

The seismic structure of Iceland

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Abstract — We review here the contributions that seismology has made toward our understanding of Iceland, the best-studied of ridge-centered hotspots. Iceland is the subarial part of the Greenland-Iceland-Faeroe Ridge, a band of thickened basaltic crust, 25–30 km, stretching across the North Atlantic that represents the product of hotspot volcanism from the early Tertiary to present. Crustal thickness increases from 11 km on the Reykjanes Ridge section of the Mid Atlantic Ridge south of Iceland, to 35–40 km in central Iceland, and then decreases to 9 km on Kolbeinsey Ridge to the north. The correlation of crustal thickness, topographic elevation and gravity indicate that the crust is in isostatic equilibrium. Crustal seismic velocities are comparable to those in normal oceanic crust, except that Oceanic Layer 3, thought to represent gabbroic cumulates, is disproportionately thick. Judging from the generally low shear wave attenuation observed almost everywhere in Iceland, the crust is mostly subsolidus, although tomographic imaging of both shear velocity and V_P/V_S ratio indicate that hotter conditions occur in the lower crust along the neovolcanic zones (the expression of the Mid Atlantic plate boundary in Iceland). Actual basaltic melt has only been observed in shallow (2–3 km) crustal magma chambers located beneath central volcanoes, where it causes local zones of intense shear wave attenuation and delayed P wave arrival times. These magma chambers are often underlain by seismically-fast domes thought to represent gabbroic cumulates. While the crust must contain plumbing systems that deliver melt from the asthenospheric mantle to these shallow magma chambers, they have not been observed. Much of the crust, especially outside the neovolcanic zones, appears to be underlain by a 30–50 km thick mantle 'lid' of seismically-fast, relatively cool (subsolidus) mantle lithosphere. The mantle beneath this lid is dominated by a cylindrically-shaped zone of low seismic velocities (both compressional and shear) that has a diameter of 100–200 km and is centered beneath the Vatnajökull region of southeastern Iceland. Regional tomography indicates that this feature extends down to at least 350 km. Measurements of seismic anisotropy, which provides information on mantle-flow directions, provide some evidence for mantle upwelling in this region, but the pattern of flow does not have the radial pattern expected of simple upwelling. Instead, mantle flow appears to be channeled in some way by structures associated with the plate boundary. Whether the low velocity anomaly represents a deeply-rooted plume beneath Iceland is currently hotly-debated. The key issue of whether the low-velocities extend to depths deeper than 350 km has been addressed with global tomography, but results appear equivocal: some authors see little evidence in the global tomography for it extending deeper than the bottom of the transition zone (670 km), but others see it connecting with larger-scale low-velocity features in the lower mantle.

INTRODUCTION

The world's many hotspots, regions of unusually intense and persistent volcanism, occur in a wide variety of tectonic settings. Some, such as the Hawaiian-

Emperor hotspot, occur within old, otherwise-stable oceanic lithosphere, and are associated with tracks of islands and seamounts. Others, such as the Yellowstone hotspot, occur within continents, and interact

strongly with the continent's thick, sialic crust to produce explosive volcanism. Still others, like the Iceland hotspot, are part of a young, actively-spreading mid-ocean ridge system that would be volcanically-active even were the hotspot absent. J. Tuzo Wilson (1963) was the first to postulate a relationship between these hotspots and heat and melt delivery by mantle convection. This idea was further developed by Morgan (1971), whose plume-model envisioned hotspots sitting above narrow zones of convective upwelling that originated in the lower mantle, perhaps even the core-mantle boundary. Subsequently, much effort has been directed towards understanding the physical structure and chemical processes occurring in hotspots, and to linking this information to specific styles of mantle convection. The case for at least some hotspots being caused by deeply-rooted plumes is now very compelling. Yet whether such plumes are responsible for all, or even most, hotspots is less clear.

Iceland is the best-studied of the ridge-centered hotspots, and among the best-studied of any kind. It is long-lived and associated with voluminous basaltic volcanism, having been present on the northern Mid-Atlantic Ridge since at least the early Tertiary, judging from the band of thickened sea floor crossing the North Atlantic that is called the Greenland-Iceland-Faroe Ridge (White, 1989). We review here the seismological research that has contributed to defining the nature of this hotspot and identifying the processes occurring within it. As will become apparent, seismology has contributed to a host of hotspot issues, such as the configuration of the magma plumbing that is responsible for delivering basaltic melt to the surface and to the shallow crust; the structure of the basaltic crust that is the end-product of hotspot volcanism; the way that the Mid-Atlantic Ridge is interacting with the hotspot, and, finally to determining the style of mantle convection beneath Iceland and whether it conforms to the plume model. As will also become apparent, seismology has provided few definitive answers, so far. Even after nearly fifty years of research, many fundamental issues remain unresolved (or at least are hotly-debated) and will require further research.

EARLY REFRACTION EXPERIMENTS AND TELESEISMIC STUDIES IN ICELAND: THE INITIATION OF THE CRUSTAL THICKNESS DEBATE

Seismic refraction and wide-angle reflection experiments have played an important role in determining the crustal structure of Iceland (Figure 1). The first refraction experiment in Iceland, a 250 km long profile in NW-Iceland, revealed compressional velocities of 3.69–6.71 km/s in the uppermost 17.8 km, increasing to 7.38 km/s at a depth of 18–28 km (Båth, 1960). Båth divided the crust into three layers, determined its thickness to be 27.8 km and pointed out that the compressional velocity of 6.71 km/s (layer 2) agreed well with crustal velocities elsewhere in the N-Atlantic. However, velocities in his third layer corresponded to those of the deeper layer of the Mid-Atlantic ridge, which Ewing and Ewing (1959) considered to be made up of a mixture of oceanic crust and mantle and which are now thought to be gabbroic cumulates.

Tryggvason (1962) used surface wave dispersion analysis to divide the Icelandic crust into a 10 km thick, two-layer stack overlying a half-space with a compressional velocity of 7.4 km/s and a shear wave velocity of 4.3 km/s. He concluded that the half-space represented anomalously low-velocity 'mantle'. Tryggvason's velocity structure was in fact in good agreement with Båth's refraction result, but he gave a significantly different interpretation of material at 10 km depth and deeper, calling it 'mantle' rather than 'Oceanic layer 3B'. These two papers thus set the stage of a long-running controversy about the seismic structure of Iceland: Is the crust thin or thick? Is the mantle extremely slow (and therefore partially-molten)?

Pálmason (1963, 1971), divided Iceland into 5 layers, numbered 0–4, with Layer 4 ($V_p=7.2$ km/s; 8–16 km depth) being anomalous (low-velocity) mantle. Zverev *et al.* (1976) used data from the North-Atlantic Seismic Project (NASP) to analyse the crustal thickness below the Iceland-Faroe-Shetland region, and data from two stations in E-Iceland and one station at the eastern margin of the Northern Volcanic

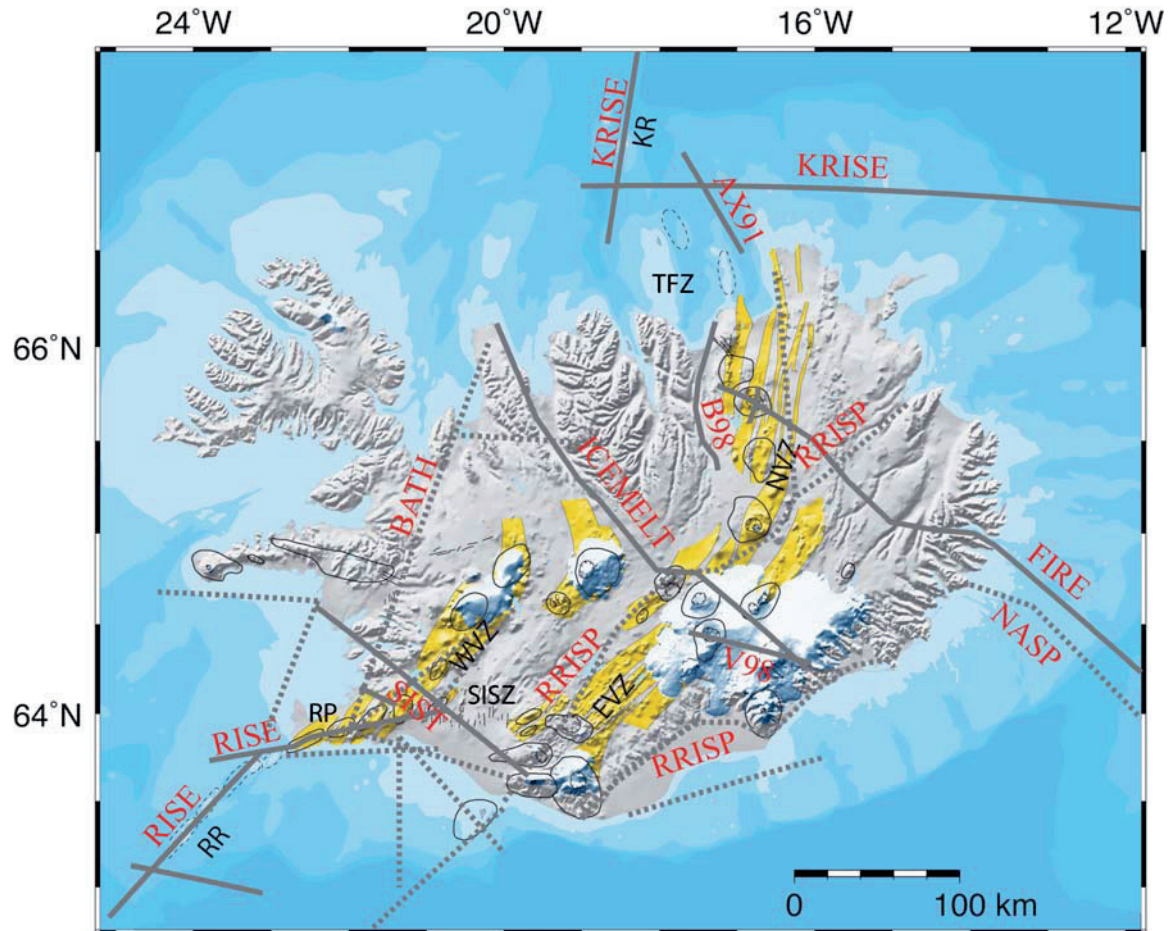


Figure 1. Refraction profiles longer than 50 km on a tectonic map of Iceland (Einarsson and Sæmundsson, 1987). Individual plate boundary segments are indicated: RR Reykjanes Ridge, RP Reykjanes Peninsula, WVZ Western Volcanic Zone, SISZ South Iceland Seismic Zone, EVZ Eastern Volcanic Zone, NVZ Northern Volcanic Zone, TFZ Tjörnes Fracture Zone, KR Kolbeinsey Ridge. – *Staðsetningar bylgjubrotsmælilína lengri en 50 km á korti Páls Einarssonar og Kristjáns Sæmundssonar (1987) af eldstöðvakerfum innan gosbelta Íslands.*

Zone (NVZ) to infer a crustal thickness of 40–60 km beneath Iceland. Zverev *et al.* (1976) claimed that crustal velocities below 8 km/s belonged to the crust but did not detect reflections, such as P_mP or S_mS , from the crust-mantle boundary (Moho). Analysing the same NASP data, Bott and Gunnarsson (1980) detected Moho reflections with apparent phase velocities

of 7.8 km/s, at about 23 km depth below the eastern coast deepening to 27 km depth at the eastern margin of the NVZ. They also observed a crustal thickness of 35 km beneath the Iceland-Faroe Ridge, an east-west oriented band of elevated sea floor located to the east of Iceland. As their data came from only two stations, 35 km apart, it is understandable that they did

not attempt to revise Pálmason's (1971) and Angenheister *et al.*'s (1979) model of Iceland, with its 10 km thick crust. They left as unexplained why the crustal thickness along the Iceland-Faroe ridge, which is trace of Iceland hotspot, should be three times thicker than above the Iceland hotspot itself.

The 1977 Reykjanes Ridge Iceland Seismic Experiment (RRISP-77) was historically important, because it was the first experiment to define the transition from the oceanic crust of the Reykjanes Ridge south of Iceland to the anomalously thick, subaerial crust of Iceland itself. The RRISP-77 dataset, especially as interpreted by Angenheister *et al.* (1979; 1980) and Gebrande *et al.* (1980), played a major role in the development of the first comprehensive model of the crust-mantle transition beneath Iceland and its relation to the Iceland mantle plume. These authors favored a thin crust / hot mantle interpretation of Iceland's seismic structure.

Around this time, the newly-developed Worldwide Standardized Seismographic Network (WWSSN) led to significant improvements in global earthquake monitoring (e.g. better locations and origin times). Mantle structure could be probed using P and S waves from distant earthquakes (teleseisms). P wave traveltimes from teleseisms to stations in Iceland were shown to be systematically long, compared to the global average (Tryggvason, 1964), implying a broad region of anomalously slow mantle beneath Iceland. Tryggvason's (1964) model had slow (7.4 km/s) compressional velocities extending down to a depth of 234 km beneath Reykjavík. Similarly, Francis (1969), on the basis of anomalously high V_p/V_s ratios, found the low-velocity region to extend down to 250 km depth. Both authors included Pálmason's Layer 4 ($V_p = 7.2$ km/s; 8–16 km depth) to be the top part of this much larger region of low-velocity mantle. Thus, the teleseismic data were seen as consistent with Iceland having a relatively thin (10–15 km) crust. The very low (<7.4 km/s) mantle compressional velocities cannot be explained by the effect of high temperatures, alone. Thus, one important implication of this velocity model was that the mantle was partially-molten, perhaps with melt concentrations as high as 10%, and that this melt was present at relatively shallow

(10–15 km) depths beneath Iceland. Gebrande *et al.* (1980) interpreted late shear wave arrivals observed on RRISP profiles as verifying the presence of melt, arguing that the high compressional to shear (V_p/V_s) ratio of 1.96 could not be explained by the effect of temperature, alone.

This model of a relatively thin, 10 to 15 km thick and thickening with age, oceanic crust overlying a very hot and partially molten mantle, is not attributable to any single author. Instead, over a period of 10–15 years, seismological (Tryggvason, 1962, Angenheister *et al.*, 1980, Gebrande *et al.*, 1980, Tryggvason *et al.*, 1983), geological (Sæmundsson 1979), geothermal (Pálmason, 1971, 1973, 1986; Pálmason and Sæmundsson, 1979), and magnetotelluric (Beblo and Björnsson, 1980, Eysteinsson and Hermance, 1985) data were used to explore how the processes of plate-tectonic spreading and hotspot magmatism interacted.

Angenheister *et al.*, (1980), noting discontinuous second arrivals from one of the RRISP shots, mentioned an alternative interpretation of the data, i.e., that the second P arrival, with an apparent velocity of 7.8 km/s (Gebrande *et al.*, 1980), may indicate a crustal thickness of 30 km under Iceland. Pavlenkova and Zverev's (1981) reinterpretation of the RRISP and NASP-data favored a crustal thickness of 30 km beneath Iceland. However, the alternative model of Angenheister *et al.* (1980), and the "lower crust" interpretation of Pálmason's Layer 4 by Zverev *et al.* (1976), based on NASP-data, and by Pavlenkova and Zverev (1981), based on both RRISP and NASP-data, was largely ignored in the subsequent literature.

The difference in data interpretation, discussed above, persisted in the literature as later surveys (Zverev *et al.*, 1976; Angenheister *et al.*, 1980; Gebrande *et al.*, 1980; Bjarnason *et al.*, 1993) as well as reinterpretations of Pálmason's data (Flóvenz and Gunnarsson, 1991) all confirmed earlier velocity measurements (compressional velocities of 7.0–7.4 km/s at depths of 10–30 km), but were interpreted in contradicting ways. Some researchers interpreted compressional velocities higher than 7.0 km/s as anomalously low-velocity "mantle" (Pálmason, 1963, 1971; Gebrande *et al.*, 1980; Flóvenz and Gunnarsson, 1991),

while others interpreted them as lower crustal velocities (Zverev *et al.*, 1976; Pavlenkova and Zverev, 1981; Bjarnason *et al.*, 1993; Staples *et al.*, 1997; Menke *et al.*, 1996; 1998; Darbyshire *et al.*, 1998; Smallwood *et al.*, 1999; Weir *et al.*, 2001).

THE THIN-CRUST VERSUS THICK-CRUST DEBATE SUBSIDES WITH NEW DATA

The question of the depth to mantle is, in principle, equivalent to the question of the thickness of the crust. Yet the two questions have had a rather decoupled history in Iceland, owing to the somewhat different analytical approaches to the underlying seismic datasets. The idea of crustal thickness is most sensible when a sharp increase in velocity, or Moho, marks the boundary between crust and mantle. Hence, crustal thickness measurements are to some degree related to the detection of the Moho-reflected compressional and shear phases, P_mP and S_mS . The South Iceland Seismic Tomography (SIST) project, a dense, digital seismic refraction profile across SW-Iceland revealed a clear wide-angle P_mP reflection from a reflector at a depth of 20–24 km, which Bjarnason *et al.* (1993) interpreted as Moho. Still, the controversy regarding the proper interpretation of Pálmason's Layer 4 spawned a related controversy regarding whether or not the reflected phases really represented reflections from the Moho, as contrasted to some deeper interface within the mantle (Bjarnason *et al.*, 1994; Gudmundsson, 1994).

Menke *et al.*'s. (1996) reinterpretation of the RRISP-77 profiles identified clear Moho-reflected phases (both P_mP and S_mS) analogous to the P_mP phase Bjarnason *et al.* (1993) had observed in SW-Iceland. Furthermore, these authors showed that the late shear wave arrivals observed on RRISP profiles were the Moho-reflected S_mS phase, and not S, thus eliminating the need to invoke high V_p/V_s ratios (and correspondingly low shear velocities) in their explanation. It thus became clear that the RRISP data had indeed provided evidence of a crustal thickness close to 30 km in central Iceland as had previously been argued by some authors. More recent refraction surveys

have confirmed the existence of prominent P_mP and S_mS reflections, with Moho ranging in depth from 11 km at 62°40'N on zero age crust at the Reykjanes Ridge to 21 km in Southwest Iceland (Weir *et al.*, 2001). Similar crustal thickness (~20 km) is observed just north of the Hengill central volcano, within the Western Volcanic Zone, as below the Krafla central volcano, in the Northern Volcanic Zone (Bjarnason *et al.*, 1993; Brandsdóttir *et al.*, 1997; Staples *et al.*, 1997). A marked crustal thickening has been observed towards the center of the Iceland mantle plume (Menke *et al.*, 1996; 1998; Darbyshire *et al.*, 1998; Allen, 2001) where the crust is about twice as thick (38 km) as beneath the Hengill and Krafla central volcanoes (Figure 2). The crust also thickens with age and is close to 35 km thick beneath the ~15 million year old central eastern fjords (Staples *et al.*, 1997). Gravity modelling shows these crustal models to be consistent with the observed gravity field (Staples *et al.*, 1997; Darbyshire *et al.*, 1998; Smallwood *et al.*, 1999; Weir *et al.*, 2001; Gudmundsson, 2003). Crustal thickness correlates with topographic elevation in a way that is consistent with an Airy-type isostatic model of density compensation (Menke, 1999; Gudmundsson, 2003).

Iceland sits on the slow spreading Mid-Atlantic Ridge. The Iceland region falls at the extreme end of such ridges due to its interaction with the Iceland mantle plume. Iceland has a crustal thickness which is 3-5 times that of 'normal' oceanic crust; it has permanent magma chambers (discussed below) which have so far only been found in one region along the Mid-Atlantic Ridge itself (Sinha *et al.*, 1998). However, the overall pattern of crustal structure and the processes that create it closely follow the oceanic norm.

THE SIGNIFICANCE OF SHEAR-WAVE DATA

Shear wave measurements are particularly relevant to the question of the presence of melt. Shear waves cannot propagate through a completely molten layer and are attenuated very rapidly in regions of partial melt. For instance, using Mavko's (1980) estimate of the shear wave quality factor, $Q_S=2$, for a 3% partial

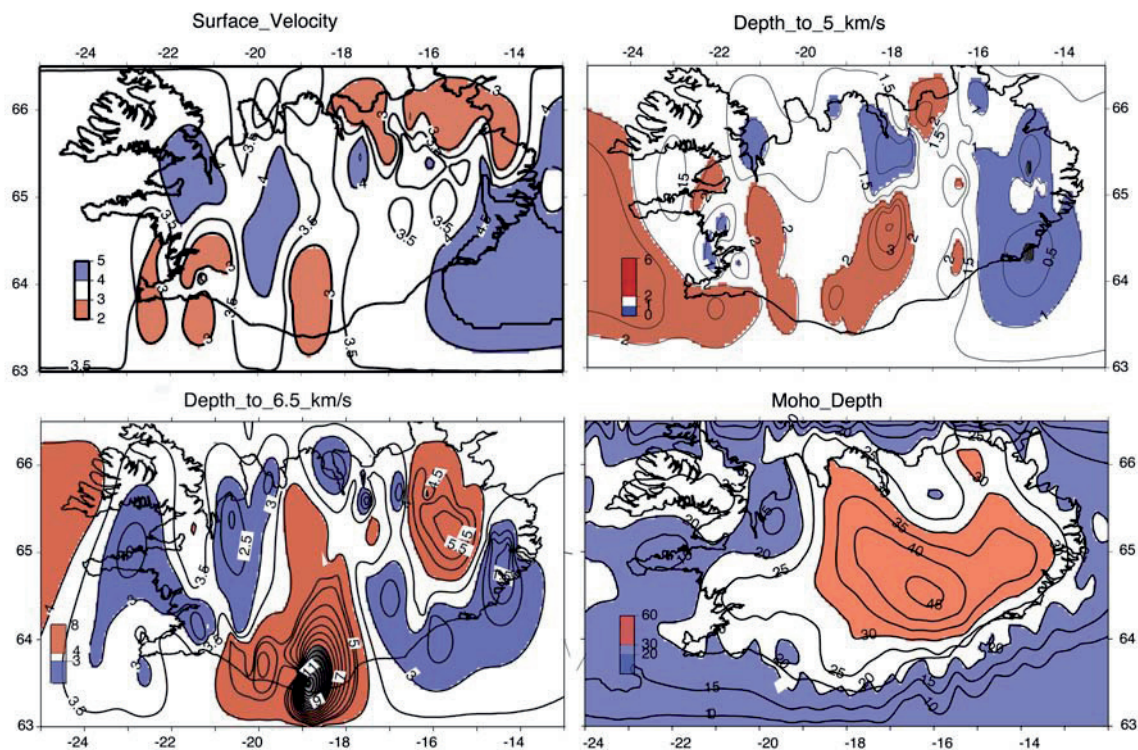


Figure 2. Crustal structure of Iceland. (Top left) Surface compressional velocity (mean velocity, 3.65 km/s). (Top right) Depth below sea level to 5.0 km/s isovelocity surface is a proxy for depth to bottom of extrusive volcanics (mean depth, 1.6 km). (Bottom left) Depth below sea level to 6.5 km/s isovelocity surface (mean depth, 3.6 km). Maps created by regridded data compiled by Pálmason (1971), supplemented with results of more recent refraction lines (SIST, KATLA, FIRE, ICEMELT, B96 and RISE, Figure 1). (Bottom right) Depth to Moho (mean depth, 22 km). Maps created by regridding data were compiled by Menke (1999) and Allen (2001). – *Yfirlitsmynd af uppbyggingu jarðskorpu Íslands sem sýnir yfirborðshraða P-bylgna, dýpi á 5 km/s og 6,5 km/s jafnhraðafleti og heildarþykkt.*

melt, the amplitude of a 10 Hz shear wave would be reduced by an order of magnitude after traveling only 0.5 km. At hot ($>800^{\circ}\text{C}$) but subsolidus temperatures, shear wave attenuation should also be high, with quality factors as low as $Q_S=20$. The ratio of compressional to shear wave velocity, V_p/V_s , which is diagnostic of both high temperature and the presence of partial melt, should be considerably higher than the 1.76 expected for a normal solid, reaching values in excess of 1.9 near the solidus.

From the earliest refraction experiments, regional shear waves with turning points in the mid- to lower crust (5–15 km depth), have been observed in Iceland. Pálmason (1963) reported V_p/V_s ratios of 1.79, 1.85, and 1.80 for crustal Layers 1, 2 and 3, respectively, in northern and eastern Iceland. These values are somewhat higher than in western (1.72–1.75) (Tryggvason, 1962), southwestern (1.74–1.79) (Menke *et al.*, 1994; 1996; Weir *et al.*, 2001), central (1.76–1.78) (Darbyshire *et al.*, 1998) and northeastern Iceland (1.78)

(Staples *et al.*, 1997). The earlier measurements by Gebrande *et al.* (1980) (excluding those based on misidentified S_mS phases) are similar, with $V_p/V_s=1.76$ out to a range of 120–140 km, corresponding to waves turning in the mid-crust.

One would expect that the V_p/V_s ratio would increase with depth in the crust, owing to the high geothermal gradient. Broadband waveform inversions of local earthquakes in Iceland, find that the V_p/V_s ratio, averaged for all of Iceland, increases from 1.78 at the surface to about 1.92 at 35 km depth, with the overall gradient in V_p/V_s being about 0.004 per km (Allen 2001; Allen *et al.*, 2002a). This change is primarily driven by the increase of V_p with depth, since V_s remains approximately constant (3.7–3.8 km/s) for depths deeper than about 10 km. Tomographic reconstruction of lateral variations in shear velocity in the mid-crust (i.e. 10 km depth), using waveforms of regional earthquakes, strongly correlate with crustal age, with the neovolcanic zones being as much as 0.25 km/s slower and the old areas of north western Iceland being as much as 0.25 km/s faster than the average value of 3.6 km/s for that depth (Allen, 2001; Allen *et al.*, 2002a). A similar range of velocities is also at near-Moho (30 km) depths, with the slowest values being concentrated in a small region beneath northwestern Vatnajökull and the fastest regions again being in northwestern Iceland. Analogous tomographic inversions based on Rayleigh wave dispersion (Li and Detrick, 2006) show a similar pattern, though with a somewhat smaller range in variability.

The observed crustal thickness decrease with increasing distance from the center of the Iceland plume corresponds to a decrease in potential temperature of the mantle upwelling, from about 1500°C for ~22 km thick crust to about 1375°C for 12 km thick crust (Bown and White, 1994). Recent measurements of seismic thickness agree with measurements of igneous thickness deduced from rare earth element inversions of the extrusive volcanics (Nicholson *et al.*, 1991; White *et al.*, 1992, 1995). Using the model of Bown and White (1994), a 15–38 km thick crust beneath Iceland implies temperatures in excess of 100–300°C for the plume head if formed by passive adiabatic decompression melting of the mantle and includ-

ing a component of active convection in the plume core. Such a temperature anomaly in the mantle is in agreement with 'hot-narrow' plume models derived from teleseismic travel-time studies; i.e. models with about 4% reduction in S-velocity and 2% in P-velocity within a 100 km wide, cylindrical plume conduit below a wider plume head above 200 km depth (Wolfe *et al.*, 1997; Allen *et al.*, 1999; 2002b). The mantle temperature and oceanic crustal thickness decrease to normal (~1350°C and 7–8 km respectively) within 1000–1200 km from the plume centre.

SHALLOW MAGMA CHAMBERS BENEATH ICELAND

Iceland is transected by a multi-branched, slow spreading plate boundary which, due to its interaction with the Iceland mantle plume, contains many tectonic features of a fast spreading ridge. The divergent plate boundary is made up of volcanic systems (axial rift segments) consisting of central volcanoes and their associated fissure swarms and is characterized by large variations in upper crustal compressional velocity. Early refraction experiments were not designed to image active central volcanoes, partly because of higher attenuation in these regions which made it more difficult using the insensitive analog equipment of that time. Prior to the RRISP-77 survey, most of the profiles were also less than 60 km long corresponding to raypaths down to a maximum depth of 6–10 km. No shallow crustal magma chambers were identified prior to the Krafla rifting episode which commenced in December 1975 with a small eruption in the center of the Krafla caldera followed by several lateral intrusion events and eruptions, which activated a 80 km long section of the Northern Volcanic Zone (Björns-son *et al.*, 1977). The Krafla magma chamber is currently one of the best surveyed magma chambers along a slowly-divergent plate boundary.

Einarsson (1978) was the first to use local earthquakes recorded during an inflation period, in order to delineate two regions of high shear wave attenuation within the Krafla caldera. He inferred the existence of a shallow crustal magma reservoir at a depth of approximately 3 km. This magma chamber is smaller

than the caldera, about 2 km by 7 km, with the long axis oriented E-W and divided near its top. Its maximum depth of roughly 7 km was not well constrained. Brandsdóttir and Menke (1992) performed waveform studies on earthquakes which originated within the Krafla caldera during a brief inflation period in July 1988 in order to study the thickness of the magma chamber. They identified reflections from beneath the magma chamber which indicated that it was less than 1 km thick in the central-northern part of the caldera, but they could not constrain its lateral dimensions. An undershooting experiment across the Krafla central volcano revealed a velocity structure characterized by large variations in compressional velocity. A 40 km wide high-velocity dome extends from the lower crust (11–14 km depth) beneath the volcano narrowing upward (Figure 3). The magma chamber sits at its top near 3 km depth. It is defined by both 0.2–0.3 s compressional wave delays and shear wave shadowing to be 2–3 km N-S, 8–10 km E-W, and 0.7–1.8 km thick in the central part of the Krafla caldera (Brandsdóttir *et al.*, 1997). Another undershooting experiment revealed the existence of a magma chamber, 5 km across, at 3 km depth beneath the Katla central volcano (Figure 4) (Gudmundsson *et al.*, 1994).

Doming of the lower crust is observed beneath the central volcanoes (Pálmason, 1971; Flóvenz og Gunnarsson, 1980; Gudmundsson *et al.* 1994; Brandsdóttir *et al.* 1997). Where detected, 1–2 km thick, localised magma chambers sit on top of these high-velocity domes, at around 3 km depth, i.e. approximately at the level of buoyant equilibrium for basaltic melt within the crust. The domes likely consist of olivine-rich cumulates, formed at the base of magma chambers and subsequently advected downward and outward during the spreading process. The magma chambers in Iceland are thus different from those on the Mid-Atlantic Ridge itself (Sinha *et al.*, 1998) where a thin axial melt lens is underlain by a zone of lower velocities, interpreted as crystalline mush containing partial melt. The height of the high-velocity domes most likely reflects their evolutionary stage and their volume is indicative of their lifespan. For instance, the 33 km wide Krafla dome may be at least 1.8 million years old based on a 0.9 cm/year half spreading rate (Brands-

dóttir *et al.*, 1997). The high-velocity anomaly is also observed, but weaker, beneath the neighboring Peistareykir and Fremri-Námur central volcanoes. Krafla, being at a higher evolutionary stage than its neighbors, may therefore be a more significant source of lower crustal rocks (Menke *et al.*, 1998). The crustal magma chambers are intermittently replenished during rifting episodes and act as buffers on the magma upwelling from the mantle as they regulate the magma residence time within the plumbing system, the rate and volume of magma output, and alter the composition of the magma. Central volcanoes thus play a major role in crustal genesis along the plate boundary.

THE MODERN IMAGE OF THE CRUST AND UPPERMOST MANTLE BENEATH ICELAND

Today, our image of the Icelandic crust is that of a relatively cool (subsolidus), 20–40 km thick crust, dotted by high-velocity upper crustal domes beneath both extinct and active eruptive centers, and with many of the active eruptive centers containing small, shallow magma chambers. Whereas large variations are observed in upper crustal velocities (0–15 km), lower crustal velocities are more uniform (Pálmason, 1971; Allen *et al.*, 2002a) and characterized by large-scale undulations in the vicinity of the neovolcanic zones (Figure 5). The upper/lower-crustal interface is defined by an abrupt change in velocity gradients from $0.25\text{--}0.57\text{ s}^{-1}$ to $0.024\text{--}0.05\text{ s}^{-1}$ (Flóvenz, 1980; Staples *et al.*, 1997; Weir *et al.*, 2001). Intermediate gradients of 0.15 s^{-1} are also observed beneath the central part of the Reykjanes Peninsula (Weir *et al.*, 2001).

Local earthquake depths are in accordance with the lower crust being ductile as they are mostly confined to the uppermost 10 km (Einarsson, 1991; Stefansson *et al.*, 1993). Stefansson *et al.* (1993) reported hypocenters down to 6–8 km depth beneath the plate boundary at Hengill deepening to 12 km along the South Iceland Seismic Zone. Deeper events observed locally beneath active volcanoes, Eyjafjallajökull (Gudmundsson *et al.*, 2001) and in the vicinity of Askja (Soosalu *et al.*, 2007) have been attributed

to magma upwelling. Assuming that the deepest hypocenters observed so far mark the brittle-ductile transition, one can on the basis of all earthquakes recorded in Iceland during the last 25 years, infer that the seismogenic crust has a maximum thickness of 12–15 km and is generally thinner within the Neovolcanic Zones.

Observed shear wave attenuation beneath Iceland is also much lower than would be expected for a near-solidus lower crust. Menke and Levin (1994) reported lower crustal shear wave quality factors (Q_S) of 150–225 for paths across central western Iceland and $Q_S=300$ for paths in southwestern Iceland, values which indicate that the temperature of the lower crust is at least 250–325°C below the solidus of gabbro. Regions of extremely high shear wave attenuation appear to be very localized. Thus, for example, S_mS phases are clearly visible for some paths through the crustal volume below the Krafla magma chamber, even though the more shallow diving S waves are

severely attenuated by the magma chamber itself.

As identification of mantle refracted phases (P_n and S_n) requires data from denser arrays than is provided by the Iceland seismic network, a limited number of mantle phases have been observed beneath central Iceland. A few regional earthquakes recorded on temporary arrays indicate that most of the energy is reflected at the crust/mantle boundary and only a small fraction is transmitted into the mantle. Bjarnason *et al.* (1993) reported an apparent Moho refractor velocity of 7.6–7.7 km/s beneath SW-Iceland whereas Menke *et al.* (1998) observed apparent P_n and S_n velocities of 8.00 ± 0.10 km/s and 4.31 ± 0.04 km/s, respectively ($V_P/V_S=1.85 \pm 0.02$), beneath central Iceland indicating that the uppermost mantle is also sub-solidus and P_n and S_n velocities of 7.66 ± 0.04 km/s and 3.96 ± 0.04 km/s along the Kolbeinsey Ridge, offshore N-Iceland ($V_P/V_S=1.93$). Allen *et al.*, (2002b) using 40 S_n paths from six regional earthquakes and teleseismic surface waves obtained S_n phase veloci-

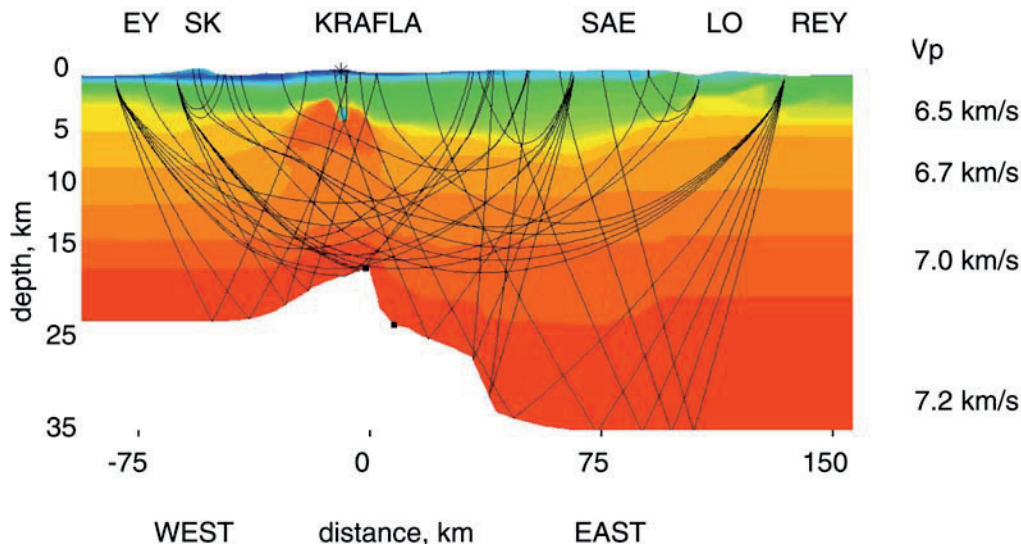


Figure 3. Crustal structure across the Krafla volcanic system along the FIRE refraction profile, NE-Iceland. Adapted from Brandsdóttir *et al.* (1997). See Figure 1 for location of FIRE profile. – *Uppbygging jarðskorpunnar á FIRE mæliníunni yfir Kröflu og Norðausturland. Staðsetning línunnar er sýnd á 1. mynd.*

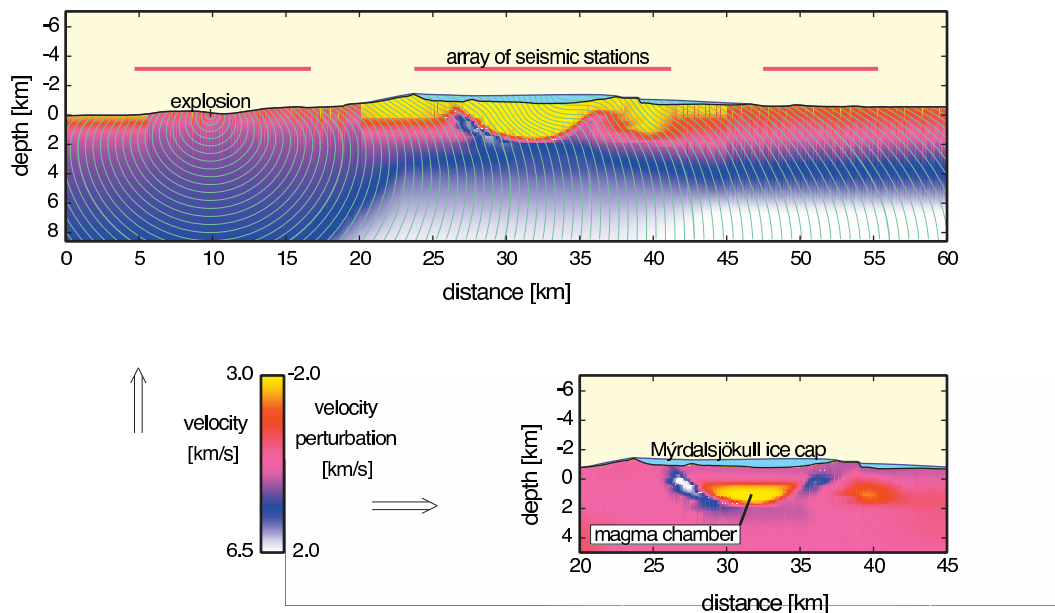


Figure 4. Crustal structure across the Katla volcanic system, S-Iceland. Adapted from Gudmundsson *et al.* (1994). See Figure 1 for location of profile. – *Uppbygging jarðskorpunnar undir Kötlu og Mýrdalsjökli. Staðsetning mælingunnar er sýnd á 1. mynd.*

ties ranging from 3.9–4.5 km/s at 50 and 100 km depth beneath Iceland with an average of 4.1 km/s. Interestingly, observed S_n velocities across western Iceland are significantly lower, in spite of similar crustal age, than across eastern Iceland. As shear S_n velocity of 4.18 km/s has been thought to characterize peridotite just below solidus (Nishimura and Forsyth, 1989), so S_n velocity values for the mantle offshore northeastern Iceland and beneath western Iceland indicate hotter and perhaps even partially molten conditions. Hence the top of melting may be at shallower depths there.

Although the refraction measurements may be biased by dipping mantle structure and the surface wave inversions by upper mantle velocity anisotropy they nevertheless provide some insight into the thermal state of the uppermost mantle indicating that a high-velocity mantle lid may be separating the crust from

the melting region beneath central and eastern Iceland. These results are also in agreement with Menke and Sparks's (1995) prediction of a 'mantle-lid' above the zone of partial melt. Their model places the top of mantle melting about 25 km below Moho. Further support for a mantle lid is presented by Rayleigh-wave inversions providing higher mantle velocities (4.0–4.1 km/s) in the uppermost mantle (40–60 km) beneath Eastern and Western Iceland and lower velocities under central Iceland (Li and Detrick 2006).

Using more than 3000 Rayleigh waveforms recorded at stations in Iceland, Greenland and Europe, Pilidou *et al.* (2004) imaged a roughly circular, 1000 km wide plume head at 75 km depth beneath Iceland spreading out at 100 km depth to 2000 km along axis while narrowing to 600 km across axis. Their plume has a nearly constant S_V velocity of 4.1 ± 0.05 km/s. Using a higher resolution dataset from stations in Iceland Li and Detrick (2006) observed a low-velocity zone (S_V velocity of 3.9–

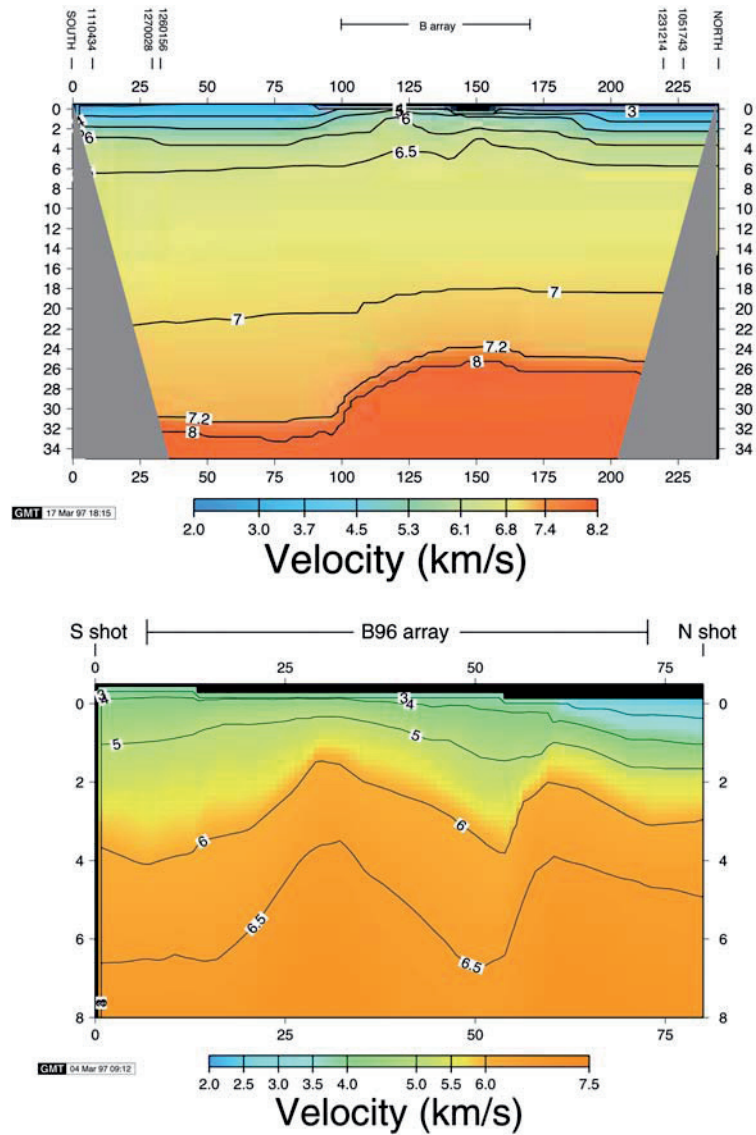


Figure 5. Crustal structure along the B96 refraction profile, N-Iceland, based on both refraction and earthquake data (Top). A close-up of uppermost crustal structure along the seismic array (Below). Adapted from Menke *et al.* (1998). See Figure 1 for location of B96 profile. – Uppbygging jarðskorpunnar á B96 mælilínunni eftir Bárðardal og norður á Tjörnes. Staðsetning mælilínunnar er sýnd á 1. mynd.

4.1 km/s) at 60–120 km depth beneath the volcanic zones, interpreted as mantle material being channelled along the divergent plate boundary. Rayleigh-wave inversions reveal an average mantle shear velocity structure beneath Iceland compatible with velocities at the East Pacific Rise (Nishimura and Forsyth, 1989) and Reykjanes Ridge (Gaherty, 2001).

Shear wave splitting measurements of teleseismic SKS waves reveal seismic anisotropy caused by a complex upper mantle flow beneath Iceland. The pattern of anisotropic fast directions, which are proxies for mantle flow directions, are not consistent with either a corner-flow model of mantle upwelling that would be characteristic of a normal mid-ocean ridge or a radial-divergence pattern that would be characteristic of a plume. Splitting patterns at the ICEMELT and HOTSPOT stations exhibit both azimuthal and vertical variations that are being interpreted as caused by lithospheric plate shear relative to return mantle flow (Bjarnason *et al.*, 2002), plume-influenced mantle being preferentially channelled along the divergent plate boundary at depths greater than 50 km and N-S oriented mantle flow at greater depths (Li and Detrick, 2003) and northwestward flow from the upwelling axis beneath Iceland (Xue and Allen, 2005). Thus all three of these studies invoke splitting observations inconsistent with radial flow of material beneath Iceland in favour of different versions of ridge channelled flow to explain the pattern of anisotropy.

THE ENIGMA OF HOW ICELAND'S STEEP MOHO TOPOGRAPHY IS SUSTAINED

The Moho descends from a depth of 11–14 km along the northernmost Reykjanes Ridge to about 20 km in southwest Iceland (Bjarnason *et al.*, 1993) and 39 km in central Iceland (Darbyshire *et al.*, 1998) over a total distance of only 250 km, implying an average dip of about 6°. Similar dips occur offshore northern Iceland where the crust thickens from 9–12 km along the southernmost Kolbeinsey Ridge to 25 km on land (Hooft *et al.*, 2006). Locally dips are even higher, with Menke's (1999) Moho depth compilation implying dips of 10–12° in a few places (e.g.

the southern end of the eastern neovolcanic zone, Figure 2). One might expect that this Moho topography would tend to collapse over time, especially given the low crustal viscosities implied by the shallow maximum depth of earthquakes (Stefansson *et al.*, 1993) and by the crust's rapid geodetic response to rifting events (Pollitz and Sacks 1996), in manner similar to the now well-documented collapse that is occurring in Tibet (Royden and Burchfiel, 1987). Such collapse, if it were occurring, might be expected to be associated with extensional faulting around the periphery of Iceland, yet no such pattern is observed.

THE PLUME DEBATE; WHAT WE SEE AND WHAT WE DON'T

Improved seismic instrumentation and advanced computing have facilitated the development of new data analysis techniques for the mantle structure beneath Iceland. These tomographic imaging efforts fall into two broad classes: whole-mantle tomography using a global distribution of earthquakes and seismometers, where the Iceland region is just one small part of the overall image; and regional tomography using a global distribution of earthquakes but seismometers confined to the Iceland region. Each technique has advantages and disadvantages. Whole-mantle tomography is better able to correct for the biasing effect of mantle heterogeneities that are outside of the Iceland region, but may have a resolution controlled by poorly-observed parts of the Earth that are far from Iceland. Regional tomography is more affected by unmodeled near-source mantle heterogeneity, and is limited to the upper mantle (typically <600 km), but usually has higher resolution at least in the upper 300–400 km.

All major whole-mantle and regional tomographic studies indicate that the mantle beneath Iceland contains a low-velocity structure that is roughly cylindrical in shape and with a sub-vertical long axis that corresponds to the plume. However, the shape of the plume differs considerably between these different studies, often in ways that have major impact on geodynamic interpretations of plume dynamics, and these differences have engendered substantial debate. Key to

unravelling these issues is understanding which plume characteristics are tightly-constrained by the tomography, and which are merely not-inconsistent with them.

Whole-mantle tomographic studies differ in their support for a deep-mantle origin and shape of the Iceland plume. Bijwaard and Spakman (1999) use whole-mantle P-wave traveltime tomography to map the low-velocities below Iceland, and infer an unstable, bent plume with a diameter ~ 500 km, rising from a broad root zone (>1000 km) near the core-mantle boundary and culminating in a plume top with a diameter of 1200 km. In contrast, Montelli *et al.* (2004), using finite-frequency tomography, argued that the plume extends at most to 1000 km depth. Also using finite-frequency sensitivity kernels to image the upper mantle structure Hung *et al.* (2004) illuminated the mantle low velocity structure to extend at least to the base of the transition zone. Additional shear-wave tomography (Montelli *et al.*, 2006) revealed a separate plume within the lower mantle (below 1450 km) suggesting a pulsating plume beneath Iceland in line with Helmberger *et al.* (1998) and Zhao (2001), petrology (O'Connor *et al.*, 2000) and geodynamical modeling (Ito, 2001). Zhao (2001) imaged an hourglass shaped upper-mantle plume beneath Iceland, narrowest at the transition zone and a lower mantle plume (1800–3000 km) deflected westwards, towards Greenland. Both Zhao (2001) and Montelli *et al.* (2006), detect a plume head within the upper-mantle being channelled along the Reykjanes Ridge and Kolbeinsey Ridges. Zhao (2004, 2007) argues for some connection between the Iceland and Jan Mayen plumes, suggesting that they are fed by the same deep-mantle source. On the other hand, no such connection is evident in Montelli *et al.*'s (2006) images.

Regional tomography has been applied to Iceland starting with an early study by Tryggvason *et al.* (1983), which was based on P wave traveltimes measured off paper seismographic recordings, but which clearly detected the low-velocities in the shallow Icelandic mantle and demonstrated that they were concentrated beneath the center of the island and extended to several hundred kilometers depth. Two notable studies using modern, digital seismometers

are those of the ICEMELT experiment (Wolfe *et al.*, 1997) and the HOTSPOT experiment (Allen, 2001; Allen *et al.* 2002a). These studies show clear compressional and shear low-velocity anomalies within the uppermost mantle, centered beneath the Vatnajökull region of E-Iceland, with the percent reduction in shear velocity being about twice that of compressional velocity. The depth extent of the upper-mantle plume, thinning of the mantle transition zone beneath Iceland (Shen *et al.*, 1998) and the tilt of the plume (Shen *et al.*, 2002) has, however, been questioned (Foulger *et al.*, 2001; Du *et al.*, 2006).

Determinations of the shape of the plume are hampered by the limited resolution of regional tomography. Menke (2005) examined the generic problem of imaging plume-like structures with regional tomography. He argues that, even under ideal circumstances the depth-distribution of velocities within a more-or-less cylindrical plume is very poorly constrained. Allen and Tromp (2005) examine the specific case of the Icelandic plume, and especially the observational geometries used by the HOTSPOT experiment and the ICEMAN velocity model (Allen 2001) that resulted from it. They verified that ICEMAN is one of a set of possible solutions, and furthermore show that all acceptable solutions must have a plume that extends to a depth of at least 350 km and must have horizontal widths between 100 and 200 km. The velocity contrast of the plume model reflects its width, with the 200 km wide plumes requiring compressional and shear wave anomalies of only -2% and -4%, respectively, but the 100 wide plumes requiring very much larger anomalies of around -10%. Thus while the evidence for a plume-like low-velocity structure in the shallow mantle beneath Iceland is very strong, its depth extent and whether it is indeed contiguous with deeper mantle structure is as yet unclear.

FUTURE TASKS

The structure of the lower crust and uppermost mantle beneath Iceland is still poorly resolved. Shear-wave splitting times of 0.1–0.3 s obtained from local earthquakes originating within the uppermost 10 km of the crust, and caused by near-surface tension cracks (Menke *et al.*, 1994) are still being used as a proxy

for whole crustal contribution to observed splitting of teleseismic events. We need to assess the magnitude of anisotropy within the lower crust, as might be caused by the effect of lower-crustal flow on pyroxene alignment, and upper mantle lid beneath Iceland as well as effects of a multi-branched plate boundary in order to obtain more detailed constraints on upper mantle flow. Furthermore, as results from experimental studies indicate that water induces fabric transitions within the olivine lattice (Jung and Karato, 2001; Kaminski 2002) and melt percentage as well as strain influences their orientation as well (Holtzman *et al.*, 2003) we need to assess the amount of water and state of stress within the uppermost mantle. A broadband ocean bottom seismic survey within the North Atlantic near Iceland is needed in order to put further constraints on various physical properties of the mantle beneath the North Atlantic, compatible with seismic and geodynamic constraints. Only through these efforts we will be able to address many fundamental questions of mantle dynamics.

We have successfully monitored the accumulation of magma in shallow crustal reservoirs and the movement of this magma along the plate boundary during periods of rifting and magmatism. However, we have not been able to identify magma feeding zones within the lower crust (although recently recorded deep earthquakes beneath the Askja central volcano and Upptyppingar are thought to be caused by melt movement (Soosalu *et al.*, 2007; Jakobsdóttir, 2008) or deeper seated batches of magma. The mechanism of melt transport through the lower crust thus remains a mystery. Is it continuous with time or are we looking at intermittent melt injections in association with episodic rifting of the plate boundary? The effects of repeated ridge jumps towards the center of the Iceland plume on crustal generation are also difficult to encompass with the currently available data. Finally, the geodynamic and geochemical interactions between the plume and the segregated plate boundary need to be studied in both space and time. Together with Iceland, the Reykjanes Ridge and the Kolbeinsey Ridge provide a promising natural laboratory for developing a comprehensive model of the dynamics of hotspot-ridge interactions.

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ÁGRIP

Uppbygging jarðskorpu Íslands

Hér eru raktar kenningar um uppruna og þróun jarðskorpunnar undir Íslandssvæðinu, sem byggðar voru á mismunandi túlkun mæligagna úr bylgjubrots- og jarðskjálftamælingum. Sagan spannar fimm áratugi. Fyrstu bylgjubrotsmælingarnar á Íslandi voru gerðar sumarið 1959 (lína BATH, 1. mynd) með það að markmiði að skýra hvort undir landinu lægju leifar af meginlandi eða hvort Íslandsskorpan væri svipuð úthafsskorpunni. Á þessum tíma höfðu dýptar- og endurkastsmælingar, ásamt hitaflæðismælingum og kjarnatöku víða um úthöfn staðfest tilvist úthafshryggjanna og gjörbylt heimsmynd jarðfræðinnar. Árið 1959 birtist einnig fyrsta kortið af upptökum jarðskjálfta á Mið-Atlantshafshryggnum og grein um bylgjuhraða í úthafsskorpunni í Norður-Atlantshafinu, m.a. sunnan og austan Íslands.

Niðurstöður línu BATH, leiddu í ljós svipaðan hljóðbylgjuhraða og í úthafsskorpunni umhverfis landið, en bentu til þess að Íslandsskorpan væri mun þykkari. Endurköst á 28 km dýpi voru talin marka botn þriggja laga jarðskorpu en úthafsskorpan er að meðaltali um 8 km þykk. Í kjölfar frekari mælinga reis ágreiningur um hvort neðsta lagið tilheyrði jarðskorpunni eða möttlinum, hvort undir Íslandi væri afbrigðilegur lággraðamöttull undir 10–15 km þykkri jarðskorpu eða hvort Íslandsskorpan, þrátt fyrir að vera úthafsskorpa, væri álfka þykk og meginlandsjarðskorpa vegna aukins uppstreymis möttulefnis. Þessi ágreiningur leystist ekki fyrr en með tilkomu stafrænna gagna, rúmlega þremur áratugum seinna.

Í dag ríkir almenn sátt um þykkt jarðskorpunnar undir Íslandi og ágreiningurinn beinist að möttlinum. Bylgjubrotsmælingar síðustu tvo áratugi sýna að skorpan er þykkust undir miðju landsins, allt að 40 km, en þynnst, um 20 km, undir gosbeltunum (2. mynd), í góðu samræmi við bergfræðileg líkön.

Pykkun jarðskorpunnar er almennt talin afleiðing aukins uppstreymis möttulefnis (möttlulstróks). Rekja má áhrif möttlulstróksins frá Íslandi eftir Íslands-Grænlands- og Íslands-Færeyjahryggnum, aftur til opnunar Atlantshafsins fyrir um 60 milljónum ára. Einnig um 1000 km suður eftir Reykjaneshrygg, og meira en 500 km norður eftir Kolbeinseyjahrygg. Takmörkuð gögn og aðferðir til greininga á uppsprettu stróksins koma þó enn í veg fyrir að hægt sé að greina neðri hluta hans, þ.e. neðan 670 km dýpis.

Ítarlegri mynd af uppbyggingu Íslands gefur til kynna að þrátt fyrir herra hitastig í neðri hluta skorpunnar og efsta hluta möttulsins undir gosbeltunum er kvika ekki til staðar í greinanlegu magni utan megineldstöðvanna. Bylgjubrotsmælingar hafa leitt í ljós grunnstæð kvikuhólf undir Kröflu (3. mynd) og Kötlu (4. mynd) ofan á háhraðastrompum (innskotum) úr neðri hluta skorpunnar. Grunnstæð kvikuhólf hafa enn fremur mikil áhrif á hitastigul jarðskorpunnar og kynda háhitasvæði landsins. Innan strompanna nær eðlisþyngri kvika til yfirborðs en utan þeirra. Breytileika í uppbyggingu efri hluta jarðskorpunnar má enn fremur rekja til þeirra gosbelta sem skorpan myndaðist í, hvort þar hafi verið rekbelti og hversu nálægt megineldstöð skorpan hafi orðið til (5. mynd).

Mikilvægum spurningum varðandi eðliseiginleika jarðskorpunnar og möttulsins er þó enn ósvarað. Afla þarf frekari gagna til þess að hægt sé að meta hitaástand neðri hluta skorpunnar, hversu hátt hlutfall kviku situr innan hennar og hvernig hún hnígur með tíma. Samtúlka þarf fyrirbyggjandi niðurstöður ýmissa jarðefna-, og jarðeðlisfræðilegra mælinga til þess auka skilning okkar á því hvernig kvikan rís frá möttlum, um skorpuna, til yfirborðs. Þróa þarf nýjar aðferðir til greininga á kvikustreymi. Ætla verður að þessum spurningum verði ekki svarað nema með samtúlkun gagna af ólíkum sviðum jarðvísinda.

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