Icelandic-type crust

G. R. Foulger, Z. Du and B. R. Julian

Department of Geological Sciences, University of Durham, Durham DH1 3LE, UK. E-mail: G.R.Foulger@durham.ac.uk
Institute of Theoretical Geophysics, University of Cambridge, Cambridge DB2 3EQ, UK. E-mail: Zhijun.Du@itg.cam.ac.uk
US Geological Survey, 345 Middlefield Rd, Menlo Park, CA 94025, USA. E-mail: julian@usgs.gov

SUMMARY
Numerous seismic studies, in particular using receiver functions and explosion seismology, have provided a detailed picture of the structure and thickness of the crust beneath the Iceland transverse ridge. We review the results and propose a structural model that is consistent with all the observations. The upper crust is typically 7 ± 1 km thick, heterogeneous and has high velocity gradients. The lower crust is typically 15–30 ± 5 km thick and begins where the velocity gradient decreases radically. This generally occurs at the \( V_p \) ∼ 6.5 km s\(^{-1}\) level. A low-velocity zone ∼10 000 km\(^2\) in area and up to ∼15 km thick occupies the lower crust beneath central Iceland, and may represent a submerged, trapped oceanic microplate. The crust–mantle boundary is a transition zone ∼5 ± 3 km thick throughout which \( V_p \) increases progressively from ∼7.2 to ∼8.0 km s\(^{-1}\). It may be gradational or a zone of alternating high- and low-velocity layers. There is no seismic evidence for melt or exceptionally high temperatures in or near this zone. Isostasy indicates that the density contrast between the lower crust and the mantle is only ∼90 kg m\(^{-3}\) compared with ∼300 kg m\(^{-3}\) for normal oceanic crust, indicating compositional anomalies that are as yet not understood. The seismological crust is ∼30 km thick beneath the Greenland–Iceland and Iceland–Faeroe ridges, and eastern Iceland, ∼20 km beneath western Iceland, and ∼40 km thick beneath central Iceland. This pattern is not what is predicted for an eastward-migrating plume. Low attenuation and normal \( V_p/V_s \) ratios in the lower crust beneath central and southwestern Iceland, and normal uppermost mantle velocities in general, suggest that the crust and uppermost mantle are subsolidus and cooler than at equivalent depths beneath the East Pacific Rise. Seismic data from Iceland have historically been interpreted both in terms of thin–hot and thick–cold crust models, both of which have been cited as supporting the plume hypothesis. This suggests that the plume model for Iceland is an a priori assumption rather than a hypothesis subject to testing. The long-extinct Ontong–Java Plateau, northwest India and Paraná, Brazil large igneous provinces, beneath which mantle plumes are not expected, are all underlain by mantle low-velocity bodies similar to that beneath Iceland. A plume interpretation for the mantle anomaly beneath Iceland is thus not required.

Key words: crust, Iceland, mantle, plume, receiver functions, seismology.

INTRODUCTION
The term ‘Icelandic-type crust’ was first proposed by Bott (1974) who noted that the crust beneath the Iceland transverse ridge differs fundamentally from both oceanic and continental crust. He pointed out that individual layers are generally equivalent to those of normal oceanic crust, but are much thicker and more variable. This was founded on seismic work performed both in Iceland and on the Iceland–Faeroe ridge, commencing with the pioneering work of Båth (1960), who measured long explosion seismic profiles in western Iceland.

A great deal of work has been done subsequently to map the structure and thickness of Icelandic-type crust. The data quality and processing techniques have progressively improved. In the case of early explosion profiles, data processing used traditional methods based on first-arrival times only, and models comprised stacks of homogeneous layers. Such models provide only a first-order description of structure. As is the case for oceanic crust in general, layers that are clearly and unambiguously laterally extensive must be defined in terms of velocity gradient.

Icelandic-type crust is laterally inhomogeneous, as has been shown by modern experiments involving dense arrays of threecompartment seismic stations and waveform modelling, tomographic and receiver-function processing techniques (e.g. Bjarnason et al. 1993; Miller et al. 1998b; Du & Foulger 1999). Horizontal velocity gradients commonly exceed vertical gradients, velocities characteristic of the lower crust may become shallower by >10 km over short distances, and even approach the surface. Low-velocity
zones (LVZs) are common and structural variations occur on all spatial scales. It is questionable whether the term ‘Mohorovicic discontinuity’ (Moho) is appropriate for the crust–mantle boundary beneath much of Iceland, which comprises a transition zone several kilometres thick throughout which seismic velocities are gradational. Such structure is predicted by petrogenesis models. Laterally continuous intervals of constant seismic velocity, and the seismological base of the crust, are probably variable petrologically, and this must be considered if seismic results are to contribute to understanding the geology of Icelandic-type crust and the processes that create it.

The assumption that Iceland is underlain by a mantle plume has constrained interpretations of crustal studies for the last three decades, despite the fact that extremely different crustal models have been proposed. In the 1980s Icelandic-type crust was considered to be as thin as ~10–15 km and to be underlain by partially molten upper mantle at temperatures of ~1100 °C (the ‘thin–hot’ crust model). With the acquisition of better data, this was superseded in the 1990s by a model involving a crust ~20–40 km thick with anomalously low temperatures in the lower crust (the ‘thin–cold’ crust model). Both of these models have been interpreted as supporting the plume hypothesis.

In this paper we review the structure of Icelandic-type crust as revealed by a suite of seismological methods, and propose a seismic and petrological model that is consistent with all the observations. We compare this model to both normal-thickness oceanic crust and oceanic large igneous provinces (LIPs) and critically evaluate the results within the framework of the plume hypothesis.

THE ICELANDIC TRANSVERSE RIDGE

Subsequent to the opening of the North Atlantic ~54 Ma, the ~1200 km long Icelandic transverse ridge formed as a result of locally enhanced magmatic activity beneath a section of the mid-Atlantic ridge (MAR) several hundred kilometres in north–south extent (Fig. 1). The ridge formed in the neighbourhood of a major, east–west-trending structure that separates regions with contrasting tectonic histories to its north and south. This structure comprises a complex of fracture zones that have been active at different times (Bott 1985). Those parts of the transverse ridge that are currently submarine formed subaerially and subsequently cooled and subsided below sea level (e.g. Bott 1974; Nilsen 1978). Only Iceland is now above sea level. The transverse ridge is asymmetric about Iceland. The Iceland–Faeroe ridge is longer than, and differently orientated from, the Greenland–Iceland ridge as a result of westward migration of the MAR and a change in spreading direction ~44 Ma (Bott & Gunnarsson 1980).

Because it is subaerial, more is known concerning the geology and tectonics of Iceland than other parts of the MAR. Some 30 spreading segments are exposed, and two fracture zones that offset to the east that part of the spreading plate boundary between the Reykjanes ridge to the south and the Kolbeinsey ridge to the north. In south Iceland the active spreading zone comprises the Western and Eastern Volcanic Zones (WVZ, EVZ), between which lies an east–west-oriented shear zone, the South Iceland Seismic Zone (SISZ). In the north, spreading occurs along the single Northern Volcanic Zone (NVZ). In central Iceland the WVZ, EVZ and NVZ are connected by the Middle Volcanic Zone (MVZ).

The geological and tectonic structure of Iceland is complicated because spreading has occurred about a parallel pair of ridges since ~26 Ma (Bott 1985), and ridge migrations both to the east and west have occurred (e.g. Saemundsson 1979; Foulger 2002). Ex-
Figure 1. Map of the Icelandic transverse ridge showing bathymetric contours and tectonic features in Iceland. The neovolcanic zones are outlined. Spreading segments (volcanic systems) are shown in grey and glaciers in white. RR, Reykjanes ridge; KR, Kolbeinsey ridge; WVZ, Western Volcanic Zone; EVZ, Eastern Volcanic Zone; NVZ, Northern Volcanic Zone; MVZ, Middle Volcanic Zone; SISZ, South Iceland Seismic Zone; WF, Western Fjords; SP, Snaefellsnes Peninsula. The NASP, FIRE, ICEMELT, B96, RRISP, SIST, RISE and SIGMA 1 explosion seismology profiles are labelled. The star indicates the location of the Reydarfjordur shot discussed in the text. Thin lines indicate the profiles of Bøth (1960) and dashed lines indicate the profiles of Palmason (1971) that were longer than 100 km. Circles containing numbers indicate seismic stations of the Iceland Hotspot Project. Circles with three-letter labels indicate other broadband seismic stations in Iceland.
to the upper extrusive sequence and the underlying sheeted dyke complex, including the zeolite and greenschist metamorphic facies. Layer 3 corresponds to gabbro intrusives, grading down through plagiograniyte, olivine gabbro and pyroxenite to wehrlite, troctolite and dunite. The upper part of this sequence corresponds to amphibolite metamorphic facies and the lower part to granulite facies. The seismic Moho is thought to correspond to the boundary between overlying troctolite and underlying dunite (e.g. Salisbury & Christensen 1978; Christensen & Smewing 1981; Dilek et al. 1998).

The ‘petrological Moho’ recognized in ophiolites is the transition from dunite to underlying tectonized harzburgites. The separation of the seismological and petrological Mohos may be 0–1.5 km, and thus the dunite layer may be up to 20 per cent of the total crustal thickness. Anisotropy in \( V_p \) may be as large as \( \sim 8 \) per cent in ophiolitic harzburgites as a result of preferential olivine crystal orientation, with velocities varying from approximately 7.8 to 8.5 km s\(^{-1}\) with propagation direction.

**Oceanic crust beneath the mid-Atlantic ridge and the North Atlantic**

The structure of the oceanic crust beneath the North Atlantic, and particularly the crest and flanks of the MAR and the transition to Icelandic-type crust, has been the subject of numerous explosion seismic-refraction and some surface wave studies (e.g. Talwani et al. 1971; Fowler 1976, 1978; Bunch 1980; Bunch & Kennett 1980; Goldflam et al. 1980; Jacoby & Girardin 1980; Keen et al. 1980; White 1984; Ritzert & Jacoby 1985; Smallwood et al. 1995; Richardson et al. 1998; Smallwood & White 1998; Weir et al. 2001). The crust beneath the North Atlantic is best modelled as an upper crust with high velocity gradients of \( 1–2 \) s\(^{-1}\), and a lower crust with low gradients of \( \sim 0.1 \) s\(^{-1}\). Reports of the nature of the crust–mantle transition are variable. Some studies report a sharp Moho with a velocity discontinuity as large as \( V_p = 0.8 \) km s\(^{-1}\), with \( V_p \) increasing from \( \sim 7.2 \) to 7.8 km s\(^{-1}\) (Table 1). Others report a gradational crust–mantle transition zone with a thickness of up to 15 per cent of the total crustal thickness. It is not clear to what extent the variability in these reports results from different experimental conditions, and whether the crust–mantle transition may be gradational everywhere. The picture is complicated further by the use of widely differing definitions of ‘upper-mantle’ velocities, from \( V_p = 7.2 \) km s\(^{-1}\) at the ridge crest (e.g. Fowler 1976) to \( V_p = 8.2 \) km s\(^{-1}\) beneath the ridge flank (e.g. Bunch 1980).

Structure beneath the MAR crest contrasts with that beneath the flanks. Maximum velocities of \( V_p \sim 7.2 \) km s\(^{-1}\) are reported, which have been interpreted as being caused by anomalous upper mantle
Table 1. Summary of the crustal subdivision schemes and main results of recent explosion seismology studies of Icelandic-type crust. Velocities ($V_p$, $V_s$) are given in km s$^{-1}$, velocity gradients (grad) in s$^{-1}$ and depths and thicknesses in kilometres.

<table>
<thead>
<tr>
<th>Project name</th>
<th>Seismic definition of the upper crust</th>
<th>Seismic definition of the mid-crust</th>
<th>Seismic definition of the lower crust</th>
<th>$V_p/V_s$</th>
<th>Moho velocity discontinuity</th>
<th>Mantle velocity</th>
<th>thickness of the upper crust</th>
<th>crust–mantle boundary depth</th>
</tr>
</thead>
<tbody>
<tr>
<td>(Flovenz 1980)</td>
<td>Grad $\geq 0.57$</td>
<td>Not defined</td>
<td>Zero or low gradient, $V_p \geq 6.5$</td>
<td>n/a</td>
<td>n/a</td>
<td>$V_p \geq 6.9$</td>
<td>5–6</td>
<td>n/a</td>
</tr>
<tr>
<td>(Angenheister et al. 1980; Gebrande et al. 1980)</td>
<td>RRISP</td>
<td>$V_p \geq 6.5$</td>
<td>&quot;</td>
<td>n/a</td>
<td>n/a</td>
<td>$V_p \geq 6.9$</td>
<td>3–8</td>
<td>10–15</td>
</tr>
<tr>
<td>(Bjarnason et al. 1993)</td>
<td>SIST</td>
<td>$V_p &lt; 5.0$</td>
<td>$V_p = 5.0–6.5$</td>
<td>Jump of $V_p \approx 7.2/7.25$ to 7.5, rising to $V_p = 7.7$ just below discontinuity.</td>
<td>$V_p \geq 7.5$</td>
<td>0.7–3</td>
<td>20–24 beneath the WVZ and SISZ</td>
<td></td>
</tr>
<tr>
<td>(Staples et al. 1997)</td>
<td>FIRE</td>
<td>Grad $\geq 0.57$</td>
<td>Not defined</td>
<td>Basal crustal $V_p$, estimated by downward extrapolation, jump from $V_p = 7.1$ to 7.9 beneath NVZ, jump from $V_p = 7.35$ to 8.0/8.3 beneath E Iceland</td>
<td>$V_p = 7.9$ under NVZ, $V_p = 8.0/8.3$ under E Iceland</td>
<td>Up to $\sim 4$ km beneath the NVZ and E Iceland</td>
<td></td>
<td></td>
</tr>
<tr>
<td>(Darbyshire et al. 1998)</td>
<td>ICEMELT</td>
<td>Grad $\geq 0.2$, $V_p \geq 6.4$, variable gradients</td>
<td>$V_p = 6.9–7.1$, Grad $&lt; 0.03$</td>
<td>$1.76–1.78$</td>
<td>$V_p = 7.2–7.3$ estimated from downward extrapolation</td>
<td>Not estimated</td>
<td>5–10</td>
<td>25 beneath Skagi, 40 beneath NW Vatnajokull</td>
</tr>
<tr>
<td>(Menke et al. 1998)</td>
<td>B96</td>
<td>Not defined</td>
<td>Not defined</td>
<td>1.76</td>
<td>Not determined</td>
<td>$V_p = 7.7$ (eq from N) and 8.0 (eq from S)</td>
<td>Not given</td>
<td>25–31 beneath W flank of NVZ</td>
</tr>
<tr>
<td>(Smallwood et al. 1999)</td>
<td>FIRE</td>
<td>High grad</td>
<td>Not defined</td>
<td>Low grad, $V_p \leq 7.1–7.3$</td>
<td>Jump in $V_p$ of $\sim 7.3–7.7$</td>
<td>$V_p = 7.7–7.9$</td>
<td>$\leq 10$</td>
<td></td>
</tr>
<tr>
<td>(Du &amp; Foulger 1999; Du et al. 2002)</td>
<td>The Iceland</td>
<td>$V_s \leq 3.7$ (or $V_p \leq 6.5$)</td>
<td>Not defined</td>
<td>$V_s \leq 4.1$ (or $V_p \leq 7.2$)</td>
<td>Variable</td>
<td>$V_s &gt; 4.1$ (~equivalent to $V_p &gt; 7.2$)</td>
<td>5–11</td>
<td>24–38</td>
</tr>
<tr>
<td>(Weir et al. 2001)</td>
<td>RISE</td>
<td>High grad</td>
<td>Not defined</td>
<td>$V_p \geq 6.8$, Grad $\sim 0.05$</td>
<td>$1.78$</td>
<td>$V_p = 7.2$ at base from downward extrapolation, velocity jump not estimated</td>
<td>Not estimated</td>
<td>5–10</td>
</tr>
</tbody>
</table>
Early explosion seismology studies of the crust beneath the Iceland transverse ridge used P-phase arrival times. These, along with an early surface wave study, were interpreted in terms of simple layered models (Fig. 3a) (Báth 1960; Tryggvason 1962; Palmason 1971; Zverev et al. 1976; Bott & Gunnarsson 1980). In a pioneering experiment in 1959, Báth (1960) recorded data on two unreversed explosion profiles ~150 and ~250 km long in western and central Iceland (Fig. 1). He interpreted the times of first arrivals in terms of a three-layered model and observed reflections he concluded to come from the Moho at a depth of ~28 km. Tryggvason (1962) constructed standard surface wave group-velocity dispersion curves for Love and Rayleigh waves from 20 earthquakes along the mid-Atlantic ridge up to ~1000 km from Iceland, and also interpreted the results in terms of a three-layered structure. Palmason (1971) gathered data on ~40 profiles up to ~150 km long covering most of Iceland (Fig. 1), and processed the data using conventional seismic refraction theory. He emphasized the lateral heterogeneity of the Icelandic crust, and developed layered models involving average velocities and interface depth ranges. In 1972 Zverev et al. (1976) and Bott & Gunnarsson (1980) shot the ~450 km long NASP profile, which extended almost the entire length of the Iceland–Faeroe ridge and was recorded at stations on the Faeroe Islands, Iceland, and the ridge itself (Fig. 1). They studied variations in the depth to the mantle beneath the Iceland–Faeroe ridge using the τ–p method and modified time-term analysis.

There is little resemblance between the models proposed (Fig. 3a). For the shallow crust, where velocities are less than ~6.0 km s\(^{-1}\), a great variety of characteristic layer velocities have been proposed. In addition, two fundamentally different interpretations were proposed for the deepest, highest-velocity material sampled, with material with \(V_p = 7.2–7.4\) km s\(^{-1}\) being variously attributed to the lower crust (Báth 1960) or the upper mantle (Tryggvason 1962; Palmason 1971). Báth (1960) assigned this material to the lower crust because reflections were observed from a deeper interface, which provided a candidate for the Moho. In contrast, Tryggvason (1962) and Palmason (1971) did not observe deeper layers or interfaces and because of this attributed the material with \(V_p = 7.2–7.4\) km s\(^{-1}\) to the upper mantle by analogy with work on mid-ocean ridges (MORs) elsewhere (e.g. Fowler 1976). The Iceland–Faeroe ridge lacked the ambiguous \(V_p = 7.2–7.4\) km s\(^{-1}\) layer directly overlying material with \(V_p = 7.8\) km s\(^{-1}\) at depths of 27–42 km (Zverev et al. 1976; Bott & Gunnarsson 1980) (Fig. 3a). This structure naturally invited placement of the crust–mantle boundary between these two layers.

**EARLY SEISMIC STUDIES OF ICELANDIC-TYPE CRUST**

Early explosion seismology studies of the crust beneath the Iceland transverse ridge used P-phase arrival times. These, along with an early surface wave study, were interpreted in terms of simple layered models (Fig. 3a) (Báth 1960; Tryggvason 1962; Palmason 1971; Zverev et al. 1976; Bott & Gunnarsson 1980). In a pioneering experiment in 1959, Báth (1960) recorded data on two unreversed explosion profiles ~150 and ~250 km long in western and central Iceland (Fig. 1). He interpreted the times of first arrivals in terms of a three-layered model and observed reflections he concluded to come from the Moho at a depth of ~28 km. Tryggvason (1962) constructed standard surface wave group-velocity dispersion curves for Love and Rayleigh waves from 20 earthquakes along the mid-Atlantic ridge up to ~1000 km from Iceland, and also interpreted the results in terms of a three-layered structure. Palmason (1971) gathered data on ~40 profiles up to ~150 km long covering most of Iceland (Fig. 1), and processed the data using conventional seismic refraction theory. He emphasized the lateral heterogeneity of the Icelandic crust, and developed layered models involving average velocities and interface depth ranges. In 1972 Zverev et al. (1976) and Bott & Gunnarsson (1980) shot the ~450 km long NASP profile, which extended almost the entire length of the Iceland–Faeroe ridge and was recorded at stations on the Faeroe Islands, Iceland, and the ridge itself (Fig. 1). They studied variations in the depth to the mantle beneath the Iceland–Faeroe ridge using the τ–p method and modified time-term analysis.

There is little resemblance between the models proposed (Fig. 3a). For the shallow crust, where velocities are less than ~6.0 km s\(^{-1}\), a great variety of characteristic layer velocities have been proposed. In addition, two fundamentally different interpretations were proposed for the deepest, highest-velocity material sampled, with material with \(V_p = 7.2–7.4\) km s\(^{-1}\) being variously attributed to the lower crust (Báth 1960) or the upper mantle (Tryggvason 1962; Palmason 1971). Báth (1960) assigned this material to the lower crust because reflections were observed from a deeper interface, which provided a candidate for the Moho. In contrast, Tryggvason (1962) and Palmason (1971) did not observe deeper layers or interfaces and because of this attributed the material with \(V_p = 7.2–7.4\) km s\(^{-1}\) to the upper mantle by analogy with work on mid-ocean ridges (MORs) elsewhere (e.g. Fowler 1976). The Iceland–Faeroe ridge lacked the ambiguous \(V_p = 7.2–7.4\) km s\(^{-1}\) layer directly overlying material with \(V_p = 7.8\) km s\(^{-1}\) at depths of 27–42 km (Zverev et al. 1976; Bott & Gunnarsson 1980) (Fig. 3a). This structure naturally invited placement of the crust–mantle boundary between these two layers.
Apart from the observation of velocities in the range $V_p = 7.2$–7.4 km s$^{-1}$ beneath Iceland, only two other ubiquitous features were reported. These are the extreme variation in velocity, both vertically and horizontally, in the upper crust (where $V_p$ is typically observed to be $\approx 6.5$ km s$^{-1}$), and the contrasting homogeneity of the underlying lower crust.

## RECENT SEISMIC STUDIES OF ICELANDIC-TYPE CRUST

Since 1977, numerous long explosion-seismology profiles have been measured using densely distributed, three-component seismic stations and, latterly, GPS timing. Digital recording has aided phase recognition by enabling easy filtering of the noisy data typically collected in the North Atlantic, and facilitated the application of modern data analysis techniques such as waveform modelling. Experiments have spanned contrasting tectonic regions and focused on determining lateral variations in structure (Figs 1 and 3b).

As soon as waveform modelling was applied to explosion seismology data from Iceland it became clear that Icelandic-type crust is most naturally subdivided on the basis of velocity gradient. Flovenz (1980) reinterpreted explosion seismograms gathered by Palmason (1971) using synthetic seismograms, and was the first to emphasize that a simple layered model is inconsistent with observed amplitudes. He divided the crust into two layers on the basis of velocity gradient (Fig. 3b, Table 1). The upper crust exhibits high velocity gradients of $\sim$0.57 s$^{-1}$ beneath the $V_p \sim 3.5$ km s$^{-1}$ horizon. The lower crust beneath the $V_p \sim 6.5$ km s$^{-1}$ velocity horizon, typically at 5–6 km depth, has a much smaller velocity gradient.

Lateral heterogeneity in the upper crust is extreme. Velocities of $V_p = 6.5$ km s$^{-1}$ may occur close to the surface beneath volcanoes (e.g. Emarsi 1978; Foulger & Toomey 1989; Toomey & Foulger 1989; Arnott & Foulger 1994; Gudmundsson et al. 1994; Foulger et al. 1995; Miller 1996), and ground deformation indicates the presence of shallow magma chambers (e.g. Björnsson 1985; Sigmundsson et al. 1992). Velocities of $V_p \sim 6.5$ km s$^{-1}$, probably associated with gabbroic intrusions, occur within $\sim$2 km of the surface, and low-velocity magma accumulations exist locally in the upper crust. Horizontal velocity gradients may exceed vertical gradients, and so schemes that use absolute velocities to define laterally extensive layers are unsuitable for Icelandic-type crust.

Considerable debate has revolved around whether the $V_p = 7.2$–7.4 km s$^{-1}$ material should be attributed to the lower crust, as originally proposed by Båth (1960) (the thick–cold crust model), or to the upper mantle, as suggested by later workers (the thin–hot crust model). The 1977 800 km long RRISP land–sea explosion profile extended along magnetic anomaly 5 (10 Ma) on the southeastern flank of the Reykjanes ridge and crossed Iceland along the EVZ and the NVZ (Aangenheister et al. 1980; Gebran et al. 1980) (Fig. 1). The first-arrival times were forward modelled by ray tracing through a structure divided into three layers, each with a characteristic velocity gradient (Fig. 3b, Table 1). Both offshore and beneath Iceland, the shallowest levels exhibit high velocity gradients, and deeper levels, where velocities exceeded $V_p \sim 6.5$ km s$^{-1}$, have much lower gradients. Offshore the crust was found to be $\sim$10 km thick, beneath which $V_p$ rises steeply to $\sim$7.8 km s$^{-1}$, a velocity that was attributed to the mantle. No such discontinuity or high velocities were found beneath Iceland, however. Instead, the data were interpreted as indicating that velocities did not exceed $V_p = 7.6$ km s$^{-1}$ in the upper 50 km.

Like Palmason (1971), Gebran et al. (1980) attributed material with $V_p = 7.0$–7.6 km s$^{-1}$ beneath Iceland to anomalously hot upper mantle. Both Palmason (1971) and Gebran et al. (1980) were aware of the extensive debate regarding the assignation of this material, and that it had been attributed variously to crustal layer 3B or upper mantle. Gebran et al. (1980) remark that it is ‘more or less a question of definition’ whether it is attributed to the lower crust or the upper mantle. They favoured an anomalous upper-mantle interpretation because:

1. they did not observe any widespread deeper discontinuity, reflector or feature that was a candidate for the Moho or the crust–mantle boundary;
2. teleseismic $P$-wave delays in Iceland had been shown to be compatible with low upper-mantle velocities extending from the surface down to over 200 km, and with the observed Bouguer gravity low over Iceland (Tryggvason 1964; Bott 1965; Long & Mitchell 1970);
3. velocities in the range $V_p = 7.2$–7.6 km s$^{-1}$ had been reported for MOR axial zones, and attributed to crust–mantle mixing or to partial fusion in the upper mantle (e.g. Ewing & Ewing 1959; Bott 1965; Oxbergh & Turcotte 1968; Vogt et al. 1969; Fowler 1976);
4. temperatures of 1000–1100 °C, and thus partial melt, were predicted at depths of 10–20 km, where velocities of $V_p = 7.0$–7.4 km s$^{-1}$ were encountered. At such temperatures, the seismic velocity in gabbros and metagabbros is much less than $V_p = 7.0$–7.4 km s$^{-1}$, whereas these velocities agree well with partially molten mantle peridotite. Four types of observation had been interpreted as indicating high temperatures and partial melt, which were expected beneath a hotspot and within a mantle plume such as was assumed to underlie Iceland.

(a) High surface temperature gradients in Iceland of $\sim$50–100 °C km$^{-1}$ away from geothermal areas, which predict supra-solidus temperatures at $\sim$10–20 km depth if extrapolated linearly downward (Palmason 1974; Flovenz & Saemundsson 1993).

(b) Shear waves from RRISP explosions that penetrated the $V_p = 7.0$ km s$^{-1}$ material were attenuated, suggesting partial melt (Gebran et al. 1980).

(c) A high average $V_p/V_s$ of 1.96, consistent with partial melt, was deduced from RRISP data for trains that travelled in the $V_p > 7.0$ km s$^{-1}$ material beneath Iceland (Gebran et al. 1980).

(d) Magnetotelluric measurements over much of Iceland detect a widespread, high-conductivity layer, interpreted as a $\sim$5 km thick interval of high-degree ($\sim$10–20 per cent) partial melt at 10–20 km depths (e.g. Hermance & Grillot 1974; Beblo & Björnsson 1978; 1980; Beblo et al. 1983; Eysteinsson & Hermance 1985).

The thin–hot crust model was challenged by Björnason et al. (1993), who shot the 170 km long South Iceland Seismic Transect (SIST) from the west coast of Iceland across the WVZ, the SISZ and the EVZ (Fig. 1). The seismic stations were sufficiently dense for high-resolution tomography of the upper crust and lower-resolution sampling to $\sim$25 km depth, where structure was modelled using forward ray tracing. This study confirmed the lateral inhomogeneity of the upper crust and the gradational velocity structure of the entire crust (Fig. 3b, Table 1). It also detected an apparent velocity of $V_p \approx 7.7$ km s$^{-1}$ for a reflecting horizon at a depth of 20–24 km, thereby becoming the first study to repeat the observation of Båth (1960). This part of the profile was unreversed (i.e. an explosion at only one end of the profile sampled the volume of interest), and thus this velocity is only the correct mantle velocity if there is no dip on the refracting horizon. The horizon was interpreted as a sharp Moho across which the velocity jumps from $V_p = 7.25$ to 7.5 km s$^{-1}$, rapidly rising to 7.7 km s$^{-1}$ below the reflector. The
agreement with the result of Båth (1960), who obtained a depth of 28 km for his reflective horizon, is impressive considering the early date of that experiment. Bjarnason et al. (1993) point out that similar reflections observed in the RRI SP data beneath the NVZ and eastern Iceland, and attributed by Gebr and et al. (1980) to thin, high-velocity lenses, could also be reflections from a horizon \( \sim 30 \) km deep.

The interpretation of Bjarnason et al. (1993) implied that material with velocity up to \( V_p = 7.5 \) km s\(^{-1}\) is gabbroic and not mantle peridotite. On the basis of seismic velocity they estimated the temperature at the reflector to be 600–900 °C, which is below the solidus for gabbro. This implies that it does not contain pervasive melt as predicted by the thin–hot crust model, although Gudmundsson (1994) pointed out that this interpretation was not safe on the basis of seismic velocities alone, given the uncertainties.

Subsequent reappraisal of the non-seismic evidence for subsolidus temperatures in the \( V_p = 7.0–7.5 \) km s\(^{-1}\) material suggested that some data may be compatible with subsolidus temperatures (Menke & Levin 1994; Menke & Sparks 1995; Menke et al. 1995, 1996). Linear downward extrapolation of surface geothermal gradients predicts temperatures corresponding to the brittle–plastic transition (\( \sim 400–650 \) °C for the lithology and strain rates expected under Iceland) at the seismogenic base, which varies in the range \( \sim 6–14 \) km (e.g. Stefansson et al. 1993; Foulger 1995). However, linear extrapolation is only valid in the steady-state conductive regime in the absence of heat sources. The thermal state of the shallow upper crust is influenced by hydrothermal circulation, and the geothermal gradient may be smaller in the lower crust (Menke & Sparks 1995).

Shear waves from earthquakes at ranges of up to \( \sim 350 \) km exhibit little attenuation on passage beneath Iceland at 10–20 km depth and minimum values of \( Q_p \geq 100 \) and \( Q_s \geq 250 \) are inferred for the lower crust beneath central and southwest Iceland (Menke & Levin 1994; Menke et al. 1995). Values of \( Q_s = 20 \) are expected for near-subsolidus gabbro and \( Q_s \geq 40 \) for peridotite at 950 °C. The observations suggest maximum temperatures for the lower crust in general of 700–950 °C, substantially below the solidus of both gabbro and peridotite at \( \sim 20 \) km depth. Menke et al. (1996) suggest that the low values of \( V_p \) and \( V_p/V_s \) ratios of 1.96 deduced from the RRI SP data (Gebr and et al. 1980) were erroneous and result from uncorrected source static anomalies and misidentification of S–S' as S. Reappraisal of those data suggests more normal values of \( V_p/V_s \) in the lower crust, consistent with subsolidus temperatures and the results of later explosion profiles (e.g. Staples et al. 1997; Darbyshire et al. 1998). Interpretation of magnetotelluric data suffers from poor resolution and the ratio of depth: thickness of low-resistivity layers is ambiguous. The magnetotelluric data from Iceland could be fit by a thin, low-resistivity layer or a thick higher-resistivity layer, and do not strongly constrain structure at depth beneath Iceland.

Subsequent to this revision in thinking, several additional long, high-quality explosion profiles were measured in Iceland (Fig. 1). The 1994 Faeroe–Iceland Ridge Experiment (FIRE) profile comprised over 60 large explosions, with 100 three-component recorders arrayed along the Iceland–Faeroe ridge and crossing eastern Iceland and the NVZ (Staples et al. 1997). The 310 km long 1995 ICEMELT profile extended from Skagi and traversed central Iceland to the southeast coast (Darbyshire et al. 1998) (Fig. 4). This profile included \( \sim 60 \) three-component stations and six large explosions. In 1996 the \( \sim 75 \) km long, B96 profile probed the western flank of the NVZ (Menke et al. 1998). The Reykjanes peninsula and Reykjanes ridge immediately offshore were investigated by the \( \sim 220 \) km long RISE profile, which involved \( \sim 40 \) stations on land and 27 ocean-bottom seismometers (Weir et al. 2001). The data from all of these profiles were modelled using forward ray tracing.

All modern explosion seismology studies confirm the structure proposed by Flovenz (1980), with shallow, \( V_p \approx 6.5 \) km s\(^{-1}\) material having large vertical velocity gradients (typically \( \sim 0.25 \) s\(^{-1}\)) and great lateral heterogeneity, and deeper material having velocity gradients of magnitude lower (typically \( \sim 0.024 \) s\(^{-1}\)) (e.g. Fig. 4). Despite such observations, schemes for subdivision on the basis of absolute velocity are nevertheless often presented (Table 1). These schemes vary between studies, may require modification even along a single profile and are thus of limited use. Deep, post-critical reflections of variable clarity are reported from all the profiles and are generally attributed to a sharp Moho, although Staples (1997) suggested on the basis of reflectivity modelling that the FIRE data may be satisfied by a transition zone up to \( 2 \) km thick. Only on two unversed profiles have mantle-diving rays been detected, both within or close to Tertiary blocks (Bjarnason et al. 1993; Menke et al. 1998), and so clear, unambiguous measurement of mantle velocity using explosion seismology has not been achieved. Synthetic seismograms calculated by the reflectivity method were used to estimate the amplitude of the velocity discontinuity in a number of cases. Although the mantle is expected to exhibit up to 0.7 km s\(^{-1}\) anisotropy in \( V_p \) (e.g. Salisbury & Christensen 1978; Christensen & Smewing 1981), no systematic mantle velocity variation with azimuth is discernible. The main conclusions from these profiles are summarized in Table 1.

A recent surface wave study investigated the structure of the crust beneath Iceland using partitioned waveform inversion (Allen et al. 2002). The frequency window used was 0.03–0.1 Hz, corresponding to wavelengths of \( \sim 40–120 \) km and the inversion was constrained by the estimates of crustal thickness from explosion seismology. The large velocity gradients in the upper few kilometres and lower gradients beneath were confirmed. The agreement with mantle velocities determined by explosion seismology (and receiver functions, see below) is much poorer. For the mantle, \( V_s \sim 3.7–4.0 \) km s\(^{-1}\) (corresponding to \( V_p \sim 6.5–7.05 \) km s\(^{-1}\)) was determined, along with \( V_p/V_s \sim 1.92 \) at 35 km depth. This is close to the value of 1.96 determined by Gebr and et al. (1980) from RRI SP data and shown to be erroneous by Menke et al. (1996) using the same data, which brings into question the validity of the surface wave results.

Seismic and gravity data have been jointly modelled both to test the seismic results (Staples et al. 1997; Weir et al. 2001) and to produce pan-Iceland crustal models (e.g. Darbyshire et al. 2000b). The models developed place density anomalies in the upper mantle, which, if interpreted as variations in temperature, require large anomalies of \( \sim 600 \) °C beneath the Reykjanes ridge (Weir et al. 2001) and 700 °C beneath the NVZ (Staples et al. 1997). Darbyshire et al. (2000b) assembled all explosion seismology and receiver function results then available and used the gravity field to estimate the crustal thickness beneath regions unsampled seismically (Fig. 5). They concluded that the crustal thickness varies from \( \sim 40 \) km beneath northwest Vatnajökull to \( \sim 15 \) km in the southwest. They further concluded that the gravity data require density anomalies in the mantle of up to \( 110 \) kg m\(^{-3}\) a few tens of kilometres wide beneath the rift zones and northwest Vatnajökull, corresponding to lateral density anomalies of \( \sim 3 \) per cent and temperature anomalies of \( 450–700 \) °C. Allen et al. (2002) present a crustal thickness model broadly similar to that of Darbyshire et al. (2000b) from joint inversion of surface waveforms and topography, assuming Airy isostatic compensation.

© 2003 RAS, GJI, 155, 567–590
Figure 4. Top, crustal model derived from the ICEMELT explosion seismology profile (Fig. 1). The thick lines indicate locations of reflective horizons interpreted as the Moho. Bottom, 1-D velocity profiles at various offsets along the profile relative to an explosion in central Iceland (from Darbyshire et al. 1998).

The structure beneath the Greenland–Iceland and Iceland–Faeroe ridges has also been investigated in recent explosion seismology experiments. The FIRE profile extended the entire length of the Iceland–Faeroe ridge, ~25 km north of the earlier NASP profile (Richardson et al. 1998; Smallwood et al. 1999) (Fig. 1). The crust was found to have a similar structure to that beneath Iceland, with an upper crust ~10 km thick with high velocity gradients underlain by a lower crust with low-velocity gradients. Velocities at the base of the crust were $V_p = 7.1–7.3$ km s$^{-1}$ and the depth to the base of the crust was fairly uniform at ~30 ± 3 km, shallowing to ~25 km close to Iceland. In contrast with Iceland, reflections that could be interpreted as coming from a Moho were not observed, but mantle-diving rays were clear and constrained the upper-mantle velocity by $V_p = 7.7–7.9$ km s$^{-1}$. The observation of mantle-diving rays indicates a positive velocity gradient in the uppermost mantle beneath the Iceland–Faeroe ridge, and the rarity of such rays beneath Iceland suggests negative velocity gradients there. It is not clear how diving rays can exist beneath the Iceland–Faeroe ridge in the absence of post-critical reflections.

The structure of the Greenland–Iceland ridge was investigated by the ~300 km long SIGMA 1 profile (Holbrook et al. 2001) (Fig. 1). There, the crust was >30 km thick velocities attributed to the lower crust as high as $V_p = 7.4–7.5$ km s$^{-1}$.

**Structure of Icelandic-type crust from receiver functions**

A homogenous suite of crustal structures from receiver functions covering most of Iceland at regular intervals has recently been completed using data from the Iceland Hotspot Project (Du & Foulger 1999, 2001; Du et al. 2002). This project gathered data from a network of regularly spaced broad-band seismometers that
were operated during 1996–1998 (Fig. 1), and yielded a large suite of well-recorded teleseismic earthquakes (Foulger et al. 2001).

The receiver function method uses shear waves generated by mode conversion from compressional waves arriving as P phases from distant earthquakes. It is based on the assumption that for steeply arriving teleseismic wave trains, the vertical-component seismograms approximate the incident P waveforms, whereas horizontal-component seismograms are dominated by shear waves generated by mode conversion at interfaces in the structure beneath the station (Langston 1979). The vertical-component seismogram is deconvolved from the two horizontal-component seismograms, yielding the response function for the crust beneath the station. This is inverted to determine the shear-velocity structure by iteratively perturbing an initial model. The method works best for shallow structure, for which the mode-converted signals are best recorded at the surface.

Three significant problems must be overcome if reliable results are to be obtained. The first is the velocity–depth ambiguity. Receiver functions constrain the absolute velocities only weakly, and thus independent constraints are required. Du & Foulger (1999, 2001) and Du et al. (2002) obtained absolute-velocity starting models using phase-velocity dispersion of surface waves between station pairs, which provide local estimates of the average $V_s$ structure of the crust, and waveform modelling of large regional events. The second problem is the sensitivity of the result to the starting model. Starting models close to the final model are needed to avoid convergence to a local rather than a global minimum. Du & Foulger (1999, 2001) address this problem using broad suites of initial models to initialize inversions and progressively refining the preferred starting model. Du et al. (2002) improved on this approach using a genetic algorithm, which searches a large model space and finds solutions near to the global minimum. The third problem, that of the high microseismic noise level, is particularly serious in the case of data recorded in Iceland. Du & Foulger (1999, 2001) and Du et al. (2002) suppressed noise by stacking earthquakes from similar backazimuths and filtering the data using a Gaussian low-pass filter with a corner frequency of 1.2 Hz, which corresponds to shear wavelengths of $\sim$2–4 km. Structures on the scale of $\sim$2 km at shallow depth and $\sim$2.5 km in the deep crust are resolvable with such data. Smaller features are smoothed vertically on this spatial scale.

Du & Foulger (1999, 2001) and Du et al. (2002) present detailed results for 31 stations covering much of Iceland uniformly. The first-order results are (Fig. 6):

(1) the crust beneath Iceland naturally divides into an upper crust, which is laterally heterogeneous and characterized by high vertical velocity gradients, and a lower crust, which is less heterogeneous, and characterized by low-velocity gradients (Fig. 6a);

(2) LVZs are common, particularly in the lower crust, and are most extreme and best constrained beneath the MVZ where almost the entire lower crust forms a spatially extensive, strong, coherent LVZ with velocities as low as $V_s = 3.4$ km s$^{-1}$ (corresponding to $V_p = 6.0$ km s$^{-1}$) at depths of $>20$ km (Fig. 6b);

(3) the transition from velocities characteristic of the crust ($V_s \leq 4.1$ km s$^{-1}$, $V_p \leq 7.2$ km s$^{-1}$) to velocities characteristic of the mantle ($V_s \sim 4.45$ km s$^{-1}$, $V_p \sim 7.8$ km s$^{-1}$) occurs over a depth
interval that may be relatively small (e.g. 3 km; Fig. 6a) or up to a few kilometres thick (e.g. 7 km; Fig. 6c).

Du & Foulger (1999, 2001) and Du et al. (2002) map variations in the depths to the $V_s = 3.7$ and 4.1 km s$^{-1}$ velocity contours ($V_p = 6.5$ and 7.2 km s$^{-1}$, assuming $V_p/V_s = 1.76$), these being typical values used for the bases of the upper and lower crusts in other studies (Fig. 7). In particular, there is fairly good agreement between the $V_s = 4.1$ km s$^{-1}$ horizon and the reflective horizon assigned to the Moho in explosion seismology experiments. Using these definitions, the upper crust varies in thickness from $\sim 6$ to 11 km (Fig. 7a). It is exceptionally thick in the extreme south of Iceland where a value of 11 km was obtained for station 22 (Figs 1 and 7a). The unusually thick upper crust there was noted in earlier studies (e.g. Flovenz 1980) and its cause is unknown. Apart from this anomalous region, the upper crust has a rather uniform thickness of $\sim 7 \pm 1$ km. The small variations observed do not correlate in any obvious way with surface tectonics, except for a tendency for the upper crust to be slightly thicker beneath the Trollaskagi block in north Iceland.

The depth to the $V_s = 4.1$ km s$^{-1}$ velocity contour, assigned to the base of the lower crust, varies from $\sim 20$ to 40 km (Fig. 7b). The crust is thickest where the MVZ, NVZ and EVZ converge in central Iceland, which is also coincident with the general centre of the $\sim 120$ mGal Bouguer gravity low over Iceland. The crust thins rapidly to the southwest and is shallowest there and in western Iceland. Corresponding shallowing is not observed in the north and east, where a minimum crustal thickness of $\sim 27$ km is detected.

Additional receiver function results for eight stations in central and north Iceland are reported by Darbyshire et al. (2000a). They used single earthquakes and different starting models, and suppress structure required by parts of the receiver function wave train later inaccessible way with surface tectonics, except for a tendency for the upper crust to be slightly thicker beneath the Trollaskagi block in north Iceland.

The depth to the $V_s = 4.1$ km s$^{-1}$ velocity contour, assigned to the base of the lower crust, varies from $\sim 20$ to 40 km (Fig. 7b). The crust is thickest where the MVZ, NVZ and EVZ converge in central Iceland, which is also coincident with the general centre of the $\sim 120$ mGal Bouguer gravity low over Iceland. The crust thins rapidly to the southwest and is shallowest there and in western Iceland. Corresponding shallowing is not observed in the north and east, where a minimum crustal thickness of $\sim 27$ km is detected.

Additional receiver function results for eight stations in central and north Iceland are reported by Darbyshire et al. (2000a). They used single earthquakes and different starting models, and suppress structure required by parts of the receiver function wave train later inaccessible way with surface tectonics, except for a tendency for the upper crust to be slightly thicker beneath the Trollaskagi block in north Iceland.

The depth to the $V_s = 4.1$ km s$^{-1}$ velocity contour, assigned to the base of the lower crust, varies from $\sim 20$ to 40 km (Fig. 7b). The crust is thickest where the MVZ, NVZ and EVZ converge in central Iceland, which is also coincident with the general centre of the $\sim 120$ mGal Bouguer gravity low over Iceland. The crust thins rapidly to the southwest and is shallowest there and in western Iceland. Corresponding shallowing is not observed in the north and east, where a minimum crustal thickness of $\sim 27$ km is detected.

Additional receiver function results for eight stations in central and north Iceland are reported by Darbyshire et al. (2000a). They used single earthquakes and different starting models, and suppress structure required by parts of the receiver function wave train later inaccessible way with surface tectonics, except for a tendency for the upper crust to be slightly thicker beneath the Trollaskagi block in north Iceland.

The depth to the $V_s = 4.1$ km s$^{-1}$ velocity contour, assigned to the base of the lower crust, varies from $\sim 20$ to 40 km (Fig. 7b). The crust is thickest where the MVZ, NVZ and EVZ converge in central Iceland, which is also coincident with the general centre of the $\sim 120$ mGal Bouguer gravity low over Iceland. The crust thins rapidly to the southwest and is shallowest there and in western Iceland. Corresponding shallowing is not observed in the north and east, where a minimum crustal thickness of $\sim 27$ km is detected.

Additional receiver function results for eight stations in central and north Iceland are reported by Darbyshire et al. (2000a). They used single earthquakes and different starting models, and suppress structure required by parts of the receiver function wave train later inaccessible way with surface tectonics, except for a tendency for the upper crust to be slightly thicker beneath the Trollaskagi block in north Iceland.

The depth to the $V_s = 4.1$ km s$^{-1}$ velocity contour, assigned to the base of the lower crust, varies from $\sim 20$ to 40 km (Fig. 7b). The crust is thickest where the MVZ, NVZ and EVZ converge in central Iceland, which is also coincident with the general centre of the $\sim 120$ mGal Bouguer gravity low over Iceland. The crust thins rapidly to the southwest and is shallowest there and in western Iceland. Corresponding shallowing is not observed in the north and east, where a minimum crustal thickness of $\sim 27$ km is detected.

Additional receiver function results for eight stations in central and north Iceland are reported by Darbyshire et al. (2000a). They used single earthquakes and different starting models, and suppress structure required by parts of the receiver function wave train later inaccessible way with surface tectonics, except for a tendency for the upper crust to be slightly thicker beneath the Trollaskagi block in north Iceland.

The depth to the $V_s = 4.1$ km s$^{-1}$ velocity contour, assigned to the base of the lower crust, varies from $\sim 20$ to 40 km (Fig. 7b). The crust is thickest where the MVZ, NVZ and EVZ converge in central Iceland, which is also coincident with the general centre of the $\sim 120$ mGal Bouguer gravity low over Iceland. The crust thins rapidly to the southwest and is shallowest there and in western Iceland. Corresponding shallowing is not observed in the north and east, where a minimum crustal thickness of $\sim 27$ km is detected.

Additional receiver function results for eight stations in central and north Iceland are reported by Darbyshire et al. (2000a). They used single earthquakes and different starting models, and suppress structure required by parts of the receiver function wave train later inaccessible way with surface tectonics, except for a tendency for the upper crust to be slightly thicker beneath the Trollaskagi block in north Iceland.

The depth to the $V_s = 4.1$ km s$^{-1}$ velocity contour, assigned to the base of the lower crust, varies from $\sim 20$ to 40 km (Fig. 7b). The crust is thickest where the MVZ, NVZ and EVZ converge in central Iceland, which is also coincident with the general centre of the $\sim 120$ mGal Bouguer gravity low over Iceland. The crust thins rapidly to the southwest and is shallowest there and in western Iceland. Corresponding shallowing is not observed in the north and east, where a minimum crustal thickness of $\sim 27$ km is detected.

Additional receiver function results for eight stations in central and north Iceland are reported by Darbyshire et al. (2000a). They used single earthquakes and different starting models, and suppress structure required by parts of the receiver function wave train later inaccessible way with surface tectonics, except for a tendency for the upper crust to be slightly thicker beneath the Trollaskagi block in north Iceland.© 2003 RAS, GJI, 155, 567–590

Figure 6. Examples of end-member structures determined from receiver functions. (a) Station 6, from earthquakes approaching from the NE backazimuth (from Du & Foulger 1999), (b) station HVE, from earthquakes approaching from the SW backazimuth (from Du & Foulger 2001), (c) station 8, from earthquakes approaching from the north backazimuth (from Du & Foulger 1999).

Differences in methodology

The formal errors associated with the receiver functions, using the method of Du & Foulger (1990, 2001) and Du et al. (2002) are typically up to $\sim 2.5$ km in depth (the thickness of the thickest layer modelled) and $\pm 0.1$ km s$^{-1}$ in $V_s$ in the deeper part of the crust (Fig. 6). The formal error in modern explosion seismology studies of the depth to the deep reflector is $\pm 2$ km (e.g. Darbyshire et al. 1998). However, intrinsic problems in all the methods used cause much larger variations in the results than the simple formal misfits to the data imply.

The receiver-function method contrasts fundamentally with explosion seismology. A receiver function is derived from waves with frequencies of the order of 0.1 Hz, and thus samples an area some tens of kilometres in diameter around a station. It is most sensitive to reflections caused by large variations in impedance (the product of density and shear wave velocity), comparatively insensitive to absolute velocities, and its sensitivity is largely independent of depth within the crust. It can detect velocity reversals as easily as velocity increases with depth. Explosion seismology, on the other hand, involves much higher-frequency waves, samples relatively narrow zones along linear profiles and is most sensitive to laterally averaged absolute compressional-wave velocity at a few particular depths where the rays corresponding to observed waves have
turning points. It is also sensitive to major zones of rapid or discontinuous velocity increase with depth, which focus rays to produce large, later-arriving, post-critically ‘reflected’ waves, and to the structure of shallow regions, for which the ray density tends to be high and wave amplitudes are large. It is relatively insensitive to structure in the deeper crust, which is sampled only by small, later-arriving waves, and it is insensitive to velocity reversals, which lead to traveltime curves the structural interpretation of which is mathematically ambiguous, even for 1-D models and complete and error-free data. For most of the explosion profiles in Iceland, velocity near the base of the lower crust is not constrained, but estimated by downward extrapolation of velocity gradients (e.g. Staples et al. 1997; Darbyshire et al. 1998). Furthermore, the error in the depth to the base of the crust obtained by combining explosion seismic results and gravity data (Fig. 5; Darbyshire et al. 2000b) is large and difficult to quantify because of the serious problem of correctly assigning density.

These fundamental differences in methods mean that receiver-function and explosion-seismology results are not directly comparable. The $V_s$ models obtained from receiver functions cannot be related in a simple way to the $V_p$ models obtained from explosion seismology if there are variations in the $V_p/V_s$ ratio or in the relative attenuation of compressional and shear waves. These are, furthermore, expected consequences of variable petrology, anisotropy, partial melting and the presence of fluids and hydrothermal systems. A clear illustration of this point is provided by Darbyshire et al.
(2000a), who used receiver functions from station REN (Fig. 1) to determine structure beneath the Krafla central volcano, near the northern end of the FIRE explosion profile. They detected a LVZ 5 km thick at the top of the lower crust. This volume was well sampled by crossing rays of the FIRE profile (Staples et al. 1997), but synthetic seismogram modelling showed that even in this, a part of the crust best sampled by explosion seismology, that method could not detect such a LVZ if it exists (Darbyshire et al. 2000a).

Both the receiver-function and explosion methods suffer from inherent ambiguities. The receiver-function model space is large (roughly 25 free parameters) and must be searched efficiently to ensure that the starting model is close to the global minimum. Receiver functions depend on structure in a complicated way, so sophisticated methods such as genetic algorithms must be used to search for models that are globally optimal. Explosion seismology observations are usually analysed by forward modelling techniques, which do not provide estimates of uncertainties in the derived models. The ambiguity problem with explosion data is most severe in the deeper crust.

Similarities in the results

Notwithstanding the fundamental differences in methodology, essentially all modern seismic studies of Icelandic crust agree that high velocity gradients and extreme heterogeneity characterize the shallow crust, which is a few kilometres thick, and that velocity gradients an order of magnitude smaller and a greater degree of homogeneity characterize the lower crust, which is $\sim$15–30 km thick. There is no evidence that upper Icelandic crust compact with age as normal oceanic crust does, since no thinning of the upper crust from younger to older parks of Iceland is observed. The thickness of the upper crust, and the depth to the crust–mantle boundary determined from receiver functions are in general agreement with the results from explosion seismology, considering the significantly different definitions of these features in common use (Table 1). The similarities in the pan-Iceland maps of crustal thickness from receiver functions (Fig. 7b), and explosion seismology/gravity (Darbyshire et al. 2000b) (Fig. 5) are:

1. the crust is thinnest beneath central Iceland, where it is up to $\sim$40 km thick;
2. crustal thickness is asymmetric north–south and east–west;
3. the crust thins to 20–25 km beneath south and west Iceland, but is $\sim$30 km almost to the coast beneath much of north Iceland; and
4. the crust is $\sim$30 km thick or more beneath eastern Iceland, where it is significantly thicker than crust of the same age in western Iceland.

Differences in the results

Receiver functions typically yield more complex structures than explosion seismology, with velocity reversals on a wide range of scales being common. Significant LVZs occur in the lower crust in approximately half of the $\sim$80 receiver functions of Du & Foulger (1999, 2001) and Du et al. (2002). These vary from being thin to occupying most of the lower crust. Explosion seismology usually cannot constrain such velocity reversals. Models without reversals may fit the explosion data, but the data cannot rule out reversals. LVZs may be more pronounced in $V_\rho$ than in $V_s$, and thus represent zones of high $V_\rho/V_s$. A comprehensive map of $V_\rho/V_s$ across Iceland could significantly improve our knowledge of lower crustal LVZs. However, although $V_\rho$ and $V_s$ have been well investigated over Iceland separately, using explosion seismology and receiver functions, respectively, such a map cannot yet be produced because the methods differ so fundamentally that the combined errors would be larger than the variations. LVZs are not generally modelled for explosion seismology data and thus the full range of models compatible with the data is not known. The most significant LVZ detected by receiver functions occupies an area of $>10$ 000 km$^2$ of the MVZ and has velocities up to $\sim$8 per cent low throughout the depth interval $\sim$20–35 km (Du & Foulger 1999, 2001; Du et al. 2002). The ICEMELT profile passed near the periphery of this zone and an apparent shadow zone may be seen in the published sections, though it was not interpreted (Darbyshire et al. 1998). The possibility that this LVZ in $V_s$ is also associated with a LVZ in $V_\rho$ has thus not yet been investigated.

On high-quality explosion seismology profiles, clear post-critical reflections are usually detected from depths of $\sim$20–40 km. These are generally interpreted as indicating a simple, relatively sharp, laterally extensive Moho with a velocity jump of $V_\rho = 0.45$–0.95 km s$^{-1}$ (corresponding to $V_s = 0.25$–0.55 km s$^{-1}$, assuming $V_\rho/V_s = 1.76$). Such a discontinuity is not detected by the receiver functions, however, which indicate that a gradational crust–mantle boundary several kilometres thick is common (Fig. 6). This is an important point, since receiver functions are powerful for detecting sharp reflective discontinuities if they exist.

Note that variations in the maps of total crustal thickness include:

1. the locus of thickest crust determined using receiver functions is centred on the confluence of the MVZ, NVZ and EVZ (Foulger et al. 2000, 2001), whereas explosion seismology/gravity suggests it underlies northwest Vatnajokull;
2. the crust beneath the Tertiary surface rocks east of Vatnajokull is significantly thinner in the receiver-function model than in the explosion seismology/gravity model (cf. Figs 7b and 5).

Reconciling the receiver function and explosion seismology results

To reconcile the receiver-function and explosion-seismology results, we investigated the most significant differences in the models: sharp first-order Moho discontinuities and the absence of crustal LVZs in published explosion-seismology models. Comparing published explosion seismic data with theoretical seismograms computed using a modern, full-wave reflectivity method (Kennett 1983; Randall 1989), we found that the interpretation of seismic-refraction data is not unique; there is a trade-off between crustal thickness and the sharpness of the Moho discontinuity. The largest possible crustal thicknesses correspond to sharp first-order Moho discontinuities, and transitional zones several kilometres thick are compatible with smaller crustal thicknesses and higher wave speeds in the lower crust. Furthermore, the explosion data are consistent with the presence of crustal LVZs, and in many cases contain strong evidence for them.

A transitional crust–mantle boundary

Receiver functions rule out a sharp Moho discontinuity with $\Delta V_\rho \gtrsim 0.5$ km s$^{-1}$ ($\Delta V_s \gtrsim 0.3$ km s$^{-1}$) occurring over a depth interval of less than 2 km because it would produce strong near-vertical reflections that are not observed. This is shown by synthetic receiver-function modelling that shows that a strong, sharp discontinuity would generate clear $Ps$ and $PpPms$ converted phases, which are not
observed (Fig. 8). Seismic refraction data also place constraints on the sharpness of the Moho for similar reasons: sharp discontinuities produce larger pre-critical reflected waves. Fig. 9 shows observed seismograms for the Reydarfjordur shot of the FIRE experiment (Staples et al. 1997) (Fig. 1), and compares them with synthetic seismograms computed using the reflectivity program of Kennett (1983) and Randall (1989). A model with a sharp Moho at 29 km depth was used (Fig. 10, dashed line), which is the approximate depth to the crust–mantle boundary determined by receiver functions in the area. The later-arriving \( PmP \) phase on the theoretical seismograms is the most prominent arrival at distances beyond approximately 65 km, whereas on observed records it is essentially undetectable at distances of less than 100 km. The Reydarfjordur data are quite consistent, however, with a transitional boundary at this depth, such as that shown as a solid line in Fig. 10. The seismograms computed from this model (Figs 11f and g) strongly resemble the observed data (Figs 11a and b), with the \( PmP \) phase being the dominant arrival at distances beyond 100 km, and essentially undetectable at shorter distances.

It is possible to fit the observed amplitude–distance relation of these data fairly well with a sharp Moho if it is 35 km deep, as is illustrated in fig. 13 of Staples et al. (1997), and if the wave speed is higher in the lower crust (so the velocity discontinuity at the Moho is smaller). However, this depth is significantly larger than that found from receiver functions in this area (Fig. 7b), and is not compatible with those data.

**Mid-crustal LVZs**

The upper- and lower-crustal LVZs common in models derived from receiver functions are not ruled out by explosion seismology, which in fact provides evidence in support of such features. Apparent shadow zones starting at distances of 60–120 km are visible in several published seismogram record sections (e.g. Staples et al. 1997, Figs 4, 6, 7 and 8), and can be explained by LVZs in the lower crust. Both ray-theoretical traveltine curves and full-wave reflectivity seismograms agree well with the observations. If LVZs are not accounted for in interpreting explosion seismology data, then the depths to deeper discontinuities and the magnitudes of their velocity jumps are overestimated. Such errors may partly account for the differences in crustal thicknesses obtained from explosion seismology and receiver functions. In particular, LVZs are expected beneath the major volcano cluster under northwest Vatnajokull, and are required there by receiver functions (Du & Foulger 2001).
Figure 9. Effect of sharp first-order Moho discontinuities at different depths on seismic waves. All plots have the same scale and reducing velocity. Seismograms are plotted with gain proportional to the epicentral distance. (a) Observed vertical-component seismograms for the Reydarfjordur shot on the FIRE profile, filtered with a pass band from 3 to 8 Hz (Staples et al. 1997, Fig. 4a). (b) The same, filtered with a pass band from 0.5 to 2 Hz (Staples et al. 1997, Fig. 4b). (c) Theoretical velocity seismograms computed using the reflectivity method (Kennett 1983; Randall 1989) for an impulsive source at 30 m depth in the model shown as a dashed line in Fig. 10. (d) The same, with the effects of the source time function and the low-pass filter of (b) represented by a Gabor wavelet with parameters $f_m = 2.5$ Hz, $\gamma = 4$, $\nu = 0$, $t_i = 0.5$ s (Cerveny 2001, Section 6.1). A sharp Moho at 29 km predicts large reflected later arrivals at distances less than 100 km that are not observed.

Fig. 11 illustrates the compatibility of the Reydarfjordur shot data with a model having an upper-crustal LVZ (Fig. 10). Curvature of the first-arrival traveltime curves at distances of less than approximately 50 km implies that wave speeds increase rapidly in the upper ~5 km. For the Reydarfjordur profile, $V_p$ reaches 6 km s$^{-1}$ at the unusually shallow depth of approximately 1 km (on other profiles it is a few km deeper) and a further rapid increase to 6.5 km s$^{-1}$ occurs at 5 km depth. This increase causes a triplication in the traveltime curve, with large later arrivals from approximately 20 to 40 km distance, and a new branch becomes the first arrival at approximately 40 km. The large later arrivals of the triplication also occur in the surface-reflected PP phase, which are clear on the theoretical seismograms at approximately 55–90 km distance (Fig. 11f). These waves are also detectable on the low-pass filtered observed seismograms, but they are weak, probably because of heterogeneity near the surface reflection point. The new first-arrival branch has small amplitudes, and cannot be seen beyond approximately 60 km, where a larger branch, delayed by approximately 0.3 s, begins. Such observations
Anelastic wave attenuation

The theoretical seismograms shown in Figs 9(c), (d), 11(f) and (g) were computed using anelasticity models based on the work of Menke & Levin (1994) and Menke et al. (1995). In the upper 4 km $Q_p = 60$ and $Q_s = 100$, and both values increase with depth, reaching values of $Q_p = 800$ and $Q_s = 1500$ at 12 km depth. In the lower crust, Menke et al. (1995) infer slightly increased shear wave attenuation ($Q_s = 800$) but negligible compressional-wave attenuation ($Q_p > 5000$). However, we find that such values predict PnP amplitudes much larger than are observed. Below 16 km we use $Q_p = 150$ and $Q_s = 500$, values that give amplitudes similar to those observed.

In the NVZ, Staples et al. (1997) observe PnP-like arrivals at shorter distances and earlier times than elsewhere, and they interpret these in terms of crustal thinning from 34 to 20 km. The resulting model predicts diving waves (Pn) as first arrivals where none are observed, from which Staples et al. (1997) infer extraordinarily high attenuation ($Q_p < 20$) in the mantle. An alternative explanation is that the observed waves are reflected from thin LVZs at approximately 20 km and not from the crust–mantle boundary. This would also explain the lack of the ~40 mGal gravity anomaly predicted by such an extreme mantle upwelling. This must otherwise be cancelled out by invoking strong density anomalies in the mantle, which imply an implausible 700 °C temperature anomaly, and for which there is no independent evidence.

A structural model that satisfies all seismic data

As illustrated above, the explosion seismology data are compatible with the receiver functions. The receiver functions require LVZs in the crust, and the explosion data show evidence of these in the shallow crust and are insensitive to such zones in the lower crust. Receiver functions rule out a sharp velocity discontinuity at the base of the crust, and a crust–mantle transition zone that is either gradational or comprises thin, alternating, high- and low-velocity layers is consistent with both types of data. The structural resolution of the receiver functions at the base of the crust is ~2.5 km and for a zone of thin layers the inversion process will yield a smoothed profile with velocities averaged on that scale. Such a structure was suggested for a reflective horizon detected by explosion seismology in the lower crust west of the SW tip of the Reykjanes peninsula by Weir et al. (2001). This boundary was not identified as the Moho only because of gravity modelling and the observation of deeper reflectors. If the crust–mantle transition zone does contain such fine structure, the reflections detected by explosion seismology profiles might then come from either the shallowest high-velocity layer or from constructive interference of reflections throughout much of the stack.

Fig. 12 illustrates the first-order structural models proposed for Icelandic-type crust from explosion seismology data (Fig. 12a) and receiver functions (Fig. 12b). These figures show those features of the crust in Iceland reported from most locations, and local detail is omitted. They are similar except for the nature of the crust–mantle transition zone, illustrating that the first-order results from the two methods are similar. Fig. 12(c) shows a schematic, composite model that combines the features observed in Icelandic-type crust at various localities. This figure is not intended to indicate any one structure, nor the structure everywhere, but to summarize the main features that occur, that are compatible with both the receiver functions and explosion data. This model contains the following features.

1. The upper crust is typically 7 ± 1 km thick, with extremal values of 5.5 and 11 km at various places in Iceland. The velocity gradients are variable, but typically high. High-velocity zones (HVZs) occur in some areas, often near extinct central volcanoes, and there velocities may be as high as ~6.0 km s$^{-1}$ at shallow depth.
Figure 11. Comparison of observed and theoretical seismic data for a model with a mid-crustal LVZ and a transitional Moho discontinuity. (a) and (b) Observed seismograms, as in Figs 9(a) and (b), (c) ray-theoretical seismograms for the 2-D model of Staples et al. (1997), (d) paths of seismic rays considered by Staples et al. (1997) in interpreting data, (e) ray-theoretical $P$-wave traveltimes for the 1-D crustal model shown as a continuous line in Fig. 10, computed using the method of Julian & Anderson (1968), (f) and (g) theoretical velocity seismograms for the same model, computed as in Figs 9(c) and (d).

(2) The lower crust is typically 15–30 ± 5 km thick, with extremal values of 13 and 32 km. Its average velocity gradient is an order of magnitude lower than that of the upper crust. Its top is usually at approximately the $V_p = 6.5 \text{ km s}^{-1}$ level, but this is not everywhere the shallowest occurrence of such a high velocity since extreme HVZs occur at some locations in the upper crust. The lower crust may contain LVZs, which are usually thin and with minor wave speed anomalies, but may be thick and with major wave speed anomalies, e.g. beneath central Iceland (Fig. 6b).

(3) The crust–mantle transition occurs over a zone typically 5 ± 3 km thick, but which may be thinner or thicker than this from place.
to place. The average velocity increases from $V_p \sim 7.2$ to 8.0 km s$^{-1}$ within this transition, which may be gradational or a stack of thin layers with alternating high and low velocities.

(4) The top of the upper mantle may have a negative velocity gradient beneath Iceland, and a positive velocity gradient beneath the Iceland–Faeroe ridge.

**PETROLOGICAL COMPOSITION OF ICELANDIC-TYPE CRUST AND THE CRUST–MANTLE BOUNDARY**

**The upper crust**

The upper Icelandic crust probably corresponds to layers 0–2 in oceanic crust and ophiolites. Its vertical and lateral heterogeneity probably results from its composition of a mélangé of lava flows, subsided hyaloclastites, intrusions of all geometries and small, local magma bodies, in particular beneath central volcanoes (e.g. Foulger & Toomey 1989; Foulger et al. 1995). Its petrological variability may result from the remelting of subsided, hydrated basalts (Oskarsson et al. 1982) or from source heterogeneity (Foulger et al. 2002). The average seismic velocity increases rapidly downwards because of compaction due to increasing overburden pressure and mineralization. These progressively close cracks and pores, and increase the metamorphic grade from zeolite through chlorite-epidote and greenschist to amphibolite facies (e.g. Oskarsson et al. 1982; Flovenz & Gunnarsson 1991). It has been suggested that the base of the oceanic upper crust, at $\sim 3$ km depth, represents the depth at which all cracks and pores are closed as a result of overburden pressure. This is unlikely, however, since open cracks exist in Icelandic crust down to at least 7 km, where pressures are much greater. This is known from non-double-couple earthquakes, the source of which processes involve the opening and closing of cracks (Foulger 1988; Foulger et al. 1989; Miller et al. 1998a,b). The base of the upper crust is also unlikely to correspond to a single metamorphic horizon, such as the transition from amphibolite to granulite facies, because then a systematic thinning of the upper crust over the rift zones, where geothermal gradients are high, would be predicted, but this is not observed (Fig. 7a). By analogy with ophiolites, the base of the upper crust probably corresponds to the transition from basalt to gabbroic petrology (Mutter & Mutter 1993).

**The lower crust**

The Icelandic lower crust probably corresponds to oceanic layer 3, which in ophiolites is a largely gabbroic sequence in the amphibolite to granulite metamorphic facies. Epidote is probably also important in the lower crust (Christiansen & Wilkins 1982; Flovenz & Gunnarsson 1991). The velocity which marks the top of the lower crust, $V_p \sim 6.5$ km s$^{-1}$, is approximately the upper limit of velocities in unfractured basalt (Mutter & Mutter 1993). However, increasing pressure in basalt alone is not sufficient to explain higher velocities. A change in petrology, probably to gabbro, is required, and velocities deep in the lower crust of $V_p \gtrsim 7.1$ km s$^{-1}$ require an ultramafic component.

The LVZs that are a significant feature of the lower crust may result from variations in temperature, petrology, metamorphic facies or the presence of fluids, especially magma, and such variations are also required to explain the reflective horizons occasionally observed in the lower crust (e.g. Weir et al. 2001). The low attenuation observed for shear waves in the lower crust permits only small percentages of pervasive partial melt (Menke & Levin 1994). Slightly higher shear wave attenuation observed beneath central Iceland compared with south Iceland may indicate temperatures $\sim 75^\circ$C hotter beneath central Iceland (Menke & Levin 1994), which would correspond to a reduction in $V_s$ of $\sim 1.5$ per cent (Humphreys & Dueker 1994). This is less than the maximum of $\sim 8$ per cent observed. The remaining anomaly could be explained by less than 1 per cent partial melt, which would depress $V_s$ much more than $V_p$. $V_p/V_s$ might then be as high as $\sim 1.9$ beneath the MVZ, but its value in that region has not been measured to date.
The crust–mantle boundary

The seismological Moho was first observed beneath Europe, where it comprises a velocity discontinuity of $\sim$1.0 km s$^{-1}$ in $V_p$. A seismic discontinuity, or rapid increase in velocity, is also widespread at the base of the oceanic crust, but is usually weaker, typically $\sim$0.6–0.8 km s$^{-1}$ in $V_p$. Laboratory experiments on ophiolites suggest that the oceanic seismological Moho corresponds to the boundary between layered gabbros with some interlayered peridotites above, and harzburgite and dunites below (Fig. 2).

Petrological models predict that, if oceanic crust is formed from partially melting mantle peridotite, a high-velocity, dunite-rich material is deposited at the base of the crust (e.g. Farth et al. 1996). For peridotite, the greater the depth of melting, the more potential olivine is contained in the melt (e.g. Herzberg & O'Hara 1998). Low-degree partial melting of peridotites at depths of $\geq$70 km produces picritic melt, which contains more olivine than the tholeiitic lavas erupted at the surface (e.g. Stolper 1980), and higher MgO content (15–18 wt per cent MgO, compared with 6–8 wt per cent in surface-erupted tholeiites). These differences require that significant quantities of olivine and MgO are removed from the melt by fractional crystallization before eruption at the surface. Fractionation is expected to occur preferentially near the base of the crust, a major change in composition, density and rheology and thus a potential impediment to upward movement of dense melts. Petrological models therefore predict a distinct, dunite-rich, high-velocity layer at the base of the crust, comprising olivine, augitic pyroxene, and, at depths shallower than $\sim$30 km, variable amounts of plagioclase. Kelemen et al. (1997) have suggested that replacive dunites mark conduits for focused porous flow of MORB melt in the mantle, and provide the geochemical isolation from surrounding peridotite that is required by observations that residual peridotites are not in equilibrium with MORB.

The gradational removal of plagioclase with depth through its combination with olivine to form clinopyroxene would also contribute to gradational downwards velocity increase with depth. The high-velocity basal crustal layer may thus comprise a pile of thinner layers with various petrological compositions, including both crustal cumulates and interleaved mantle peridotites (Cox 1980; Kelemen & Aharonov 1998, and references therein) throughout which velocities rise from lower-crustal to upper-mantle values. Such interleaving is observed in the field, e.g. in the Rum intrusion and the Great Dyke of Zimbabwe (Huppert & Sparks 1980; Singh & McKenzie 1993). The layer is expected to exhibit seismic velocities in the range $V_p \sim 7.5–7.9$ km s$^{-1}$. The absence of such a layer would require that the melting that supplied crustal material occurred at depths of $\geq$30 km, which is not possible since it would then be occurring within the $\sim$30 km thick crust itself. The model proposed here also predicts olivine gabbroic intrusions with $V_p \sim 6.8–7.0$ km s$^{-1}$ in the upper crust, which are indeed observed in Iceland.

Such a petrological model accounts for the broad, first-order features of the gradational crust–mantle transition zone beneath Iceland, if it is assumed that the thick Icelandic crust is formed from essentially similar but scaled-up processes to normal oceanic crust. If Icelandic-type crust is up to $\sim$40 km thick, and if the thickness of the basal zone is $\sim$20 per cent of the overlying crust, its thickness would increase from approximately 3 to 8 km between the coast and central Iceland. If it is assumed that melting occurs beneath the crust, the average melting depth will be greatest beneath the thickest crust, and the model would predict that a slightly thicker crust–mantle transition zone would occur there. Plagioclase, which has relatively low velocities, would be expected to form a component in the crust–mantle transition zone shallower than $\sim$30 km. This would tend to suppress the seismological contrast between the crust–mantle transition zone and the overlying lower crust beneath thinner parts of the Icelandic crust, and reduce the apparent thickness of the transition zone. It is not clear why the phase change associated with the disappearance of plagioclase is not associated with a clear seismological horizon beneath Iceland or elsewhere in thick oceanic crust.

If the crust varies from $\sim$20 km to $\sim$40 km thick, the composition of lowest crustal material must also vary laterally. Pyroxene hornfels facies assemblages, in which plagioclase is stable with olivine, will not be stable in those parts of the crust $\geq$30 km deep, where the olivine and plagioclase will combine and a pyroxene granulite will result. If temperatures are low at the base of the crust, garnet might also be present in varying amounts, forming garnet-granulites. Hornblende might also occur if sufficient water is present. Such variations in petrology affect seismic velocity and may contribute to the gradational nature of the crust–mantle boundary.

The gradational crust–mantle transition zone suggests a downwards increase in abundance of high-velocity rocks such as dunite and harzburgite compared with lower-velocity pyroxenites, wehrlites and troctolites rather than an abrupt transition from one petrology to another. Such a model also helps to explain the anomalously low density contrast of 2–4 per cent between the lower crust and mantle predicted by isostasy, which is consistent with only minor changes in composition between the crust and mantle (Menke 1999; Darbyshire et al. 2000b). Assuming maximum basal velocities of $V_p = 7.1–7.3$ km s$^{-1}$, Menke (1999) finds that the effects of reasonable mantle thermal, entrainment and depletion effects cannot fully account for such a small density contrast with the crust. Simply put, this means that if a model is assumed that comprises a lower crust, separated from the upper mantle by a sharp, or thin transition zone, isostatic considerations require that the density of the lower crust is inexplicably high. A crust–mantle transition zone $\sim$20 per cent of total crustal thickness and with an average velocity of $V_p \sim 7.5$ km s$^{-1}$ would account for much of the unexplained discrepancy. A small crust–mantle boundary density contrast would further encourage polybaric crystalization and the distribution of olivine throughout a relatively large depth range as melt ascends. Like the lower-crustal LVZs, the slight velocity reversal beneath the crust–mantle transition zone shallower than $\sim$30 km is not explained. A crust–mantle transition zone suggested by the rarity of observed mantle-diving rays in Iceland might be explained by a high geothermal gradient at the top of the mantle underneath much of Iceland. However, there is no direct seismic evidence from anomalously high $V_p/V_s$ ratios or attenuation for melt at the base of the crust or in the crust–mantle transition zone. The apparent lack of a velocity reversal in the topmost mantle beneath the Iceland–Faeroe ridge (Bott & Gunnarsson 1980; Smallwood et al. 1999) is consistent with the lithospheric mantle there having cooled as it was transported away from the ridge by plate motion.

It has recently been suggested that much of the great crustal thickness at Iceland may result from the melting of eclogite in the Caledonian suture, still remaining in the upper mantle beneath Iceland, and not entirely from peridotite partial melt (Foulger 2002; Foulger et al. 2002, 2003). In this case, large quantities of fractionated dunite at the crust–mantle transition would not be expected. An alternative model to explain the gradational crust–mantle transition zone and the unusually small density contrast between the lower crust and the mantle might then be gradation between crustal and mantle rocks over a depth interval of tens of kilometres rather than the few kilometres proposed for oceanic crust (Cannat 1996). Such a diffuse gradation might be a consequence of

© 2003 RAS, GJI, 155, 567–590
the exceptionally large and vigorous flux of melt from the mantle, and raises the question of what the true relationship is between the seismological and the magmatic thickness of Icelandic-type crust.

**VARIATION IN CRUSTAL THICKNESS BENEATH THE ICELAND TRANSVERSE RIDGE**

Because the crust–mantle transition zone is gradational, and undetected LVZs and multiple reflectors may exist in the deep crust, there is likely to be significant uncertainty in estimates of crustal thickness both in the receiver function and explosion seismology results. Nevertheless, it is tempting to view maps of crustal thickness over the Iceland transverse ridge as snapshots of the top of a plume beneath central Iceland or northwest Vatnajökull that delivers an anomalously large amount of melt and creates the great thickness of crust there. However, careful consideration of the observations reveals difficulties with this interpretation. Since crustal spreading is ongoing, if a plume underlies central Iceland, the crust would be expected to have a uniform thickness of ~40 km parallel to the spreading direction. This is not the case. The crustal thickness falls rapidly to ~25 km within ~200 km west of the thickest spot and to ~30 km at a similar distance to the east. Beneath the Greenland–Iceland ridge the crust is ~33 km thick and beneath the Iceland–Faeroe ridge it is ~25–30 km thick. Furthermore, there is no evidence from seismic attenuation, petrology or geochemistry for high temperatures beneath central Iceland compared with coastal regions.

The exceptionally large crustal thickness beneath central Iceland detected by receiver functions is best explained as a foundered oceanic microplate (see Foulger 2002, 2003 and references therein). In the region that is currently north Iceland, a second rift zone, parallel to the MAR, formed ~26 Ma. An oceanic microplate ~100 km wide containing Icelandic-type crust up to ~33 Myr old was captured between this and the MAR. Tectonic reconstructions show that this microplate currently underlies central Iceland. Surface lava flows have thus been added to the existing crust since ~26 Ma, and the oldest exposed rock at the surface is currently ~12 Myr old (Saemundsson 1979). Lava accumulation rates near the rift zones in Iceland are typically 1–2 km Ma⁻¹ and thus a growth in thickness of ~10 km of crust originally ~30 km thick is in accord with models of crustal growth in Iceland. This model explains the extensive LVZ in the lower crust since it suggests that the upper part of the lower crust is subsided, former upper crustal material.

Other possible explanations for the exceptionally thick crust beneath the centre of Iceland include a very recent increase in magma production locally beneath central Iceland. This would imply an ephemeral process that has not previously occurred since the opening of the Atlantic 54 Ma and is inconsistent with a progressively declining plume. It would also imply that the observed ‘plume-like’ pattern of crustal thickness is a fortuitous consequence of our happening to look at the moment of maximum melt production in the history of the hotspot, and that it would not have been observed had crustal thickness been mapped at any other time. Shallowing of the seismological base of the crust with age might also explain the observations. Kinematic modelling predicts that the crust cools as it subsides and is transported laterally away from the spreading axis (e.g. Palmason 1980; Menke & Sparks 1995). The apparent additional ~10 km of crustal thickness beneath central Iceland might then be an artefact of locally high temperatures and partial melt that depresses velocities in the uppermost mantle to crustal values. Estimates of seismic attenuation argue against this, however. Staples (1997) has suggested that the crust may be ~40 km thick everywhere beneath the Icelandic transverse ridge but that the lowest ~10 km of crust transforms from gabbro to higher-velocity garnet granulite over tens of millions of years, and is mistakenly identified as upper mantle. The thick crust does not, however, underly the currently active rift zones everywhere, and so this argument would apply to crust formed at the centre of Iceland only.

The east–west asymmetry in crustal thickness about central Iceland (Figs 5 and 7b) argues against the theory that a plume has migrated eastwards from beneath Greenland since the opening of the North Atlantic (e.g. Vink 1984; Lawver & Muller 1994). This theory predicts that a plume underlay north–west Iceland ~20 Ma, subsequently migrated SE across Iceland, and now underlies Vatnajökull. Such a model would predict thicker crust in the wake of the plume beneath the Greenland–Iceland ridge and north–west Iceland, and thinner crust beneath eastern Iceland and the Iceland–Faeroe ridge, ahead of the plume. Such a pattern is not observed. The crust beneath the Greenland–Iceland and Iceland–Faeroe ridges is approximately equally thick. In Iceland, the observed pattern is opposite to that predicted, with thinner crust beneath western than eastern Iceland. Heat flow is also lower west of the Reykjanes ridge than east of it, also arguing against an easterly migrating plume (Stein & Stein 2003). These results are more consistent with a model whereby the source of excessive melt in the Iceland region has been beneath the MAR since the opening of the North Atlantic (e.g. Bott 1985; Foulger et al. 2000, 2001, 2003; Foulger 2002, 2003).

The asymmetry of crustal thickness about central Iceland probably results from crustal accretion having occurred along a parallel pair of rift zones since ~26 Ma (Bott 1985; Foulger 2002). The crust beneath the Western Fjords and eastern Iceland formed ~10–15 Ma, but along two widely separated rift zones. This is clear from the fact that the ~15 Ma isochrons in Iceland are currently ~200 km further apart than they are across the Reykjanes and Kolbeinsey ridges. Thus, although there is currently a single spreading zone in north Iceland (the NVZ), this has probably only been the case for the last ~7 Myr (Jancin et al. 1985; Foulger 2002). That the crust beneath the Western Fjords and western Iceland is ~5 km thinner than corresponding crust beneath eastern Iceland suggests that the western rift zone was less productive than the eastern zone. This is in keeping with its subsequent extinction in favour of the eastern rift zone (the NVZ), and the subsidiary nature of crustal accretion in western Iceland today.

The crustal thickness models based on explosion seismology show thinning of the crust to ~20 km to the north and south of central Iceland, locally beneath the EVZ (Darbyshire et al. 2000b) and NVZ (Darbyshire et al. 2000b). The gravity field shows no anomalies that support such thinning (Eysteinsson & Gunnarsson 1995). In order to model an essentially flat gravity field with a seismic structure involving a severely updoming mantle, Staples et al. (1997) had to invoke compensating low crustal and mantle densities for which there is no independent evidence. Such thinning is not expected for the following reasons. The EVZ is highly volcanically active, and formed at ~2 Ma in pre-existing crust. Only ~10 km of crustal widening will have occurred subsequently, assuming spreading to have been shared with the WVZ. Additional volcanic activity at this location would be expected to slightly thicken the crust by the addition of surface lava flows and intrusions. In the case of the NVZ, locally thin crust is at odds with the agreement of all crustal thickness studies that the crust on the flanks is thicker. Anomalously thin crust beneath the current NVZ would require that magmatism beneath the NVZ has been significantly reduced for the last several Myr. There is no independent evidence for such waning in activity.
**Discussion**

Critical to any discussion of crustal thickness in Iceland is a clear definition of what is meant by the base of the crust. The Mohorovicic discontinuity was first observed seismically beneath Europe, where it comprises a strong discontinuity of $\sim 1.0$ km s$^{-1}$ in $V_p$ (Mohorovicic 1909). It is often assumed to be sharp, despite the fact that pre-critical reflections are then predicted but not observed. Its definition is often adapted to whatever is observed in a particular study area, and it is sometimes even used to describe changes in velocity gradient without velocity discontinuities. Vague usage of the term, coupled with the assumption that such a structure separates the crust and mantle everywhere, potentially obscures the geology by substituting classification for understanding. Despite the simple, elegant crustal subdivision proposed by Flovenz (1980), and widely accepted as the most natural for oceanic crust in general, multilayer subdivisions for Icelandic-type crust based on absolute velocity that fit the data poorly are still commonly adopted (e.g. Bjarnason et al. 1993; Darbyshire et al. 1998).

Seismic models of Icelandic-type crust and underlying upper mantle have evolved from an original thick model where velocities of $V_p \leq 7.2$ km s$^{-1}$ were attributed to crust (Báth 1960; Zverev et al. 1976) through a thin–hot crust model, where material of $V_p \geq 7.0$ km s$^{-1}$ was attributed to partially molten upper mantle at temperatures of $\sim 1100$ °C (Tryggvason 1962; Palmason 1971; Angenheister et al. 1980; Gebran 1980) and back to a thick–cold crust model, with material of $V_p \sim 7.2$ km s$^{-1}$ attributed to sub-solidus crust with temperatures of $\leq 900$ °C (e.g. Bjarnason et al. 1993; Staples et al. 1997; Darbyshire et al. 1998; Menke et al. 1998). It is difficult to imagine more dissimilar interpretations for geophysical data, and this provides an interesting cautionary tale. Both the thin–hot crust model and the thick–cold crust model have been widely interpreted as providing evidence for a plume. This suggests that no crustal structure would be considered inconsistent with this hypothesis, and that an Icelandic plume is an a priori assumption rather than a hypothesis.

The key factor in the recent rejection of the thin–hot crust model was the presence of a Moho reflector first recognized by Báth (1960). In addition, “mid-crustal” reflectors have also been observed, e.g. in the lower crust offshore and to the west of the Reykjanes peninsula (Weir et al. 2001). It is possible that the mid-crustal reflector would have been interpreted as the Moho had the deeper reflector not been observed, and this raises the question regarding the continuity of the deep reflectors interpreted as a Moho beneath Iceland. These are observed on a piecemeal basis and, where the segments detected are spatially close and at similar depths, it is reasonable to propose that they are parts of a continuous structure. However, where they are widely separated spatially, and at radically different depths, such an interpolation is less safe. The deeper part of the crust and the crust–mantle transition zone have considerable thickness and probably comprise a stack of subhorizontal layers with significant velocity contrasts, some of which may be reflective sills or petrologically contrasting layers in the lower crust. Later reflections may be difficult to observe, since they would be masked by earlier reflections. Icelandic crust may thus contain more than one strong reflective horizon and the shallowest everywhere may not comprise a single, laterally continuous horizon.

The model of Staples et al. (1997), for example, has a Moho that shallows from $\sim 35$ km beneath eastern Iceland to $\sim 19$ km beneath the NVZ, with a dip of up to $\sim 35\$° 2-D gravity modelling of this structure predicts a gravity anomaly of $\sim 50$ mGal, which is not observed, a temperature contrast of $\sim 700$ °C, and requires a special explanation for how such extreme lateral variability could be maintained mechanically. No such thinning is observed beneath the WVZ. It seems paradoxical that the crust beneath the NVZ, thought to be more productive of melt than the WVZ, should be thinner. A model in which a deeper reflector is laterally continuous beneath a shallower reflector under the NVZ, and crustal thickness varies only mildly, might be consistent with the seismic data and more geologically plausible. Such a deeper reflector was detected nearby in the RRISP data (Bjarnason et al. 1993). Dramatic local variations in crustal thickness are not observed beneath the Greenland–Iceland and Iceland–Faeroe ridges.

The model whereby the thickest crust beneath central Iceland is interpreted as a location of enhanced melt production associated with a plume (Darbyshire et al. 2000b; Allen et al. 2002) is at odds with seismic-wave attenuation studies of central Iceland (Menke & Levin 1994), which show that the lower crust is relatively cold, and substantially below the solidus of gabbro at 30–40 km depths. Since Icelandic crust is much thicker than crust at MORs, this implies that the top 40 km is colder over the presumed plume centre beneath MORs. The mantle beneath is not likely to be as hot as under normal MORs at equivalent depths since the lowermost crust would then be extensively molten. Uppermost mantle velocities estimated from explosion seismology lie in the range $V_p = 7.7–8.3$ km s$^{-1}$, compared with uppermost mantle velocities beneath MORs of $V_p = 7.3–7.8$ km s$^{-1}$ for crust–mantle transition zones. The seismic observations thus provide no evidence for anomalously high temperatures in the uppermost mantle beneath Iceland either. Menke et al. (1998) suggested that a subsolids ‘lid’ overlies an underlying hot mantle plume. An alternative interpretation of the observations is that they provide no support for, but evidence against, a hot plume.

From isostatic considerations, Menke (1999) concluded that the crust–mantle density contrast beneath Iceland is $39 \pm 12$ kg m$^{-3}$. Several studies have jointly modelled explosion seismology crustal structures and gravity assuming crust–mantle density contrasts of $\sim 200$ kg m$^{-3}$ (Staples et al. 1997; Darbyshire et al. 2000b; Weir et al. 2001). Local mantle density deficiencies of up to $\sim 3$ per cent have been proposed to account for discrepancies between the two data sets beneath the rift zones and central Iceland. However, a gradational crust–mantle transition zone that varies in thickness by $\sim 5$ km beneath Iceland, or a thick zone of crust–mantle mixing, could account for lateral variations in gravity of up to $\sim 20$ mGal, and LVZs in the upper and lower crust could account for an additional several tens of mGal. It is thus not clear to what extent density deficiencies in the mantle are required.

It has been suggested that $^3$He/$^4$He culminates in central Iceland and that this indicates the centre of a plume (Bredell et al. 2000). However, a radial geometry of $^3$He/$^4$He would not be expected for an eastward-migrating plume. Instead, higher $^3$He/$^4$He would be expected throughout western and central Iceland than in eastern Iceland. Such an asymmetric pattern is not observed. Measurements of helium isotope ratios in rocks dredged from the Greenland–Iceland and Iceland–Faeroe ridges would contribute valuable additional data to this debate, and to the debate concerning the depth of origin of anomalous $^3$He/$^4$He values (Foulger & Pearson 2001).

It is interesting to speculate to what extent the Iceland volcanic province resembles other oceanic LIPs. High-quality seismic data constraining lower-crustal and basal-crustal structure are available from the Ontong–Java Plateau (Furumoto et al. 1970, 1976; Richardson et al. 2000), the Kerguelen Islands (Req et al. 1990), the Madagascar ridge (Sinha & Louden 1981) and Hawaii (Watts et al. 1985). In all cases high-velocity basal crustal layers are detected up to 30 per cent as thick as the crust (Table 2).
suggest that the structure of Icelandic-type crust is generically similar to that of thick oceanic crust elsewhere. It is not clear to what extent its location over a spreading plate boundary affects its crustal structure, but formation at ridges has been suggested for oceanic LIPs in general (Anderson 1994).

It is interesting to note that there are several reports of plume-like uppermantle low-velocity bodies in places where plumes are thought to have once existed but now are long extinct, or where plumes are not expected. In addition to being underlain by thick crust with high basal velocities, the mantle beneath the Ontong–Java Plateau contains a body with a velocity anomaly as strong as $V_p = -5$ per cent throughout the depth range $\sim 50–300$ km (Richardson et al. 2000). This crust and mantle structure is almost identical to that observed beneath Iceland (Foulger et al. 2000, 2001) and yet volcanism at the Ontong–Java Plateau ceased $\sim 90$ Ma and there is no surface heat-flow anomaly. Mantle low-velocity bodies have been detected beneath the Paraná flood basalts, South America (VanDecar et al. 1995) and northwest India (Kennett & Widiantoro 1999), which ceased to be active $\sim 80$ and $\sim 65$ Ma, respectively. These observations suggest that a high-temperature interpretation of the mantle low-velocity body beneath Iceland is non-unique and that composition and very low-degree partial melt may be important.

Notwithstanding the extensive work done in Iceland, a number of key problems still remain unsolved. The thick-crust model for Iceland, which is supported by petrology and marine heat-flow measurements (Stein & Stein 2003), seems at odds with the hypothesis that central Iceland is underlain by a hot plume. On the other hand, Iceland is clearly a site of enhanced melt production, and the most natural model for such productivity in the absence of a large temperature anomaly is anomalous mantle composition (Foulger 2002). The question remains unanswered, however, by what structures and mechanisms melt is transported from its zone of origin in the mantle up through tens of kilometres of crust to shallow holding chambers or directly to the surface.

ACKNOWLEDGMENTS

We thank Mike O’Hara, Don Anderson, Warren Hamilton, Jerry Winterer, Niels Oskarsson, Olafur Flovenz, Dave Hill, Walter Mooney, Bob White and Rob Staples for helpful discussions. This research was supported by Natural Environment Research Council (NERC) grant GR3/10727 and a Sir James Knott Foundation fellowship held by GRF. Reviews by two anonymous reviewers improved the manuscript.

This paper is dedicated to the memory of Professor Ron Girdler, whose unconditional enthusiasm, support and encouragement inspired us, and who will be remembered with gratitude and affection.

REFERENCES


© 2003 RAS, GJI, 155, 567–590


