

An Iceland hotspot saga

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Abstract — *Is Iceland a hotspot, with ridge-centered plume? In Iceland vigorous volcanism has built up a plateau 3.0 km higher than at a normal mid-ocean ridge with 3 to 4 times thicker crust than average oceanic crust. This volcanism can be associated with anomalous volcanism for 56–61 Ma in the form of aseismic ridges that stretch across the North Atlantic Ocean through Iceland, i.e. the Greenland-Iceland-Ridge (GIR) and the Faeroe-Iceland-Ridge (FIR). Iceland is a “meltspot” and an hotspot and the GIR and FIR may be hotspot trails. The trends or age progressions of the GIR and FIR are too uncertain to conclude if the Iceland hotspot can be a fixed reference point. There is a large seismic low-velocity anomaly (LVA) in the mantle under Iceland at least down to 400–450 km depth and with globally low velocities down to ~200 km depth. The center of the LVA is at 64° 40' N and 18° 10' W between the glaciers Hofsjökull and Vatnajökull. The shape of the LVA is approximately that of a cylinder in the depth range 100–450 km, but at certain depths elongated in the north-south direction. The LVA extends at least up to 30–40 km depth beneath Central Iceland and the rift zones. The shallower part of the LVA (i.e. above ~150 km depth) extends at least ~700 km outside of Iceland to the southwest, along the Reykjanes Ridge. The LVA has been numerically modelled with geodynamic methods by several authors as a ridge-centered convecting plume. They try to fit crustal thickness of the Iceland hotspot and neighbouring ridge, and the magnitude and shape of the LVA. The latest of these models find a best fit: A plume 135–150°C hotter than background mantle, retaining in general 1% partial melt in a maximum ~90 km thick melting zone, but reaching up to 2–3% partial melt in the shallowest mantle. The rest of produced melt goes into forming the crust. Considerable work has been carried out on various plume models to explain these and other observations in Iceland, but the models are still some way from reaching a mature state. As long as important observations are lacking and some key questions remain unanswered, alternatives to the plume model or more realistic variants of it in a larger tectonic framework, including heterogeneous mantle, should not be discouraged.*

INTRODUCTION

Hotspot definition

Hotspots are places with vigorous volcanism over millions or tens of millions of years. They warm up the crust in their locality. The main volcanic activity is confined to a relatively narrow area, usually under 500 km in diameter, that builds up a region approximately oval in shape. It is often proposed that hotspots may be related to vigorous vol-

canic activity over much broader regions, i.e. large igneous provinces. Hotspots place their mark on the plate with a trail of past locations of their volcanism, called a hotspot trail or track (aseismic ridge or volcanic chain). In the oceans larger hotspots build up a high plateau over their surroundings, with thicker crust. Hotspot volcanism is mostly basaltic in the oceans, but a mix of rhyolite and basalt on the continents. Hotspots are here defined as a crustal feature,

but in the literature their deeper source is sometimes also referred to as a hotspot. There is consensus that the sources of hotspots are in the mantle, but not of their actual nature or depth extent. It is even debated if hotspots are hot spots at all (Foulger and Anderson, 2005), whereas all agree that they are regions of increased melt production, "meltspots". The number of defined active hotspots has gradually risen to over 40, varying in ages from 5–200 Ma (Steinberger, 2000). Hotspot proliferation has been criticised.

The deep plume hotspot-source hypothesis

The Canadian geoscientist J. Tuzo Wilson (1908–1993) first proposed in a number of articles (e.g. Wilson, 1963) that a hotspot source is deep in the mantle and probably moves slower than the overriding plate (i.e. rheological lithosphere). He suggested that hotspot sources provide a fixed reference frame for the plates, and that trends and age progression of hotspot trails indicate the absolute motion of the plates during geological time. The American geophysicist W. J. Morgan proposed that hotspot sources are upwellings of hot buoyant rock rising all the way from the lower mantle, and called them "Convection plumes from the lower mantle" (Morgan, 1971). As the plume material reaches a shallow depth in the mantle it starts to partially melt, because of the high temperature of the plume and the positive relation between solidus temperature and pressure (i.e. lowered solidus temperature at lower pressure causing decompression melting). The melt has lower density than the surrounding mantle and percolates upwards. Part of the melt that reaches the crust rises all the way to the surface, which may supply vigorous volcanism; the rest is intruded at depth. With the assumption that the hotspot sources are fixed in the mantle, the absolute motions of the Earth's plates have been calculated from the directions and age progressions of several hotspot trails. One of the arguments for the fixity of plumes relative to the plates comes from the large difference in viscosity of the different mantle layers. The asthenosphere, a relatively thin shallow zone in the mantle (~160 km thick on average in the depth range 80–220 km), decouples the plates from the rest of the upper mantle. It is approximately 10 times less viscous than the upper mantle beneath

it and 1000 times less viscous than the lower mantle (Steffen, 2007). Hence the general movement of the mantle under the asthenosphere, and lateral movements of plumes should be slower than those of the plates. Early on it was realised that not all hotspot sources are fixed with respect to each other although in many cases they do move slower than the plates (Molnar and Atwater, 1973; Molnar and Francheteau, 1975). There are, however, hotspots which are relatively stable with respect to each other. They are Pacific hotspots called the Hawaii group, Northeast-Atlantic hotspots (the Iceland group) and African, Atlantic and Indian ocean hotspots (the Tristan group) (Pilger, 2003; mesoplates in Wikipedia). Through Cenozoic time (since 65 Ma) misfits in hotspot traces of the Hawaiian (Pacific) group and hotspots in the Tristan (Indo-Atlantic) group imply relative motion of 5–80 mm/yr between the groups, with the lower number in more recent times (Raymond *et al.*, 2000). A global hotspot-fixed reference frame with respect to the mantle is not as fixed as previously thought, and further research is needed in this field. However, it is still an important tool together with magnetic stratigraphy of the oceans for plate reconstruction (i.e. the history of the movements and changes in the size of the plates). The relative movement of the Iceland group with respect to the other two large groups is not much reported on in the literature, probably due to rather large uncertainties in the hotspot tracks within the Iceland group.

Other hotspot-source hypotheses

Anderson (2000), later joined by Foulger (2002), have led a vigorous campaign opposing the lower mantle plume source of hotspots, and even mantle plumes in general. They prefer to call hotspots "meltspots". They argue that mid-plate volcanic island chains are loci of maximum strain in the plates, and that heat in the asthenosphere can vary considerably producing shallow "pseudo-plume-like" structures without deep thermal instabilities, i.e. small scale convection movements without deep thermal sources. They also argue that the mantle is intrinsically heterogeneous (agreed by many), and hence variably fertile for melt production. Where a fertile mantle and large strain areas in plates coincide, meltspots will form. It follows

that if hotspot sources are confined to the shallow low-viscosity asthenosphere then it is unlikely that hotspots can provide a mantle-fixed reference frame.

It seems, at least in theory, that meltspots cannot escape being crustal hot spots unless they are vigorously cooled. Melt is an efficient transporting medium for bringing heat up towards the surface from the mantle. If small-scale convections in the asthenosphere exist that "... give rise to pseudo-plume-like structures...", a phenomenon that the no-plume advocates have not excluded (Anderson, 2000), they are in my opinion thermal mantle plumes, only shallow ones.

Other hypotheses for hotspot sources include meteorite impacts, that would provide additional heat and maybe different chemistry in the asthenosphere, causing hotspot melting. This hypothesis does though not work well for continental hotspots, which have too thick lithosphere for even a large meteorite to penetrate into the asthenosphere.

HISTORY AND MORPHOLOGY OF THE ICELAND HOTSPOT

From the beginning, Iceland and Hawaii have been included among major hotspots of the Earth (Wilson, 1963). The Iceland hotspot (I-hotspot) has all the main characteristics of a hotspot. Despite this it is not usual in all respects: It has an active oceanic spreading-ridge passing right through it (Figure 1). Due to this location the I-hotspot sits in the middle of its hotspot trail rather than at the end of its trail. The trail extends across the North Atlantic Ocean, from Iceland to the Faeroe Islands, the Faeroe-Iceland ridge (rise) on the Eurasian plate, and from Iceland to East Greenland, the Greenland-Iceland ridge (rise) on the North American plate (Figure 1). This interpretation makes the I-hotspot at least as old as the opening of the North Atlantic Ocean in this area, at least 56 Ma based on oceanic magnetic anomalies (Nunns *et al.*, 1982), and perhaps 5 Ma older if associated with the general opening of the Atlantic Ocean north of 55°N (Jolley and Bell, 2002). The geological definition of the current center of the I-hotspot should naturally coincide with the highest volcanic activity in the past few million years and the highest plateau in the cen-

tral Iceland rift system, beneath northwest Vatnajökull glacier (Sæmundsson, 1979). Later we will see that the locus of the hotspot source in the mantle is at 64°40'N and 18°10'W, between the glaciers Hofsjökull and Vatnajökull. The shape of the volcanic edifice that the I-hotspot has built in the last ~25 Ma is oval with approximately 500 km N-S minor and 800 km E-W major axis, or 0.3 million km² in area including the Iceland shelf. The high plateau of Iceland stands ~3.0 km above the -2.3±0.3 km average depth of hotspot-unaaffected young crust farther south in the North Atlantic Ocean (Malinverno, 1990). To assess the crustal (melt) production in the I-hotspot, it is necessary to know the average crustal thickness in Iceland. That is still being debated. Assuming neither the thickest crustal models (Darbyshire *et al.*, 1998; Allen *et al.*, 2002a; Fedorova *et al.*, 2005) nor the thinnest crustal model (Björnsson *et al.*, 2005), an intermediate 25–26 km average crustal thickness (Bjarnason and Schmeling, 2007) yields 2.4 times higher crustal production in the I-hotspot than the average of the Atlantic Ocean needed to fill up the diverging plate boundary. This comparison assumes 3.0 cm total average spreading rate and 7 km thick crust on average in the North Atlantic Ocean. With an even lower spreading rate in the northern part than in the southern part of the North Atlantic Ocean, this greater crustal production naturally piles up a thick crust, 3.6 times thicker than global average oceanic crustal thickness (White *et al.*, 1992), closer to that of continental crust (Mooney and Meissner, 1991) (Figure 2).

There are hypotheses that the I-hotspot may be older than the 56–61 Ma rifting of the North Atlantic Ocean. Lawver and Müller (1994) associate large basalt provinces in the Qeqertarsuaq (Disko) area and further north in West Greenland, dated 59.5–62 Ma (Larsen *et al.*, 1992, Saunders *et al.*, 1997), with the I-hotspot and even 120 Ma basalt on Ellesmere Island to the northwest of Greenland. The problem with their hypothesis (i.e. for 120 Ma old I-hotspot) is that there are large stretches along the trail that are either covered with ice or show no sign of hotspot trail at all. The flood basalts in West Greenland are possibly only related to the opening of a failed proto-Atlantic Ocean to the west of Greenland.

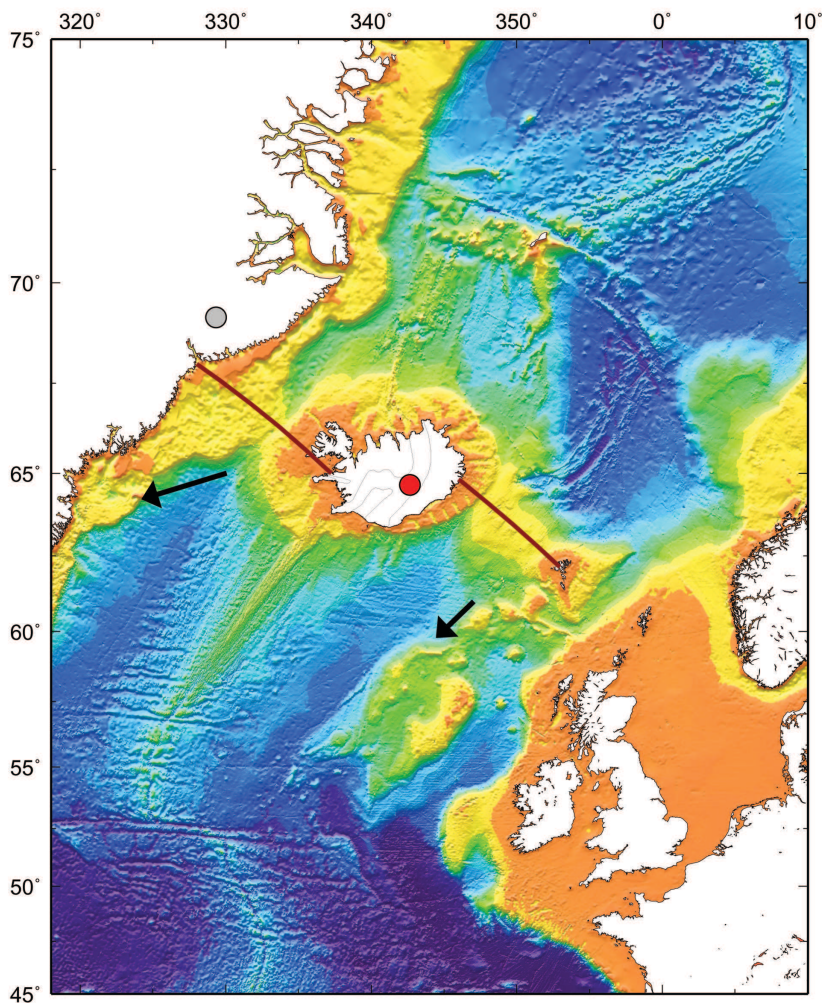


Figure 1. Bathymetry map showing the Iceland hotspot on the mid-Atlantic ridge. Red circle shows the center of the hotspot source today, and grey circle its position at 40 Ma according to Lawver and Müller (1994). The aseismic ridges between Greenland and Iceland (GIR) and the Faeroes and Iceland (FIR) are possibly hotspot tracks (red lines). Absolute plate motions of the North American and Eurasian plates are shown with arrows with velocities 26 mm/yr and 15 mm/yr respectively from the HS3-Nuvel1A model (Gripp and Gordon, 2002). The HS3-Nuvel1A fails to reproduce a hotspot trail with the trend of the GIR. – *Kort sem sýnir útbreiðslu íslenska heita reitsins á Atlantshafshryggnum. Rauður hringur sýnir hvar miðja uppsprettu reitsins er nú. Grár hringur sýnir ætlaða staðsetningu hans fyrir 40 milljónum ára. Óvirku hryggirnir, Grænlands-Íslands hryggurinn (GÍH) og Færeyja-Íslands hryggurinn (FÍH), eru hugsanlega spor (eftirstöðvar) eftir heita reitinn (rauðar línur). Rekstefna Norður-Ameríku flekans og Evrópu-Asíu flekans er sýnd með örvum með rekhraðanum 26 mm/ári og 15 mm/ári samkvæmt líkani HS3-Nuvel1A. Þetta líkan, sem gerir ráð fyrir því að heitir reitir í Kyrrahafinu séu fastir viðmiðunarpunktar jarðar, getur ekki skýrt stefnu GÍH sem spor eftir íslenska heita reitinn.*

It has been proposed that during the opening of the North Atlantic Ocean the I-hotspot source was influential in the formation of the North-Atlantic-Igneous-Province (NAIP) (White and McKenzie, 1995; Saunders et al., 1997; Storey *et al.*, 2007). The NAIP is one of the major large igneous provinces (LIP) of the world. It covers a 1.3 million km² area of flood basalts extending from eastern Canada and western Greenland to the British Isles, and up and down the Norwegian and East-Greenland shelves (Saunders *et al.*, 1997). The volcanic activity of the I-hotspot during the last ~25 Ma is not strictly part of the NAIP. It lacks several features of a LIP: It does not have the generally short formation time (a few million years) of very large volume of magmatic rocks. The largest flood basalts of the I-hotspot are rather small, e.g. the largest Holocene lava-flow is “only” ~21 km³ (minimum number) (Hjartarson, 1988), compared to sizes of LIP lava-flows that can be one or two orders of

magnitude larger. However, over tens of millions of years the I-hotspot has built its own igneous province that is approximately a quarter of the area of the NAIP.

The Iceland hotspot trails

The fixed-hotspot reference frames for global plate motions are mainly based on hotspots in the Pacific Ocean (Gripp and Gordon, 1990, 2002). The more recent model of these authors called HS3-Nuvel1A, predicts motion of the North American (NA) and Eurasian (EU) plates in the neighbourhood of the I-hotspot as shown in Figure 1. The vectors of plate motion around the I-hotspot show that both plates have a net west- and southward motion, the NA plate with higher speed. The EU plate has however higher southward velocity than the NA plate, with 6 mm/yr relative speed between them. The HS3-Nuvel1A model averages plate motion over the last ~5.8 Ma. It is an inter-

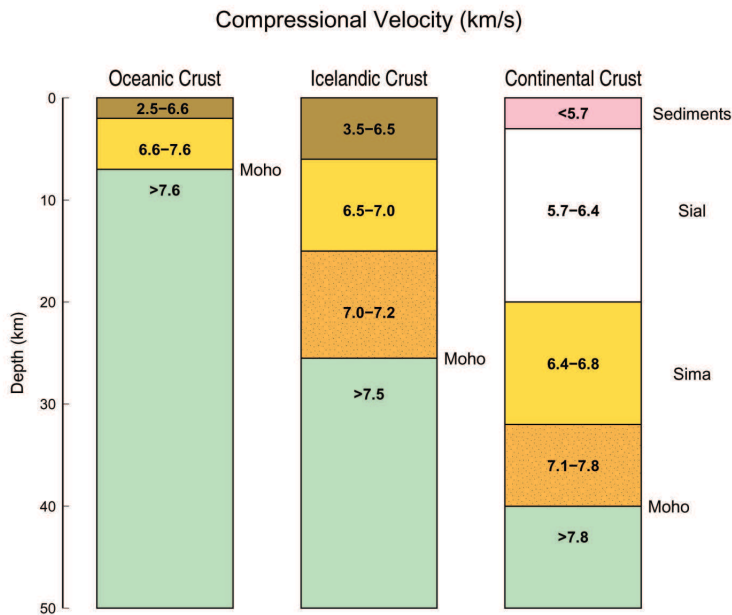


Figure 2. Iceland crustal structure shown in comparison with average oceanic and average continental crust. The 25–26 km average thickness of the crust of Iceland (Bjarnason and Schmelting, 2007) is a reduction from the thickest crustal models with ~30 km average thickness. The brown layers corresponds to the extrusive and bitrusive upper and middle crust, yellow and orange to the intrusive lower crust. –*Gerð jarðskorpu Íslands borin saman við meðaltals-skorpugerð úthafanna og meginlandanna. Hér er miðað við að meðalþykkt íslenskrar skorpu sé 25–26 km, en það er þynnra en ýmis önnur líkön gera ráð fyrir (~30 km). Brúnu lögin eru efri og miðskorpa gerð úr gosbergi og innskotsbergi. Gulu og appelsínugulu lögin eru neðri skorpa gerð úr djúpbergi.*

esting question whether the I-hotspot trails fit the predicted plate motions of the HS3-Nuvel1A model. The model may be limited in two respects for the I-hotspot region: The time it applies to is only 1/10 of the opening time of the North Atlantic Ocean, and the I-hotspot may not be fixed in the Pacific hotspot frame. The third possibility is that a hotspot-fixed reference is not appropriate for the I-hotspot, i.e. that the I-hotspot is a wandering hotspot, with highly deformed plume head at depth, influenced by the plate boundary in this region (Mihalffy *et al.*, 2008). The trend of the Faeroe-Iceland ridge (FIR) is relatively well constrained, but not the trend of the Greenland-Iceland ridge (GIR) (Figure 1). In spite of this uncertainty we can measure from Figure 1 that the FIR has drifted south by $\sim 12 \pm 3$ mm/yr with respect to the GIR during the opening of the North Atlantic Ocean, or with approximately double the rate of the HS3-Nuvel1A model. According to the HS3-Nuvel1A model both the NA and EU plates should be drifting to the south compared to a hotspot fixed reference frame.

A partial remedy for these inconsistencies would be some southward component of motion of the I-hotspot. The trend of the FIR agrees with that, but the trend of the GIR implies a northern component of motion of the NA plate relative to the I-hotspot, in opposition to the HS3-Nuvel1A model (Pacific hotspot-fixed). The distance from the I-hotspot center to similar-aged basalts on the Faeroe Islands is shorter than that to East Greenland, which qualitatively agrees with a westwards drifting plate boundary. According to the hotspot models HS2-Nuvel1 (Gripp and Gordon, 1990) or HS3-Nuvel1A this drift is 16–18 mm/yr to the west. With the I-hotspot source shifted ~ 150 km to the east of the main trend of the ridge system between 55° – 72° N, and using the Gripp and Gordon values, the hotspot source was below the large-scale trend of the MAR plate boundary 9 Ma ago (i.e. ridge segment that has not jumped to the east). If one extends the oceanic magnetic lineations (midway between anomaly 5 and 5B) southwest of Iceland through the hotspot center, then the large-scale trend of the MAR plate boundary over-rode the hotspot source 12 Ma ago. Vink (1984) proposed that by 20 Ma the I-hotspot source had already moved to the east

of the large-scale trend of the MAR plate boundary. At least this play with numbers gives the indication that the I-hotspot source has been under or very near to Iceland during its total 15 Ma sub-aerial geological history. Hardarson *et al.* (1997) propose that in the last ~ 15 Ma there have been 2 ridge relocations to east in North Iceland, and 3 ridge relocation to the east in South Iceland, with the third one still in progress (i.e. the temporally overlapping West and East Volcanic Zones). The oldest proposed sub-aerial location of the ridge prior to ~ 15 Ma, was somewhere on the current Westfjords shelf. The above discussion implies that the I-hotspot is not well fixed relative to the Pacific hotspot group, and Norton (2000) showed that the I-hotspot is not fixed to the Indo-Atlantic hotspot group either.

The proposed hotspot source has been thought to be the cause of the change in strike of the extensional features across central Iceland, a feature which can be seen throughout the on-land geological time (Sæmundsson, 1979). Vink (1984) suggested that prior to 20 Ma, when the I-hotspot source was located to the west of the plate boundary, it fed laterally the closest ridge segment on the Mid-Atlantic Ridge, with hot and melt-rich mantle creating thicker crust and plateau there, and the apparent hotspot tracks of the GIR and FIR. Morgan (1983), using magnetic stratigraphy, expressed the opinion that during the time 80–30 Ma the motion of the NA plate between Greenland and Norway is quite uncertain, a warning that the history of I-hotspot track should not be extrapolated too far into the past.

The I-hotspot source has much wider contemporary effect, which is e.g. seen in shallow residual depth and excess crustal thickness along the MAR southwest of Iceland. This effect gradually tapers out some 1400 km away from the plume center in Iceland at the Charlie-Gibbs Fracture Zone (White, 1992). The Kolbeinsey ridge extending ~ 800 km to the NNE of the I-hotspot center is also affected in terms of shallow ridge depth and excess crustal thickness (Hoofst *et al.*, 2006), and probably somewhat further north along the Mohns ridge. The effect to the north of Iceland maybe though be an interplay between two hotspot sources, that of the I-hotspot and

the smaller Jan Mayen hotspot, as proposed by Vink (1984) and modelled by Steinberger (2000).

THE ICELAND HOTSPOT SOURCE

Seismic view of the mantle under Iceland

The traditional definition of a mantle plume has been discussed above. In short, a plume is a rather narrow vertical flow of hot material from the lower mantle (Morgan, 1971). As the plume reaches the top of the upper mantle it spreads out laterally. The top of the plume beneath the lithosphere is referred to as a plume head. The deeper part is called plume conduit or stem. The effect of a mantle plume on seismic waves passing through it is to slow them down and attenuate due to the higher temperature of the plume rocks and partial melt estimated to form in the last ~ 100 km of the plume ascent.

It was discovered by means of a limited seismic network in the past that teleseismic waves are delayed (1.6 s P-wave delay) under Iceland compared to Greenland and continental Europe. This delay could only be explained with low velocity in the upper mantle, down to a couple of hundred kilometers depth (Tryggvason, 1964; Long and Mitchell, 1970). With an improved local seismic network a compressional wave low-velocity anomaly (LVA) was mapped in 3 dimensions in the upper mantle under central Iceland down to 375 km depth (Tryggvason *et al.*, 1983). Later experiments designed specially to image the upper mantle under Iceland with modern broadband seismometers in the ICEMELT (Bjarnason *et al.*, 1996) and HOTSPOT experiments (Foulger *et al.*, 2000) improved the accuracy of the mapped teleseismic P-wave velocity structure with better fit to the travel time data and added images of the S-wave velocity structure (Wolfe *et al.*, 1997; Foulger *et al.*, 2000, 2001; Allen *et al.*, 2002b). These tomographic images resolve well the structure in three dimensions between ~ 75 –400 km depth, with networks that spanned across Iceland. There are some variations in the detailed shape of the LVA between the different authors, but the predominant feature has the form of a chimney or a cylinder. The LVA does not

have a sharp lateral boundary. It is presented as a proportional velocity variation across Iceland, and has the largest velocity reduction within a region 200–300 km in diameter within the well-resolved depth range with magnitudes $\sim 2\%$ for P-wave and $\sim 4\%$ for S-wave velocities. These images are somewhat limited in representing the structure only as velocity variations rather than absolute mantle velocities. The center of the LVA fluctuates a bit laterally with depth, and there are differences in the P-wave and S-wave velocity structures, but the main LVA locus is between the glaciers Hofsjökull and Vatnajökull ($64^{\circ}40'N$ and $18^{\circ}10'W$), or ~ 50 km to the southeast of the geographical center of Iceland. At 125 km depth Wolfe *et al.* (1997) observe the LVA to approximately follow the position of the neovolcanic zones, while Foulger *et al.* (2001) see it more in the form of a cylinder. Below 250 km depth Foulger *et al.* (2001) observe the LVA with a north-southerly elongated shape. The broadening of the LVA with depth seen by Wolfe *et al.* (1997) and Allen *et al.* (2002b) may be an artifact of the ray geometry (Wolfe *et al.*, 2002).

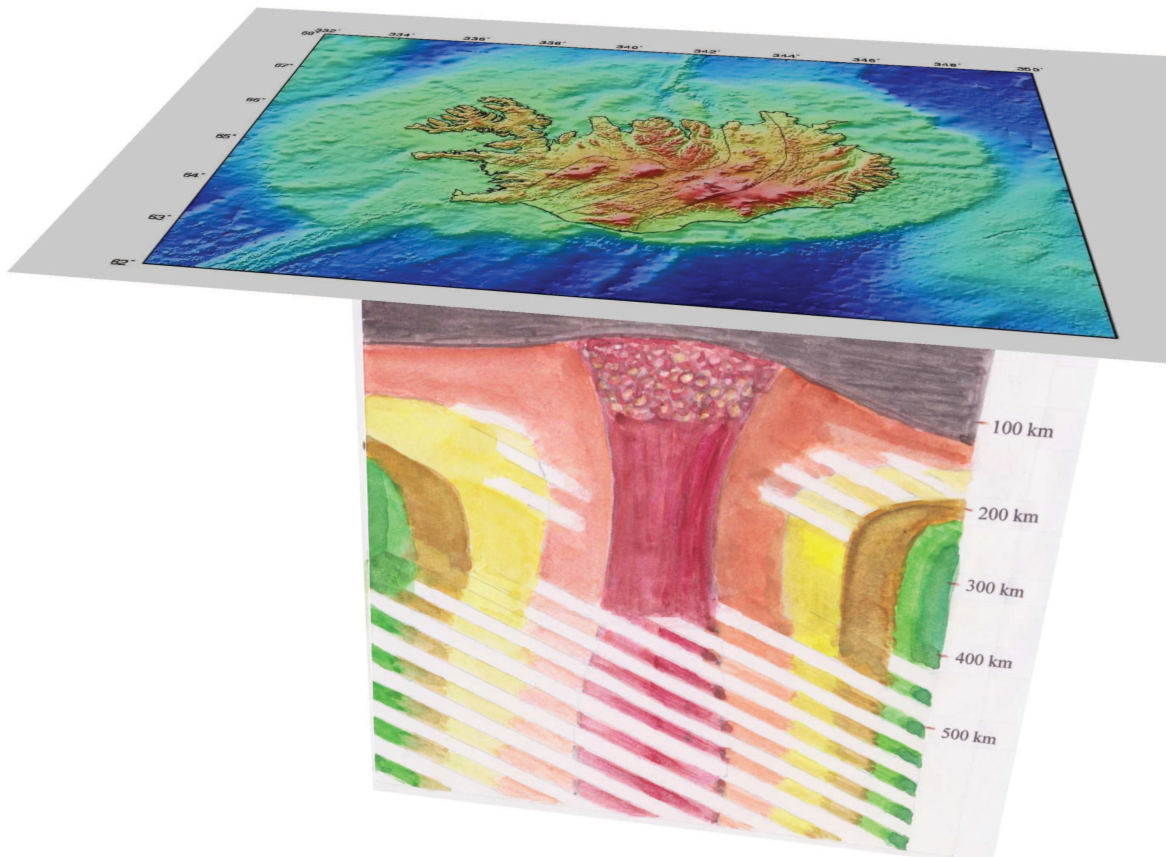
Due to the limited size of Iceland, none of these seismic experiments “see” under the bottom of the LVA, although there is disagreement on that (Keller *et al.*, 2000; Foulger *et al.*, 2001; Wolfe *et al.* 2002). Even though the picture of the LVA is not clear below 400–450 km depth, indirect evidence has been presented for the thermal anomaly, and hence the LVA, to extend through the boundary of the upper and lower mantle at 660 km depth. This evidence is in the form of thinner upper mantle “transition zone”, i.e. the distance between the velocity discontinuities at 410 km and 660 km depths, by ~ 20 km compared to the average Earth, which would happen with anomalously higher temperature (Shen *et al.*, 1998). Further analysis of an augmented dataset puts the thinned “transition zone” 100 km south of the hotspot center, indicating a tilted mantle plume (Shen *et al.*, 2002). The thinned “transition zone” is not regarded as strong enough evidence for the community to make everyone agree that the LVA extends into the lower mantle. These findings may need a better calibration on a global scale. Could e.g. local difference in mantle composition affect this thickness?

Several seismologists have commented on the presence or absence of a lower mantle extension of the LVA under Iceland. Bijwaard and Spakman (1999) interpret a bent narrow plume from the core-mantle boundary up to Iceland in form of a weak velocity reduction in their global P-velocity mantle tomography model. It is hard to agree with them that the anomaly can be interpreted as narrow (i.e. ≤ 500 km), because by inspecting their results, the anomaly is considerably larger than 500 km. Similar feature is seen by Zhao (2004). Ritsema *et al.* (1999) and Montelli *et al.* (2004) see no significant continuation of Iceland's LVA into the lower mantle to the core-mantle boundary, but both see that kind of continuation for some other hotspots in the world. The resolution of the global or regional seismic models of the lower mantle is generally not of the order of few hundred km, as needed to detect a narrow plume in the lower mantle. More targeted studies can hopefully resolve this problem, but we should expect the global models at least to agree on the broader anomalies, or to have good explanations why they do not.

Does it look like a plume?

The teleseismic body waves certainly see a structure that does look like a plume conduit in the depth range ~ 100 – 400 km (Figure 3). They do however not resolve clearly a structure that looks like a plume head. Assuming a plume head under Iceland, this lack of detection could happen if the plume head is either rather thin (few tens of kilometers) and wide, i.e. extending outside of Iceland. The plume head could also be in the top 100 km of the mantle, where the ICEMELT and HOTSPOT body wave data have less resolution in 3 dimensions. The reason for less shallow resolution of the body waves is the limited number of recordable earthquakes across the whole network within regional distance (≤ 2000 km) during the recording time of the experiments and possibly also a denser network may be needed to record crossing rays at shallow depth. The teleseismic body waves can determine the average velocity above ~ 100 km depth, and they do indicate that the largest velocity variation is in the mantle above 100–125 km depth (Wolfe *et al.*, 1997; Foulger *et al.*, 2000).

Figure 3. A cartoon of the mantle low-velocity anomaly (LVA) under Iceland, which looks like a plume. The top layer (dark brown) with variable thickness between 30 and 120 km is the lithosphere. It is double as thick under the Eastfjords and the East Iceland shelf (~ 120 km) compared to the Westfjords and their shelf (~ 60 km). The upper part of the lithosphere is the crust and its lower part is cooled mantle. Under Central Iceland and the volcanic zones the major part of the lithosphere (~ 30 km) is the crust. The LVA does not have a sharp lateral boundary. The region coloured red and orange below 120 km depth contains 2/3 of the velocity reduction across Iceland. Partial melting starts at ~ 120 km depth and extends up to the lithosphere (pink-yellow bubbles). This region is referred to as the melting zone with 1–3% retained partial melt and 135–150 °C higher temperature than background mantle. Geodynamic models favour a plume with a plume head (like the head of a mushroom), but the lower border of this feature is not well constrained with seismology. – *Teiknimynd af lághraðasvæði í möttlinum (LSM) undir Íslandi sem lítur út eins og möttulstrókur. Efsta lagið (dökkbrúnt) er steinhvel 30–120 km þykkt, skilgreint með jarðskjálftamælingum. Það er athyglisvert að steinhvelið er u.þ.b. helmingi þykkara undir Austfjörðum og landgrunni þeirra (~ 120 km) heldur en undir Vestfjörðum og landgrunni þess (~ 60 km). Efsti hluti steinhvelsins er jarðskorpa og neðri hlutinn er kældur möttull. Undir miðju Íslandi og gosbeltunum er meginhluti steinhvelsins jarðskorpa. LSM hefur ekki skörp lárétt skil en svæði litað rautt og appelsínugult neðan 120 km dýpis afmarkar 2/3 hraðalækkunar undir Ísland. Hlutbráðnun möttulsins byrjar á ~ 120 km dýpi og nær upp undir steinhvelið (gul-bleikar bólur). Þetta er svo kallað bræðslusvæði möttulsins. Það inniheldur 1–3% hlutbráð og er 135–150 °C heitara en bakgrunns möttullinn. Möttulstrókar sem reiknaðir hafa verið með affræðilegum aðferðum eru í laginu eins og hattsvæppir. Neðri hluti hattlögunar LSM hefur ekki verið nákvæmlega ákvarðaður með jarðskjálftamælingum.*



This most dramatic landscape of seismic velocity structure under Iceland has been analysed with surface waves from 100–200 km depth upward (Bjarnason and Sacks, 2002; Allen *et al.* 2002b; Li and Detrick, 2003; Bjarnason and Schmeling, 2007). These authors all use different analytical techniques, with Allen *et al.* (2002b) and Li and Detrick (2003) using either only Love or only Rayleigh waves. These are surface waves with different polarity, which can reveal quite different velocity structures in case of anisotropy. It is therefore not surprising that the results vary more than for the deeper part. Allen *et al.* (2002b) did a 1-D velocity model for the whole of Iceland, but Bjarnason and Schmeling (2007) a series of 1-D models for different cross-sections through Ice-

land. Li and Detrick (2003) mapped Rayleigh wave phase velocity variations across Iceland, which are in effect like several 2-D horizontal cross-sections through the structure at variable depth.

There are two main questions to address with the surface waves: a) What are the absolute velocities of the LVA? – The surface waves feel very low mantle velocity with peaks of velocity reduction of 10–15% in the top 50 km of the mantle compared to absolute velocities of global models (Allen *et al.* 2002b; Bjarnason and Schmeling, 2007). The vertically polarised shear wave velocity is as low as 4.03 ± 0.03 km/s at 50 km depth and even as low as 4.10 ± 0.05 km/s at 110 km depth (Bjarnason and Sacks, 2002; Allen *et al.*, 2002b; Bjarnason and

Schmeling, 2007). These are globally low seismic velocities, comparable to the lowest velocities found under the Pacific region (Nishimura and Forsyth, 1989). b) Is there a structure that resembles a plume head? – Allen *et al.* (2002b) present their model with a clear evidence of a plume head extending under all of Iceland and its shelves from approximately the crust-mantle boundary to 200 km depth. This result is obtained by subtracting a constant velocity generally seen in continental lithosphere from their 1-D average shear velocity model for Iceland. This presentation is rather artificial, because there is no sign of a 2-D or 3-D structure resembling a plume head in their surface or body wave data. A plume head is a 3-D structure. The meaning of their presentation is rather that the asthenosphere or the top part of the mantle LVA has very low velocity, which is the answer given to question a). However, surface wave measurements extending from Iceland and ~700 km offshore to the southwest, show ~600 km wide 2-D LVA zone centered on the Reykjanes Ridge, which is interpreted as a part of a plume head emanating from the Iceland plume center (Delorey *et al.*, 2007).

Li and Detrick (2003) see the LVA elongated along the rift zones and extending double the width of the rift zones to the west between 60–100 km depth, but more circular at shallower depth centered between Hofsjökull and Vatnajökull. Of these authors, Bjarnason and Schmeling (2007) may give the sharpest resolution of the lithosphere-asthenosphere boundary, and map it in 2-D from east to west across Iceland with a series of 1-D models. They define this boundary seismically, while e.g. Sigmundsson (1991) defines it as a rheological boundary, and the two may not coincide exactly. Bjarnason and Schmeling (2007) observe significant asymmetry in lithospheric thickness: The lithosphere is at least 100 km thick under East Iceland, ~60 km thick under Northwest Iceland, and 20–35 km thick under Central Iceland. Hence, if the upper part of the asthenosphere is a plume head it is quite asymmetric. Still at ~125 km depth Bjarnason and Schmeling (2007) don't see to the bottom of an hypothesised plume head under West and Northwest Iceland.

The surface wave analysis ties Iceland's mantle LVA better with the rest of the world. It has globally a low value. A plume-like conduit has been clearly mapped with body waves, but the shape of a plume head has not been fully constrained. Probably a considerably wider zone has to be imaged with seismic methods to constrain a plume head, and comparison be made with regular ridge structure for reference.

Geodynamic modelling of the Iceland hotspot and plume

Several seismologists have estimated the excess mantle temperature and some the partial melt needed to explain the mantle LVA under Iceland, giving the temperature range <200°C to 300°C (Wolfe *et al.*, 1997; Bijwaard and Spakman, 1999; Foulger *et al.*, 2001). Geodynamicists, however, are the real experts in this matter.

Iceland's LVA, crustal thickness and certain geochemical constraints have been numerically modelled with geodynamic methods by several authors as a ridge-centered upwelling plume (Ribe *et al.*, 1995; Ito *et al.*, 1996; Ruedas *et al.*, 2004; Kreutzmann *et al.*, 2004). Ruedas *et al.* (2007) give a very broad review of these studies and of many of the relative geophysical and geochemical constraints. The geodynamic models use basic laws of physics like conservation of energy and mass, rheology laws, and geochemical empirical relations to simulate a convecting plume in the upper mantle. They try to fit crustal thickness of the hotspot and of the neighbouring ridge, the magnitude and shape of the mantle LVA, and in some cases topography and gravity. All of these models produce a plume head as the shallowest feature of the plume. As has been mentioned, seismologists have not constrained the mantle plume head and the dust has not yet settled regarding the crustal thickness of Iceland. Crustal thickness is a key parameter in these models. These and other uncertainties project into the geodynamic models.

Ito *et al.* (1996) made a model for a bit narrower plume conduit (radius=60 km) than the seismologists have been observing and another model for a broader plume. Their narrow plume model has excess temperature of 170°C with melt up to 2–2.9%

in some places. Ruedas *et al.* (2004) and Kreutzmann *et al.* (2004) include all the newest results from the ICEMELT and HOTSPOT seismic experiments, and find the best plume model with 135–150°C excess temperature and maximum 1% melt retained in the plume mantle, with melting starting at ~120 km depth. There is a trade-off between excess mantle temperature and melt retained in the plume mantle in fitting the LVA, but crustal structure helps in constraining it. The models of Ruedas *et al.* (2004) and Kreutzmann *et al.* (2004) are possibly too simple in allowing only constant maximum retained melt in the mantle. Bjarnason and Schmeling (2007) have concluded that the best parameters to explain the ratio of compressional and shear velocity in the shallowest mantle at ~30 km depth are 2–3% partial melt, 1400±50°C rock temperature, and 0.3 (30%) melt depletion of the mantle (i.e. the proportion of volume removed from the mantle with melting).

CONCLUSIONS

Iceland is the most anomalous structure along the whole Atlantic Ocean Ridge in terms of high topography and crustal thickness. It is a melt- and hotspot, and has been concurrent with the opening of the North Atlantic Ocean for 56–61 Ma. The Iceland hotspot does not seem to be a fixed point in the Pacific or Indo-Atlantic hotspot reference frames. The Iceland hotspot source is clearly seen in the upper mantle at least down to 450 km depth as a seismic low-velocity anomaly (LVA). The LVA and the thick hotspot crust have been geodynamically modelled. The models explain these observables in terms of a thermal mantle plume. The more recent of these models simulate a plume with 135–150°C higher temperature than background mantle, and 1% partial melt retained in the mantle, but with maximum 2–3% partial melt in the shallowest part of the ~90 km thick melting zone. A lower mantle source of the Iceland hotspot has neither been proved or disproved. Plume-alternative hypotheses exist (Foulger *et al.*, 2005), but they have not been formulated geodynamically.

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ÁGRIP

Er Ísland heitur reitur (hotspot) sem dregur orku sína frá möttulstróki undir úthafshrygg? Þarna hefur öflug eldvirkni myndað hásléttu sem rís 3,0 km upp yfir úthafshrygginn í Norður-Atlantshafi, og jarðskorpu sem er 3–4 sinnum þykkari heldur en venjulegur úthafshryggur hefur. Þessa umfram eldvirkni Íslands má rekja 56–61 milljón ára aftur í tímann í óvirkum hryggjum sem teygja sig eftir Norður-Atlantshafinu, þ.e. í Grænlands-Íslands hryggnum (GÍH) og Færeyja-Íslands hryggnum (FÍH). Ísland er heitur reitur, svæði með óvenju mikinn kvikubúskap (meltspot), og áðurnefnda hryggi má líta á sem spor eða eftirstöðvar heita reitsins (hotspot trails). Stefna GÍH og FÍH og aldursdreifing bergs í þeim er hins vegar óviss til þess að slá megi því föstu að íslenski heiti reiturinn sé einn af föstum viðmiðunarpunktum jarðarinnar. Undir Íslandi er stórt svæði sem tefur jarðskálftabylgjur meira en gerist annars staðar vegna lághraðasvæðis í möttlinum (LSM). LSM nær a.m.k. niður á 400–450 km dýpi. Miðja þess er staðsett 64°40'N og 18°10'V, það er á milli Hofsjökuls og Vatnajökuls. Í stórum dráttum hefur LSM sívalningslöggun á dýpinu 100–450 km, en það teygist úr svæðinu í N-S stefnu sums staðar á þessu dýpi. Undir miðju Íslands og undir gosbeltunum nær LSM að komast næst yfirborði jarðar eða upp að 30–40 km. Grynri hluti LSM (þ.e. ≤150 km) nær a.m.k. 700 km út fyrir Ísland eftir Reykjaneshryggnum. Með affræðilegum reiknilíkönnum hefur verið hermt eftir áhrifum möttulstöks. Nýjustu líkönin og þau sem herma best eftir LSM og þykkt jarðskorunnar undir heita reitnum og nálægum úthafshryggjum reikna með flæði efnis sem er 135–150°C heitara en bakgrunnsmöttullinn. Í 90 km þykku (lóðrétt) bræðslusvæði stróksins er að meðaltali 1% hlutbráð. Mest nær hlutbráðin 2–3% í grynsta hluta svæðisins. Umframbráðin stígur upp og myndar jarðskorp-

una. Möttulstrókslíkön hafa þróast, en hafa ekki enn náð nægilegri fullkomnun. Á meðan ekki hefur verið lagður grunnur að öllum grundvallareiginleikum möttulstróka, ætti ekki að letja hugmyndavinnu sem sækir á önnur mið til þess að útskýra uppruna heitra reita. Möttulstrókslíkön framtíðarinnar taka væntanlega mið af stærri tectónískri heild og mismunandi efnafraeðilegri gerð möttulsins.

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