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Reply

Ingi T. Bjarnason

Department of Terrestrial Magnetism, Carnegie Institution of Washington, Washington, D. C.

William Menke

Lamont-Doherty Earth Observatory of Columbia University, Palisades, New York

Ólafur G. Flóvenz

Orkustofnun, Reykjavik, Iceland

The central question discussed by *Gudmundsson* [this issue] can be succinctly stated: "Is the temperature of the shallowest upper mantle of Iceland at the peridotite solidus" (nominally, $1200^{o}C$). The traditional view, as developed by numerous authors during the 1970s and early 1980s (reviewed by *Palmason* [1986]) and to which Gudmundsson ascribes, is that it is supersolidus and partially molten. For instance, *Palmason* [1981, 1986] uses a numerical model of crustal accretion and cooling to predict that the solidus is at a depth of 7 km beneath the volcanic zone (the Icelandic analog of a mid-ocean ridge), descending to 15 km for 5 Ma crust.

Our tentative hypothesis, as expressed by *Bjarnason et al.* [1993], is that lower crustal and shallowest upper mantle temperatures, at least in southwest Iceland, are substantially cooler, ~$300^{o}C$ below the peridotite solidus. We presented two observations that support this view:

The first observation is that compressional velocities are in the range of 7.0-7.2 km/s at depths 10-20 km, a region that we interpret as being part of the lower crust (on the basis of it being above a prominent 20-24 km deep reflector that we interpret as Moho). This interpretation differs from previous work [*Gebrande et al.*, 1980; *Hermance*, 1981; *Hersir et al.*, 1984], which assumes a considerably thinner, 10-15 km thick crust. The previous models interpreted the material with 7.0-7.2 km/s compressional velocity as being peridotite mantle at supersolidus temperature with large amounts (>10%) of partial melt present. Our new interpretation models the 7.0-7.2 km/s material as being gabbroic lower crust, without much partial melt, as the compressional wave seismic velocity of unmelted gabbro is ~7.17 km/s in the 10-20 km depth range (D. Mackensie, personal communication, 1993), close to the observed value. The compressional velocity decreases by 1-2% (0.07-0.14 km/s) per percent of melt, depending upon whether the melt is distributed in tubules [*Mavok*, 1980] or on crystal faces [*Faul et al.*, 1994]. Hence while the observed lower crustal compressional velocity might be compatible with very small (~1%) melt fractions, it is not compatible with the large (>10%) melt fraction hypothesized previously [*Gebrande et al.*, 1980; *Hermance*, 1981; *Hersir et al.*, 1984]. Indeed, these large melt fractions would cause a compressional velocity inversion in the lower crust and produce a prominent seismic shadow zone. Such a feature is not observed in our refraction lines, nor in any others of which we are aware.

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We are not the first authors to postulate a thick crust under Iceland. *Bath* [1960] gives a 27.8 km crustal thickness, and *Pavlenkova and Zverev* [1981] favor 30 km crustal thickness. A crustal thickness estimate, also in the 30 km range, is also available for the geologically related Iceland-Faero ridge (the Tertiary trail of the Iceland hotspot) [*Bott*, 1971]. However, all these estimates are based on interpretations of sparse analog data that do not show clear Moho reflections and hence must be treated cautiously.

Our second relevant observation is a relatively fast mantle refractor velocity. This refractor, associated with seismic rays that turn in the mantle directly beneath the western volcanic zone, has an apparent velocity of 7.74 ± 0.05 ($1\ \sigma$) km/s. We argued that this velocity was faster than what might be expected from a supersolidus mantle (i.e., >$1200^{o}C$), on the basis of laboratory measurements of the effect of temperature on the seismic velocity of peridotite [*Murase and Kushiro*, 1979; *Sato et al.*, 1989a].

We would like to emphasize that the crustal and uppermost mantle velocity structure presented by *Bjarnason et al.* [1993] is for southwest Iceland. It is clear from previous work that all of Iceland does not have a laterally homogeneous lower crustal structure. For example, a clear 7.7 km/s refractor is not observed across central Iceland at the cross-over distance that we observed in southwest Iceland [*Gebrande et al.*, 1980], which may indicate that the crust is thicker there. We would also like to point out that our measurements in southwest Iceland do not have the resolution to determine precisely the nature of the crust-mantle boundary directly under the western volcanic zone, as Gudmundsson assumes in his comments. The mantle reflector is observed on either side of the volcanic zone but with only few bounce points directly under it. Similarly, most of the turning points of the 7.74 ± 0.05 km/s refractor are to the east of the volcanic zone.

Gudmundsson comments on the uncertainties in measurements of the 7.74 ± 0.05 refractor velocity and argues that the true velocity might be lower than our unreversed measurement. We think that the measured apparent velocity might actually be a slight underestimate of the true velocity, as it is the downdip measurement of the refractor velocity.

Gudmundsson's main comment is concerned with showing that a slightly lower mantle refractor velocity than we observe may actually be consistent with an uppermost mantle and lower crust at solidus temperature, by developing what he considers the best chemical composition of the upper mantle under Iceland and applying an equation of state that relates composition, pressure, and temperature to compressional velocity. Our opinion is that while Gudmundsson uses perfectly plausible estimates, the level of certainty in either the composition of the Icelandic mantle or the equation of state is just not good enough for this sort of argument to be decisive. Furthermore, the relationship between temperature and compressional velocity is, as of now, just too imprecise for any conclusion to be treated as anything more than very tentative.

A review of the literature reveals that numerous other geophysical data are relevant to the question of lower crustal temperatures, with some of the data being interpreted as supporting either the near solidus or subsolidus lower crust in Iceland.

The two most decisive arguments for a near solidus (~1200°C) lower crust come from magnetotelluric (MT) measurements of crustal-mantle electrical resistivity and borehole measurements of temperature gradient.

A pervasive feature of almost all published magnetotelluric (MT) soundings for Iceland is a minimum in apparent resistivity at periods of about 100 s. This minimum implies a low electrical resistivity layer in the depth range of 10-20 km, increasing in depth away from the volcanic rift zones [*Beblo and Bjornsson*, 1978, 1980; *Hersir et al.*, 1984; *Eysteinsson and Hermance*, 1985; *Eysteinsson*, 1994]. While the data do not constrain the thickness of this layer (it trades off with resistivity), they do require it to be in the 10-20 km depth range. This low electrical resistivity zone has been interpreted as being due to partial melt, as partial melt significantly reduces the electrical resistivity of crustal and mantle rocks [*Hermance*, 1979]. The low resistivity has been associated with the 7.0-7.2 km/s seismic velocity material and used to support the interpretation that it is a partially molten mantle material. However, it is known that in some parts of continental crust similar low resistivity layers are found and interpreted as high conductivity graphite layers [*Duba et al.*, 1994] or interconnected saline pore fluids [*Hyndman et al.*, 1993]. If a similar explanation could apply to the low resistivity anomaly in Iceland, then the thick crust model will be consistent with the MT data. We do not, however, exclude that some melt exists in the lower crust within the framework of the thick crust model. Hence the MT data might still be explained by melt in the crust, if it is interconnected over large areas.

Heat flow data of Iceland have recently been reevaluated [*Flovenz and Saemundsson*, 1993]. If data from active geothermal convection systems are excluded, a systematic decrease in heat flow with distance from the rift axis is observed just as in the normal ocean ridge case. Temperature gradients, as measured in shallow (<2 km) boreholes, often exceed 100°C/km. One-dimensional extrapolation of these temperatures to depth indicate that the gabbro solidus is at depth of 10 km beneath the volcanic zones, declining to 25 km at the eastern and westerns coasts of Iceland [*Flovenz and Saemundsson*, 1993]. Analysis of possible variations of thermal properties of rock with depth rather indicates a decrease in thermal conductivity than an increase [*Seipold*, 1992]. Such a decrease in thermal conductivity would lead to an increase in temperature gradient with depth and give shallower depth to solidus temperature than would be expected from linear extrapolation of near-surface gradients. In the thin crustal model the near linear interpolation of temperature gradients gives 1200°C at the top of the low resistivity layer. The coherency of the MT and heat flow data is commonly regarded as strong support for the thin and hot crust in Iceland.

The two most decisive arguments for a subsolidus (~900°C) lower crust come from the maximum depth of earthquakes and more so from shear wave attenuation.

The maximum hypocentral depth of earthquakes is controlled by the brittle-ductile transition, which is temperature dependent. In continental ("granitic") rocks, the transition temperature is observed to be about 350°C [*Chen and Molnar*, 1983]. Data on rheology properties of gabbro at lower crustal conditions are very scarce, and we hope the rock mechanics community will supply us with more in the near future. However, we can get a tentative estimate of the transition temperature in southwest Iceland by applying recent rheology measurements on microgabbro by *Wilks and Carter* [1990]. Microearthquakes in southwest Iceland are concentrated in a 10-km-wide shear zone (called the South Iceland Seismic Zone) and show a significant increase in hypocentral depth (from approximately 7 km to 14 km) with distance from the Western Volcanic Zone [*Stefansson et al.*, 1993; *Bjarnason and Einarsson*, 1991]. Using strength measurements on microgabbro [*Wilks and Carter*, 1990] and assuming a 2 cm/yr plate velocity across the 10-km-wide South Iceland Seismic Zone, we obtain a transition temperature of 600°C. The compressional velocity at the maximum depth of earthquakes is in the range 6.9-7.0 km/s. An upper limit for the strength of gabbro can be calculated by assuming that its strength is mainly governed by the strength of the pyroxenes. A flow model of clynopyroxenite gives an upper transition temperature of 760°C [*Boland and Tullis*, 1986]. Hence, tentatively, we can say that 7.0 km/s seismic material has only reached about half to two third the solidus temperature of gabbro.

A hot, but still subsolidus material will not completely block shear waves. Nevertheless, it will substantially attenuate them, since the shear wave quality factor Q_S of gabbro at its solidus is $Q_S < 10$ [*Kampfmann and Berckhemer*, 1985]. Actually, very impulsive shear waves are observed at regional distances in Iceland [see *Stefansson et al.*, 1993, Figure 5]. *Menke and Levin* [1994a,b], using data from earthquakes occurring in the South Iceland Seismic Zone as recorded by the South Iceland Lowland (SIL) network, report a lower crustal shear wave quality factor of $Q_S \sim 300$. Shear waves from recent earthquakes in north Iceland that propagated deeper into the lower crust under central Iceland to SIL stations and broadband Carnegie station in south Iceland show a somewhat higher degree of attenuation ($Q \sim$ 100-200 ± 15) [*Bjarnason et al.*, 1994]. Over 95% of the total distance along the ray paths are within the 7.0-7.2 km/s seismic material and show normal compressional to shear velocity ratios of 1.76 ± 0.02. If the 7.0-7.2 km/s material were peridotite at supersolidus temperature (>1200°C), then we would expect very high attenuation ($Q_S \sim 10$) and very high ratios (>1.8) along these ray paths. Laboratory measurements of shear wave Q_S of gabbro indicate that $Qs \sim$ 100-200 corresponds to the 750-800 ± 50°C temperature range [*Kampfmann and Berckhemer,* 1985]. These average temperatures are probably underestimated, because the laboratory measurements were done at atmospheric pressure. Using the Q temperature and pressure relation for peridotite of *Sato et al.* [1989b] gives corresponding temperature range of 910-970 ± 33°C for $Q_S \sim$ 100-200. These temperatures are probably an overestimate as peridotite is less attenuating then gabbro at high temperatures [*Kampfmann and Berckhemer*, 1985].

We note that these path-averaged attenuation measurements do not rule out "patches" of partial melt in the lower crust. A path-averaged value of $Q_S = 100$ can also be caused by propagation through a ~10-km-wide region of partial melt, with the rest of the path being through cooler material with $Q_S = 160$.

We have studied all four of these lines of evidence, and believe all of the underlying data to be accurate. There are linear temperature gradients exceeding 100°C/km in Iceland. The apparent resistivity measurements do have a minimum that implies a low resistivity zone. Earthquake hypocenters do occur at depths to 12-14 km in southwest Iceland. Shear waves do propagate with little attenuation through the lower crust. However, obviously, the previous interpretations of some (if not all) of these data must be incorrect, since they have been used to argue for wildly different thermal models. An accurate depiction of the thermal state of Iceland must reconcile all these data, however contradictory they now seem.

We do not feel that this reply is the proper forum to begin such a discussion, which must necessarily review a substantial amount of old data. Indeed, while we hold that our hypothesis of a generally subsolidus lower crust in Iceland is at least plausible (but by no means proven), a variety of other tools, including three-dimensional seismic tomography of the deep crust and seismic surface wave imaging of the lower crust and upper mantle, will be needed before we completely understand how melt is extracted from the mantle and cools to form the Icelandic crust. Surface wave imaging will be particularly important for finding low-velocity zones in the lower crust and upper mantle, since body waves have low sensitivity to such structures.

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I. T. Bjarnason, Department of Terrestrial Magnetism, Carnegie Institution of Washington, 5241 Broad Branch Road, N.W., Washington, DC 20015.

Ó. G. Flóvenz, Orkustofnun, Grensásvegi 9, 108 Reykjavik, Iceland.

W. Menke, Lamont-Doherty Earth Observatory, Palisades, NY 10964.

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