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Crustal structure above the Iceland mantle plume imaged by the ICEMELT refraction profile

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SUMMARY

The crustal structure of central Iceland is modelled using data from a 310 km long refraction profile shot during summer 1995. The profile traversed Iceland from the Skagi Peninsula on the north coast (surface rocks of age 8.5–0.8 Myr) to the southeast coast (surface rocks of age 8.5–3.3 Myr), crossing central Iceland (surface rocks of age 3.3–0 Myr) over the glacier Vatnajökull, below which the locus of the Iceland mantle plume is currently centred. The crustal thickness is 25 km at the north end of the profile, increasing to 38–40 km beneath southern central Iceland. The upper crust is characterized by seismic *P*-wave velocities from 3.2 to approximately 6.4 km s^{-1}. At the extreme ends of the profile, the upper crust can be modelled by a two-layered structure, within which seismic velocity increases with depth, with a total thickness of 5–6 km. The central highlands of Iceland have a single unit of upper crust, with seismic velocity increasing continuously with depth to almost 10 km below the surface. Below the central volcanoes of northern Vatnajökull, the upper crust is only 3 km thick. The lower-crustal velocity structure is determined from rays that turn at a maximum depth of 24 km below central Iceland, where the seismic velocity is 7.2 km s^{-1}. Below 24 km depth there are no first-arriving turning rays. The Moho is defined by *P*- and *S*-wave reflections observed from the shots at the extreme ends of the profile. *P*- to *S*-wave velocity ratios give a Poisson's ratio of 0.26 in the upper crust and 0.27 in the lower crust, indicating that, even directly above the centre of the mantle plume, the crust is well below the solidus temperature.

Key words: crustal structure, Iceland, mantle plume, seismic refraction.

1 INTRODUCTION

Iceland was created by the interaction between the Mid-Atlantic Ridge and the Iceland mantle plume. The plume causes thermal uplift of a region at least 1000 km in radius, and the elevated mantle temperatures [approximately 150 °C (White & McKenzie 1995; White *et al.* 1995) to 300 °C (Wolfe *et al.* 1997)] in the plume core cause considerably more melting than below a normal mid-ocean ridge, resulting in a region of anomalously thickened crust in Iceland. The Mid-Atlantic Ridge is a slow-spreading ridge; in Iceland the average full spreading rate is 18 mm yr^{-1} (DeMets *et al.* 1994).

The ridge axis in Iceland is expressed as a set of three volcanic rift zones, composed of central volcanoes transected by rifts and fissure swarms (Fig. 1). In the south of Iceland, there are two subparallel volcanic zones. The Reykjanes Ridge spreading centre is first expressed on land in the far south-west of Iceland, and continues northeastwards as the Western Volcanic Zone (WVZ). This is offset from the Eastern Volcanic Zone (EVZ) by a transform region, the South Iceland Seismic Zone (Fig. 1). It appears that the EVZ is taking over as the main locus of spreading in southern Iceland from the WVZ (Jóhannesson 1980), propagating slowly southwards and becoming more active with time. In northern Iceland, one spreading centre, the Northern Volcanic Zone (NVZ), is currently present. The NVZ is bounded to the north by the Tjörnes Fracture Zone, which links the northern Icelandic rift system to the Kolbeinsey Ridge, the active offshore spreading centre extending northwards from Iceland.

The positions of the volcanic rift zones in Iceland have shifted throughout the island's lifetime. (Saemundssson 1979; Helgason 1984, 1985; Hardarson *et al.* 1997; Smallwood *et al.* 1999). An initial ridge jump, prior to 16 Ma, brought the spreading axis to western Iceland, where the oldest surface rocks, 16 Myr in age (Saemundsson 1979), are found. The northern section of the ridge then jumped eastwards, presumably in order to follow the track of the mantle plume, initiating the now extinct Húnaflói–Skagi Volcanic Zone

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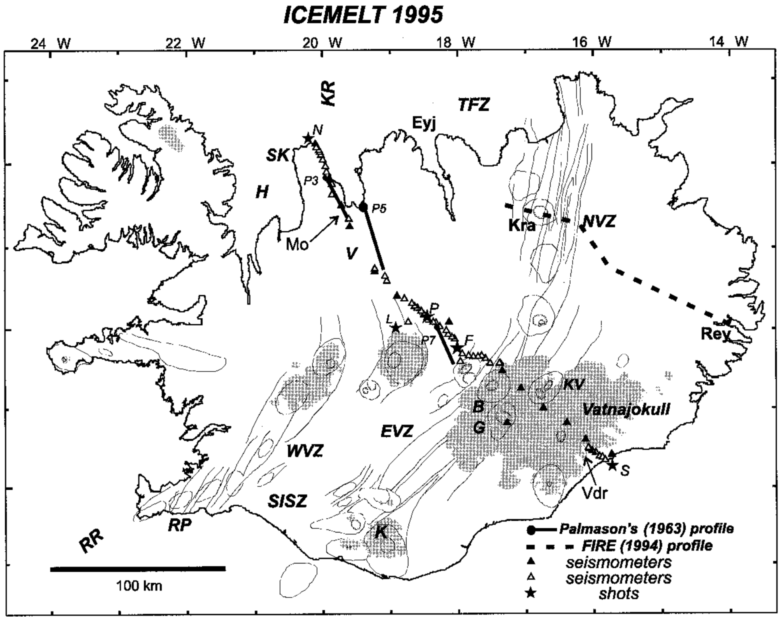

Figure 1. Map of Iceland showing the locations of the shots and seismic stations used in the ICEMELT refraction profile, overlaid on rift zones (thin lines), central volcanoes (ovals) and glaciers (shaded). Two sets of seismometers are shown; the solid triangles represent stations deployed for several months and the open triangles represent stations deployed for the duration of the controlled-source experiment only. Heavy solid lines show the three refraction profiles of Pálmason (1963) which were used to augment the ICEMELT data. Shots for these profiles are shown by filled circles. The 1994 FIRE profile (Staples *et al.* 1997) is shown by a heavy dotted line. N: North shot (off north coast of Skagi); P3, P5, P7: Pálmason's (1963) profiles; L: 'Lava' pool shot; P: 'Pond' shot P; F: Fjördungsvatn shot F; S: South shots; NVZ: Northern Volcanic Zone; EVZ: Eastern Volcanic Zone; WVZ: Western Volcanic Zone; SISZ: South Iceland Seismic Zone; TFZ: Tjörnes Fracture Zone; RR: Reykjanes Ridge; KR: Kolbeinsey Ridge; RP: Reykjanes Peninsula; SK: Skagi Peninsula; V: Vesturdalur; H: Húnaflói; Eyj: Eyjafjördur; Rey: Reydarfjördur. Central volcanoes: B—Bárdarbunga, G—Grímsvötn, KV—Kverkfjöll, K—Katla, Kra—Krafla, Mo—Molduxi (extinct), Vdr—Vedurárdalur (extinct).

(HSVZ) (see Fig. 1 for the position of the present Skagi Peninsula and Húnaflói). The most recent jump of the northern spreading centre was at 7–6 Ma, when the ridge jumped eastwards from the HSVZ to its present position. The HSVZ ceased activity at approximately 3 Ma (Helgason 1984, 1985).

During the last few decades, the nature of the Icelandic crust has been the subject of much study and debate. Geophysical investigations have resulted in two main hypotheses about Iceland's crustal structure. One model has a thin, hot crust (10–15 km) overlying anomalously slow mantle, perhaps with a pervasive region of partial melt at the base of the crust (Gebrande *et al.* 1980; Beblo & Björnsson 1978, 1980; Beblo *et al.* 1983). The alternative model has a cooler crust, with temperatures at least 200–300 °C below the basalt and gabbro solidi, and a Moho depth varying between 20 and 40 km below the island (Zverev *et al.* 1976; Bjarnason *et al.* 1993; Staples *et al.* 1997; Menke *et al.* 1998).

The 310 km long ICEMELT refraction profile, part of the larger ICEMELT experiment, uses diving rays and wide-angle reflections recorded by land-based seismometers to establish a 2-D model of the seismic velocity structure of the crust and uppermost mantle beneath central Iceland. Of particular interest is the crustal thickness directly above the plume centre.

2 PREVIOUS GEOPHYSICAL STUDIES OF ICELAND

Early seismic profiles carried out in Iceland led to two possible interpretations of the Icelandic crust, a 'thick' crust model and a 'thin' crust model. Båth (1960) resolved a three-layered crust of thickness 28 km, with the deepest layer having seismic velocities of 7.4 km s^{-1}. Tryggvason (1962) used surface-wave dispersion to model the crust as a two-layered, 10 km thick structure underlain by a half-space of seismic velocity 7.4 km s^{-1}. It was Pálmason's work in the 1960s and 1970s (1963, 1971) that led to a long-accepted layered model for the crust. He divided the crust into four layers comprising the upper and lower crust, underlain by rocks he called 'layer 4', with a seismic velocity of 7.4 km s^{-1}. This 'layer 4' was interpreted as anomalously slow, partially molten mantle material based on the extrapolation of high surface-temperature gradients measured in Iceland.

The thin crust model, underlain by anomalous mantle of *P*-wave velocity 7.0–7.4 km s^{-1}, was used as the main interpretation of Iceland's structure for many years. Angenheister *et al.* (1980) and Gebrande *et al.* (1980) interpreted results from the RRISP (Reykjanes Ridge Iceland Seismic Profile) data as attributable to a crust of thickness 10–15 km, although they noted that, were the 7.0–7.4 km s^{-1} layer taken as being part of the crust, the depth to the Moho would be approximately 30 km. The extent of the 'anomalous mantle' was taken as evidence for a pervasive region of partial melt below the crust. Flóvenz (1980) and Flóvenz & Gunnarsson (1991) also supported the thin crust model, although they began to modify Pálmason's layered model. Flóvenz (1980) showed that record-section amplitudes were more consistent with a continuously increasing seismic velocity with depth than with a structure of a few constant-velocity layers. He divided the Icelandic crust into

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two parts—the upper crust, with high velocity gradients, and the lower crust, with a small or zero velocity gradient.

This model, with a 10–15 km thick crust underlain by a pervasive region of partially molten mantle, was supported by the interpretation of data from other geophysical studies of Iceland. Beblo & Björnsson (1978, 1980), Beblo *et al.* (1983), Hersir *et al.* (1984), and Eysteinsson & Hermance (1985) measured the electrical resistivity structure of Iceland using the magnetotelluric method. They found a widespread high-conductivity layer below Iceland, except in the southeast and at the west coast. The depth to this layer varied from 10 km below the rift axis to 25 km below the oldest crust. This high-conductivity layer was interpreted as partially molten basalt at the base of the crust, an interpretation consistent with the extrapolation of surface heat-flow data. Flóvenz & Saemundsson (1993) reported temperature gradients of 50–150 °C km^{-1} in Icelandic boreholes. Extrapolating this gradient downwards and assuming an increase in thermal conductivity with depth, they predicted partially molten material at a depth of 10–30 km.

Several arguments were put forward during this period suggesting that the Icelandic crust was thick and subsolidus in nature. The NASP (North Atlantic Seismic Project) wide-angle seismic experiment of 1976 profiled the Faroe–Iceland Ridge and the northeast of Iceland. Results from this profile (Zverev *et al.* 1976) suggested that the crust of eastern Iceland was at least 40 km thick, with mantle velocities of 7.8–8.0 km s^{-1} below the Moho.

More recently, the SIST (South Iceland Seismic Tomography) profile across the Western Volcanic Zone yielded data that was used to support the thick crust theory. Bjarnason *et al.* (1993) observed strong wide-angle reflections from several shots, which they interpreted as reflections from the Moho at 20–24 km depth. They measured an apparent sub-Moho velocity of 7.7 km s^{-1} and lower-crustal velocities up to 7.2–7.25 km s^{-1}, and therefore interpreted Pálmason's layer 4 as part of the lower crust. In the light of these new results, Menke *et al.* (1996) studied the RRISP data and reinterpreted them to give a crustal thickness of 30–35 km below central Iceland.

Other seismological data appeared to support the model of a relatively cool, thick crust. Menke *et al.* (1995) measured values of the seismic Q-factor for rays turning at depths of 12–20 km beneath the Western Volcanic Zone [interpreted as the mid–lower crust by Bjarnason *et al.* (1993)]. The Q values found were of the order of several hundred, suggesting that if the lower crust is gabbroic in nature, its temperature is approximately 700–775 °C, well below the dry gabbro solidus. Menke *et al.* (1995) noted that no known petrology could explain these Q values for a temperature greater than 900 °C. These results would rule out a region of pervasive partial melt below Iceland, at least along the measured seismic paths.

In 1994, the Faroe–Iceland Ridge Experiment (FIRE) profiled the Faroe–Iceland Ridge and the eastern part of Iceland in considerable detail. Staples *et al.* (1997) interpreted high-quality wide-angle seismic arrivals, including Moho reflections, to give a crustal thickness of 19 km in the rift zone below Krafla, thickening to almost 35 km in the Tertiary region of eastern Iceland. Mantle velocities were modelled as approximately 7.9 km s^{-1} below the rift, increasing to as much as 8.2 km s^{-1} in eastern Iceland (based on amplitude comparisons with synthetic seismogram models). The Poisson's ratio of 0.27 inferred from observations of P- and S-wave traveltimes also suggested a subsolidus crust.

The B96 profile (Menke *et al.* 1998), shot along the western edge of the Northern Volcanic Zone, suggested crustal thicknesses between 25 and 31 km, thickening towards the plume centre below northern Vatnajökull. Traveltime ratios of P and S waves again suggested a subsolidus crust with no pervasive regions of partial melt.

Of interest to this study, in addition to the overall crustal thickness and temperature measurements, is the detailed structure of Iceland's central volcanoes, and also the structure of the Iceland mantle plume itself, particularly its influence on lithospheric structure and shallow temperature variations. It has long been noted that the crustal structure beneath the central volcanoes is different from that observed elsewhere in Iceland. Pálmason (1963, 1971) and Flóvenz (1980) noted that the upper crust was significantly thinned below the central volcanoes. More recently, studies of Katla (Gudmundsson *et al.* 1994), Krafla (Brandsdóttir *et al.* 1997) and volcanoes of the Northern Volcanic Zone (Menke *et al.* 1998) and Western Volcanic Zone (Bjarnason *et al.* 1993) have revealed thinning of the upper crust, attributed to 'domes' of material of elevated seismic velocity directly beneath the central volcanoes. At the tops of these domes, small magma chambers at depths of $\sim$3 km have been inferred beneath Katla and Krafla.

The Iceland mantle plume has been mapped by Tryggvasson *et al.* (1983) and Wolfe *et al.* (1997) using tomographic analysis of teleseismic earthquake arrivals. A cylindrical zone of low mantle velocities, of approximately 150 km in radius, is found to extend from less than 100 km depth to at least 400 km depth, centred below the volcanoes of northwestern Vatnajökull in central Iceland, above which the ICEMELT profile passes. Analysis of the velocity perturbation by Wolfe *et al.* (1997) suggests a mantle temperature anomaly of 200–300 °C in the plume core.

3 THE ICEMELT REFRACTION EXPERIMENT

The ICEMELT refraction line traversed Iceland from the northern tip of the Skagi Peninsula on the north coast of Iceland, crossing the central highlands and the glacier Vatnajökull and terminating on the southeast coast of the island (Fig. 1). Up to 60 land-based instruments recorded each of six explosive shots (dynamite), which ranged in size from 25 to 400 kg. Three shots were fired offshore, and the three land shots were fired in lakes in the central highlands of Iceland (see Table 1).

The experiment used a wide variety of instruments (see Table 1), mostly Refteks, with extra land coverage provided by Scintrex PRS4 and Cambridge DSR data loggers. Most of the instruments were deployed for the duration of the controlled-source experiment only, with a more sparse array collecting earthquake data over a period of months before and after the shots were fired. GPS timing was used for the explosive shots and for many of the receivers. Where instruments were calibrated with a different timing system, the data were corrected for the offset between this system and GPS. Locations of shots and seismic stations were measured using multiple GPS readings. Location errors for the stations and the inland shots were of the order of a few tens of metres, while the

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Table 1. Shot and station information for the ICEMELT refraction profile.

Shot name	Size (kg)	Location	Water depth (m)
North	400	tip of Skagi peninsula	45
South	325	southeast coast	45
'Small'	25	southeast coast	49
Fjördungsvatn	150	Sprengisandur, central highlands	1.2
'Lava'	25	northern highlands, offline	2.5
'Pond'	25	northern highlands, online	1.5

Data logger	Number	Seismometers
Reftek 72A-03	25	Willmore MkIIIA[1], Kinemetrics SV-1/SH-1[2]
Reftek 72A-07/08	14	Guralp CMG-3T[3], Guralp CMG-3ESP[3], Teledyne S13[1] Mark L22D[1], Kinemetrics SV-1/SH-1[2]
Scintrex PRS4	15	Mark L22D[1]
Cambridge DSR	7	Mark L4-C[1]

[1] Short-period sensor
[2] Intermediate-period sensor
[3] Broad-band sensor

offshore shots were more accurately located by differential GPS. System timing errors were kept to less than the sampling rate of the recording instruments.

Data from the seismic stations were gathered into seismic record sections (Fig. 2). Despite poor weather conditions during the shooting period, the signals recorded by the seismic stations are of good quality, with arrivals visible up to 300 km from the largest shots, and up to 100 km from the smaller inland shots.

4 DATA AND MODELLING

The upper crust, characterized by high velocity gradients (>0.2 s^{-1}) and *P*-wave velocities of 6.4 km s^{-1} or less, is sampled by arrivals from all four of the principal on-line shots, while the lower crust is sampled by arrivals from the end shots only. Clear impulsive arrivals from crustal diving waves, P_g, are usually seen out to approximately 100 km offset. Crustal diving rays observed at larger offsets tend to be rather more emergent in nature. The record sections for stations recording the northern and southeastern shots show a low-amplitude wide-angle reflected arrival, which we have interpreted as the Moho reflection, P_mP. This phase can be observed in record sections filtered with a wide frequency passband, but is clearer at the large offsets when the data are bandpass filtered between 1.0 and 5.0 Hz (Fig. 3).

A closer study of the upper-crustal arrivals shows that the modelling of the upper crust is best achieved by using a continuously increasing velocity with depth (although abrupt changes in velocity gradient appear to exist at the northern and southern ends of the profile), similar to that proposed by Flóvenz (1980). Shot P, in particular, shows a smooth and rather symmetrical traveltime curve for the first arrivals both to the northwest and to the southeast of the shot (see Fig. 2). In contrast, while the Fjördungsvatn shot (F) shows a similar pattern of first arrivals to the northwest, arrivals with a velocity characteristic of the lower crust are significantly advanced in time, suggesting a pronounced thinning of the upper crust below northern Vatnajökull.

The data used to resolve the upper-crustal seismic velocity structure were augmented by the use of three of Pálmason's (1963) refraction profiles. These lay sufficiently close to the ICEMELT line to be used directly in the crustal model. The profiles used were 3 (Tindastóll), 5 (Blönduhlid) and 7 (Laugafell–Sprengisandur), denoted P3, P5 and P7 in the figures. The P5 profile was particularly useful as it covered a region where ICEMELT station coverage was sparse (Figs 1 and 4).

Arrival times of P_g and P_mP phases were picked by hand from the record sections. Fig. 5 shows picks for the more emergent arrivals observed at offsets greater than 100 km. A crustal model was generated using the RAYINVR computer algorithm of Zelt & Smith (1992). The initial stages of analysis used forward modelling by 2-D ray tracing, comparing synthetic traveltimes generated by the algorithm to the hand-picked traveltimes. A damped least-squares inversion was then performed to refine the model further. The inversion was stable in most cases, but was not used for the refinement of the upper crust below Vatnajökull, where the significant change in velocity structure was most effectively modelled by pinching out units of intermediate crustal velocity, a modelling method that makes the inversion scheme unstable.

A total of 181 arrival-time picks were used to model the structure of the crust, giving an rms traveltime residual between picked and calculated arrivals of 81 ms. This traveltime residual is slightly less than the average uncertainty in the picks of arrivals from the record sections. The normalized chi-squared value, representing the misfit function of calculated traveltimes to the data, was 1.31. Fig. 6 shows the rays traced by the RAYINVR algorithm through the upper-crustal model. Figs 7 to 10 show record sections for each of the four principal ICEMELT shots, with the corresponding rays traced by RAYINVR. Calculated traveltimes are superimposed on the record sections. Fig. 11 shows the final crustal model for the ICEMELT profile, together with the crustal model for the FIRE 1994 profile (Staples *et al.* 1997) for comparison. The FIRE profile crossed the northeast of Iceland, approximately 120 km north of the plume centre (see Fig. 1). Fig. 12 shows 1-D velocity profiles along the ICEMELT line.

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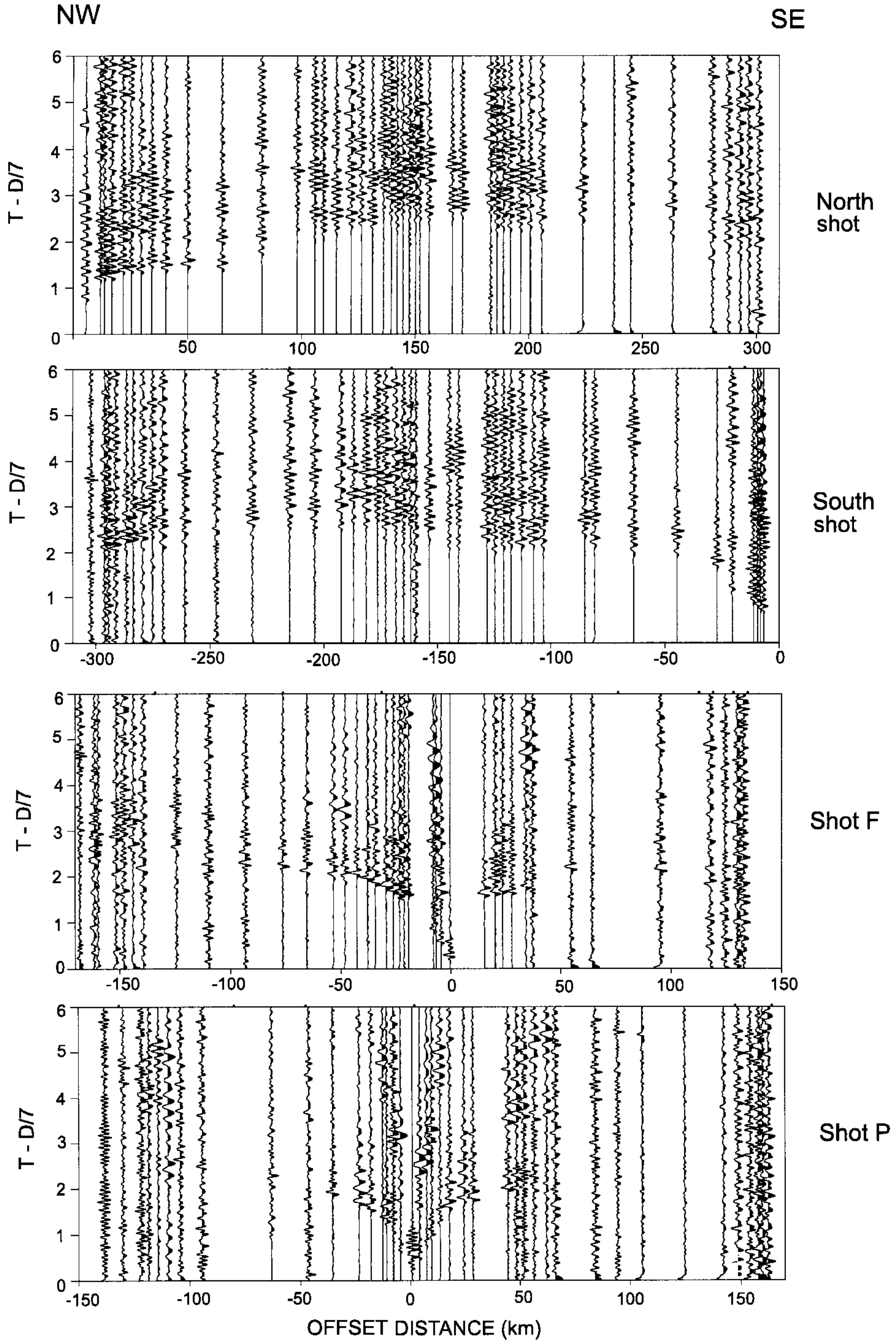

Figure 2. Trace-normalized record sections for the four main ICEMELT shots used in the analysis. The reduction velocity is 7.0 km s^{-1}and traces are bandpass filtered from 2.5 to 15.0 Hz.

5 THE UPPER CRUST

We refer to the region with seismic velocities increasing from 3.1 km s^{-1} to a maximum of 6.4 km s^{-1} as the upper crust. This part of the crust is also characterized by high velocity gradients (>0.2 s^{-1}).

The upper crust along the profile is, in the main, well constrained by the four principal shots and the extra data from Pálmason's (1963) profiles. Shot and station spacings mean that the best constraint is possible in the central highland region, with the poorest ray coverage in the Vatnajökull area, due to the sparser shot and station spacings. In the Vatnajökull region, only the largest-scale features of the upper crust may be resolved.

The near-surface rocks (depths less than 1 km) sampled by the profile vary in velocity. At distances along the profile from −150 to −50 km (Skagi and Vesturdalur), near-surface velocities are 3.13–3.18 km s^{-1}, while in the central highlands (−40 to +50 km) the velocities are 3.20–3.30 km s^{-1}. The highest surface velocities of 3.48 km s^{-1} are measured at the

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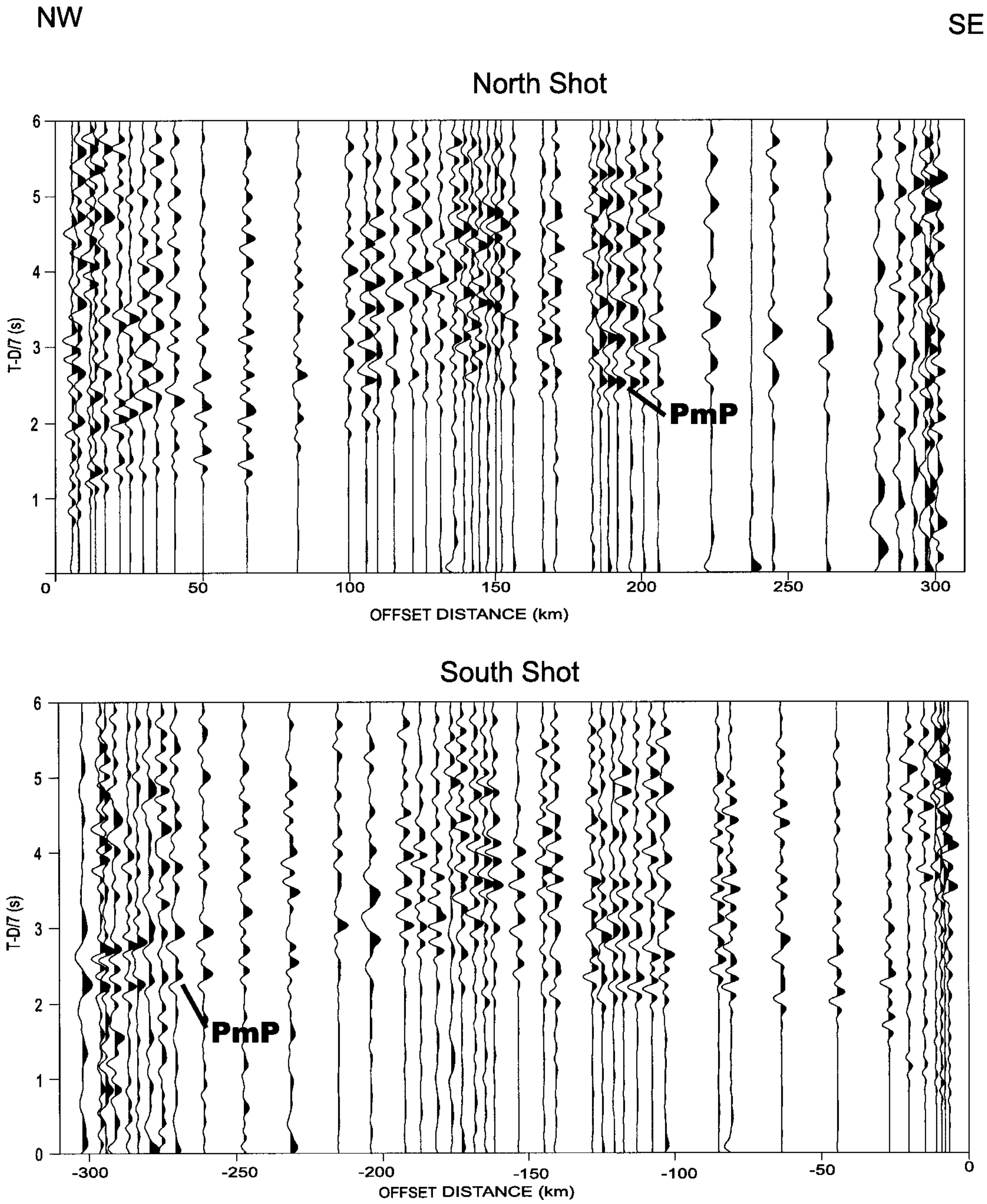

Figure 3. Trace-normalized record sections for the North and South shots, bandpass filtered from 1.0 to 5.0 Hz. Reduction velocity is 7.0 km s^{-1}. P_mP, the Moho reflection, is labelled.

southeast coast in Kálafellsdalur (+140 to +170 km along the profile). Following Gudmundsson *et al.* (1994), the seismic velocity through the Vatnajökull ice cap (+60 to +140 km) is taken as 3.60 km s^{-1}.

The traveltime curves measured from the record sections are most effectively fitted by an upper-crustal model with seismic velocity increasing smoothly with depth. Where discontinuities or unusually high velocity gradients are seen, these are modelled by a thinning or pinching out of the appropriate seismic velocities. We observe a considerable variation in upper-crustal structure along the profile.

From Skagi to northern Vesturdalur (−150 to −80 km), the upper crust can be divided into two units, the relative thicknesses of which change along the profile. The total thickness of the upper crust in this region is approximately 5 km, with minor variations in the base topography. The base of the upper crust is best modelled by a small velocity discontinuity, although this feature is not directly resolved, and is defined by seismic velocities of 6.4 km s^{-1}.

In central Iceland, from southern Vesturdalur to Fjördungsvatn (−70 to +40 km), the upper crust is characterized by a continuous increase in seismic velocity with depth. The upper-crustal thickness increases gradually from 5 km at Vesturdalur (−70 to −60 km) to a little over 10 km below the inland shot 'P' (zero offset). South of Shot P, the thickness of the upper crust decreases to approximately 5 km beneath the

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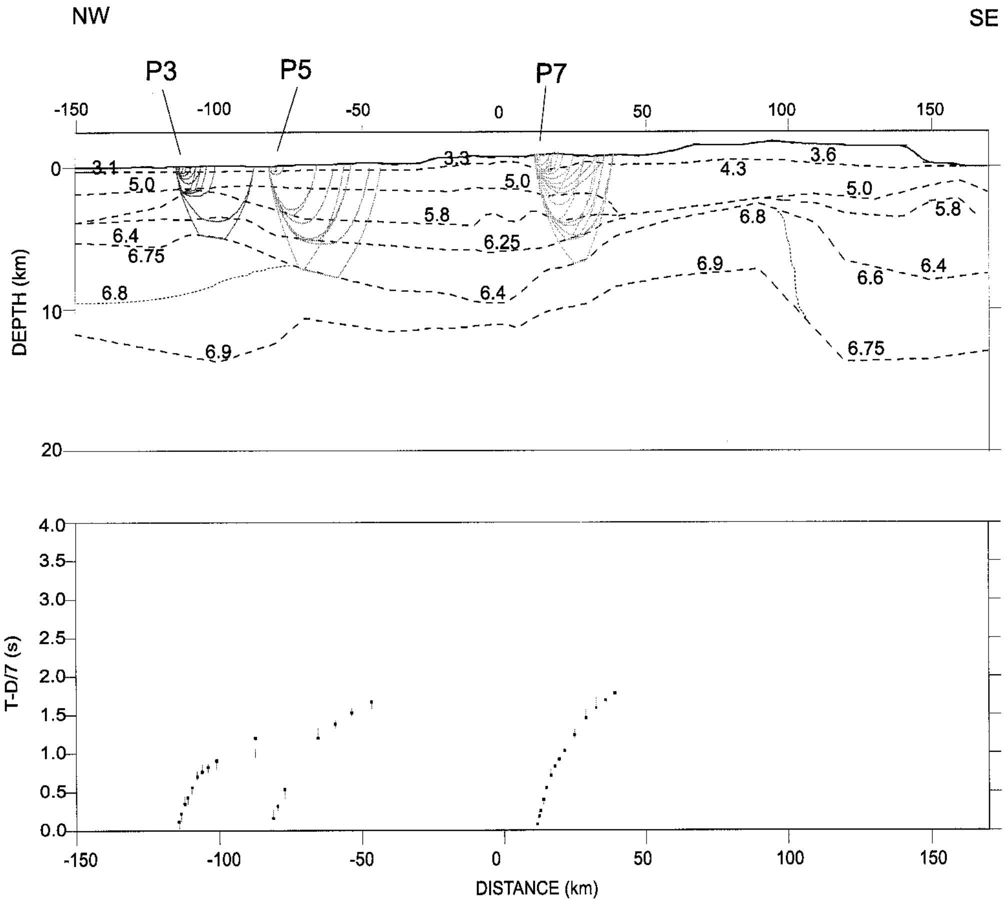

Figure 4. RAYINVR plot showing rays traced for the data from the three Pálmason (1963) profiles. The bottom plot shows the picked traveltimes as small vertical bars of size proportional to pick uncertainty, and the traveltimes calculated by the RAYINVR code are superimposed as small black squares. Dashed lines on the ray-tracing plot represent isovelocity contours in the model.

Fjördungsvatn shot (F; offset 33 km). In this region, the base of the upper crust is characterized by a seismic velocity of almost 6.6 km s^{-1}.

From Fjördungsvatn to central Vatnajökull (+40 to +100 km), the profile passes over a region containing several active central volcanoes above the centre of the mantle plume. Here, the nature of the upper crust is very different to that below other sections of the profile. To the southeast of the Fjördungsvatn shot (F), arrivals with seismic velocities characteristic of the lower crust occur early in the seismic section. The upper crust appears to thin greatly below the central volcanoes with minimum depth to the base at 2.6 km below northern Vatnajökull (+90 km along profile). Upper-crustal velocities are low, with the region of velocities of 5.5–6.2 km s^{-1} either greatly thinned or completely absent. This result agrees with previous findings (Pálmason 1971; Flóvenz 1980; Toomey & Foulger 1989; Bjarnason *et al.* 1993; Gudmundsson *et al.* 1994; Staples *et al.* 1997; Brandsdóttir *et al.* 1997), which note that upper- and lower-crustal velocities are elevated beneath numerous central volcanoes. Similarly, we interpret the elevation of velocities in the mid–upper crust below two extinct central volcanoes on the profile, the Molduxi volcano (−120 to −90 km) and the Vedurárdalur volcano (140 to 165 km) (Jóhannesson & Saemundsson 1998), to be caused by the same crustal processes as are currently at work below the active Vatnajökull volcanoes (see Fig. 1 for the locations of the volcanoes).

At the southeastern end of the profile, the upper-crustal structure is similar to that at the northern end of the profile. The upper crust is separable into two units, one with seismic velocities increasing from 3.5 to 5.9 km s^{-1} and the second with seismic velocities increasing from 6.2 to 6.4 km s^{-1}. The base of the upper crust lies at 5–6 km depth, with little variation in depth.

The upper crust has been defined by Flóvenz (1980) and later authors as the region of high velocity gradients in the Iceland crust. Flóvenz (1980) reported velocity gradients of 0.5–0.6 s^{-1}. Here, a typical velocity gradient (measured in the continuous region of the northern highlands) is approximately 0.21 s^{-1}.

6 THE MID-CRUSTAL REGION

Below the base of the upper crust is a region with seismic velocities between 6.6 and 6.9 km s^{-1}, with variable velocity gradients along the profile. The depth range of this region varies along the profile; it lies between the base of the upper crust and a depth of 9–12 km. The velocities are characteristic of lower-crustal material by conventional definitions, but we

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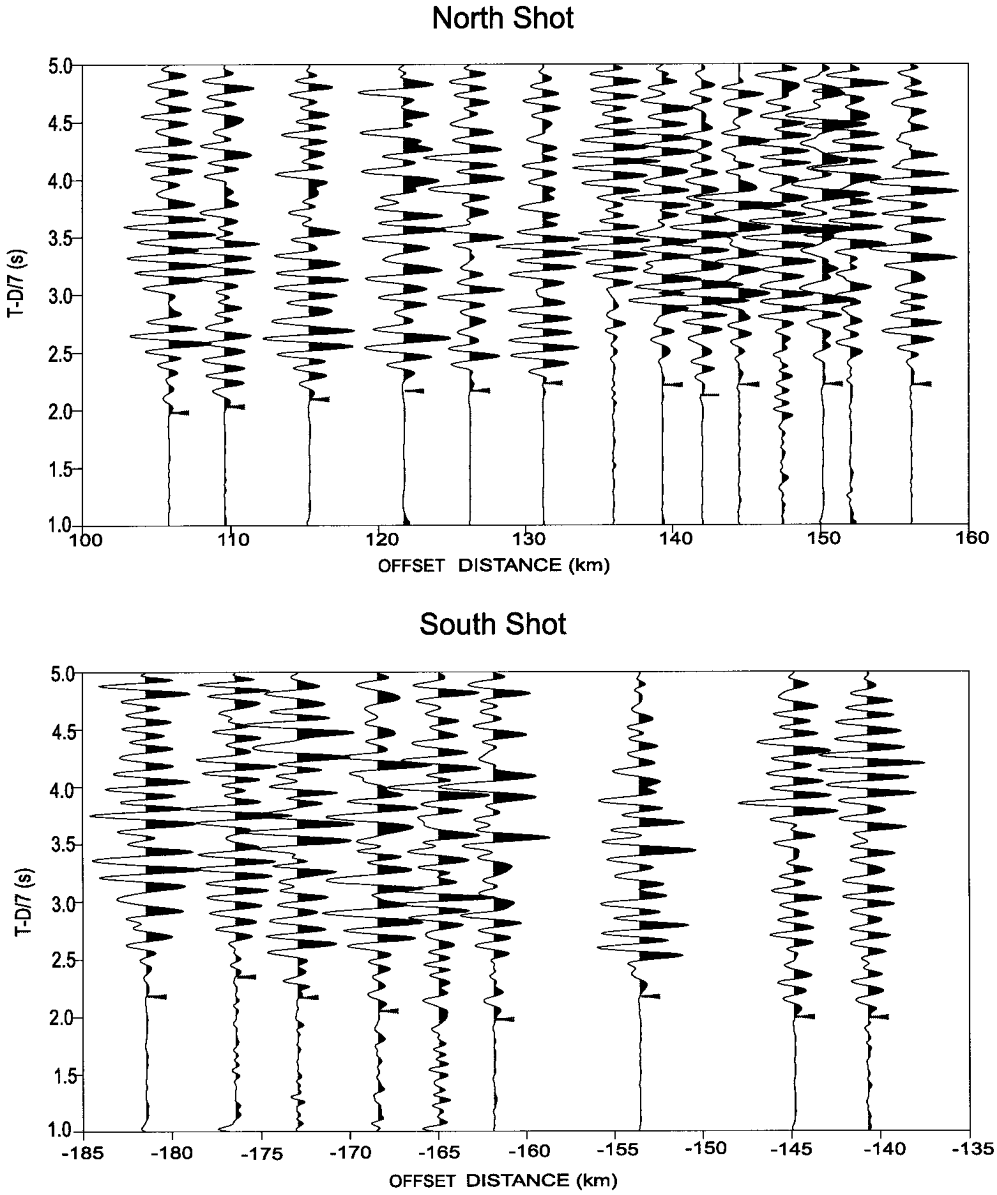

Figure 5. Part of the trace-normalized record sections for the North and South shots, showing picks (arrowheads) made for the emergent arrivals between 100 and 200 km offset.

describe this region separately due to the variable nature of the velocity gradients.

From Skagi to central Vesturdalur (-150 to -80 km), seismic velocities increase from 6.75 to 6.85 km s^{-1} within the mid-crustal region. The low velocity gradient of 0.017 s^{-1} at the tip of Skagi (-150 km) is characteristic of lower-crustal material, but the velocity gradient of 0.060 s^{-1} in this zone below Vesturdalur (-80 km) is intermediate between upper- and lower-crustal velocity gradients.

From Vesturdalur to central Vatnajökull (-80 to $+100$ km), seismic velocities within the mid-crustal region increase from 6.8 to 6.9 km s^{-1}. Here, the crustal velocity gradient varies from 0.021 s^{-1} below Vatnajökull (measured at a distance of $+90$ km) (this value is typical of lower-crustal velocity gradients) to 0.067 s^{-1} below shot P, where the upper crust is thickest.

The nature of the mid-crustal region below southern Vatnajökull and the southeastern end of the profile ($+100$ to $+170$ km) is slightly different from that found elsewhere along the profile. The seismic velocity at the upper boundary is 6.6 km s^{-1}, increasing to 6.75 km s^{-1} at the lower boundary. Crustal velocity gradients, at 0.036 s^{-1}, are intermediate between those characteristic of the upper crust and those observed in the lower crust.

7 THE LOWER CRUST AND MOHO

The lower crust is characterized by high seismic velocities and a low velocity gradient (<0.03 s^{-1}). In the ICEMELT model, lower-crustal velocities increase from 6.9 to 7.1 km s^{-1} at the northwestern end of the profile to 7.2 km s^{-1} in the central and southeastern sections of the model, with a velocity gradient

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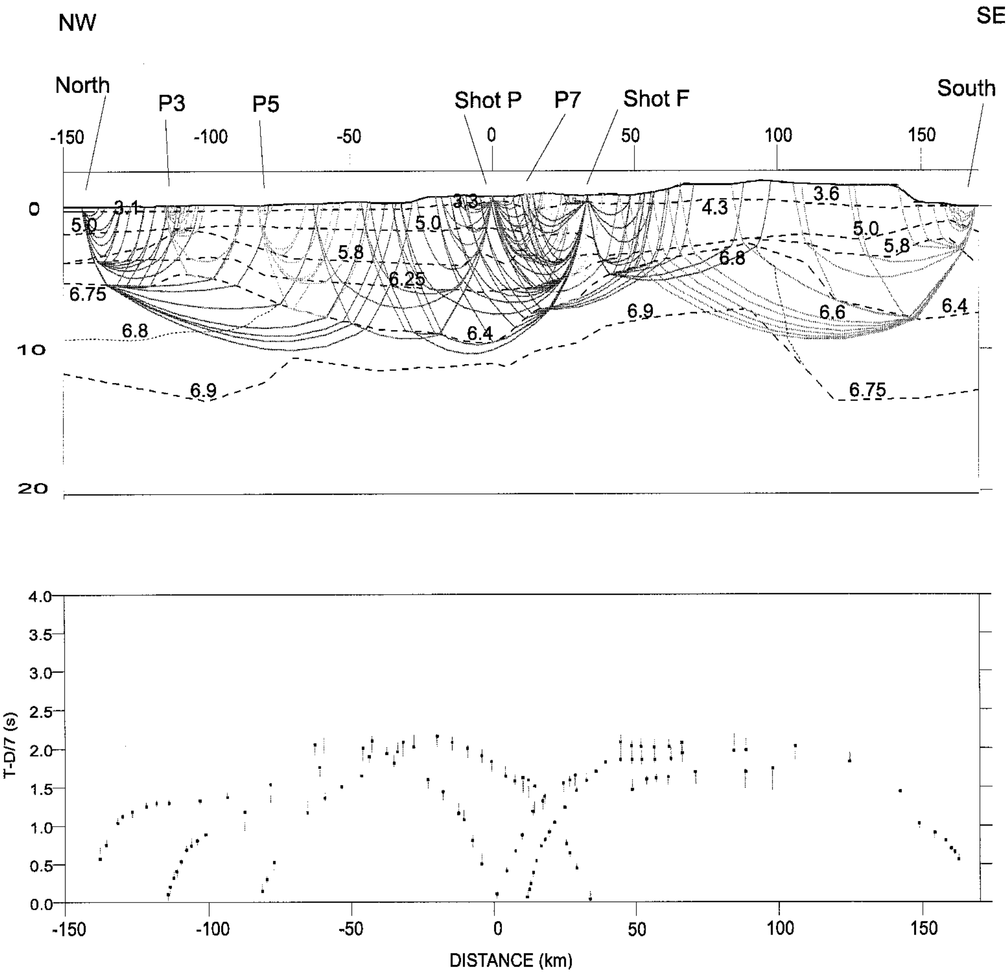

Figure 6. RAYINVR plot showing ray coverage of the upper crust of Iceland along the ICEMELT profile. Shot locations are labelled. Picked traveltimes are shown as vertical bars on the bottom plot, with size proportional to pick uncertainty, and traveltimes calculated by RAYINVR are superimposed as small black squares. Dashed lines on the ray-tracing plot represent isovelocity contours in the model.

between 0.015 and 0.021 s^{-1}, an order of magnitude smaller than that found in the upper crust. Diving rays from the northern and southeastern offshore shots resolve the velocity structure of the lower crust down to 24 km depth below central Iceland, where the seismic velocity is almost 7.2 km s^{-1}. Below this depth, the seismic velocity is assumed to remain constant down to the base of the crust. An alternative assumption would be to extrapolate the velocity to the base of the crust using the measured lower-crustal velocity gradients, which would give a seismic velocity of 7.35 km s^{-1} at the base of the crust of central and southeastern Iceland. The effects of the two different models on Moho depth have been examined; the difference between the two models is negligible in the southern half of the model. The assumption of constant velocity from the 7.2 km s^{-1} isovelocity contour is favoured as it provides a better fit to the Moho reflection than the alternative.

The velocities in the deepest part of the lower crust are higher than those observed in standard oceanic crust (Sinha *et al.* 1981; White & McKenzie 1995) but are thought to be typical of lower-crustal material found above mantle plumes (Sinha *et al.* 1981; White & McKenzie 1995) and have been observed from several refraction profiles in Iceland that sample the lower crust (e.g. Gebrande *et al.* 1980; Bjarnason *et al.* 1993). These high-velocity rocks are thought to contain a higher proportion of MgO than standard oceanic lower-crustal rocks (Bjarnason *et al.* 1993; White & McKenzie 1995; Staples *et al.* 1997).

The seismic Moho is resolved by P_mP reflections from both of the large offshore shots. In the northwest, the Moho lies at 25 km depth, increasing southeastwards to a maximum depth of 40 km below northwestern Vatnajökull (+30 to +50 km). The great Moho depth under Vatnajökull suggests that the crustal thickening caused by enhanced melt production at the plume centre far outweighs the crustal thinning associated with the rift zones (*cf*. crustal thickness measurements: Bjarnason *et al.* 1993; Staples *et al.* 1997). The crustal thickness of 25 km below the Skagi Peninsula in the north is similar to that observed by Menke *et al.* (1998) and Staples *et al.* (1997) on the western flank of the Northern Volcanic Zone.

Mantle velocities cannot be constrained using the refraction line data, as no P_n (mantle diving ray) arrivals were observed. The P_n phase has previously been observed in Iceland; Bjarnason *et al.* (1993, 1994) reported a refracted arrival of apparent velocity 7.7 km s^{-1} in the SIST profile, which was interpreted as P_n. Modelling of P_mP phases inferred a Moho with a dip of approximately 2°, which suggests that the

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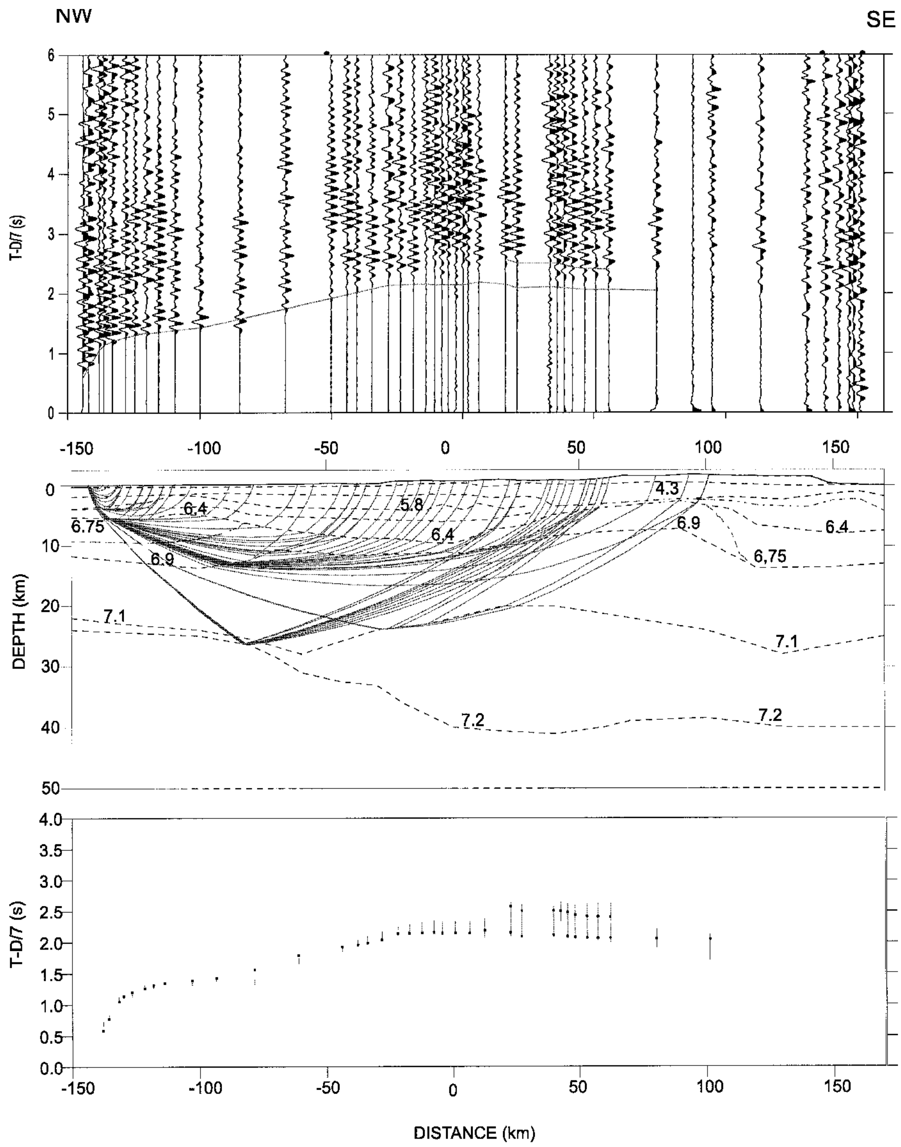

Figure 7. Top: trace-normalized record section for the North shot. Reduction velocity is 7.0 km s^{-1}, and traces are bandpass filtered from 2.5 to 15.0 Hz. Calculated traveltimes from the RAYINVR code are overlaid on the seismic traces. Middle: ray diagram for the northernmost shot showing crustal diving rays and Moho reflections. Dashed lines on the ray-tracing plot represent isovelocity contours in the model. Bottom: picked traveltimes (vertical bars) and traveltimes calculated by the RAYINVR code (small black squares).

apparent velocity of the refracted arrival is close to the velocity of the uppermost mantle below the profile. Menke *et al.* (1996) recorded an apparent P_n velocity of 8.0 km s^{-1} for a path across southern central Iceland, although the dip of the Moho is unknown in this region. The P_n arrival is generally of low amplitude; at large offsets the signal to noise ratios on the ICEMELT record sections are too low to resolve P_n.

8 MODEL RESOLUTION

It is important in a study such as this to indicate where the crustal model is well resolved and constrained by the data points. To achieve this, we use three complementary methods.

(1) Fig. 13(a) shows a plot of rays traced through the crustal model. This provides a good qualitative estimate of the regions of the model between shots and receivers illuminated by seismic energy. The best constraint is achieved where many ray paths are traced through a particular portion of the model, particularly where ray paths from different shots intersect at a wide range of angles. This occurs mostly in the depth range 1–15 km in the central section of the profile. Although much of the lower crust is sampled by several rays from P_g and P_mP phases (north and south shots), few of these ray paths intersect. Below Vatnajökull (+70 to +130 km), the upper crust is sampled by few rays, giving poor model constraint in this region. It is possible to estimate the thickness and average velocity of the upper crust in this part of the profile, but not to provide details of short-wavelength variations.

(2) To provide a more quantitative estimate of resolution in the crustal model, we follow the method of Zelt & Smith (1992). The RAYINVR model is parametrized by an arbitrary number of nodes that define the positions of model boundaries and the seismic velocities between these

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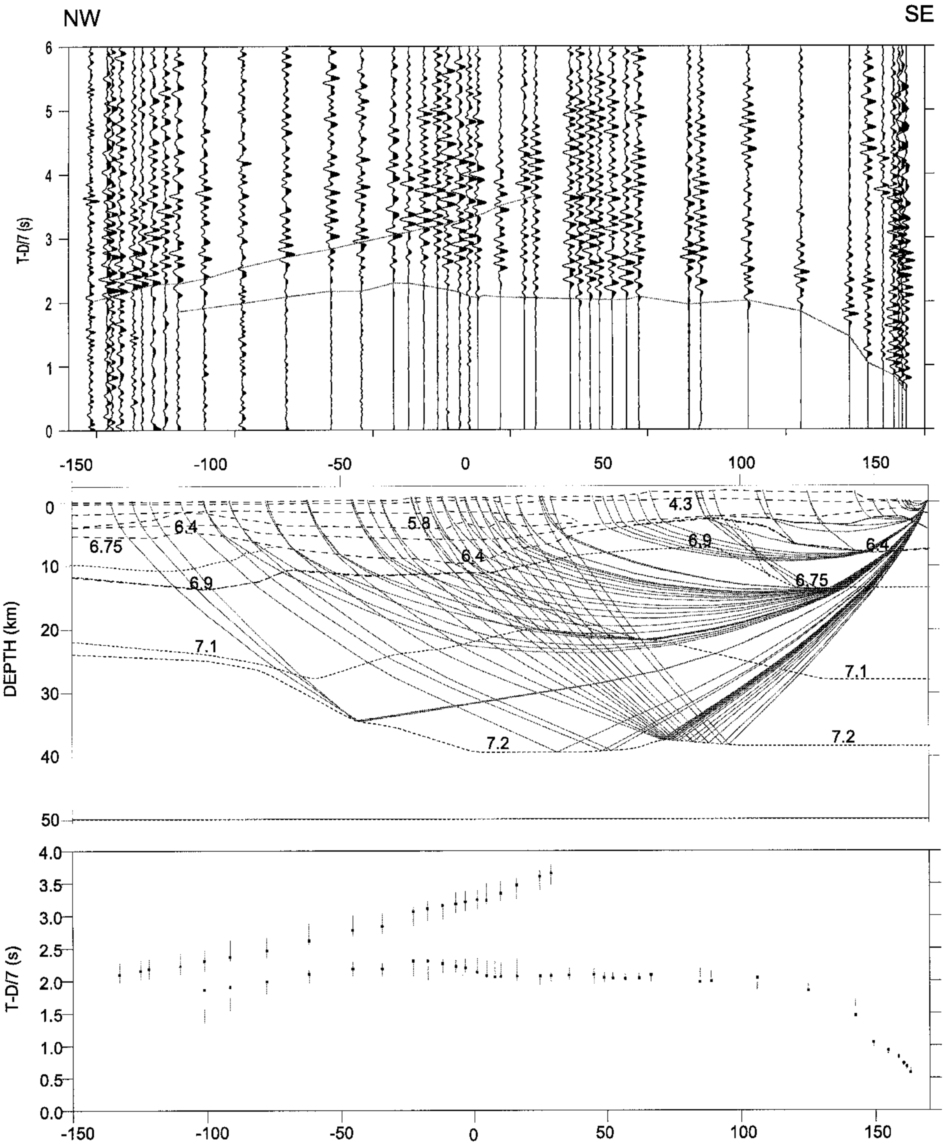

Figure 8. As for Fig. 7, but showing traces, rays and traveltimes for the South shot.

boundaries. Resolution in a region of the model is dependent on the number of rays that constrain the nodes in that region. Fig. 13(b) shows the resolution plot for the ICEMELT model achieved by following the method of Zelt & Smith (1992). According to this method, a resolution value of greater than 0.5 indicates a region of the model that is well resolved and reliable. It should be noted that this method of estimating model resolution does not take into account the improvement in constraint afforded by the presence of crossing ray paths.

The ICEMELT model shows rather poor numerical resolution values in the upper crust, even in the central section of the model, which is constrained by many crossing ray paths. This is because many nodes are required to model the short-wavelength variations in structure in order to fit the picked traveltimes. The ray density per node is therefore small in much of the upper crust. In such cases there are strong trade-offs between achieving a close traveltime fit to the data points and achieving high numerical resolution values. It is possible to decrease the spatial resolution of the model by inserting fewer nodes, thereby increasing the numerical resolution. However, this would underparametrize the model as individual traveltimes could not be fitted so well. The best numerical resolution achieved is in the middle depth range of the crust, where the traveltime picks can be fitted by a section of the model with comparatively few nodes.

(3) An estimate of uncertainty in the depth of any chosen boundary is calculated by perturbing the depth of that boundary, keeping all other parameters in the model fixed, and examining the effect on the resultant misfits between the calculated and observed traveltimes. For this study, the boundaries at the base of the upper crust and at the Moho were perturbed. Fig. 13(c) shows graphs of the chi-squared value, a statistical measure of the model misfit, against the change in depth of the interface for these two boundaries. The chi-squared value is given by

$$\chi^2 = \frac{1}{n-1} \sum_{i=1}^{n} \left(\frac{t(i)_{\mathrm{calc}} - t(i)_{\mathrm{obs}}}{u(i)} \right)^2 , \qquad (1)$$

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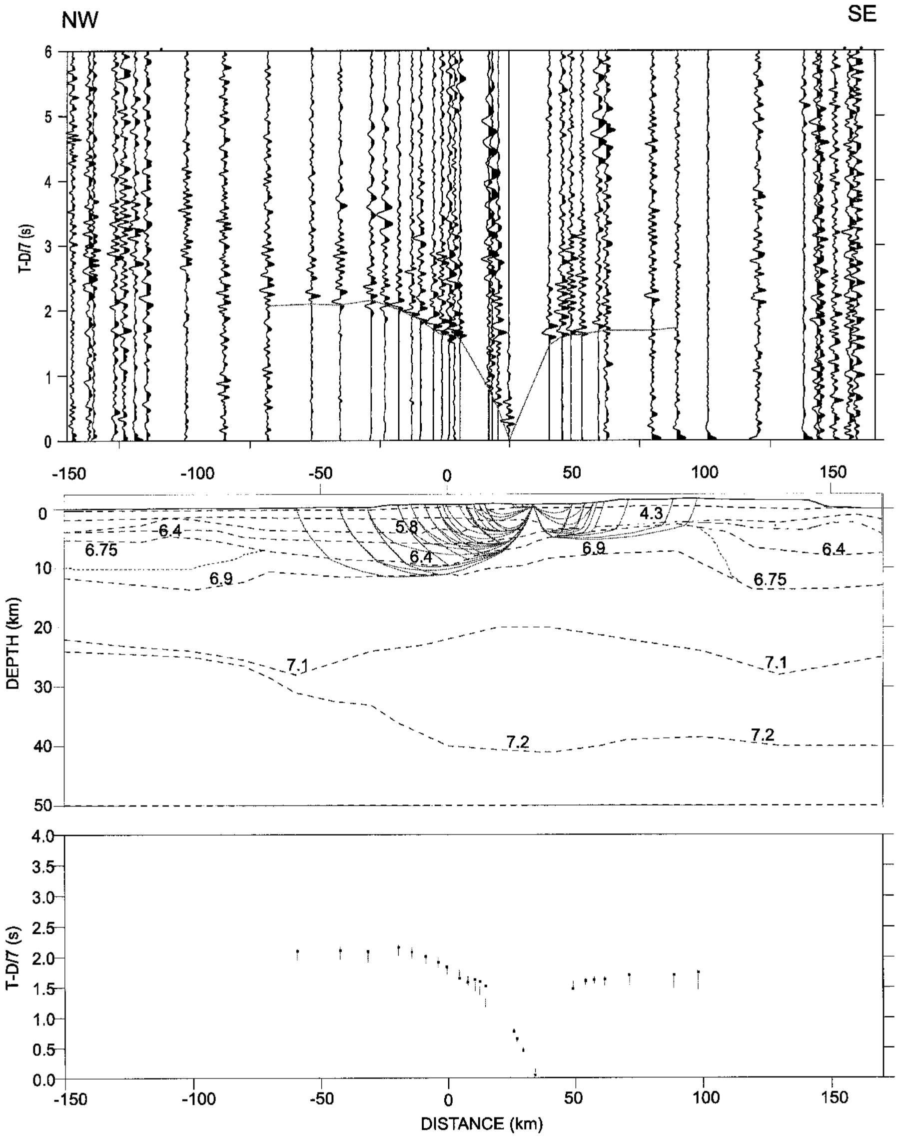

Figure 9. As for Fig. 7, but showing traces, rays and traveltimes for the Fjördungsvatn shot, F, central highlands.

where n is the number of rays traced successfully, $u(i)$ is the estimated pick uncertainty for the ith observation, $t(i)_{calc}$ is the calculated traveltime and $t(i)_{obs}$ is the observed traveltime. When the crustal model is perturbed, changes in boundary positions and velocity gradients can result in a loss of rays traced through a particular region of the model, therefore the number of data points successfully fitted is a useful indication of the success of the model. The uncertainty estimate is the value of the perturbation at which the fit to the model becomes statistically unacceptable compared to the fit afforded by the final model, quantified by an F-test (Press *et al.* 1992), which was used to place 95 per cent confidence limits on the chi-squared values and number of data points fitted. Using this test gives the following depth uncertainties:

base of upper crust—depth uncertainty is ± 0.6 km;
Moho transition—depth uncertainty is ± 2 km.

These three methods give some sense of how well resolved the ICEMELT crustal model is, but there are further factors that should also be taken into account.

(a) The resolution of the model is limited by the wavelength of the seismic waves from the shots. The dominant frequency for P_g phases was 7 Hz and for P_mP phases was 3–4 Hz, giving wavelengths of 0.5–2 km.

(b) A compromise must be reached between fitting the data points well and overparametrizing the crustal model.

(c) A further source of uncertainty in the depth of the Moho occurs due to the lack of constraint of the seismic velocity below 24 km depth, as mentioned earlier in this paper. Two possibilities were explored: an extrapolation of the lower-crustal velocity gradient below 24 km depth and a zero velocity gradient below 24 km depth. The difference between best-fitting Moho depths for these two models was less than 1 km.

Although the model appears to be poorly constrained in places, the resolution and uncertainty tests performed show that we can have confidence in the main features of the ICEMELT crustal model. The upper crust thickens to the north of the volcanic rift zone and thins significantly below

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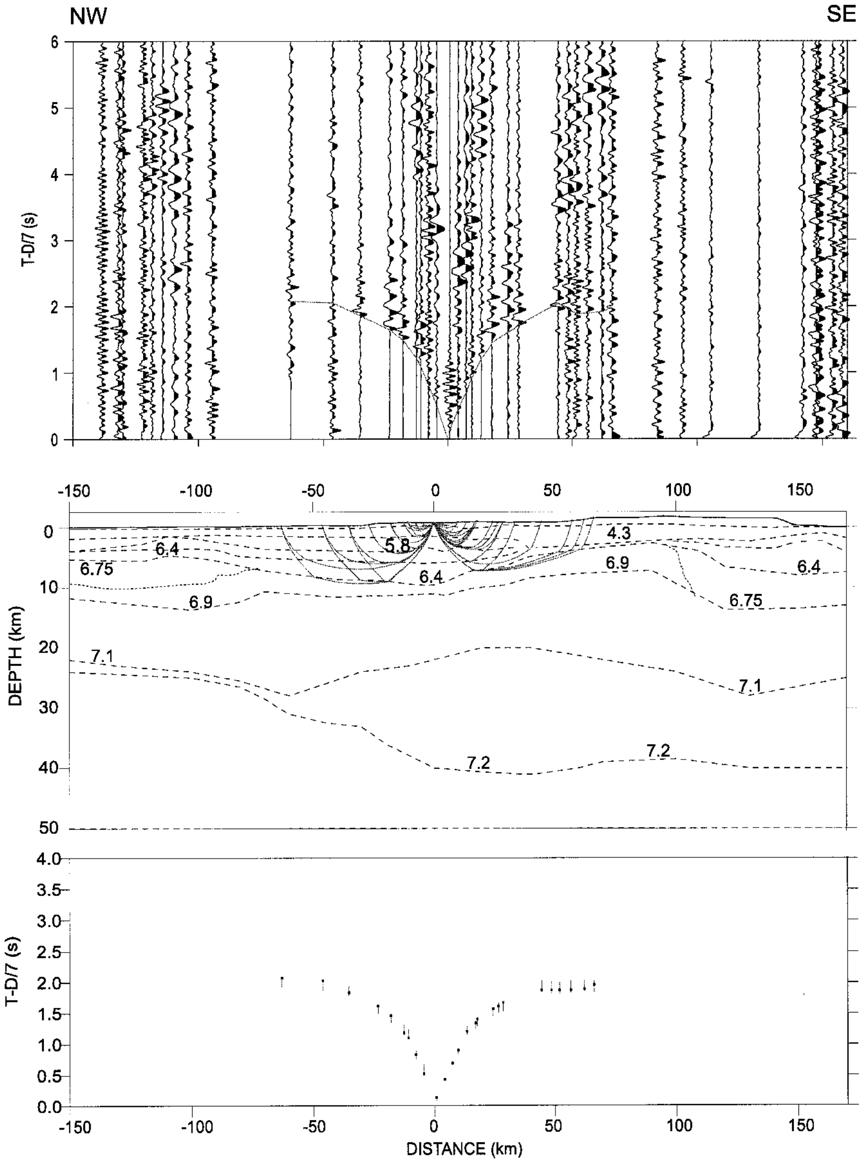

Figure 10. As for Fig. 7, but showing traces, rays and traveltimes for the 'Pond' shot, P, in the northern central highlands.

the central volcanoes of northwestern Vatnajökull. The Moho depth increases from approximately 25 km at the northwestern end of the profile to 38–40 km above the centre of the Iceland mantle plume. These variations in upper-crustal thickness and Moho depth are larger than can be attributed to errors in the data picks or in the modelling procedure.

9 COMPARISON OF *P*- AND *S*-WAVE VELOCITIES

The *S* waves recorded by the horizontal component seismometers deployed on the ICEMELT line are not, in general, of sufficiently good quality to add further information to the traveltime model already generated by using *P*-wave arrivals. However, it was possible to pick *S*-wave arrivals for all four of the principal shots and to compare their traveltimes to those of the corresponding *P*-wave arrivals (Fig. 14). *S* waves are visible with almost equal clarity on both the radial and the transverse components of the seismic receivers.

The radial component record sections shown in Fig. 14 show that the clarity of the *S*-wave arrivals varies considerably between shots. The northernmost shot shows rather poor *S* waves; arrivals are emergent out to 40 km offset, and are barely visible at greater offsets. Both highland shots show clear impulsive *S*-wave first arrivals at offsets out to 50–100 km, though the southeastern arrivals from the Fjördungsvatn shot (F) deteriorate beyond 30 km offset. The clearest *S*-wave arrivals are seen in the southeastern shot record section. Although mainly emergent in character, the arrivals can be observed out to at least 250 km offset as S_g and S_mS phases.

Traveltimes of the *S* arrivals were picked by hand. Comparison with *P*-wave traveltimes yields a direct measurement of the *P*-wave to *S*-wave seismic velocity ratio, if one assumes a constant Poisson's ratio over the ray path. From the ratio V_p/V_s, it is possible to calculate Poisson's ratio for the crust and upper mantle, and to make deductions about the properties of the rocks.

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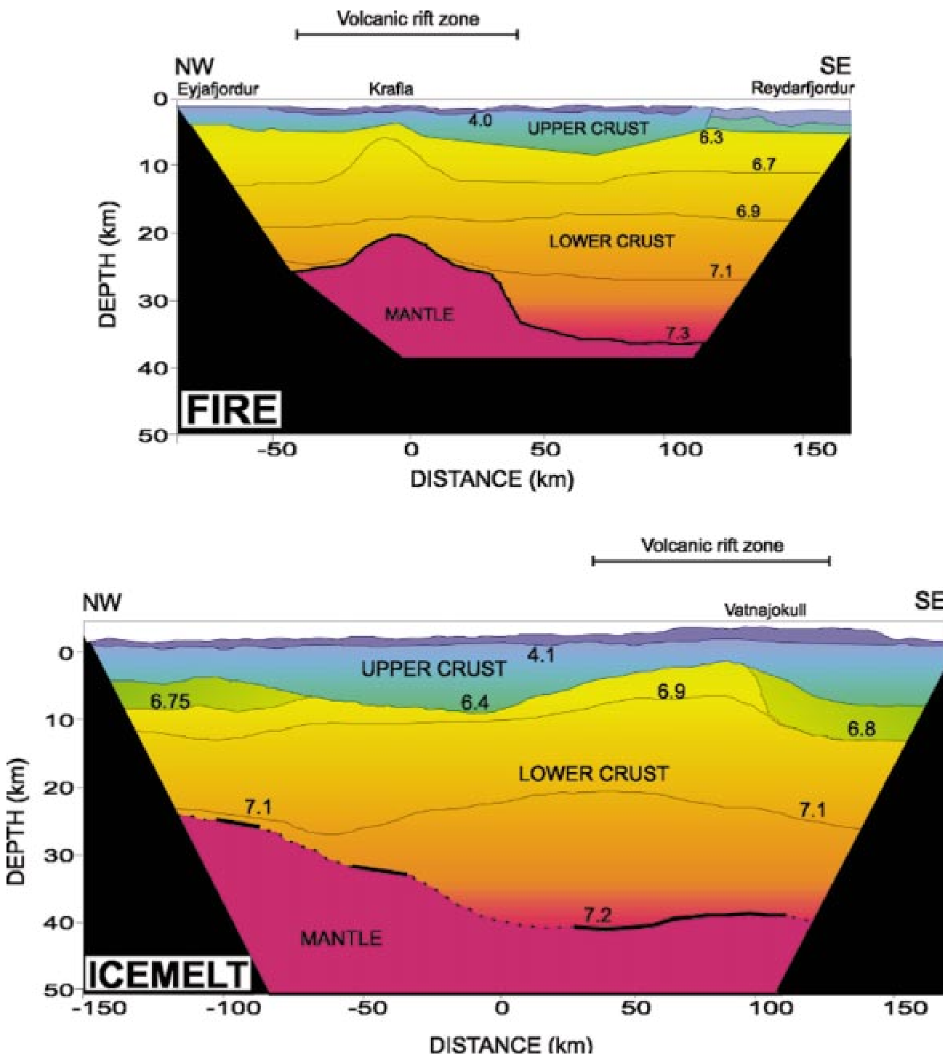

Figure 11. Top: crustal model for the Northern Volcanic Zone from the FIRE 1994 project, crossing the spreading axis at zero offset (Staples *et al.* 1997). Bottom: crustal model for the ICEMELT refraction line (this study). The model crosses the volcanic rift zone at distances along the profile between 50 and 100 km, passing directly over the plume core in this region. Solid lines on the Moho represent sections of the Moho constrained by reflected seismic energy.

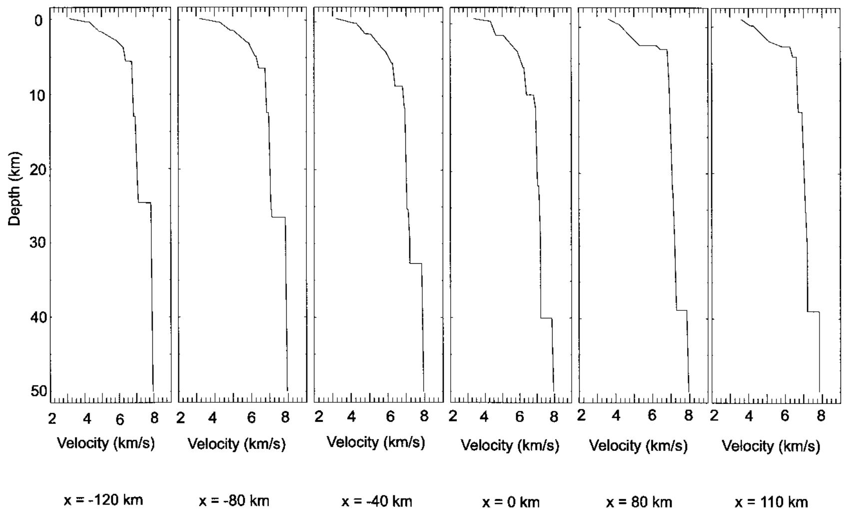

Figure 12. 1-D velocity profiles from the ICEMELT crustal model at offsets of −120, −80, −40, 0, 80 and 110 km from the 'Pond' highland shot (P).

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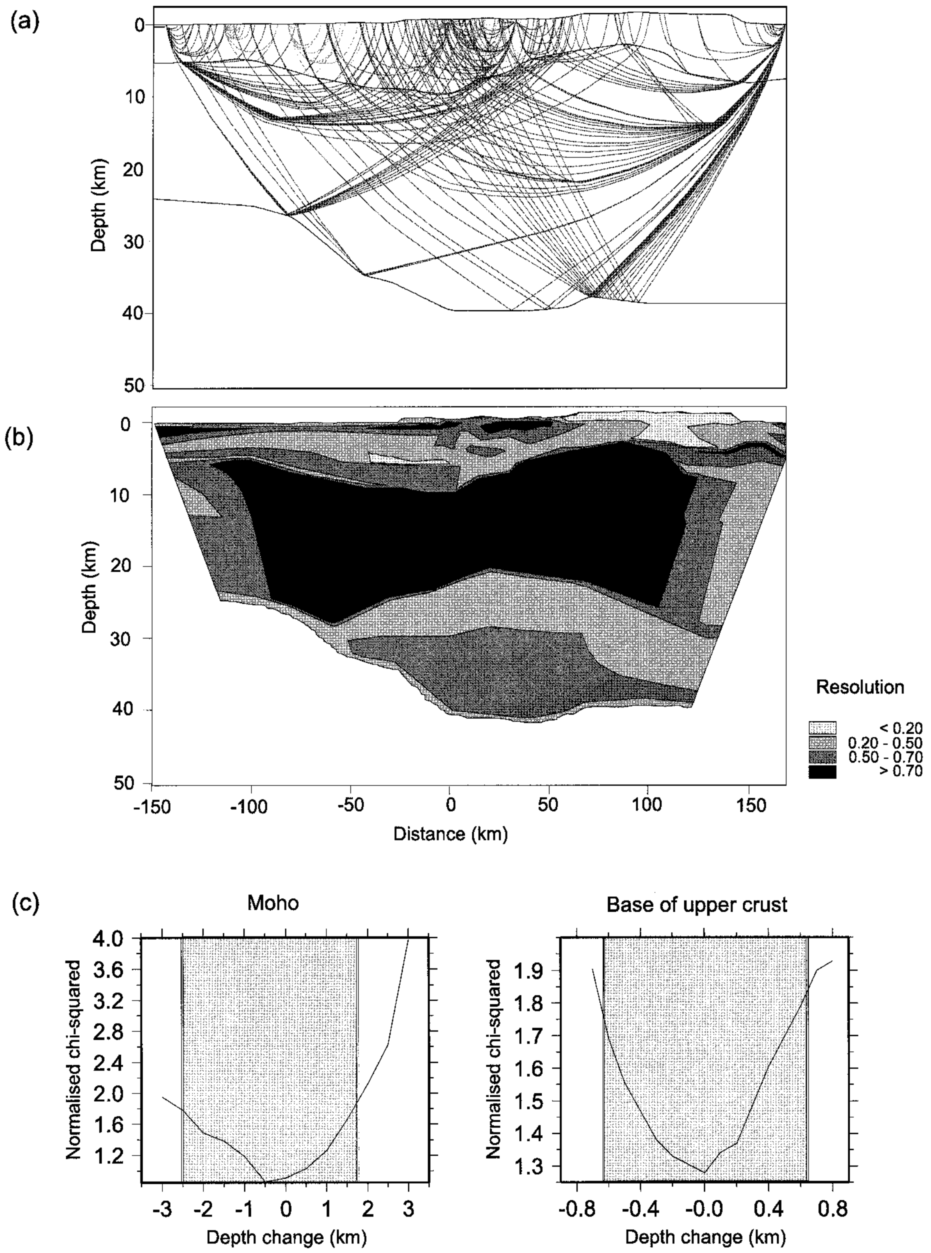

Figure 13. (a) Plot of all rays between shots and receivers traced through the ICEMELT crustal model by the RAYINVR code, using two-point ray tracing. (b) Resolution plot for the ICEMELT crustal model, using the method of Zelt & Smith (1992). Regions of the model with numerical resolution values of 0.5 or higher are said to be well resolved and reliable. (c) Plots of the chi-squared misfit parameter against boundary depth change for the Moho and for the base of the upper crust. The grey-shaded area represents depth changes within the 95 per cent confidence limits calculated by the F-test method (Press *et al.* 1992). Note that, for the Moho, the lowest chi-squared value occurs at a −0.5 km depth change. This does not, however, represent the best model fit, as the decrease in chi-squared value for this depth change is accompanied by a decrease in the number of picks fitted.

The highland shots show V_p/V_s of 1.76 ± 0.03 in the upper crust of central Iceland, while the large offshore shots, sampling the crust down to approximately 24 km depth, give an average V_p/V_s of 1.78 ± 0.03. These results yield an average upper-crustal Poisson's ratio of 0.26, while the average Poisson's ratio for the crust as a whole is 0.27.

These results compare favourably with previous measurements made in Iceland. The V_p/V_s ratio has been measured at between 1.74 and 1.79 by several authors (Pálmason 1963; Menke *et al.* 1996; Brandsdóttir *et al.* 1997; Staples *et al.* 1997). The Poisson's ratio value of 0.26–0.27 is consistent with a relatively cool crust with temperatures well below both the basalt and the gabbro solidi (Bjarnason *et al.* 1993; Menke & Levin 1994; Menke *et al.* 1996; Staples *et al.* 1997).

10 DISCUSSION

The upper-crustal structure along the ICEMELT profile shows variations that are comparable to those found by previous modern seismic profiles, and can be explained by considering

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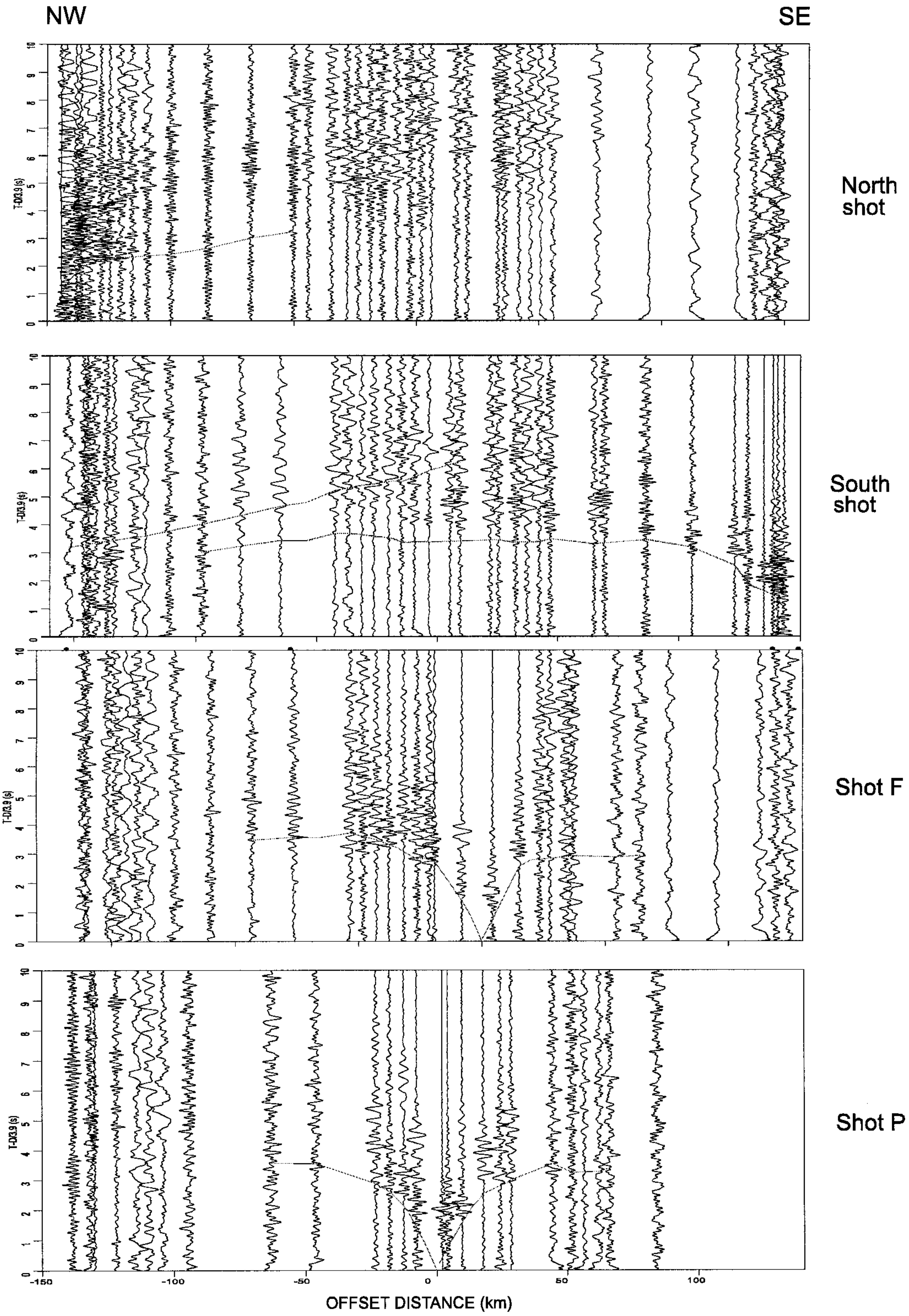

Figure 14. *S*-wave record sections for South, North and highland shots (radial-component seismograms), reduced at 3.9 km s^{-1}. Record sections are trace-normalized, and *P*-wave traveltimes multiplied by 1.78 are superimposed on the sections.

Iceland's present tectonic structure and tectonic history. The thickness of the upper crust at the extreme ends of the profile (-150 to -120 km; 120 to 170 km) is typical of old Icelandic crust (surface rocks of age >3.3 Myr). Beneath the northern highlands, the upper crust almost doubles its thickness, to approximately 10 km. To explain this significant increase we must consider the effect of past ridge jumps. The present Northern Volcanic Zone has been active for only the past 6–7 Myr, while the Eastern Volcanic Zone started to propagate southwards from the present plume centre at only 2 Ma

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(Jóhannesson 1980). Prior to 7 Ma, the Húnaflói–Skagi and Snæfellsnes volcanic zones were the active spreading centres on Iceland; Húnaflói–Skagi ceased activity at ~3 Ma (Helgason 1984, 1985). When the ridge jumps and a new rift zone starts to produce new lava flows, these flows pile up onto already established, cool upper-crustal rocks (Bjarnason *et al.* 1993; Staples *et al.* 1997). This increases the thickness of the upper crust as the old, cool lava flows can be buried by new flows without a significant increase of their seismic velocity due to metamorphism. In addition, the thick upper crust in the northern highlands may also indicate a very small amount of glacial erosion (and consequent uplift) in this region during glacial times.

While the total crustal thickness of central Iceland is well constrained by the presence of Moho reflections, it may initially be considered surprisingly thick in the light of previous modern seismic studies that have crossed the Icelandic volcanic rift zones (Bjarnason *et al.* 1993; Staples *et al.* 1997). These studies placed the Moho at 19–24 km depth below the spreading axis. Thicker crust, with a Moho depth of 35 km, is found in eastern Iceland (Staples *et al.* 1997). However, the ICEMELT profile cannot be said to pass over a conventional Icelandic spreading axis. Instead, it crosses the volcanic rift zone directly above the plume centre. No on-axis crustal thinning is observed below central Iceland; it appears that the melt production at the plume centre is so great as to outweigh the thinning effect of crustal rifting. This enhanced melt production is likely to be caused by two factors acting in tandem. First, the mantle temperature is hottest in a relatively narrow rising region, probably only about 150 km in diameter in the region of greatest melt production (50–100 km depth) (Courtney & White 1986; Watson & McKenzie 1991; White & McKenzie 1995; Wolfe *et al.* 1997). The higher mantle temperatures cause enhanced melt production as the mantle decompresses. Second, active convection in the plume core causes upwelling to occur faster than the rate that would result from purely passive upwelling in response solely to plate separation. This, too, would cause enhanced melt production in the plume core beneath Vatnajökull. Away from the narrow plume core, the mantle is likely to rise passively beneath the volcanic rift zones, reducing the amount of melt generated below these regions (White 1997).

Menke *et al.* (1998) noted that results from the B96 profile along the western flank of the Northern Volcanic Zone showed a thickening of the crust towards central Iceland. This would also appear to be the case on the rift axis itself. Bjarnason *et al.* (1993) and Menke *et al.* (1996; personal communication, 1997) noted an apparent wide-angle reflection on the record section of shot D of the 1977 RRISP seismic profile (Gebrande *et al.* 1980), which had originally been interpreted as a reflection from a thin lens of high-velocity material. In the light of more recent results, this arrival was reinterpreted as the Moho reflection, P_mP. Subsequent traveltime modelling led to a crustal thickness measurement of 35 km in the rift zone between northern Vatnajökull and the Askja central volcano in the southern part of the Northern Volcanic Zone.

The northern section of the profile runs across the east side of the extinct Húnaflói–Skagi Volcanic Zone, and it is interesting to note that the crustal thickness of 25 km in this region is similar to that modelled by Menke *et al.* (1998) from the B96 profile, which runs along the western flank of the current northern spreading centre.

P-wave to *S*-wave velocity ratios suggest that the entire crust of central Iceland is subsolidus, even though it lies directly above an active mantle plume, with temperatures elevated above those of normal mantle (White & McKenzie 1995; White *et al.* 1995; Ito *et al.* 1996; White 1997; Wolfe *et al.* 1997). The temperature anomaly of the plume centre has been estimated using a variety of methods. Results from geochemical analysis of Icelandic basalts, uplift in the Iceland region and crustal thickness measurements suggest a temperature anomaly of 150–200 °C (White & McKenzie 1995; White *et al.* 1995; White 1997.) Wolfe *et al.* (1997) inferred a mantle temperature anomaly of 200–300 °C based on velocity anomalies modelled from teleseismic tomography, taking anelastic dispersion into account.

Fig. 15 shows the change in crustal thickness and bathymetry–topography along the ridge axis from the Reykjanes Ridge, through Iceland and on to the Kolbeinsey Ridge. The effects of the narrow zone of greatly enhanced melt production in the core of the plume are clear. A narrow region of very thick crust is created (Fig. 15, top), and considerable topographic elevation occurs (Fig. 15, bottom). Away from the central core of the plume, the mantle temperature and, correspondingly, the elevation and crustal thickness, decrease gradually over a distance of more than 600 km.

11 CONCLUSIONS

The crust below the ICEMELT refraction profile is 25 km thick in the north and increases in thickness to a maximum of 38–40 km above the plume centre. This is presumed to be caused by increased melt generation due to the increased temperature of the mantle and the active convection occurring in the plume core.

Large variations in the nature and thickness of the upper crust occur along the profile. A two-layered structure, within which seismic velocities increase with depth, is found in the north and south of Iceland, with a total upper-crustal thickness of 5–6 km. In the northern highlands, the seismic velocities of the upper crust increase continuously with depth and the thickness is almost 10 km. Below the central volcanoes of northern Vatnajökull, which lie at the northernmost limit of the Eastern Volcanic Zone, the upper crust thins to less than 3 km. The seismic velocities in the mid–upper crust are also elevated beneath two extinct central volcanoes at the north and south ends of the profile.

The ratio of *P*-wave to *S*-wave velocity along the profile is approximately 1.76 in the upper crust alone and 1.78 in the crust as a whole, giving a Poisson's ratio of 0.26–0.27. This value is consistent with a subsolidus crust across the region sampled.

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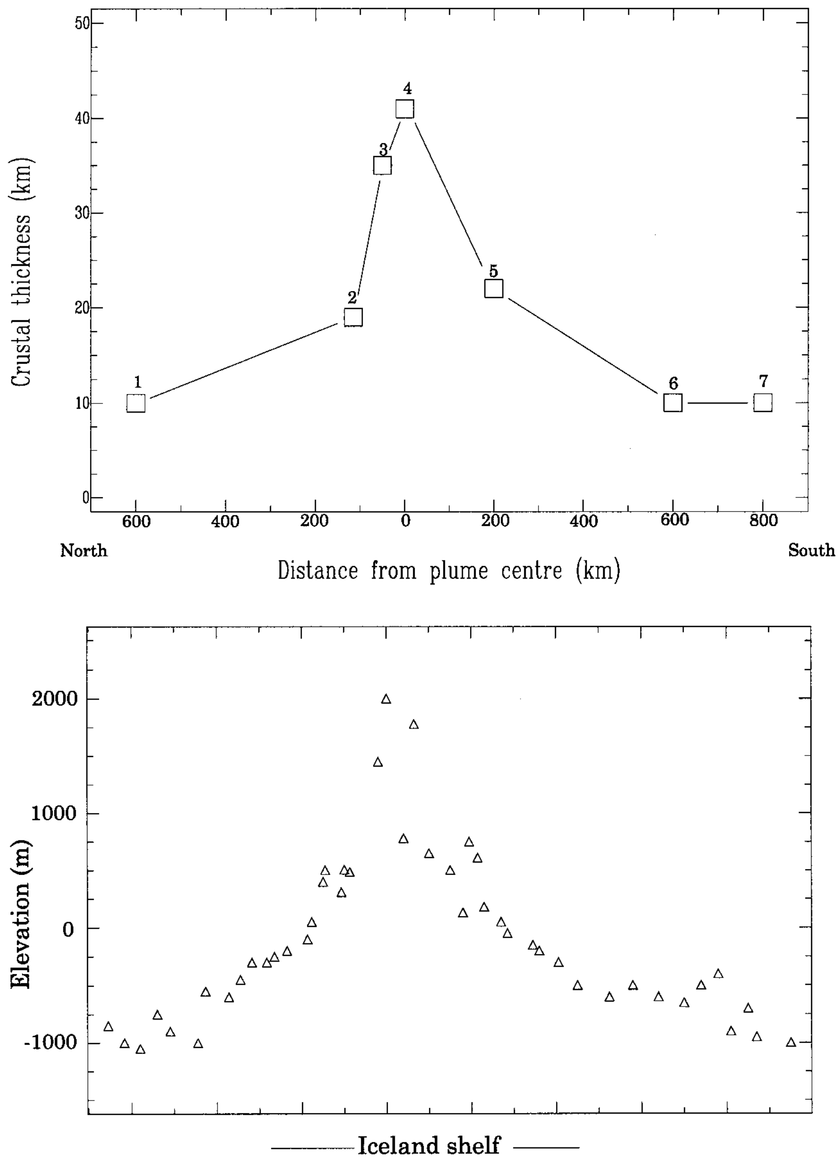

Figure 15. Top: on-axis crustal thickness in the Iceland region, plotted against distance from the plume centre (taken to be the location of the Bárdarbunga central volcano in Iceland). Results from the Reykjanes Ridge, the Kolbeinsey Ridge and Iceland itself are shown. 1. Kodaira *et al.* (1997); 2. Staples *et al.* (1997); 3. Menke *et al.* (1998); 4. this study; 5. Bjarnason *et al.* (1993); 6. Smallwood *et al.* (1995); 7. Bunch & Kennett (1980). Bottom: plot of bathymetry and topography of the axial spreading centre in the Iceland region, plotted against distance from the plume centre.

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