6.27				
				1.65	3.00				
				1.78	3.02				
				1.82	6.33				
				2.59	6.46				

icant dipping layer boundaries or large velocity anomalies occur at less than 2 km depth. However, small, local deviations from the average (calculated) traveltime curve occur. One such deviation appears 2.5 km east of the Gæsadalur shotpoint (filled circles in Figure 6a), where the P wave is advanced 0.1 s relative to the reversed profile. This anomaly is most likely caused by higher velocity material (intrusives) within the caldera than at comparable depths outside of it. Small, (0.1 s) systematic traveltime differences are also observed along the profiles. For instance, traveltimes south of Víti, on raypaths parallel to the Krafla fault swarm (filled circles in Figure 6b) are about 12% faster than those west of Víti, i.e., along raypaths perpendicular to the fault swarm.

Extremely low apparent P-wave velocities of about 1.1 km/s are observed adjacent to the shotsites. The uppermost raypaths from the Víti shotsite travel through a fumarole field and the Víti 1724 tephra layer, whereas rays from Gæsadalur go through postglacial, basaltic lava flows. The Víti tephra is over 100 m thick next to Víti [Sæmundsson, 1991] but thins rapidly to about 1 m at a distance of 1 km, north and west of Víti. The extremely low P-velocity is thus associated with the unconsolidated tephra and porous recent lava flows adjacent to the shotsites. Elsewhere in Iceland, near-surface compressional velocities range from 1.6-2.0 km/s in the Neovolcanic zones, increasing progressively with age, to 3.0-4.7 km/s in the Tertiary regions [Flóvenz and Gunnarsson, 1991]. Thus, it cannot be considered abnormal to find P-wave velocites as low as 1.1 km/s in surface layers within central volcanoes.

We first constructed a 1-D regional model, KRA93, which provided an overall fit to the traveltime data (Figure 7, Table 4). We used the subroutine TTGEN from the earthquake location program HYPOINVERSE [Klein, 1978] to calculate traveltimes. The KRA93-model fits the data considerably better than the older, KRA73-model, with one exception. The older model has a better fit to the 1994, NS-line data at ranges less than 6 km in Hlídardalur, south of Víti (Figure 6). However, both 1-D models provide very low-resolution representation of the crustal structure. A detailed examination of the record sections reveal fine-scale velocity fluctuations which are presented below.

Two P-wave shadow zones are evident on the EW profile, within the Krafla caldera. They occur at a range of 3.0 km and 5.0 km from the Víti shotsite (Figure 8). The upper shadow zone is visible on all three record sections but the lower shadow zone is not visible from the central (Leirhnúkur) shotpoint, which has poor signal-to-noise ratios at these ranges. The shadow zones are caused by two, thin, approximately flat-lying, low-velocity layers that occur at a depth of about 1300 and 1700 m (bold curve in Figure 7, EW-model in Table 4). The vertical traveltime of 0.08-0.10s through these low-velocity layers is tightly constrained by the traveltime offset. Their thickness trades off with velocity to some degree. Our best fit gives a thickness of 240 m for the shallower layer, with a compressional velocity of 3.0 km/s. Thicknesses greater than 375 m (with a velocity of 4.2 km/s) result in unacceptably little overlap in the traveltime branches. Therefore, a thickness of 375m should be considered as an upper bound. The lower layer has similar characteristics. The compressional velocity, just above and below the low velocity layers, ranges from 5.9-6.3 km/s.

The upper shadow zone is clearly absent on the NS profile, north of Víti. Although no lower seated shadow zones are detected elsewhere along the NS profile, either, the station spacing along that profile is probably too coarse to completely rule them out. In any case, observed differences between the two profiles indicate some degree of lateral heterogenity, in the uppermost 2 km within the Krafla caldera.

In order to improve the resolution of the shallow caldera structures, we created two-dimensional, NS (Figure 9) and EW (Figure 10) velocity models, using the Caress et al. [1992] RTMOD forward modelling program and a iterative, trial-and-error approach. The compressional velocity is represented in the model by continuous linear splines on a triangular mesh. We iteratively perturbed a starting, 1-D model until we are able to fit the small traveltime anomalies (Figure 11). The ray coverage, which is derived from three shotpoints on the EW profile but only a single shot on the NS profile, is not sufficient to uniquely determine the two dimensional structure. We can only claim that our final models give a general sense of the locations and amplitudes of the lateral heterogeneities.

Two main features appear on our caldera profiles: 1) Pálmason's [1963, 1971] Layer 3 (isovelocity surface 6.5 km/s) ascends beneath the caldera, where is reaches a minimum depth of 1.6 km and; 2) Two low-velocity zones, observed along the EW-profile sit above the high-velocity dome. The minimum depth (1.6 km) of Layer P-3 which we observe beneath Krafla is 2 km less than Pálmason's [1963] estimate. Comparing our 2-D, crustal model of the Krafla caldera to the regional (Mývatn) crustal structure [Pálmason, 1963] (Table 4) we find that our depth to the 6.3 km/s isovelocity surface is also considerably shallower than the >4.5 km depth which Gudmundsson et al. [1994] observed beneath the Katla caldera in S-Iceland. However, it is not odd to find Layer P-3 at a shallower depth within central volcanoes [Pálmason, 1971; Flóvenz, 1980], which could explain the different observations in the Mývatn region. The Katla volcano is situated in the propagating (non-rifting) part of the EVZ. The Layer P-2 thickness is exceptionally high in that region with the Layer P2/P-3 boundary at a depth of 7 [Pálmason, 1971] to 9 km [Flóvenz and Gunnarsson, 1991]. The crustal structure of Gudmundsson et al's [1994] off-rift volcano is therefore not directly comparable to the structure of the Krafla rift zone volcano.

Pálmason's [1971] data showed that the shallowest depth to Layer P-3 is associated with volcanic centers which appear as local high-velocity anomalies (domes) in an otherwise layered crust. Flóvenz [1980], reinterpreting Pálmason's data, reported much higher linear velocity-depth gradients within the central volcanoes and estimated the depth to the Layer P-2/P-3 boundary could be as little as 1 km. Flóvenz [1980] called the central volcanoes, referring to their high-velocity doming, "chimneys through the Icelandic crust". He associated them with Pálmason's Layer 3 which he equated to Layer 3 in the oceanic crust. Others, associated Layer P-3 with metamorphic facies of basaltic rocks [Pálmason, 1971], and low-porosity basalts with high contents of epidote [Christensen and Wilkins, 1982; Flóvenz and Gunnarsson, 1991] instead of a heterogeneous mixture of basaltic intrusives and extrusives as [Bödvarsson and Walker, 1964] or sheeted dike complex [Walker; 1975]. Walker's [1975] interpretation of the nature of Layer P-3, being made up of 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]. The composition of Oceanic Layer 2C has been confirmed by drilling. Although the IRDP-hole in Reydarfjördur did not penetrate Layer P-3, 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±0.65 km/s [Christensen and Wilkins, 1982]. The dike intensity in the IRDP-hole varied between 31 and 44% [Robinson et al., 1982]. Our dense refraction profile and the lithological cross sections of the Krafla geothermal fields confirm that Bödvarsson and Walker's [1964] interpretation of the composition of Layer 3 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 Layer 3 velocity can be ascribed to all the above mentioned, compositional variations.

The 1300 and 1700 m deep low velocity zones are not associated with shear wave shadow zones and therefore cannot represent layers of melt. Indeed, the traveltime of the S wave, which is clearly present from the Gæsadalur shotpoint, tracks that of the P wave very closely, with V_p/V_s=1.76 (Figure 12). 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 200-1100 m deep, upper zone of the Krafla geothermal field is 15% [Stefánsson, 1981] and the average porosity measured in the Reydarfjördur borehole is 9% [Jónsson and Stefánsson, 1982]. The main aquifers of the Leirbotnar geothermal field follow hyaloclastite/basalt boundaries at 400 and 800 m depth and the upper boundary of a granophyre intrusion at 1900-2100 m depth. The main aquifers of the Sudurhlídar system are connected to lateral, acid intrusions at 900-1200 m depth [Ármannsson et al., 1987]. Hence, it is plausible that the low velocity layers we observe at 1300 and 1700 m depth are associated with similar high-porosity, geothermal aquifers.

3.1 Comparison to Arnott and Foulger's model of Krafla

Our model of the compressional velocity structure in the uppermost 2 km of the Krafla caldera differs in two fundamental ways from the model proposed by Arnott and Foulger [1994a]. Their model has much higher near-surface velocities and is more laterally heterogeneous than ours. Neither of these features are observed in our data (Figure 11, dotted lines). Their model predicts traveltimes that are as much as 0.3 s lower than we observe. Their predicted traveltime curves generate much stronger, small-wavelength traveltime anomalies than we observe, e.g., the anomaly they predict at the western caldera rim (their Figure 7b), at a distance of 7-8 km on Figure 11C. Arnott and Foulger [1994a] do not discuss how velocity, depth of hypocenters, and origin time of earthquakes trade off in their inversion process. We find no way of reconciling their model with the observed traveltime data. Therefore, we conclude that their model is, unfortunately, seriously flawed.

We believe that the flaw is related to their underlying methodology, which is based on arrival time data from local microearthquakes, and which employs a simultaneous inversion for hypocentral parameters and velocity structure. Given optimum ray coverage such an inversion can indeed determine both the location and origin time of earthquakes and the three-dimensional velocity structure of the earth. Unfortunately, the spatial distribution of earthquakes in their dataset which are mainly confined to the caldera center (near Leirhnúkur), is far from optimum. Under such circumstances, velocity structure can trade off with the hypocentral depth and origin time.

This effect can be understood by considering how arrival time data are used to distinguish a deep earthquake from a shallow one. If the earth is laterally homogeneous (e.g., as at a distance of 0 to 10 km in Figure 13), then the traveltime curves from these two earthquakes differ in two ways: 1) The deeper earthquake has faster apparent velocities at short ranges; and 2) The deeper earthquake has longer traveltimes. Note, however, that the difference in traveltime is meaningful only when the origin time of the earthquakes is accurately known. If the earth is made laterally heterogeneous, with the velocity increasing with distance from the hypocenter, then the apparent velocities of the two traveltime curves become much more similar (as at a distance of 10 to 20 km in Figure 13). The difference in traveltime remains, but it can be traded off with origin time to produce identical arrival times. Thus, in our example, a 3 km deep earthquake in a laterally homogeneous earth has the same arrival time as a 1.5 km deep earthquake in a laterally heterogeneous earth (Figure 14). If widely separated hypocenters were available, then the ambiguity could be resolved, since one cannot make the velocity increase with distance from all the earthquakes simultaneously. But if they are clustered in one region, velocity can trade off with hypocenter depth.

Another point to consider is that Arnott and Foulger's [1994a] high-velocity anomalies at shallow depth will generate Bouguer gravity anomalies close to 10 mGals, using the velocity-density relationship from Christensen and Wilkens [1982], wheras the anomalies observed in the Krafla region do not exceed 5 mGals [Karlsdóttir et al., 1978], (Arnott and Foulger [1994a], Figures 3 and 7, and plate 1). As we discussed earlier, borehole cross sections show that two fairly uniform crustal sections exist within the Krafla caldera, above 2 km depth and that gabbroic intrusives are only found below 2000 m depth [Ármannsson et al., 1987].

Therefore, we believe that the high velocity anomalies that Arnott and Foulger [1994a] propose for the caldera rims, as well as their shallow (1-2 km) hypocentral depths, are artifacts of their inversion methodology.

4 Structure of the Northern Volcanic Zone and the Krafla magma chamber as determined through seismic undershooting

Mapping the regional seismic structure of the Northern Volcanic Zone (NVZ) and individual volcanic systems is a fundamental prerequisite for understanding the dynamics of crustal genesis along the divergent plate boundary, as well as the mechanism of melt extraction from the mantle plume into individual volcanic systems along the slow-spreading plate boundary.

Some seismic work has been conducted in the Northern Volcanic Zone prior to our experiment [Pálmason, 1963, 1971; Gebrande et al; 1980; Zverev et al., 1980]. Pálmason [1963] measured 11 short (<60 km) profiles within the NVZ. His profiles 13 (Kelduhverfi), 15 (Hólasandur), 16 (Mývatn-Jökulsá), and 19 (Grímstadir-Axarfjördur) are 24.2 km, 18.3 km, 29.4 km, and 31.5 km long, respectively. Therefore, they only sample the uppermost 5 km of the crust. Pálmason's profiles are located 65-70 km north, 10-20 km west, 8-12 km south and 30-50 km east of the Krafla caldera. Zverev [1980] later surveyed a reflection profile along profile 16. The RRISP-profile I runs along the NVZ where it crosses the northern flank of the Tungnafellsjökull and Bárdarbunga central volcanoes and the southern flank of the Askja central volcano, 10-20 km south-southeast of the Askja caldera. Its Soviet extension runs along the eastern border of the NVZ. The RRISP-profile II lies along the southern borders of the Katla and Öræfajökull central volcanoes [Angenheister et al., 1980; Gebrande et al., 1980], in the Eastern Volcanic Zone. Although some of these profiles lie along the flanks of central volcanoes, none of the older profiles cross the center of a volcanic system.

In the discussion below, we rely mostly on the 1994, NS and EW profiles, which cross the central part of the Krafla caldera and intersect near the Víti crater. We use sources (shots and microearthquakes) from the 50-250 distance range, which have P and S waves that dive deep (> 10 km) into the crust, cutting across the upper caldera at steep (< 45°) angles of incidence.

By mapping the Layer P-3 boundary and identifying crustal reflections across the Krafla central volcano we can, for the first time, examine in detail a section of the mid-Atlantic plate boundary in northeastern Iceland.

4.1 General features of the regional seismic data

The Krafla central volcano presents itself as a major crustal anomaly with large variations in seismic velocities and amplitudes at individual stations across the volcano.

P wave first arrival readings show a marked increase in apparent velocity in an approximately 20 km wide regional zone under the Krafla central volcano and large variations in traveltime within the Krafla caldera (Figures 15-18). We see an abrupt change in the overall appearance of the seismograms across the Krafla central volcano, with distinct changes in both the velocity and amplitude of P and S waves.

4.1.1 P waves

Very clear P waves are observed from detonations, towards the flank of the caldera facing the shotsite. At ranges where the P waves cross the flank of the caldera, the P wave apparent velocity increases from about 6.5 to 7.0 km/s. Stations located on the caldera rim are fastest, and have a abrupt advance in traveltime of about 0.1-0.2s. Within the caldera there is a pronounced decrease in apparent velocity and a significant reduction in amplitude. This amplitude reduction is observed from all azimuths (Figures 15 and 16). The caldera rim anomaly is visible at a station 65 km from the Axarfjördur shotsite on the NW line just before the onset of low P wave amplitudes, which is confined to the southern part of the caldera (Figure 15a). Similar anomalies are observed on all shotsites along the EW line (Figure 18), but not from the Askja shotsite (Figures 15b and 17), possibly due to poorer signal-to-noise ratio from that shot.

Velocity variations across the caldera are especially pronounced along the NS profile from the Langjökull earthquake (Figures 17 and 19) and on the EW profile from the Sænautavatn (SAEN) and Skjálfandi (SKJA) shotsites (Figure 18). Raypaths from the LANG earthquake (Figure 19), which sweep the Krafla caldera from WSW to ENE (Figure 1) along the NS profile, show a slight 0.1s increase in P wave traveltime at stations in Hlídardalur, south of the caldera (at 221.8-222.6 km range) followed by a marked 0.3s decrease in traveltime across the caldera (at 222.6-226.3 km range) and increasing again on the northeastern flank and eastern rim of the volcano (at 226.3-231.0 km range).

Although geothermal noise caused by near-surface boiling in the water-saturated geothermal fields affects the signal-to-noise ratio at a few stations, the region in which the amplitude reduction and P wave delays are observed is more widespread than the geothermal fields. Furthermore, the region of amplitude reduction exhibits a parallax effect. Its position on the NS profile (Figure 15) 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 EW profile (Figure 16) from a shot in the west differs by about 1.5 km from its position from a shot in the east. This effect is consistent with the amplitude reduction occuring at depths of at least 3 km, and not from shallower geothermal fields.

4.1.2 P_{ref} waves

We observe wide-angle reflections from all shotsites outside the Krafla region. This phase is particulary clear on a vertical component record section from a shot in the Skjálfandi bay in north Iceland, recorded on the EW profile (Figure 20). As noted above, we observe a marked increase in the apparent velocity of the P wave towards the Krafla caldera and large irregularities within the caldera (at 50-58.5 km range). A later arrival appears on stations east of the Krafla caldera at ranges beyond 58 km. It has a reduced traveltime (relative to 6.5 km/s) of 2.2s at 62.7 km and 1.45s at 84.6 km range, and an apparent velocity around 8.36 km/s. This phase may be analogous to the secondary phase observed by Angenheister et al. [1980] and Gebrande et al. [1980], from shotpoint D, in south-central Iceland and to the wide-angle reflection P_mP reflection observed by Bjarnason et al. [1993] in southwestern Iceland. However, a definitive identification of Moho requires a demonstration that the material beneath the reflecting interface has mantle-like (>7.6 km/s) velocities. We have not observed any phases refracted from beneath the reflecting interface, possibly because our measurements do not extend to sufficient range. Hence the interface could also represent an internal boundary within the lower crust. The interpretation of this interface will be addressed further when the complete FIRE dataset is analyzed. For now, we designate it as P_{ref}.

The reduced traveltime and clearness of the wide angle reflections vary between shotsites reflecting variations in crustal thickness along different profiles. For instance, shot E35 in Axarfjördur recorded on the 1991 NS profile, along the western side of the Krafla fault swarm (Figure 4) has a much clearer P_{ref} phase than the 1994 Axarfjördur shot which traversed the eastern side of the Krafla fault swarm (Figure 21). The P_{ref} phase on the 1991 record section (Figure 21b) has a reduced traveltime of 1.05s at 60.0 km and 0.46s at 79.0 km range, and an apparent velocity of 7.74-8.01 km/s. Although the P_{ref} reflection from the 1994 shot in Axarfjördur is not as clear (Figure 21a) as on the 1991 profile (Figure 21b) their reduced traveltimes are similar, whereas the reduced P_{ref} traveltime from the Skjálfandi shot is about 1s higher. The reduced P wave traveltime is around 1s for all three shots, making it impossible to explain the relatively large difference in P_{ref} traveltime by velocity variations in the upper crustal structure. Therefore, the observed difference in P_{ref} traveltime has to be explained by variations in thickness of the part of the crust overlying the reflecting interface. The data shows that this part of the crust is considerably thicker west of the Neovolcanic zone, raypaths from the Skjálfandi shot cross the NVZ from NW, than within it, raypaths from Axarfjördur travel along the NVZ. We will later discuss our regional fits of P and P_{ref} arrivals observed in NE-Iceland.

4.1.3 S waves

Shear waves, generated by P to S conversions adjacent to the shotsite, are often radiated from detonations in lakes and at the ocean bottom [Båth, 1960; Pálmason, 1971; Angenheister et al., 1980]. Shear waves, with raypaths that dive down to 10 km depth in the Northern Volcanic Zone, are also commonly observed on the regional seismic network in NE-Iceland. We therefore expected shear phases from our detonations and planned to use them to further constrain the thickness of shear wave attenuating regions within the Krafla caldera.

After generating record sections from each shotsite we were able to indentify shear phases from most of the detonations. Although hardly any S energy was observed on the vertical-component record section from individual shotsites, clear S phases occur on the radial horizontal-component record section from those same shotsites, indicating that the shear wave energy is indeed generated by P to S_V conversion at the shotsite. Still, we were not able to indentify shear phases from the Sænautavatn shot, recorded on the EW profile nor from the Askja and Axarfjördur shots, recorded on the NS profile. We attribute the lack of S energy from these shots to shotsite conditions as we observe S waves along their raypaths from other sites. For instance, although we do not observe shear waves from the 1994 Axarfjördur 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 NS profile. Why S waves were not genereated in lake Sænautavatn and the Askja caldera lake remains unexplained, although in the latter case strong S wave attenuation from the Askja magma chamber [Brandsdóttir et al., 1992] may be a factor.

The S wave traveltimes for the Eyjafjördur, Skjálfandi (Figure 22) and Lögurinn shotsites and the Axarfjördur 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 rays are attenuated under the Krafla caldera. Ratios in the range of 1.9-2.0 are clearly ruled out for these ranges, as they predict S arrivals several seconds later than is observed. The observation for Lögurinn is particularly interesting because it has the largest ranges (out to 114 km), which correspond to rays that bottom in the lower crust just east of the NVZ. There is no evidence for anomalously high V_p/V_s ratios in this region.

The 1993/1994 EW profile crosses both shear wave attenuation zones delineated by Einarsson [1978] within the Krafla caldera and the NS profile crosses the eastern zone. Again, we see large variations in shear wave velocities and energy within the caldera, coinciding with the observed P wave variations. The shear wave energy is fairly constant at the flank stations whereas drastic changes occur between individual caldera stations (Figures 23 and 24). For instance, S waves from the Eyjafjördur shot, west of Krafla, are easily traced out to a range of 73 km. They are completely absent at the station at 75 km, but then gradually reappear at larger ranges. They seem, however, to been attenuated substantially as they passed under the caldera, for they lack the higher (5-10 Hz) frequencies visible at shorter ranges.

The Langjökull earthquake provides some evidence that low-frequency (0.5 Hz) S waves can traverse the attenuating zone beneath the caldera, experiencing delays of about 0.5s (Figure 25). We note, however, that the wavelength of these S waves is 3-6 km, large enough that significant diffraction can occur. Hence these waves may not necessarily have traversed the central part of the attenuating region.

4.1.4 S_{ref} waves

We observe clear S_{ref} phases from two shots, Skjálfandi (Figure 22) and Eyjafjördur (Figure 26), both to the west of the EW profile, and from the earthquake in Axarfjördur, north of the NS profile (Figure 27). Whereas the crustal S-waves are being attenuated by the Krafla magma chamber, since they reach the caldera stations at a shallow angle of incidence, the deep crustal reflections (S_{ref}) are sharp, as they are sub-vertical and miss the shallow level magma chamber. Our estimates of V_p/V_s ratios for Sref are variable. V_p/V_s=1.76 seems a good estimate for Skjálfandi. On the other hand, we find V_p/V_s=1.88 for Axarfjördur. The two events have quite different S_{ref} reflection points, Skjálfandi's being west of the Northern Volcanic Zone and Axarfjördur's being along it. Therefore, the difference in ratios might plausibly be related to higher lower crustal temperatures beneath the Northern Volcanic Zone. However, this interpretation should be considered speculative until confirmed with further observations.

To our knowledge, ours is the first time that a deep crustal reflector has been identified in Iceland. However, the startling resemblence of the record section from Gebrande et al.'s [1980] E shotpoint (Figure 2) to our Skjálfandi and Eyjafjördur record sections (Figures 22 and 26) leads us to propose that the more distant shear arrival in the G profile is in fact S_{ref}, and not a 'delayed S' as those authors believed. If so, its moveout implies a crust that thickens from the northeast coast towards central Iceland.

4.2 Seismic structure of the Northern Volcanic Zone, across the Krafla central volcano

We have created two-dimensional NS (Figure 28) and EW (Figure 29) 2-D velocity models across the Krafla central volcano, using the Caress et al. [1992] RTMOD forward modeling program and a iterative, trial-and-error approach. The compressional velocity is represented by continuous linear splines on a triangular mesh. We begin with a smooth, 2-D structure based on a combination of our traveltimes, previously published crustal models by Pálmason [1963], Flóvenz and Gunnarsson [1991] and an unpublished N-Iceland crustal model, NOR, developed by the Science Institute of the University of Iceland. We iteratively perturb this starting model until we are able to fit the small traveltime anomalies (Figures 30 and 31). The ray coverage, which is derived from four shotpoints on the EW profile and two shotpoints and one earthquake on the NS profile, enables us to resolve the crustal structure of the uppermost 10-15 km of the Krafla central volcano and determine the crustal thickness at different midpoints from individual sources.

Note that the purpose of this modeling effort is to illuminate the structure of the central volcano. Our model of the structure outside the central volcano was constructed only to provide a basis for computing traveltimes, and is not well-resolved. A detailed profile of crustal structure in N Iceland must await a full analysis of the complete FIRE dataset.

Our ray coverage is only sufficient to determine features within the central volcano (i.e. between the ends of the profiles). Our final profiles show the locations and amplitudes of the large scale (> 2 km) lateral heterogeneities associated with the Krafla central volcano, and fit the overall traveltimes fairly well. The model has three main features:

1. A broad, high-velocity dome, or 'chimney' characterizes the substructure of the volcano. Outside the Krafla central volcano the 6.9 km/s isovelocity surface (the red surface in Figures 28 and 29) is at a depth of about 11 km on the NS profile and 14 km on the EW profile, dipping to 15.8 km adjacent to Krafla. Under the central volcano this isovelocity surface rises up to 3-4 km depth in a narrowing chimney. The base of the chimey is about 40 km wide at depth, narrowing substantially toward the surface. The shape of this chimney is constrained by the regional P wave traveltime data, which have a 15-20 km wide, 0.2s advance centered over the caldera. Interestingly, the widths of the upper part of the chimney are similar in both the EW (spreading parallel) and NS (plate boundary parallel) directions. The large difference between P wave apparent velocities of shots in Axarfjördur observed on the 1991 and 1994 NS profiles, whith raypaths along the Krafla fault swarm (Figures 4 and 21), indicate that the uppermost part of this chimney is very narrow - approximately the same width of the caldera.

2. The thickness of the crust above the reflecting interface increases towards the Krafla central volcano from all directions. Ths part of the crust is about 20 km thick under Bárdardalur and Fljótsdalur, west and east of the caldera, respectively, deepening to 25-28 km under the flanks of the central volcano. Thickness estimates for the region east of Krafla are tentative and will likely improve when the FIRE data from NE Iceland has been analysed.

3. The Krafla magma chamber sits at the top of the high-velocity dome. Its depth of 3-4 km is constrained mainly by the parallax effect discussed above. Its thickness of 0.75-1.8 km is constrained by the 0.2-0.3s delay (Figures 19 and 32) experienced by P waves that cross the structure, assuming that the magma velocity is 3 km/s and the background velocity 5-6 km/s. In our model we represent the magma chamber by a 2-6 km wide lens of low velocities. A more detailed model of its internal structure will require finely spaced (100 m) 2-D array data that can sort out the complex pattern of secondary arrivals observed in the caldera region.

Using Pálmason's [1971] definition of Layer 3, our model shows the Layer P-2/P-3 boundary (isovelocity surface of 6.5 km/s) at a depth of 5.0 km along the NVZ, rising to 1.6 km depth beneath the Krafla caldera (Figure 28). The P-2/P-3 boundary deepens slightly south of the Krafla central volcano, where it lies at 7 km depth adjacent to the high-velocity dome. A similar effect is seen on the EW profile, where the 6.5 km/s isovelocity surface (medium orange surface in Figures 28 and 29) lies at 4.7-5.0 km depth east and west of the NVZ deepening to 9.3 km adjacent to the Krafla high-velocity dome (Figure 29).

The P-2/P-3 boundary lies at an average depth of 5 km along both profiles, whereas the depth to the 6.9 km/s isovelocity surface (red surface in Figures 28 and 29), away from the Krafla central volcano, is greater (14 km) east and west of the NVZ, than within it (11 km). Although the EW regional model presented here has to be reviewed against the FIRE data from NE Iceland, it is of interest to note that the shape of the 6.5 and 6.9 km/s isovelocity surfaces along our EW profile is similar to the shape of the 6.3 km/s isovelocity surface in Pálmason's [1963] crustal model of the NVZ (Figure 33). Both profiles indicate that the P-2/P-3 boundary (wheather defined as 6.3 km/s [Pálmason, 1963] or 6.5 km/s [Pálmason, 1971]) deepens at the eastern border of the Northern Volcanic Zone. Pálmason's [1963] model has the 6.3 km/s isovelocity surface deepening from 2.5 km to about 4.5 km at the eastern border of the NVZ whereas in our model the 6.3 km/s surface (yellow surface in Figures 28 and 29) deepens from 3.7-4.3 km to 7 km in the same region.

There is little consensus concerning the petrology of Layer P-3 in Iceland. Many authors have debated wheather Layer P-3 consists of a heterogeneous mixture of basaltic intrusives and extrusives [Bödvarsson and Walker, 1964] and a sheeted dike complex [Walker, 1975], or metamorphic facies of basaltic rocks [Pálmason, 1963, 1971] such as low-porosity basalts with high contents of epidote [Christensen and Wilkins, 1982; Flóvenz and Gunnarsson, 1991]. Actually, these theories are not mutually exclusive. The velocity structure of the oceanic crust is more related to its *porosity and alteration state than to its igneous structure [Carlson and Herrick, 1990].* The nature of Layer 3 may vary between regions. The compressional velocity of the crust should increase with age, as the rock cools and as cracks are filled with hydrothermally-deposited minerals [Houtz and Ewing, 1976]. The gradual updip of the Layer P-2/P-3 isovelocity surface, with increasing distance from the diverging plate boundary (Figures 29 and 33), may be caused by this effect. The central volcanoes superimpose a shorter wavelength flucuation of Layer 3 depth on this overall trend, caused by the presence of high-velocity intrusives (in contrast to metamorphic facies). Our modeling and the Krafla borehole cross sections show that Layer 3 is associated with basaltic intrusives (dikes) at shallow depth within the Krafla central volcano, whereas P-velocities in 6.3-6.5 km/s range are most likely related to different intrusive/extrusive ratios along different sections of the NVZ.

5 Discussion and Conclusions

1. The crust near the Northern Volcanic Zone is at least 20-28 km thick. The crustal thickness we observe in northeastern Iceland is larger than the 22-24 km value determined previously for southwestern Iceland [Bjarnason et al., 1993]. A sharp lower crustal interface, identified through clear P_{ref} and S_{ref} lower crustal reflections, occurs at 20-28 km depths. These phases are observed on paths that have reflection points beneath the Northern Volcanic Zone, indicating that the reflecting interface in continuous across the plate boundary. The crust above this interface thickens towards the plate boundary and is thickest beneath the Krafla central volcano. Since no phases refracted from below the reflecting interface were observed, the interface cannot be definitively interpreted. It may be Moho. On the other hand, it may represent an internal boundary within the lower crust.

   Our bounds on crustal thickness is in agreement with the alternative interpretation of Angenheister et al. [1980], where the crustal thickness of Iceland is about 30 km, and overrules previous interpretations on the crustal thickness of Iceland based on linear extrapolation of geothermal gradients [Pálmason, 1971; Flóvenz and Gunnarsson, 1991]. A general thickening of the crust towards the center of Iceland has been proposed by Bödvarsson and Walker [1964]. The effect that we observe is more narrowly focused along the Northern Volcanic Zone than in their model.

2. The depth to isovelocity surface 6.9 km/s decreases under the Northern Volcanic Zone. This effect has been noted previously by Pálmason [1963] for the isovelocity surface 6.3 km/s in the NVZ (Figure 33) and by Bjarnason et al. [1993] for the Western Volcanic Zone. Pálmason [1971] found the shallowest depth to Layer P-3 associated with volcanic centers and suggested that the depth contours of Layer P-3 may be used to map volcanic centers. Our study supports this theory.

A broad, high-velocity dome (chimney) rising from the lower crust characterizes the substructure of the Krafla central volcano. We see Layer P-3 rise from 11-16 km depth up to 2-4 km depth, beneath the Krafla magma chamber, narrowing upwards, from a 40 km wide base to a 10 km wide top. The general structure of the high-velocity chimney supports the crustal accretion model proposed by Pálmason [1973; 1986] and Menke and Sparks [1995], in which crust is created at shallow depths in the volcanic zone and then subsides, by solid-state creep, to deeper depths.

3. The shear velocities in the crust, as deduced from S and S_{ref} wave traveltimes, are consistent V_p/V_s=1.76-1.79, with two exceptions: Shear waves do not propagate through the shallow magma chamber beneath Krafla at all; S_{ref} traveltimes for paths along the Northern Volcanic Zone are consistent with V_p/V_s=1.88, which may possibly imply higher lower crustal temperatures there.

4. No compressional or shear wave low velocity zones are detected in the mid-crust of northeastern Iceland, contrary to predictions based on the interpretation of high mid-crustal electrical conductivities [Beblo and Björnsson, 1980], which would place partial melt at these depths.

5. The Krafla magma chamber lies at a depth of 3 km. It has a width of 2 km in the NS direction and 3 km in the EW direction, based on the widths of the zone of P wave attenuation and delay. Its thickness of 0.75-1.8 km is constrained by the 0.2-0.3s delay experienced by P waves that cross the structure. Its width as inferred from the region of shear wave shadows is large, 3 km NS and 10 km EW, the same as inferred by Einarsson [1978] (Figures 34 and 35). These different estimates are not contradictory, but rather reflect differences in the parameters measured (Figure 36). P wave velocities are relatively insensitive to temperature until actual melting has occured, whereas strong shear wave attenuation can occur even at sub-solidus temperatures. For instance, the shear wave quality factor of gabbro is reported to be 20 at 925°C [Kampfmann and Berckhemer, 1985].

   The zone of high shear wave attenuation extends beneath the magma chamber, to the depth of 7-10 km, an effect which we ascribe to high (but subsolidus) temperatures in this region. The high attenuation zone does not extend down to Moho, as S_{ref} phases, which cross the central volcano at lower crustal depths, are detected within parts of the region where S waves have been attenuated.

   We have reexamined data from the SIST profile [Bjarnason et al., 1993], which crosses the Western Volcanic Zone (WVZ) about 10 km north of the Hengill central volcano in southwest Iceland. The S wave traveltime 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 (Figure 37). This result supports the idea that zones of attenuation are concentrated immediately beneath central volcanoes, and do not extend significantly along the plate boundary.

6. The near-surface structure of the Krafla caldera is approximately flat-lying, with only subdued lateral heterogeneities. Two thin (240-375m) 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.

7. The compressional velocity model put forward by Arnott and Foulger [1994a] for the near-surface of the Krafla caldera contains strong, shallow heterogeneities that predict traveltime anomalies not observed in our data. We conclude that this model is flawed. We ascribe the hypothesized strong lateral heterogeneity to an artifact of their tomographic imaging algorithm that arises from hypocentral depth trading off with velocity structure.

In our view, our model of the seismic structure supports the theory that the Krafla central volcano is playing 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 et al., 1995], with the bulk of the magmatic intrusion occuring with the central volcano itself. We interpret the high-velocity chimney as representing intrusions, generated in the magma chamber itself, which have subsequently been advected to greater depths (such as has investigated by Pálmason [1973; 1986] and Menke and Sparks [1995]) Two important classes of questions arise from this senario:

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 seemed to initiate in the southern part of the NVZ, near Vatnajökull and migrate north, until a mature rift zone was formed. How Krafla fits into this scenario is unclear. Sæmundsson [personal communication, 1995] gives 0.5 Ma as an upper bound for its age, so approximately five million years of geological history are unaccounted for. Certainly the 40 km width of the base of the high velocity chimney would seem to imply that that region had been a major volcanic center for far longer (2 Ma, based on a 10 mm/yr 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 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? Furthermore, what is the dynamic significance in the thickening of the crust under the central volcano? Does this, too, evolve with time?

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 wholely unconnected to the mantle - the zone of high shear wave attenuation, which we take as a proxy for temperature, seems restricted to the mid-crust 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 may be an indication that melt transport takes place episodically. We believe that the Krafla rifting episodes, which take place every 200 years or so, are a manifestation of intermittent melt injection from the mantle.

Answering these questions will require both detailed, three-dimensional crustal accretion models and a broader view of the structure of the Northern Volcanic Zone as a whole.

To conclude, our model of seismic structure of the Northern Volcanic Zone in general and the Krafla central volcano in particular reenforce the picture of Iceland as a slow-spreading oceanic ridge. Iceland falls at the extreme end of ridges, to be sure. Its crustal thickness is approximately three times that of a '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, follow closely the oceanic norm.

5.1 ACKNOWLEDGMENTS

We would especially like to thank our shotmaster, Jósep Hólmjárn, the National Energy Authority, for skillfully carrying out the key part of the fieldwork. Other members of the 1994 Krafla undershooting crew included Benedikt Halldórsson, Einar B. Pálsson, Gudmunda Sigurdardóttir, Ingthór Ingólfsson, Ravi Sethi, and Vadim Levin.

The 1993 Krafla reflection/refraction survey crew included Bjarni Einarsson, Brandur Bj. Karlsson, Michael C. Humphreys, Einar Kjartansson, Halldór Ólafsson, Jósep Hólmjárn, and Yngve Kristoffersen.

The help of the following persons is acknowledged: Ármann Pétursson, Reynihlíd; Kristján Lúdvíksson, Reydarfjördur; Fridrik Thorsteinsson, Grenivík; Arnar Sigurdsson, Húsavík; Hans A. Kristjánsson, Sandvík; Brynhildur Óladóttir, Sænautasel; Lucy Slater and Keith Priestley University of Cambridge; and Michelle Dance, Bristol University. We would also like to thank Knútur Otterstedt, the National Power Company of Iceland; Birkir Fanndal, Hinrik Árni Bóasson, Hjördís Albertsdóttir and Rakel Gudlaugsdóttir, Pétur Snæbjörnsson and other personnel at the Krafla Power Plant and Hótel Reynihlíd, Lake Mývatn for accomodating and feeding us during the fieldwork.

This research was supported by the Science Institute of the University of Iceland, the Icelandic Science Foundation, the National Power Company of Iceland and the Krafla Power Plant, the National Environmental Research Council of Britain (NERC), British Institutions Reflection Profiling Syndicate (BIRPS) Industrial Associates, The University of Cambridge, the US National Science Foundation, The Incorporated Research Institutions for Seismology (IRIS), and the Lamont-Doherty Earth Observatory.

Finally, we would like to thank Dave Caress for giving us access to his raytracing program and the Nordic Volcanological Institute, University of Iceland, University of Nevada, and IRIS for lending us their seismic equipment.

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