Geophys. J. Int. (2003) **155,** 567–590

Icelandic-type crust

G. R. Foulger,[1] Z. Du[2] and B. R. Julian[3]

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

Accepted 2003 June 10. Received 2003 June 10; in original form 2001 October 30

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 $V_p \sim 6.5$ 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 three-component 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 'thick–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-orientated 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). Extinct spreading zones lie along the northwestern edge of the Western Fjords (WF; Fig. 1), and along the Skagi peninsula. The Tertiary area between the NVZ and the extinct Skagi zone is called the Trollaskagi block and the area between the WVZ and the EVZ is called the Hreppar block. The current NVZ is a reactivation at ∼7 Ma of an older spreading zone in a similar location (Saemundsson *et al.* 1980; Hardarson & Fitton 1993), and the EVZ started to form at ∼2 Ma. The distance between the 13 Ma isochrones in the extreme east and west of Iceland is much greater than corresponds to the full spreading rate of 1.8 cm yr^{-1}, and indicates that ∼40 per cent of the width of the crust in Iceland must comprise rocks older than 13 Ma, buried beneath younger volcanics (Foulger 2002). Volcanically active zones within which crustal extension is insignificant are classified as flank zones (Saemundsson 1979) and include the Snaefellsnes peninsula and the southern part of the EVZ. Outside the active volcanic zones, Tertiary areas contain extinct dyke and fault swarms, central volcanoes and calderas, suggesting that a complex pattern of crustal accretion has always been the norm (Jóhannesson & Saemundsson 1998).

OCEANIC CRUST

Seismological models

Oceanic crust is typically ∼7 km thick but may be up to ∼35 km thick beneath oceanic LIPs. Models comprising a few homogenous layers are inadequate to explain seismic observations other than simple first-arrival times. Velocity gradients are required by amplitude observations (e.g. Helmberger & Morris 1970; Spudich & Orcutt 1980). Beneath unconsolidated surface layers, oceanic layer 2 is usually ∼3 km thick, characterized by steep velocity gradients of up to 2 s^{-1} and variable laterally. The transition to oceanic layer 3 is typically marked by a reduction in velocity gradient by an order of magnitude. Layer 3 may contain substantial LVZs.

A two-gradient velocity structure characterizes oceanic crust of all thicknesses (Mutter & Mutter 1993). The change in velocity gradient between oceanic layers 2 and 3 typically occurs at a velocity of $V_p \sim 6.5$ km s^{-1}, and layer 3 velocities generally range up to $V_p \sim 7.3$ km s^{-1}. Velocities in the range $V_p = 7.3$–7.8 km s^{-1} are considered to comprise a 'crust–mantle transition layer', and velocities of $V_p > 7.8$ km s^{-1} are characteristic of mantle rocks. Variations in crustal thickness correlate with variations in the thickness of layer 3. The nature of the crust–mantle transition is variable. It may comprise a sharp discontinuity in V_p of ∼1 km s^{-1}, a LVZ in the lower part of layer 3, a high-velocity crustal basal layer, or a gradational transition zone with increased velocity gradient and $V_p = 7.2$–7.8 km s^{-1}. It may have a laminated internal structure. The level of structural detail resolvable is limited by the seismic wavelengths used, and early models were probably oversimplified. Seismic velocities in oceanic crust increase with age, especially in the upper part of layer 2, because of compaction and infilling of pores with minerals.

Ophiolites

The correspondence of seismic velocity and petrology has been studied in ophiolites (Fig. 2). Seismic velocities are dependent on pressure, pore-fluid pressure, porosity, the presence of voids, the void aspect ratio, petrology and metamorphic facies, and thus seismic boundaries do not necessarily coincide with petrological boundaries (e.g. Salisbury & Christensen 1978). In general, however, laboratory measurements of ophiolitic rocks suggest that layer 2 corresponds

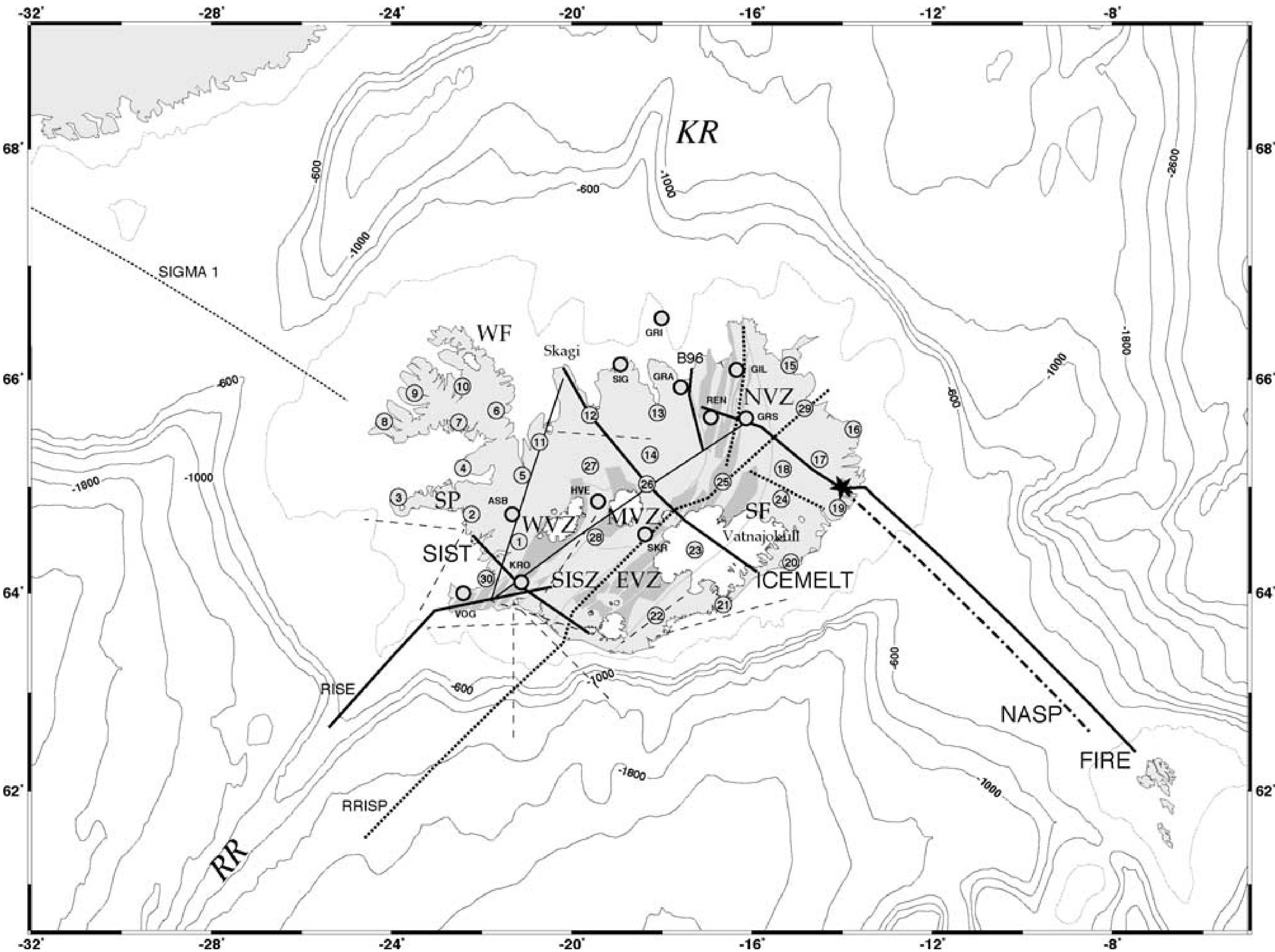

Figure 1. Map of the Icelandic transverse ridge showing bathymetric contours and tectonic features in Iceland. The neovolcanic zones are outlined. Spreading segments (volcanic systems) are shown in grey and glaciers in white. RR, Reykjanes ridge; KR, Kolbeinsey ridge; WVZ, Western Volcanic Zone; EVZ, 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 Båth (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 broad-band seismic stations in Iceland.

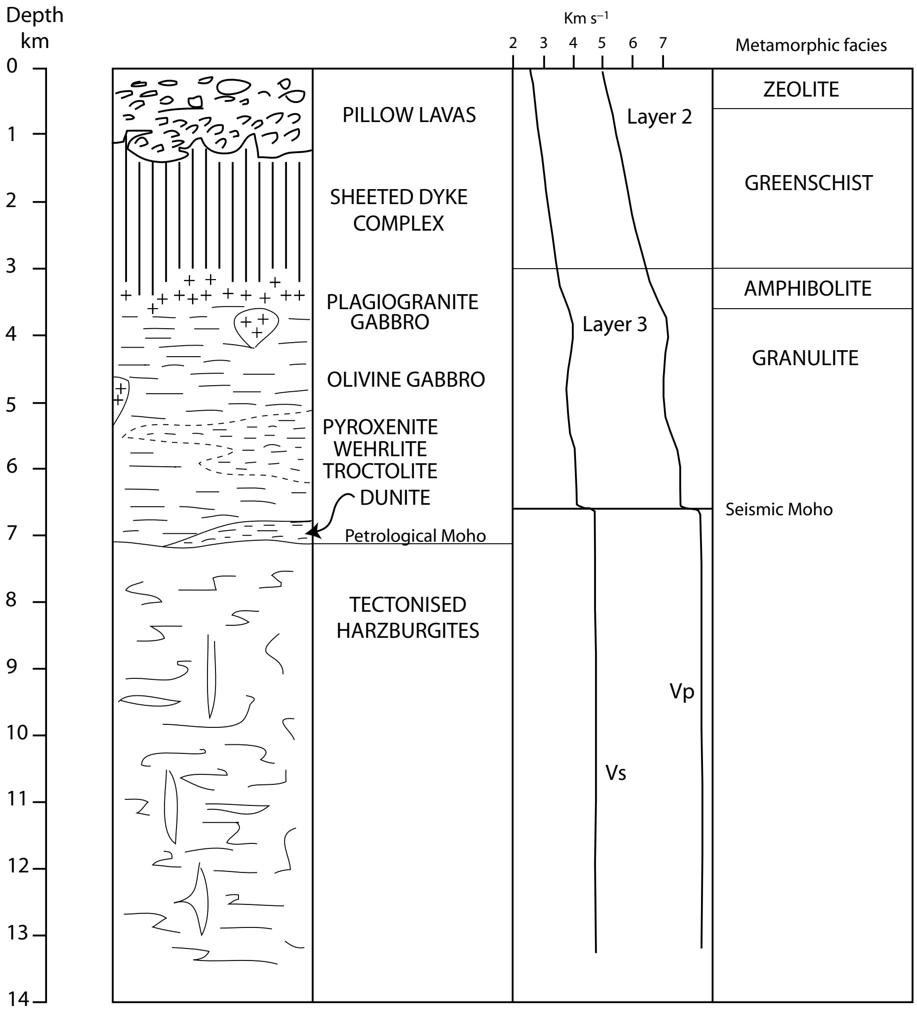

Figure 2. Schematic stratigraphic section of an ophiolite, along with petrological, seismological and metamorphic subdivisions (adapted from Christensen & Smewing 1981).

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 plagiogranite, 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 ∼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 ∼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 ∼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.

	Project name	Seismic definition of the upper crust	Seismic definition of the mid-crust	Seismic definition of the lower crust	V_p/V_s	Moho velocity discontinuity	Mantle velocity	thickness of the upper crust	crust–mantle boundary depth
(Flovenz 1980)		Grad $\gtrsim 0.57$	Not defined	Zero or low gradient, $V_p \gtrsim 6.5$		n/a	n/a	5–6	n/a
(Angenheister *et al.* 1980, Gebrande *et al.* 1980)	RRISP	$V_p \lesssim 6.5$	”	$V_p = 6.5$–6.9		n/a	$V_p \gtrsim 6.9$	3–8	10–15
(Bjarnason *et al.* 1993)	SIST	$V_p < 5.0$	$V_p = 5.0$–6.5	Low gradient material deeper than $V_p \sim 6.5$ level		Jump of $V_p = 7.2/7.25$ to 7.5, rising to $V_p = 7.7$ just below discontinuity. Mantle-diving rays	$V_p \geq 7.5$	0.7–3	20–24 beneath the WVZ and SISZ
(Staples *et al.* 1997)	FIRE	Grad $\gtrsim 0.57$	Not defined	Zero or low gradient, $V_p \gtrsim 6.5$	1.77	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	$V_p = 7.9$ under NVZ, $V_p = 8.0/8.3$ under E Iceland	Up to $\sim$4	19–35 km beneath the NVZ and E Iceland
(Darbyshire *et al.* 1998)	ICEMELT	Grad $\gtrsim 0.2$, $V_p \geq 6.4$	$V_p = 6.6$–6.9, variable gradients	$V_p = 6.9$–7.1, Grad < 0.03	1.76–1.78	$V_p = 7.2$–7.35 estimated from downward extrapolation	Not estimated	5–10	25 beneath Skagi, 40 beneath NW Vatnajokull
(Menke *et al.* 1998)	B96	Not defined	Not defined	Not defined	1.76	Not determined	$V_p = 7.7$ (eq from N) and 8.0 (eq from S)	Not given	25–31 beneath W flank of NVZ
(Smallwood *et al.* 1999)	FIRE	High grad	Not defined	Low grad, $V_p \leq 7.1$–7.3		Jump in V_p of $\sim$ 7.3–7.7	$V_p = 7.7$–7.9	$\lesssim$10	$\sim$25–30 from Iceland to the Faroe Islands
(Du & Foulger 1999, 2001; Du *et al.* 2002)	The Iceland Hotspot Project	$V_s \leq 3.7$ ($V_p \leq 6.5$)	Not defined	$V_s \leq 4.1$ ($V_p \leq 7.2$)		Variable	$V_s > 4.1$ ($\sim$equivalent to $V_p > 7.2$)	5–11	24–38
(Weir *et al.* 2001)	RISE	High grad	Not defined	$V_p \geq 6.8$, Grad ~ 0.05	1.78	$V_p = 7.2$ at base from downward extrapolation, velocity jump not estimated	Not estimated	5–10	11 at $\sim$150 km from Iceland, 21 under Reykjanes peninsula

(e.g. Fowler 1976, 1978; Bunch & Kennett 1980). Dispersion of surface waves at periods of 6–40 s suggests velocity reversals in the mantle at depths of >20 km (Jacoby & Girardin 1980; Keen *et al.* 1980). Away from the ridge crest, velocities increase, probably due to compaction, cooling and deposition of minerals in voids in the shallowest layers. Layer 3 thickens, and 'normal' uppermost mantle velocities of $V_p > 8.0$ km s^{-1} are encountered beneath crust that is a few Myr old.

Along much of the ridge between the Bight fracture zone at ∼56°30′N and Iceland, the crust is 8–11 km thick (see the references cited above). This increases relatively abruptly to ∼15 km immediately offshore of the Reykjanes peninsula in SW Iceland (Weir *et al.* 2001). Along the Reykjanes ridge, within ∼250 km of Iceland, the crust appears to thin laterally away from the ridge crest, whereas at greater distances from Iceland the crust thickens away from the ridge. Off-ridge uppermost mantle velocities are also reported to decrease on approaching Iceland, and are as low as $V_p =$ 7.7–7.9 km s^{-1} within 100 km of Iceland (e.g. Gebrande *et al.* 1980). Mantle velocities of $V_p \sim 8.2$ km s^{-1}, such as are observed at ∼7 km depth at large distances from Iceland, are not encountered shallower than 16–18 km within 100 km of Iceland.

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) (Båth 1960; Tryggvason 1962; Palmason 1971; Zverev *et al.* 1976; Bott & Gunnarsson 1980). In a pioneering experiment in 1959, Båth (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 (Båth 1960) or the upper mantle (Tryggvason 1962; Palmason 1971). Båth (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, but featured a $V_p = 6.8$ 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.

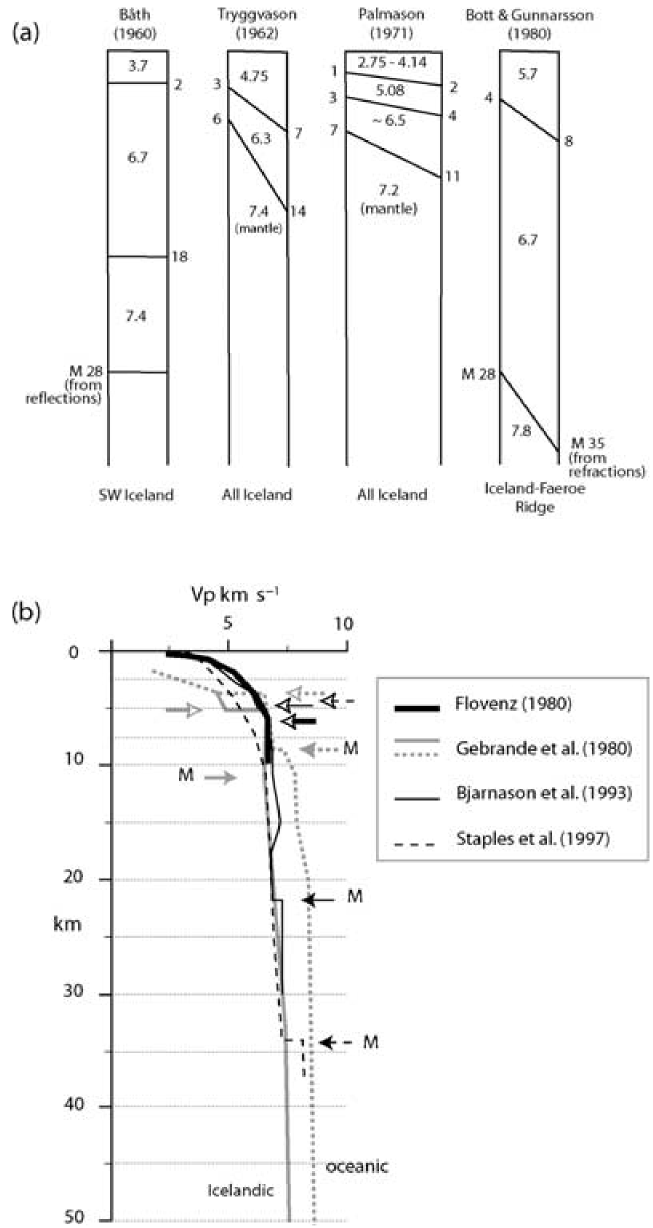

Figure 3. (a) Models of seismic crustal structure from early explosion and earthquake surface wave experiments (Båth 1960; Tryggvason 1962; Palmason 1971; Bott & Gunnarsson 1980). Sloping lines indicate thickness ranges of layers, numbers inside columns indicate average V_p in km s^{-1}, numbers outside columns indicate depths in km. reflns, reflections; refrcns, refracted arrivals; M, Moho. (b) Velocity–depth profiles for the crust in Iceland and in 10 Myr old oceanic crust south of Iceland from recent explosion seismology experiments. Open-headed arrows, estimates of the base of the upper crust; solid-headed arrows, estimates of the base of the lower crust; M, Moho identifications proposed (Flovenz 1980; Gebrande *et al.* 1980; Bjarnason *et al.* 1993; Staples *et al.* 1997).

Apart from the observation of velocities in the range $V_p = 7.2$–$7.4\ \mathrm{km\ s^{-1}}$ 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 $\lesssim 6.5\ \mathrm{km\ s^{-1}}$), 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 $\sim 0.57\ \mathrm{s^{-1}}$ beneath the $V_p \sim 3.5\ \mathrm{km\ s^{-1}}$ horizon. The lower crust beneath the $V_p \sim 6.5\ \mathrm{km\ s^{-1}}$ velocity horizon, typically at 5–6 km depth, has a much smaller velocity gradient.

Lateral heterogeneity in the upper crust is extreme. Velocities of $V_p = 6.5\ \mathrm{km\ s^{-1}}$ may occur close to the surface beneath volcanoes (e.g. Einarsson 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 $V_p \sim 6.5\ \mathrm{km\ s^{-1}}$, probably associated with gabbroic intrusions, occur within ~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 $V_p = 7.2$–$7.4\ \mathrm{km\ s^{-1}}$ 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 (Angenheister *et al.* 1980; Gebrande *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 $V_p \sim 6.5\ \mathrm{km\ s^{-1}}$, have much lower gradients. Offshore the crust was found to be ~10 km thick, beneath which V_p rises steeply to $7.8\ \mathrm{km\ s^{-1}}$, 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 $V_p = 7.6\ \mathrm{km\ s^{-1}}$ in the upper 50 km.

Like Palmason (1971), Gebrande *et al.* (1980) attributed material with $V_p = 7.0$–$7.6\ \mathrm{km\ s^{-1}}$ beneath Iceland to anomalously hot upper mantle. Both Palmason (1971) and Gebrande *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. Gebrande *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 $V_p = 7.2$–$7.6\ \mathrm{km\ s^{-1}}$ 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; Oxburgh & 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 $V_p = 7.0$–$7.4\ \mathrm{km\ s^{-1}}$ were encountered. At such temperatures, the seismic velocity in gabbros and metagabbros is much less than $V_p = 7.0$–$7.4\ \mathrm{km\ s^{-1}}$, 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 ~50–100 $^{\circ}\mathrm{C\ km^{-1}}$ away from geothermal areas, which predict supra-solidus temperatures at 10–20 km depth if extrapolated linearly downward (Palmason 1974; Flovenz & Saemundsson 1993).

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

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

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

The thin–hot crust model was challenged by Bjarnason *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 ~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 $V_p = 7.7\ \mathrm{km\ s^{-1}}$ 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 $7.5\ \mathrm{km\ s^{-1}}$, rapidly rising to $7.7\ \mathrm{km\ s^{-1}}$ 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 RRISP data beneath the NVZ and eastern Iceland, and attributed by Gebrande *et al.* (1980) to thin, high-velocity lenses, could also be reflections from a horizon ∼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 suprasolidus temperatures in the $V_p = 7.0$–7.5 km s^{-1} material suggested that those 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 (∼400–650°C for the lithology and strain rates expected under Iceland) at the seismogenic base, which varies in the range ∼6–14 km (e.g. Stefánsson *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 ∼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 $Q_s \sim 20$ are expected for near-solidus gabbro and $Q_p \sim 40$ 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 ∼20 km depth. Menke *et al.* (1996) suggest that the low values of V_s and V_p/V_s ratios of 1.96 deduced from the RRISP data (Gebrande *et al.* 1980) were erroneous and result from uncorrected source static anomalies and misidentification of *SmS* as *S*. Reappraisal of those data suggests more normal values of 1.76 for 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 ∼60 three-component stations and six large explosions. In 1996 the ∼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 ∼220 km long RISE profile, which involved ∼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, $V_p \lesssim 6.5$ km s^{-1} material having large vertical velocity gradients (typically ∼0.25 s^{-1}) and great lateral heterogeneity, and deeper material having velocity gradients an order of magnitude lower (typically ∼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 unreversed 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 ∼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 $V_p = 6.5$–7.05, assuming $V_p/V_s = 1.76$) 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 Gebrande *et al.* (1980) from RRISP 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 400 °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 ∼40 km beneath northwest Vatnajokull to ∼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 Vatnajokull, corresponding to lateral density anomalies of ∼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.

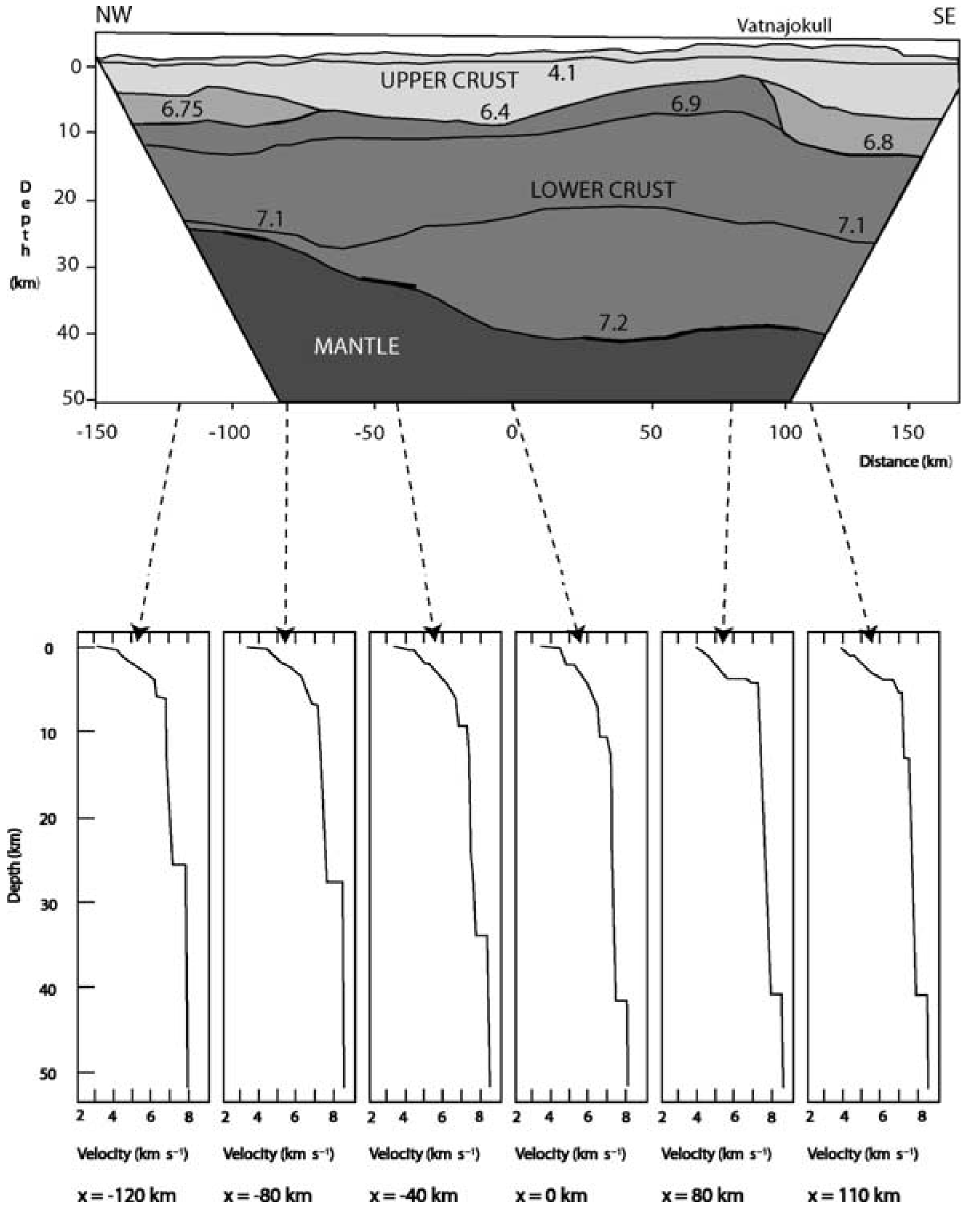

Figure 4. Top, crustal model derived from the ICEMELT explosion seismology profile (Fig. 1). The thick lines indicate locations of reflective horizons interpreted as the Moho. Bottom, 1-D velocity profiles at various offsets along the profile relative to an explosion in central Iceland (from Darbyshire *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 $\sim V_p = 7.1$–7.3 km s^{-1} and the depth to the base of the crust was fairly uniform at $\sim 30 \pm 3$ 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 to $V_p = 7.7$–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

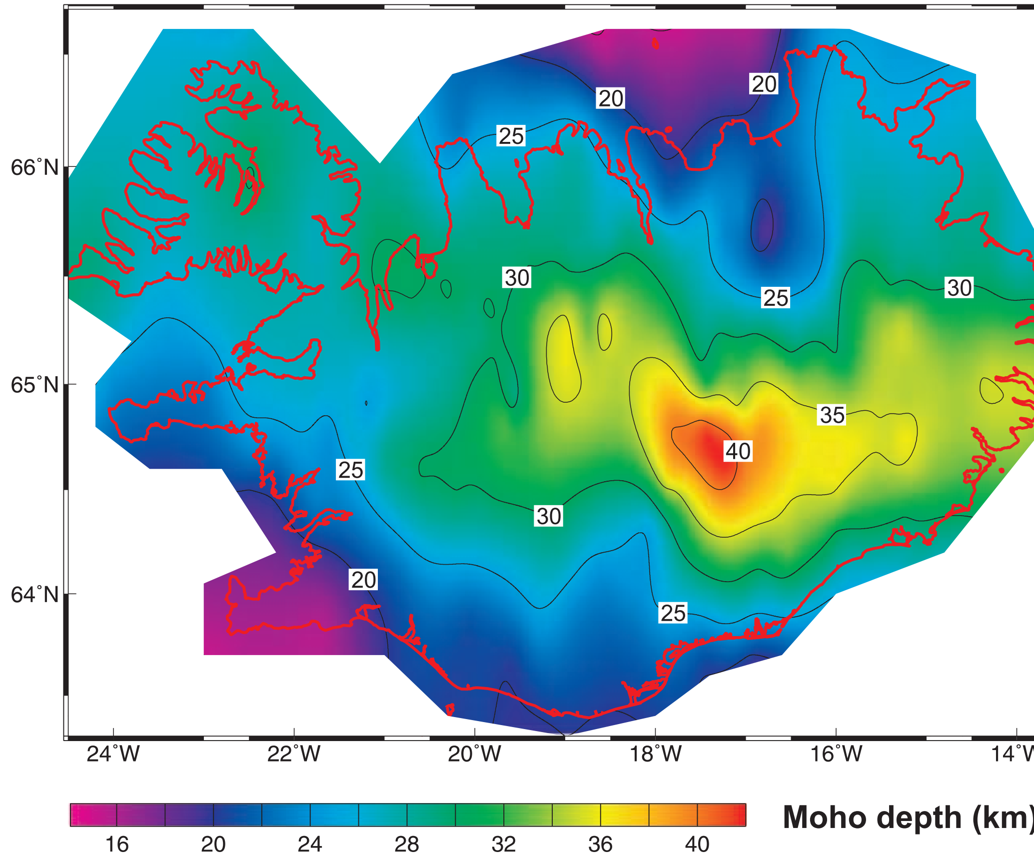

Figure 5. Contour map of total crustal thickness determined using a combination of seismic profiles, receiver functions and gravity profiles (from Darbyshire *et al.* 2000b).

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 ~2–4 km. Structures on the scale of ~2 km at shallow depth and ~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

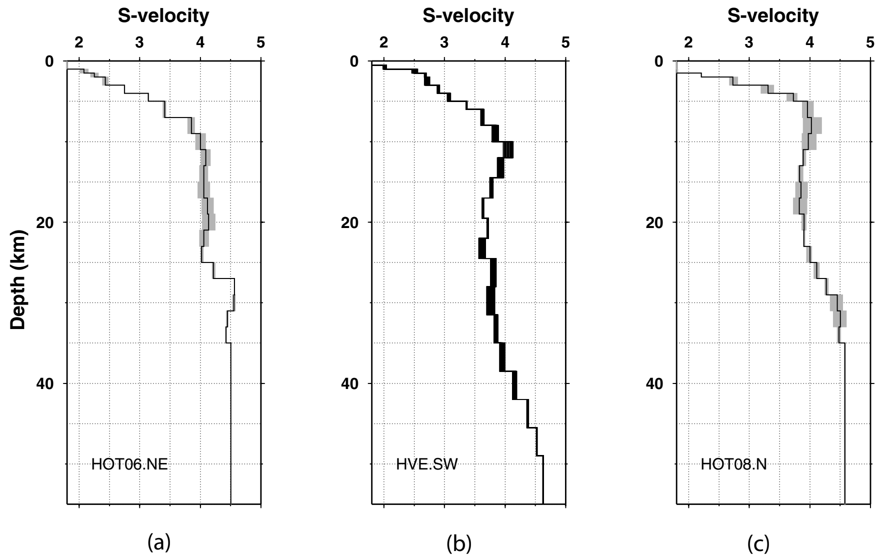

Figure 6. Examples of end-member structures determined from receiver functions. (a) Station 6, from earthquakes approaching from the NE backazimuth (from Du & Foulger 1999), (b) station HVE, from earthquakes approaching from the SW backazimuth (from Du & Foulger 2001), (c) station 8, from earthquakes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

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 ~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 ~7 ± 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 ~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 ~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 ~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 than ~11 s, corresponding to depths $\gtrsim$25 km, where they deem their data unreliable because of noise. These methodological differences result in radical differences between some of their models and those of Du & Foulger (2001), who studied some common stations. These differences illustrate the sensitivity of receiver function results to data analysis methodology and the problem of combining results obtained using different approaches. In particular, the approach of Darbyshire *et al.* (2000a) precludes detection of major LVZs such as that beneath the MVZ. The first-order features of the two sets of results agree, however. Darbyshire *et al.* (2000a) also find a distinct division of the crust into two main parts on the basis of velocity gradients. LVZs of varying thicknesses and amplitudes were detected in the lower crust, and most of the structures derived from their inversions exhibit gradational crust–mantle boundaries up to several kilometres thick.

COMPARISON OF RESULTS FROM RECEIVER FUNCTIONS AND EXPLOSION SEISMOLOGY

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 ~2.5 km in depth (the thickness of the thickest layer modelled) and ±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 ±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

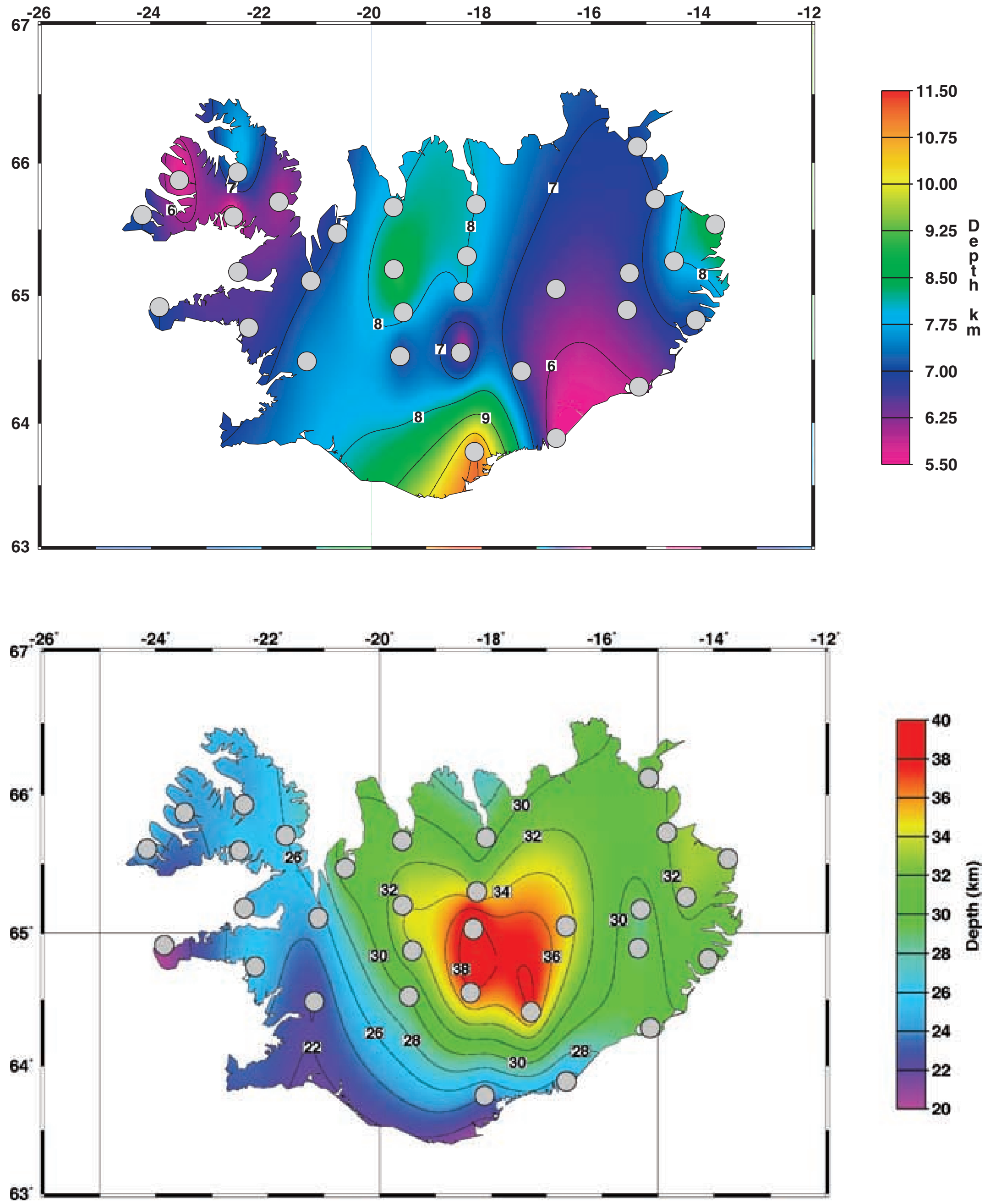

Figure 7. Contour maps showing the depth to the bases of (a) the upper crust (defined as the depth to the $V_s = 3.7$ km s^{-1} horizon) and (b) the lower crust (defined as the depth to the $V_s = 4.1$ km s^{-1} horizon) (from the receiver functions of Du & Foulger 1999, 2001; Du *et al.* 2002).

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 $\pm$ 5 km thick. There is no evidence that upper Icelandic crust compacts with age as normal oceanic crust does, since no thinning of the upper crust from younger to older parts 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 thickest 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 $\gtrsim$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_s than in V_p, and thus represent zones of high V_p/V_s. A comprehensive map of V_p/V_s across Iceland could significantly improve our knowledge of lower crustal LVZs. However, although V_p 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_p 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_p = 0.45–0.95 km s^{-1} (corresponding to V_s = 0.25–0.55 km s^{-1}, assuming $V_p/V_s = 1.76$). Such a discontinuity is not detected by the receiver functions, however, which indicate that a gradational crust–mantle boundary several kilometres thick is most 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 $\Delta V_p \gtrsim$ 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

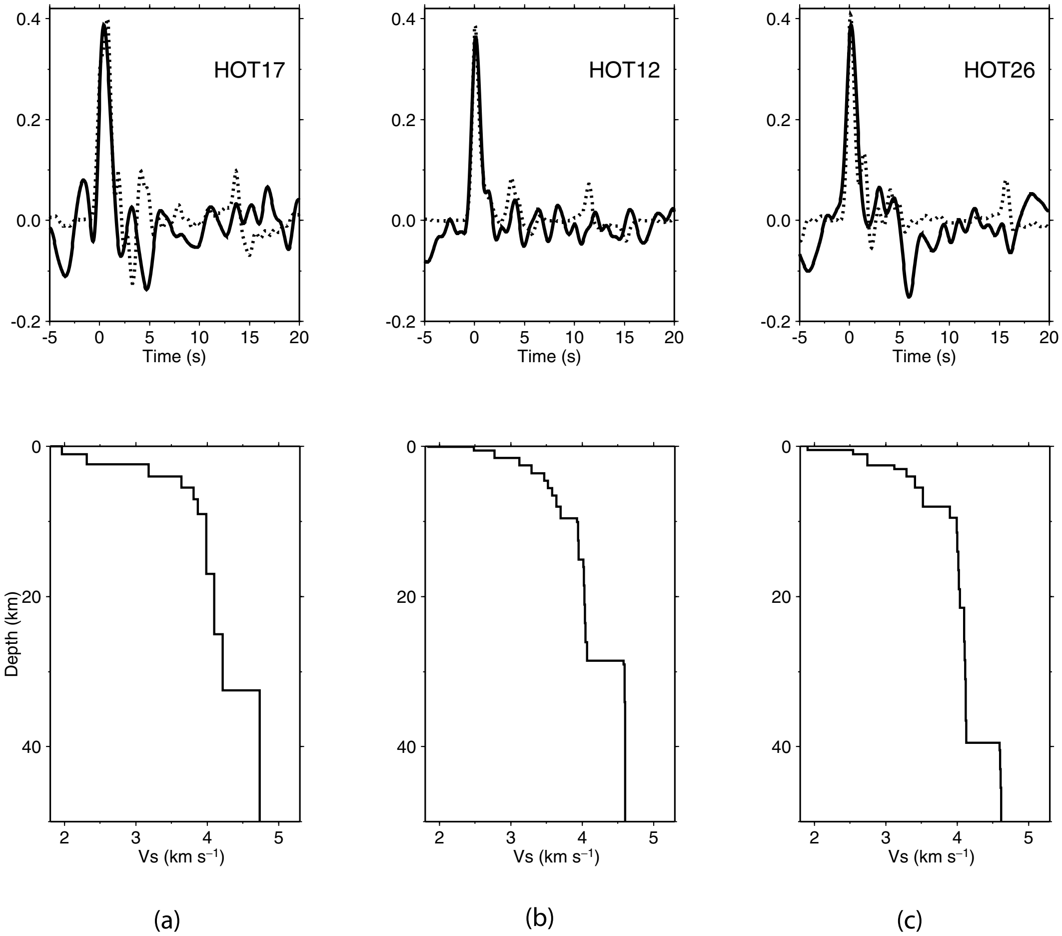

Figure 8. Upper panels, observed (solid lines) and synthetic (dotted lines) receiver functions. Lower panels, structures derived from explosion seismology profiles used to generate the synthetic receiver functions shown in the upper panels. (a) Structure from the FIRE profile (Staples *et al.* 1997) near station 17; (b) structure from the ICEMELT profile (Darbyshire *et al.* 1998) near station 12; and (c) structure from the ICEMELT profile (Darbyshire *et al.* 1998) near station 26. The prominent peaks in the synthetic receiver functions at ∼3 s are the *Ps* arrivals, and the peaks at ∼13, 11 and 16 s are the *PpPms* arrivals from the large, sharp discontinuities at 30–40 km depths. Such peaks are not observed in the data. The large trough at ∼6 s in the observed receiver function at station 26 requires a substantial low-velocity zone in the lower crust (Du & Foulger 2001).

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 traveltime 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).

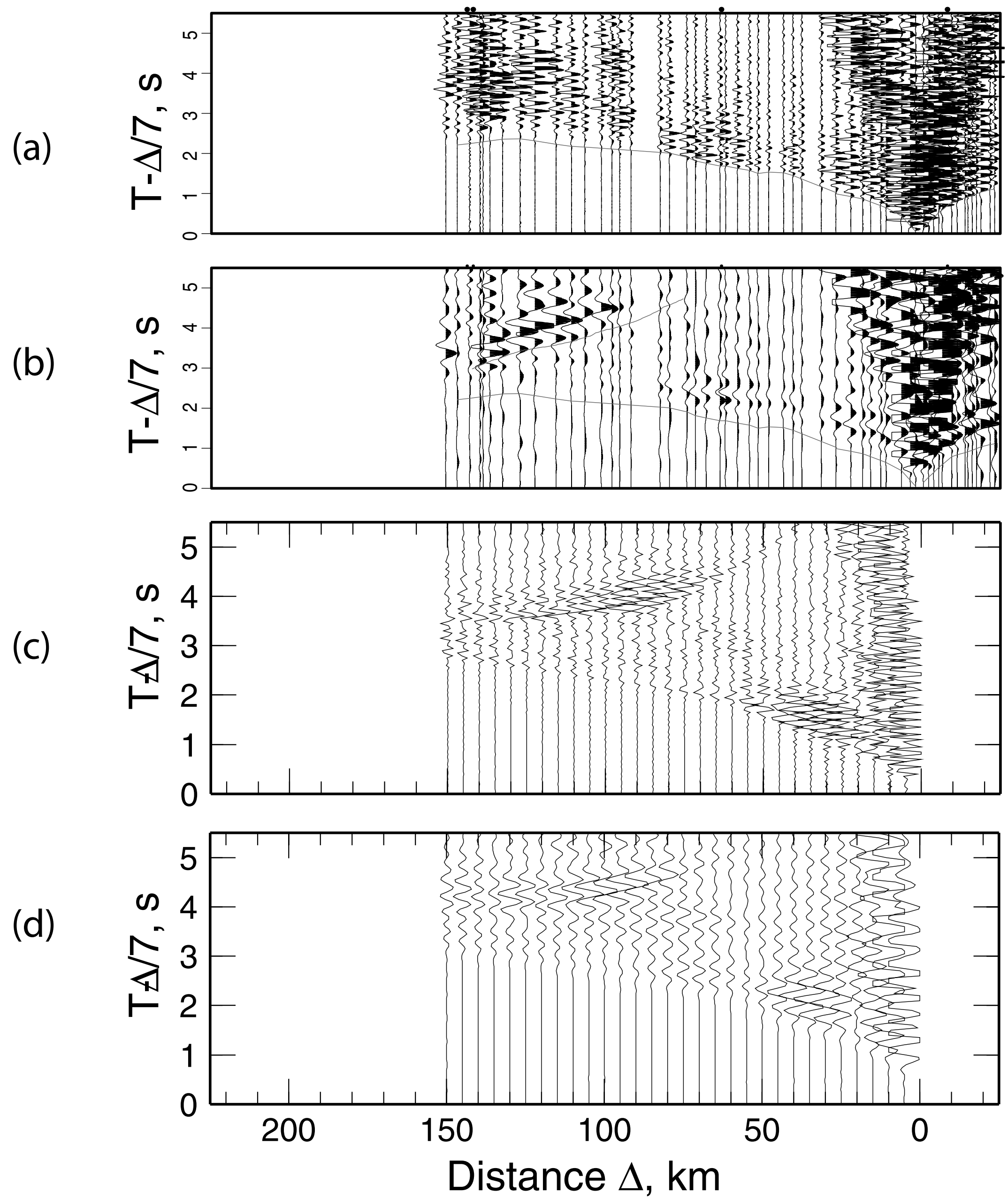

Figure 9. Effect of sharp first-order Moho discontinuities at different depths on seismic waves. All plots have the same scale and reducing velocity. Seismograms are plotted with gain proportional to the epicentral distance. (a) Observed vertical-component seismograms for the Reydarfjodur shot on the 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

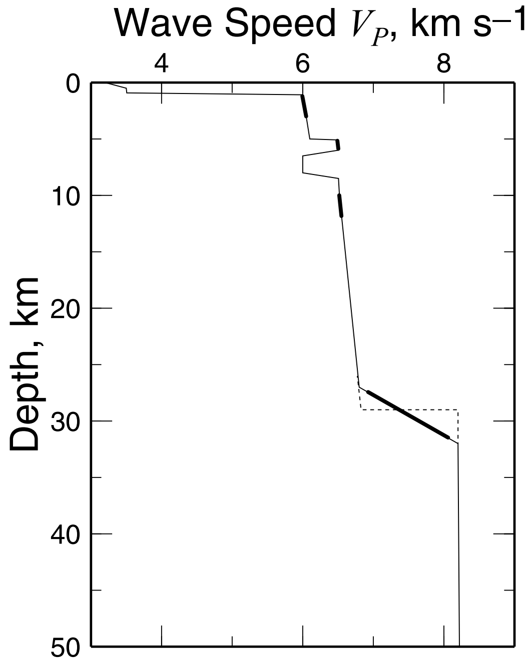

Figure 10. Compressional-wave speed V_P in a simple generic model of Icelandic-type crust, showing features for which seismic explosion (refraction and wide-angle reflection) observations typically provide control. The dashed line shows a variant with a sharp, first-order Moho discontinuity, the predictions of which are compared with observations in Fig. 9. Fig. 11 shows similar predictions for the structure indicated by the continuous line. Bold lines indicate depths from which turning rays are observable as first arrivals or large later arrivals, and for which the wave speed, and to some extent its gradient, are well constrained. The LVZ near the top of the lower crust is required by rapid amplitude decreases and traveltime offsets seen on the Reydarfjordur and other published profiles from Iceland, but its shape is not unique. The LVZ shown is at the extreme end of what is consistent with the data.

are characteristic of a shadow zone caused by an LVZ such as that in the depth interval from approximately 6 to 10 km in Fig. 10. Similar evidence of shadow zones is observed on three other shots (Staples *et al.* 1997, Figs 7, 8 and 10).

As is always the case, the velocity variation within the LVZ cannot be uniquely determined from surface-focus traveltimes, so the thickness of the zone and the large velocity contrast shown in Fig. 10 are arbitrary. Furthermore, variations in the position of the shadow zone from shot to shot indicate that the upper crust is laterally inhomogeneous. Rays bottoming between approximately 15 km and the base of the crust are not observed. They are either small later arrivals or emerge beyond the furthest seismic station. The explosion data are therefore compatible with the presence of LVZs and other features in the lower crust, but do not require them.

Waves bottoming below the crust–mantle boundary are not observed, and must be small or absent. Both for geometrical rays (Julian & Anderson 1968) and for diffracted waves (Hill 1971), amplitudes are extremely sensitive to the velocity gradient at the bottom of a ray. The absence of rays bottoming below the crust–mantle boundary is thus most simply explained by a low or negative velocity gradient, although anelastic attenuation may also play a role.

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 *PmP* 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 *PmP*-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 updoming. 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 either from 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

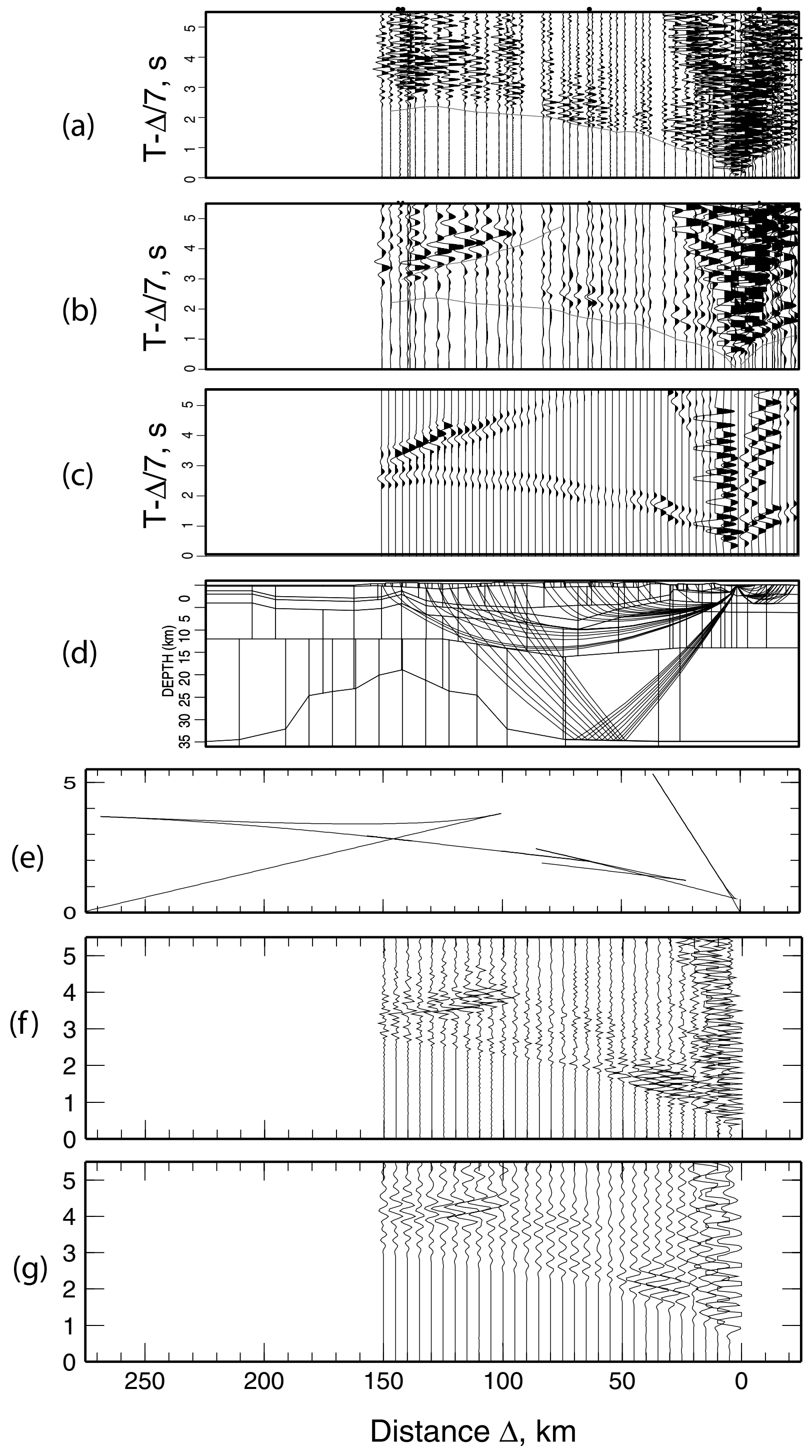

Figure 11. Comparison of observed and theoretical seismic data for a model with a mid-crustal LVZ and a transitional Moho discontinuity. (a) and (b) Observed seismograms, as in Figs 9(a) and (b), (c) ray-theoretical seismograms for the 2-D model of Staples *et al.* (1997), (d) paths of seismic rays considered 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).

(e.g. beneath station 25; Fig. 12, Du & Foulger 2001). LVZs occur occasionally (e.g. beneath station SKR; Fig. 10, Du & Foulger 2001).

(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 $V_p = 6.5$ 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

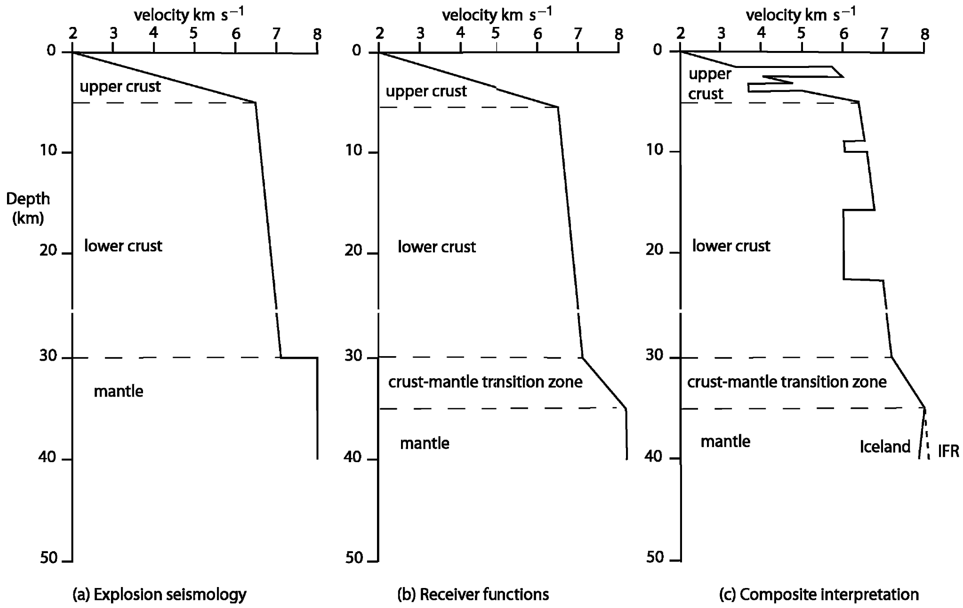

Figure 12. Simplified, schematic models of Icelandic-type crust from (a) published explosion-seismology models, (b) receiver functions. These figures show the first-order features that are reported to characterize the crust in general throughout Iceland. (c) A suggested, schematic, composite model that satisfies both kinds of observations. This model is intended to illustrate the range of features that are observed in different structures in various places in Iceland, and is not intended to correspond to any one particular location. IFR: Iceland–Faeroe ridge.

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 mélange 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 ~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 basaltic 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 (Christensen & 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 ~75°C hotter beneath central Iceland (Menke & Levin 1994), which would correspond to a reduction in V_s of ~1.5 per cent (Humphreys & Dueker 1994). This is less than the maximum of ~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 ~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. Farnetani *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 $\gtrsim$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 $\leq$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 <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 $\gtrsim$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 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 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 Vatnajokull 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 lowermost ∼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 Vatnajokull. 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 (Båth 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 >1100 °C (Tryggvason 1962; Palmason 1971; Angenheister *et al.* 1980; Gebrande *et al.* 1980) and back to a thick–cold crust model, with material of $V_p \sim 7.2$ km s^{-1} attributed to subsolidus crust with temperatures of $\lesssim 900$ °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 further detections of the deep reflector first recognized by Båth (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 subsolidus '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 89 ± 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 (Breddam *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 (Recq *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). This

Table 2. Characteristics of high-velocity basal crustal layers observed beneath some LIPs.

Location	Total crustal thickness (km)	V_p of basal crustal layer (km s^{-1})	Percentage of total crustal thickness	Source
Ontong–Java Plateau	15–40	7.6	30	Furumoto *et al.* 1970, 1976; Richardson *et al.* 2000
Kerguelan Islands	19	7.2–7.5	16	Recq *et al.* 1990
Madagasgar Ridge	25	7.6	30	Sinha & Louden 1981
Hawaiian Ridge	20	7.4–7.8	25	Watts *et al.* 1985

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 upper-mantle 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 $V_s = -5$ per cent throughout the depth range ~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 ~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 & Widiyantoro 1999), which ceased to be active ~80 and ~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–cold 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

Allen, R.M. *et al.,* 2002. Plume driven plumbing and crustal formation in Iceland, *J. geophys. Res.,* **107,** 10.1029/2001JB000584.

Anderson, D., 1994. Lithosphere and flood basalts, *Nature,* **367,** 226.

Angenheister, G. *et al.*, 1980. Reykjanes ridge Iceland seismic experiment (RRISP 77), *J. Geophys.,* **47,** 228–238.

Arnott, S.K. & Foulger, G.R., 1994. The Krafla spreading segment, Iceland: 1. Three-dimensional crustal structure and the spatial and temporal distribution of local earthquakes, *J. geophys. Res.,* **99,** 23 801–23 825.

Båth, M., 1960. Crustal structure of Iceland, *J. geophys. Res.,* **65,** 1793–1807.

Beblo, M. & Bjornsson, A., 1978. Magnetotelluric investigation of the lower crust and upper mantle beneath Iceland, *J. Geophys.,* **45,** 1–16.

Beblo, M. & Bjornsson, A., 1980. A model of electrical resistivity beneath northeast Iceland, correlation with temperature, *J. Geophys.,* **47,** 184–190.

Beblo, M., Bjornsson, A., Arnason, K., Stein, B. & Wolfgram, P., 1983. Electrical conductivity beneath Iceland—constraints imposed by magnetotelluric results on temperature, partial melt, crust and mantle structure, *J. Geophys.,* **53,** 16–23.

Bjarnason, I.T., Menke, W., Flovenz, O.G. & Caress, D., 1993. Tomographic image of the mid-Atlantic plate boundary in south-western Iceland, *J. geophys. Res.,* **98,** 6607–6622.

Björnsson, A., 1985. Dynamics of crustal rifting in NE Iceland, *J. geophys. Res.,* **90,** 10 151–10 162.

Bott, M.H.P., 1965. The upper mantle beneath Iceland, *Geophys. J. R. astr. Soc.,* **9,** 275–277.

Bott, M.H.P., 1974. Deep structure, evolution and origin of the Icelandic transverse ridge, in *Geodynamics of Iceland and the North Atlantic Area,* pp. 33–48, eds Kristjansson, L., Reidel, Dordrecht.

Bott, M.H.P., 1985. Plate tectonic evolution of the Icelandic transverse ridge and adjacent regions, *J. geophys. Res.,* **90,** 9953–9960.

Bott, M.H.P. & Gunnarsson, K., 1980. Crustal structure of the Iceland–Faeroe ridge, *J. geophys. Res.,* **47,** 221–227.

Breddam, K., Kurz, M.D. & Storey, M., 2000. Mapping out the conduit of the Iceland mantle plume with helium isotopes, *Earth planet. Sci. Lett.,* **176,** 45–55.

Bunch, A.W.H., 1980. Crustal development of the Reykjanes ridge from seismic refraction, *J. Geophys.,* **47,** 261–264.

Bunch, A.W.H. & Kennett, B.L.N., 1980. The crustal structure of the Reykjanes ridge at 59°30′N, *Geophys. J. R. astr. Soc.,* **61,** 141–166.

Cannat, M., 1996. How thick is the magmatic crust at slow spreading oceanic ridges?, *J. geophys. Res.,* **101,** 2847–2857.

Cerveny, V., 2001. *Seismic Ray Theory,* p. 713, Cambridge University Press, Cambridge.

Christensen, N.I. & Smewing, J.D., 1981. Geology and seismic structure of the northern section of the Oman ophiolite, *J. geophys. Res.,* **86,** 2545–2555.

Christensen, N.I. & Wilkins, R.H., 1982. Seismic properties, density and composition of the Icelandic crust near Reydarfjordur, *J. geophys. Res.,* **87,** 6389–6395.

Cox, K.G., 1980. A model for flood basalt volcanism, *J. Petrol.,* **21,** 629–650.

Darbyshire, F.A., Bjarnason, I.T., White, R.S. & Flovenz, O.G., 1998. Crustal structure above the Iceland mantle plume imaged by the ICEMELT refraction profile, *Geophys. J. Int.,* **135,** 1131–1149.

Darbyshire, F.A., Priestley, K.F., White, R.S., Stefansson, R., Gudmundsson, G.B. & Jakobsdottir, S., 2000a. Crustal structure of central and northern Iceland from analysis of teleseismic receiver functions, *Geophys. J. Int.,* **143,** 163–184.

Darbyshire, F.A., White, R.S. & Priestley, K.F., 2000b. Structure of the crust and uppermost mantle of Iceland from a combined seismic and gravity study, *Earth planet. Sci. Lett.,* **181,** 409–428.

Dilek, Y., Moores, E.M. & Furnes, H., 1998. Structure of modern oceanic crust and ophiolites and implications for faulting and magmatism at oceanic spreading centers, in *Faulting and Magmatism at Mid-Ocean Ridges,* pp. 219–265, eds Buck, W.R., Delaney, P.T., Karson, J.A. & Lagabrielle, Y., American Geophysical Union, Washington, DC.

Du, Z. & Foulger, G.R., 1999. The crustal structure beneath the Northwest Fjords, Iceland, from receiver functions and surface waves, *Geophys. J. Int.,* **139,** 419–432.

Du, Z. *et al.,* 2002. Crustal structure beneath western and eastern Iceland from surface waves and receiver functions, *Geophys. J. Int.,* **149,** 349–363.

Du, Z.J. & Foulger, G.R., 2001. Variation in the crustal structure across central Iceland, *Geophys. J. Int.,* **145,** 246–264.

Einarsson, P., 1978. *S*-wave shadows in the Krafla caldera in NE-Iceland, evidence for a magma chamber in the crust, *Bull. Volcanol.,* **41,** 1–9.

Ewing, J. & Ewing, M., 1959. Seismic refraction measurements in the Atlantic ocean basins, in the Mediterranean Sea, on the mid-Atlantic ridge, and in the Norwegian Sea, *Bull. Geol. Soc. Am.,* **70,** 291–318.

Eysteinsson, H. & Gunnarsson, K., 1995. Maps of gravity, bathymetry and magnetics for Iceland and surroundings, National Energy Authority, Reykjavik.

Eysteinsson, H. & Hermance, J., 1985. Magnetotelluric measurements across the eastern neovolcanic zone in south Iceland, *J. geophys. Res.,* **90,** 10 093–10 103.

Farnetani, C.G., Richards, M.A. & Ghiorso, M.S., 1996. Petrological models of magma evolution and deep crustal structure beneath hotspots and flood basalt provinces, *Earth planet. Sci. Lett.,* **143,** 81–94.

Flovenz, O.G., 1980. Seismic structure of the Icelandic crust above layer three and the relation between body wave velocity and the alteration of the basaltic crust, *J. Geophys.,* **47,** 211–220.

Flovenz, O.G. & Gunnarsson, K., 1991. Seismic crustal structure in Iceland and surrounding area, *Tectonophysics,* **189,** 1–17.

Flovenz, O.G. & Saemundsson, K., 1993. Heat flow and geothermal processes in Iceland, *Tectonophysics,* **225,** 123–138.

Foulger, G.R., 1988. Hengill triple junction, SW Iceland; 2. Anomalous earthquake focal mechanisms and implications for process within the geothermal reservoir and at accretionary plate boundaries, *J. geophys. Res.,* **93,** 13 507–13 523.

Foulger, G.R., 1995. The Hengill geothermal area, Iceland: variation of temperature gradients deduced from the maximum depth of seismogenesis, *J. Volc. Geotherm. Res.,* **65,** 119–133.

Foulger, G.R., 2002. Plumes, or plate tectonic processes?, *Astron. Geophys.,* **43,** 6.19–6.23.

Foulger, G.R., 2003. On the apparent eastward migration of the spreading ridge in Iceland, abstract in *The Hotspot Handbook, Proceedings of Penrose Conference Plume IV: Beyond the Plume Hypothesis,* Hveragerdi, Iceland.

Foulger, G.R. & Pearson, D.G., 2001. Is Iceland underlain by a plume in the lower mantle? Seismology and helium isotopes, *Geophys. J. Int.,* **145,** F1–F5.

Foulger, G.R. & Toomey, D.R., 1989. Structure and evolution of the Hengill–Grensdalur volcanic complex, Iceland; Geology, geophysics, and seismic tomography, *J. geophys. Res.,* **94,** 17 511–17 522.

Foulger, G.R., Anderson, D.L. & Natland, J.H., 2002. The melt anomaly at the Iceland volcanic province can be explained by mantle fertility in the Caledonian suture, *EOS, Trans. Am. geophys. Un. Fall Meet. Suppl.*, *Abstract*, **83,** S11A–1120.

Foulger, G.R., Anderson, D.L. & Natland, J.H., 2003. An alternative model for Iceland and the North Atlantic Igneous Province, abstract in *The Hotspot Handbook, Proceedings of Penrose Conference Plume IV: Beyond the Plume Hypothesis,* Hveragerdi, Iceland.

Foulger, G.R., Long, R.E., Einarsson, P. & Björnsson, A., 1989. Implosive earthquakes at the active accretionary plate boundary in northern Iceland, *Nature,* **337,** 640–642.

Foulger, G.R., Miller, A.D., Julian, B.R. & Evans, J.R., 1995. Three-dimensional V_p and V_p/V_s structure of the Hengill triple junction and geothermal area, Iceland, and the repeatability of tomographic inversion, *Geophys. Res. Lett.,* **22,** 1309–1312.

Foulger, G.R. *et al.,* 2000. The seismic anomaly beneath Iceland extends down to the mantle transition zone and no deeper, *Geophys. J. Int.,* **142,** F1–F5.

Foulger, G.R. *et al.,* 2001. Seismic tomography shows that upwelling beneath Iceland is confined to the upper mantle, *Geophys. J. Int.,* **146,** 504–530.

Fowler, C.M.R., 1976. Crustal structure of the mid-Atlantic ridge crest at 37°N, *Geophys. J. R. astr. Soc.,* **47,** 459–491.

Fowler, C.M.R., 1978. The mid-Atlantic ridge: structure at 45°N, *Geophys. J. R. astr. Soc.,* **54,** 167–183.

Furumoto, A.S., Hussong, D.M., Campbell, J.F., Sutton, G.H., Malahoff, A., Rose, J.C. & Woollard, G.P., 1970. Crustal and upper mantle structure of the Solomon Islands as revealed by seismic-refraction survey of November–December 1966, *Pac. Sci.,* **24,** 315–332.

Furumoto, A.S., Webb, J.P., Odegard, M.E. & Hussong, D.M., 1976. Seismic studies in the Ontong–Java Plateau, *Tectonophysics,* **34,** 71–90.

Gebrande, H., Miller, H. & Einarsson, P., 1980. Seismic structure of Iceland along RRISP-Profile I, *J. Geophys.,* **47,** 239–249.

Goldflam, P., Weigel, W. & Loncarevic, B.D., 1980. Seismic structure along RRISP—profile I on the southeast flank of the Reykjanes ridge, *J. Geophys.,* **47,** 250–260.

Gudmundsson, O., 1994. Comment on 'Tomographic image of the mid-Atlantic plate boundary in southwestern Ireland' (*sic*) by I. T. Bjarnason, W. Menke, O. G. Flovenz & D. Caress, *J. geophys. Res.,* **99,** 17 909–17 914.

Gudmundsson, O., Brandsdóttir, B., Menke, W. & Sigvaldason, G.E., 1994. The crustal magma chamber of the Katla volcano in south Iceland revealed by 2-D seismic undershooting, *Geophys. J. Int.,* **119,** 277–296.

Hardarson, B.S. & Fitton, J.G., 1993. Geochemical variation in the Tertiary basalts of Iceland, *Min. Mag.,* **58A,** 372–373.

Helmberger, D.V. & Morris, G.B., 1970. A traveltime and amplitude interpretation of a marine refraction profile: transformed shear waves, *Bull. seism. Soc. Am.,* **60,** 593–600.

Hermance, J.F. & Grillot, L.R., 1974. Constraints on temperatures beneath Iceland from magnetotelluric data, *Phys. Earth planet. Inter.,* **8,** 1–12.

Herzberg, C. & O'Hara, M.J., 1998. Phase equilibrium constrains on the origin of basalts, picrites, and komatiites, *Earth Sci. Rev.,* **44,** 39–79.

Hill, D.P., 1971. Velocity gradients in the continental crust from head-wave amplitudes, in *The Structure and Properties of the Earth's Crust.,* pp. 71–75, ed. Heacock, J.G., American Geophysical Union, Washington, DC.

Holbrook, W.S. *et al.,* 2001. Mantle thermal structure and active upwelling during continental breakup in the North Atlantic, *Earth planet. Sci. Lett.,* **190,** 251–266.

Humphreys, E.D. & Dueker, K.G., 1994. Western US upper mantle structure, *J. geophys. Res.,* **99,** 9615–9634.

Huppert, H.E. & Sparks, R.S.J., 1980. The fluid dynamics of a basaltic magma chamber replenished by influx of hot, dense ultrabasic magma, *Cont. Min. Pet.,* **75,** 279–289.

Jacoby, W.R. & Girardin, N., 1980. The evolution of the lithosphere at the southeast flank of Reykjanes ridge from surface wave data, *J. Geophys.,* **47,** 271–277.

Jancin, M., Young, K.D., Voight, B., Aronson, J.L. & Saemundsson, K., 1985. Stratigraphy and K/Ar ages across the west flank of the northeast Iceland axial rift zone, in relation to the 7 Ma volcano-tectonic reorganization of Iceland, *J. geophys. Res.,* **90,** 9961–9985.

Jóhannesson, H. & Saemundsson, K., 1998. Jardfraedikort af Islandi. Hoggun (Geological Map of Iceland. Tectonics), Náttúrfraedistofunu Islands (Icelandic Institute of Natural History), Reykjavík.

Julian, B.R. & Anderson, D.L., 1968. Travel times, apparent velocities and amplitudes of body waves, *Bull. seism. Soc. Am.,* **58,** 339–366.

Keen, C.E., Fricker, A., Keen, M.J. & Blinn, L., 1980. Reykjanes ridge crest studied by surface waves with an earthquake-pair technique, *J. Geophys.,* **47,** 265–270.

Kelemen, P.B. & Aharonov, E., 1998. Periodic formation of magma fractures and generation of layered gabbros in the lower crust beneath oceanic spreading ridges, in, *Faulting and Magmatism at Mid-Ocean Ridges*, pp. 267–289, eds Buck, W.R., Delaney, P.T., Karson, J.A. & Lagabrielle, Y., American Geophysical Union, Washington, DC.

Kelemen, P.B., Hirth, G., Shimizu, N., Spiegelman, M. & Dick, H.J.B., 1997. A review of melt migration processes in the adiabatically upwelling mantle beneath oceanic spreading ridges, *Phil. Trans. R. Soc. Lond.,* A, **355,** 283–318.

Kennett, B.L.N., 1983. *Seismic Wave Propagation in Stratified Media*, p. 342, Cambridge University Press, Cambridge.

Kennett, B.L.N. & Widiyantoro, S., 1999. A low seismic wave-speed anomaly beneath northwestern India: a seismic signature of the Deccan plume?, *Earth planet. Sci. Lett.,* **165,** 145–155.

Langston, C.A., 1979. Structure under Mount Rainier, Washington, inferred from teleseismic body waves, *J. geophys. Res.,* **84,** 4749–4762.

Lawver, L.A. & Muller, R.D., 1994. Iceland hotspot track, *Geology,* **22,** 311–314.

Long, R.E. & Mitchell, M.G., 1970. Teleseismic, *P*-wave delay time in Iceland, *Geophys. J. R. astr. Soc.,* **20,** 41–48.

Menke, W., 1999. Crustal isostasy indicates anomalous densities beneath Iceland, *Geophys. Res. Lett.,* **26,** 1215–1218.

Menke, W. & Levin, V., 1994. Cold crust in a hot spot, *Geophys. Res. Lett.,* **21,** 1967–1970.

Menke, W. & Sparks, D., 1995. Crustal accretion model for Iceland predicts 'cold' crust, *Geophys. Res. Lett.,* **22,** 1673–1676.

Menke, W., Brandsdottir, B., Einarsson, P. & Bjarnason, I.T., 1996. Reinterpretation of the RRISP-77 Iceland shear-wave profiles, *Geophys. J. Int.,* **126,** 166–172.

Menke, W., Levin, V. & Sethi, R., 1995. Seismic attenuation in the crust at the mid-Atlantic plate boundary in south-west Iceland, *Geophys. J. Int.,* **122,** 175–182.

Menke, W., West, M., Brandsdottir, B. & Sparks, D., 1998. Compressional and shear velocity structure of the lithosphere in northern Iceland, *Bull. seism. Soc. Am.,* **88,** 1561–1571.

Miller, A.D., 1996. Seismic structure and earthquake focal mechanisms of the Hengill volcanic complex, SW Iceland, *PhD thesis,* University of Durham, Durham.

Miller, A.D., Foulger, G.R. & Julian, B.R., 1998a. Non-double-couple earthquakes II. Observations, *Rev. Geophys.,* **36,** 551–568.

Miller, A.D., Julian, B.R. & Foulger, G.R., 1998b. Three-dimensional seismic structure and moment tensors of non-double-couple earthquakes at the Hengill–Grensdalur volcanic complex, Iceland, *Geophys. J. Int.,* **133,** 309–325.

Mohorovicic, A., 1909. Das Beben, *Jb. met. Obs. Zagreb (Agram.),* **9,** 1–63.

Mutter, C.Z. & Mutter, J.C., 1993. Variations in thickness of layer 3 dominate oceanic crustal structure, *Earth planet. Sci. Lett.,* **117,** 295–317.

Nilsen, T.H., 1978. Lower Tertiary laterite on the Iceland–Faeroe ridge and the Thulean land bridge, *Nature,* **274,** 786–788.

Oskarsson, N., Sigvaldason, G.E. & Steinthorsson, S., 1982. A dynamic model of rift zone petrogenesis and the regional petrology of Iceland, *J. Pet. Special Issue,* **23,** 28–74.

Oxburgh, E.R. & Turcotte, D.L., 1968. Mid-ocean ridges and geotherm distribution during mantle convection, *J. geophys. Res.,* **73,** 2643–2661.

Palmason, G., 1971. *Crustal Structure of Iceland from Explosion Seismology,* p. 187, Soc. Sci. Isl., Reykjavik.

Palmason, G., 1974. Heat flow and hydrothermal activity in Iceland, in *Geodynamics of Iceland and the North Atlantic area,* pp. 297–306, ed. Kristjansson, L., Reidel, Dordrecht.

Palmason, G., 1980. A continuum model of crustal generation in Iceland; kinematic aspects, *J. Geophys.,* **47,** 7–18.

Randall, G.E., 1989. Efficient calculation of differential seismograms for lithospheric receiver functions, *Geophys. J. Int.,* **99,** 469–481.

Recq, M., Brefort, D., Malod, J. & Veinante, J.-L., 1990. The Kerguelen Isles (southern Indian Ocean): new results on deep structure from refraction profiles, *Tectonophysics,* **182,** 227–248.

Richardson, K.R., Smallwood, J.R., White, R.S., Snyder, D.B. & Maguire, P.K.H., 1998. Crustal structure beneath the Faroe Islands and the Faroe–Iceland ridge, *Tectonophysics,* **300,** 159–180.

Richardson, W.P., Okal, E.A. & Van der Lee, S., 2000. Rayleigh-wave tomography of the Ontong–Java Plateau, *Phys. Earth planet. Inter.,* **118,** 29–51.

Ritzert, M. & Jacoby, W.R., 1985. On the lithospheric seismic structure of Reykjanes ridge at 62.5°N, *J. geophys. Res.,* **90,** 10 117–10 128.

Saemundsson, K., 1979. Outline of the geology of Iceland, *Jokull,* **29,** 7–28.

Saemundsson, K., Kristjansson, L., McDougall, I. & Watkins, N.D., 1980. K-Ar dating, geological and paleomagnetic study of a 5-km lava succession in northern Iceland, *J. geophys. Res.,* **85,** 3628–3646.

Salisbury, M.H. & Christensen, N.I., 1978. The seismic velocity structure of a traverse through the Bay of Islands ophiolite complex, Newfoundland, an exposure of oceanic crust and upper mantle, *J. geophys. Res.,* **83,** 805–817.

Sigmundsson, F., Einarsson, P. & Bilham, R., 1992. Magma chamber deflation recorded by the Global Positioning System; the Hekla 1991 eruption, *Geophys. Res. Lett.,* **19,** 1483–1486.

Singh, S.C. & McKenzie, D., 1993. Layering in the lower crust, *Geophys. J. Int.,* **113,** 622–628.

Sinha, M.C. & Louden, K.E., 1981. The crustal structure of the Madagascar ridge, *Geophys. J. R. astr. Soc.,* **66,** 351–377.

Smallwood, J.R. & White, R.S., 1998. Crustal accretion at the Reykjanes ridge, 61°–62°N, *J. geophys. Res.,* **103,** 5185–5201.

Smallwood, J.R., White, R.S. & Minshull, T.A., 1995. Sea-floor spreading in the presence of the Iceland plume: the structure of the Reykjanes ridge at 61°N, *J. Geol. Soc. Lond.,* **152,** 1023–1029.

Smallwood, J.R., Staples, R.K., Richardson, K.R. & White, R.S., 1999. Crust generated above the Iceland mantle plume: from continental rift to oceanic spreading center, *J. geophys. Res.,* **104,** 22 885–22 902.

Spudich, P. & Orcutt, J., 1980. A new look at the seismic velocity structure of the oceanic crust, *Rev. Geophys. Space Phys.,* **18,** 627–645.

Staples, R.K., 1997. Crustal structure above the Iceland mantle plume, *PhD thesis*, University of Cambridge, Cambridge.

Staples, R.K., White, R.S., Brandsdottir, B., Menke, W., Maguire, P.K.H. & McBride, J.H., 1997. Faeroe–Iceland ridge experiment 1. Crustal structure of northeastern Iceland, *J. geophys. Res.,* **102,** 7849–7866.

Stefánsson, R. *et al.,* 1993. Earthquake prediction research in the south Iceland seismic zone and the SIL project, *Bull. seism. Soc. Am.,* **83,** 696–716.

Stein, C. & Stein, S., 2003. Sea floor heat flow near Iceland and implications for a mantle plume, *Astron. Geophys.,* **44,** 1.8–1.10.

Stolper, E., 1980. A phase diagram for mid-oceand ridge basalts: preliminary results and implcations for petrogenesis, *Cont. Min. Pet.,* **74,** 13–27.

Talwani, M., Windisch, C.C. & Langseth, M.G., Jr, 1971. Reykjanes ridge crest: a detailed geophysical study, *J. geophys. Res.,* **76,** 473–517.

Toomey, D.R. & Foulger, G.R., 1989. Tomographic inversion of local earthquake data from the Hengill–Grensdalur central volcano complex, Iceland, *J. geophys. Res.,* **94,** 17 497–17 510.

Tryggvason, E., 1962. Crustal structure of the Iceland region from dispersion of surface waves, *Bull. seism. Soc. Am.,* **52,** 359–388.

Tryggvason, E., 1964. Arrival times of P waves and upper mantle structure, *Bull. seism. Soc. Am.,* **54,** 727–736.

VanDecar, J.C., James, D.E. & Assumpcao, M., 1995. Seismic evidence for a fossil mantle plume beneath south America and implications for plate driving forces, *Nature,* **378,** 25–31.

Vink, G.E., 1984. A hotspot model for Iceland and the Voring Plateau, *J. geophys. Res.,* **89,** 9949–9959.

Vogt, P.R., Schneider, E.D. & Johnson, G.L., 1969. The crust and upper mantle beneath the sea, in *The Earth's Crust and Upper Mantle*, pp. 556–617, ed. Hart, P.J., Am. Geophys. Union.

Watts, A.B., ten Brink, U.S., Buhl, P. & Brocher, T.M., 1985. A multichannel seismic study of lithospheric flexure across the Hawaiian–Emperor seamount chain, *Nature,* **315,** 105–115.

Weir, N.R.W., White, R.S., Brandsdottir, B., Einarsson, P., Shimamura, H. & Shiobara, H., 2001. Crustal structure of the northern Reykjanes ridge and Reykjanes peninsula, southwest Iceland, *J. geophys. Res.,* **106,** 6347–6368.

White, R.S., 1984. Atlantic oceanic crust: seismic structure of a slow-spreading ridge, in *Ophiolites and Oceanic Lithosphere,* pp. 34–44, eds Gass, I.G., Lippard, S.J. & Shelton, A.W., Geological Society, London.

Zverev, S.M., Kosminskaya, I.P., Krasilstchikiva, G.A. & Mikhota, G.G., 1976. *The Crustal Structure of Iceland and the Iceland–Faeroe–Shetland Region*, pp. 73–93, Soc. Sci. Isl., Reykjavik.