Reinterpretation of the RRISP-77 Iceland Shear Wave Profiles

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SUMMARY

Two shear wave profiles, E and G, collected during the 1977 Reykjanes Ridge Iceland Seismic Experiment have played an important role in models of the Icelandic crust. They were originally interpreted as indicating very low shear wave velocities and abnormally low shear wave quality factors in the 10-15 km depth range. These attributes, which are indicative of near-solidus temperatures, were used to support the hypothesis that the crust of Iceland is relatively thin (10-15 km) and underlain by partially molten material. More recent seismic data, however, contradict this hypothesis and suggest that the crust is thicker (20-30 km), and cooler. A reexamination of the RRISP-77 data indicates that the low shear wave velocities are artifacts arising from source static anomalies (in the case of Profile G) and misidentification of a secondary shear phase, S_mS, as S (in the case of the E Profile). Furthermore, the attenuation occurs at ranges when rays from the shots pass near the Askja (profile E), and Katla and Oraefajokull (Profile G) volcanoes. It may therefore have a localized source, and not be diagnostic of Icelandic crust as a whole. This new interpretation of the RRISP-77 shear wave data is consistent with models having a thick, cold crust.

INTRODUCTION

The 1977 Reykjanes Ridge Iceland Seismic Experiment (RRISP-77) (Angenheister et al. 1980) was a long-range seismic refraction experiment that probed the structure of the crust and upper mantle of the Iceland hotspot. It was historically quite important, because it was the first experiment to define the transition from the normal-thickness, submarine oceanic crust of the Reykjanes Ridge south of Iceland to the anomalously thick, subareal crust of Iceland itself.

The first comprehensive model of the structure of the Iceland hotspot that emerged in the late 1970's and early 1980's was that of a relatively thin, 10 to 15 km thick, oceanic crust overlying a very hot and partially molten mantle. The RRISP-77 dataset, especially as interpreted by Angenheister et al. (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 hotspot. This model 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 (Saemundsson 1979), geothermal (Palmason 1971, 1973, 1986; Palmason and Saemundsson, 1974), and magnetotelluric (Beblo and Bjornsson 1980, Eysteinsson and Hermance 1985) data were use to explore how the processes of plate-tectonic spreading and hotspot magmatism interacted. We cannot now review this model in its entirety, but one element of special relevance to this paper was the idea that a partially molten layer existed at the base of the crust.

Even early on, this 'hot crust' model of Iceland has had it critics. In particular, Zverev et al. (1976) and Pavlenkova and Zverev (1981) argued for a muck thicker (30 km) and cooler (<600°C) crust, on the basis of the 1972 NASP long-range refraction profile. More recent seismic data have corroborated this idea. For instance, Bjarnason et al. (1993, 1994) identified P_mP reflections from a 20-24 km deep Moho in SW-Iceland. Mid- to lower crustal velocities are in the 6.5-7.2 km/s range, which they took as typical of normal oceanic crust. Menke and Levin (1994) and Menke et al. (1995) measured lower crustal shear wave quality factors in the 100-2000 range, which they argued to be diagnostic of temperatures at least 200-300°C below solidus.

The RRISP-77 shear wave profiles were used to support the 'hot crust' idea, since they were interpreted as indicating anomalously low shear wave velocities and anomalously high shear wave attenuation in the lower crust. Our purpose here is to review these 1977 data, and to establish whether or not they permit cold lower crust. Much new knowledge, not available in 1977, can now be profitably applied to this endeavor.

Gebrande et al. (1980) describe two shear wave profiles: Profile 1 through central Iceland, from shotpoint E in the sea off the northeast coast; and Profile II along the southeast coast, from shotpoint G off the south coast (Fig. 1). Two shots, separated by about 5 km, are detonated at each shotpoint, and are recorded on an array of 'mostly horizontal' geophones. Each array is about 350 km long, with a mean station spacing of 7 km. Two record sections are constructed, one for each shotpoint, by combining the pairs of shots (Gebrande et al. 1980, their Figs. 7 and 8; also our Fig. 2). Clear shear waves are evident on the G record section out to a range of about 165 km and on the E record section out to 250 km.

The ratio of the compressional to shear wave velocities, V_p/V_s, is diagnostic of temperature, and is believed to vary from about 1.76 for unaltered, cold rocks to >2.0 for partially molten rocks. Gebrande et al. (1980) use S and P traveltime ratios, T_s/T_p, as a proxy for V_p/V_s. They measure P wave traveltimes and then compare the actual S wave arrival time to one predicted from the P wave traveltimes and T_s/T_p=1.76 (dotted lines in Fig. 2). They argue that the S wave arrives later than expected both on the G profile and for ranges greater than 140 km on the E profile. A Wadati plot (their Fig. 9) of the E profile traveltimes gives a mean ratio of V_p/V_s=1.96. Furthermore, the apparent delay and attenuation of the S wave at at ranges >140 km (especially on the E profile) is interpreted as meaning that at these ranges the S wave has dived deep enough to have encountered a highly attenuating (= hot) region in the lower crust.

We organize our reevaluation of this shear wave data into three questions:

1) Is there evidence for abnormally high V_p/V_s at short (<150 km) ranges? This is equivalent to asking why the S wave on the G profile is abnormally late, since the corresponding phase on the E profile is normal.

2) Is there evidence for abnormally high V_p/V_s at long (>150 km) ranges?

Why is the S wave on the E profile late at these ranges?

3) Is there evidence for lower crustal attenuation? Why are S waves apparently delayed and amplitudes decrease at ranges greater than about 140 km?

SHEAR WAVE PROPAGATION AT RANGES LESS THAN 150 KM

For ranges <150 km, the G profile is the only Icelandic example of unusually low-velocity S waves known to these authors (excepting some small regions of intense hydrothermal alteration, such as those described by Foulger et al. 1995). S wave propagation in southwest Iceland, as determined from 3-component digital data from the South Iceland Lowland (SIL) array (Stefansson et al. 1993), is normal. Menke et al. (1994) present P and S wave traveltimes determined from SIL micro-earthquakes locations. The T_s/T_p ratio is 1.74-1.78 for ranges out to 150 km, similar to behavior of the E profile.

We have also examined horizontal records sections from the South Iceland Seismic Tomography (SIST) profile, which crosses the mid-Atlantic plate boundary in southwestern Iceland (Bjarnason et al. 1993). Clear S waves are observed from the AK and JL shots at the extreme western and eastern ends of the array, respectively (Fig. 3). The traveltime of these S waves follow T_s=1.79 T_p with remarkable consistency, to at least 120 km range. Both the P and the S wave have the same small fluctuations in traveltime, due to lateral heterogeneties in structure.

A close examination of the G profile reveals that the S wave is 'late' (arriving after T_s=1.76 T_p) but the time delay between the predicted and actual S arrivals does not increase with range. It remains a constant 1s. Although delayed, the S wave has a normal apparent velocity, consistent with V_p/V_s=1.76. This behavior is more diagnostic of a source static anomaly (i.e. an unmodeled near-source structure which has delayed all S waves by an equal amount), or possibly to a combination of shot location and timing errors, than to an anomalously low shear wave velocity.

The corresponding P wave record section for the G profile (Fig. 4) contains features diagnostic of a source static anomaly. The seismograms from the two component shots are 0 offset by about 0.5 s from one another. Furthermore, P wave traveltime reported for stations of the permanent Icelandic network (Einarsson 1979) that are nearby Profile II are systematically advanced by 0.5 s with respect to the RRISP-77 readings. The origin of these static offsets have not been yet determined, and the issue may be difficult to resolve at this late date. Nevertheless, we feel that the anomalous behavior of the S wave on the G profile is best explained by a static anomalies and not by low shear velocities.

SHEAR WAVE PROPAGATION AT RANGES GREATER THAN 150 KM

At these large ranges, S wave propagation will be effected by the finite thickness of the crust. If the velocity gradient at the Moho is high enough, the S traveltime curve will contain a triplication, where the crustal S wave, the Moho-reflected S_mS phase, and the mantle-refracted S_n phase all arrive within a few seconds of one another.

Such triplications were discounted by Angenheister et al. (1980) and Gebrande et al. (1980), who developed velocity models with a very smooth crust-mantle transition. However, the higher resolution SIST profile recorded clear P_mP and P_n from a 20-24 km deep Moho in southwestern Iceland (Bjarnason et al. 1993). These authors do not discuss the corresponding shear phases S_mS and S_n, but such phases would likely be present, too. Indeed, a close examination of the SIST and RRISP profiles (Figs. 2 and 3) indicates that secondary shear arrivals occur in the 100-250 km distance range on these profiles (they are clearest on the SIST JL-shot and RRISP E-shot profiles).

Identification of these later arriving S phases is hampered by very limited information on the crustal thickness in central Iceland. Unfortunately, the RRISP-77 profiles do not contain clear evidence for a Moho, possibly because of the limited bandwidth of the analog recording system. However, Angenheister et al., (1980), noting discontinuous second arrivals from the RRISP shot D, mention an alternative interpretation of the data, i.e., that the second P arrival, which has an apparent velocity of 7.8 km/s (Gebrande et al., 1980), may indicate a crustal thickness of 30 km under Iceland. In our opinion, the later arrival from shot D is a Moho-reflected phase (P_mP) analogous to the phase Bjarnason et al., [1993] found in SW-Iceland and is evidence that the crustal thickness is close to 30 km under the Neovolcanic zone, in central Iceland, considerably thicker than the 22-24 km value Bjarnason et al. (1993) obtained.

Fortunately, we have discovered that seismic data are available that can corroborate this estimate. In 1991 the U.S. Geological Survey and U. Durham operated an array around the Hengill volcano in southwestern Iceland (Foulger et al. 1995). While designed to monitor local seismicity, this 33-station array also recorded several small earthquakes in central Iceland. These recordings, which span the 100-130 and 170-200 distance ranges, clearly show the phases P, P_mP, S and S_mS (Fig. 5) and P, P_n and S (Fig. 6). The moveout of the compressional wave phases (together with Bjarnason et al.'s (1993) upper-crustal velocity model) allows the crustal thickness to be estimated at 30 km (Fig. 7), and is consistent with a thicker crust in central Iceland.

We have evaluated a suite of crustal models, each based on the Bjarnason et al. (1993) compressional velocity structure and V_p/V_s=1.76, but with crustal thicknesses ranging from 25 to 40 km. The 35 km crust version fits the E profile record section reasonably well, with the 'on time' arrivals in the 0-150 km distance range corresponding to S and the 'late' arrivals in the 150-250 km range corresponding to S_mS, (Fig. 8). This fit could possibly be improved somewhat, by allowing for Moho dip and upper crustal heterogeneties. However, the very limited available data does not warrant such model complexity. The simple model is sufficient to demonstrate that the 'late' S arrivals are actually S_mS waves traveling along a longer path at normal speeds.

LOSS OF AMPLITUDE OF THE S WAVE

The amplitude of the S wave gradually decreases with range on the G profile, with the last clear first arrival being at about 165 km range. Similarly, its amplitude on the E profile is very much smaller than S_mS in the 140-250 km range. This behavior was originally ascribed to a zone of high attenuation in the lower crust (Gebrande et al. 1980). This explanation cannot be reconciled with the presence of S_mS, since that phase traverses the lower crust while retaining significant amplitude.

The actual amount of attenuation of the S wave is difficult to assess quantitatively from the RRISP-77 profiles. Gebrande et al.'s (1980) record sections appear to be 'trace-normalized' plots, which will tend to suppress the apparent amplitude of S in any distance range where it is not the largest phase. Ray theoretical calculations indicate that S_mS has a higher amplitude than S in the 100-200 km distance range, by a factor of 2-3, owing to the weak velocity gradient in the lower crust. Hence at least some of the loss in S wave amplitude may be only apparent.

Another factor may be local heterogeneties in vicinity of central volcanoes crossed by the profiles. The S wave disappears on the E profile when that profile crosses the southeastern margin of the Askja central volcano in NE-Iceland. Similarly, the S wave disappears on the G profile along raypaths that interact with the Katla and Oraefajokull central volcanoes in S-Iceland. Hence it is likely that conditions local to those volcanoes, and not regional lower-crustal attenuation, are responsible for the loss of S amplitude. The loss of amplitude on the E profile may be directly related to a shallow zone of attenuation associated with the Askja central volcano. The attenuation begins when that profile crosses the flank of the volcano, which is known to have a shallow magma chamber (Brandsdottir et al. 1992). Most ray paths probably do not intersect the magma chamber itself, but rather are affected by a wider zone of attenuation caused by the thermal anomaly that surrounds the molten zone. The relatively higher amplitude of the S_mS phase is consistent with this explanation. It traverses the region around the volcano at a more oblique angle of incidence and hence experiences less attenuation. Rays from the G profile also cross central volcanoes. They plunge deep (10-15 km) beneath the Katla central volcano in the 80-165 distance range, and cut across the shallow part of Oraefajokull for ranges greater than about 200 km. The effect of the shallow Katla magma chamber is probably minimal, as the rays are considerably deeper than depths reported for such magma chambers in Iceland (Gudmundsson et al. 1993, Einarsson 1978, Brandsdottir et al. 1995). The gradual loss of amplitudes in the 80-165 range may simply imply a low mid-crustal shear velocity gradient in the vicinity of this volcano. Oraefajokull has not been imaged seismically, but its large size, evolved rocks and recent activity make the hypothesis of a magma chamber in the crust plausible.

We have estimated the lower crustal shear wave quality factor, using S_mS/S spectral ratios measured from the Hengill array. Three seismograms of earthquake 1 (Fig. 5), all located far from the center of Hengill volcano, and each having clear S arrivals, were selected for spectral analysis. A slight decrease in spectral ratio with frequency is noted, corresponding to a 30-50% difference in quality factor (Fig. 9). This decrease is consistent with the one noted by Menke et al. (1995) for southwestern Iceland, where the shear wave quality factor decreases from 1000-2000 in the mid crust to 800 in the lower crust. The decrease may be related to an increase in temperature with depth. However, the relatively high quality factors, even in the lowermost crust, indicate that the temperature are well below the solidus even in the lowermost crust.

Differential attenuation measurements cannot be made for the RRISP-77 E profile, since the requisite digital data are not available. However, the visual appearance of the S_mS phase on the E record section (Fig. 8) is similar to its appearance on the Hengill array (Fig. 5).

SUMMARY AND CONCLUSIONS

Contrary to what was originally thought, the RRISP-77 Iceland shear wave profiles do not support the hypothesis of a layer of partial melt 10-15 km beneath Iceland. The original interpretation placed low shear wave velocities and very low shear wave quality factors in the 10-15 km depth range, and hence was consistent with a thin (10-15 km) crust underlain by partially molten material. In our reinterpretation the attenuation is related to shallow volcanism and the delay of the S wave is only apparent, being caused by a source static anomaly (on Profile G) and by misidentification of S_mS as S (on Profile E). In both cases the traveltimes are well fit by a compressional-to-shear-wave velocity ratio of about 1.76, indicative of normal shear wave velocities. This interpretation, which places the Moho at about 35 km depth, strengthens the argument of Bjarnason et al. (1993) that weak Moho-reflections are visible on some RRISP-77 profiles. Hence our reinterpretation also strengthens the case for a distinct crust-mantle transition beneath central Iceland. It also supports the idea that the crust thickens from 20-24 km in southwestern Iceland, to 30 km in central Iceland, to 35 km near the east coast.

Two existing, but more modern, datasets are compared to the RRISP-77 data.

S wave profiles from the SIST profile (Bjarnason et al. 1993) have normal shear wave traveltimes (i.e. V_p/V_s=1.79) out to at least 120 km range. This behavior is similar to what is observed on RRISP-77 profile E, and on profile G after the static anomaly is removed. Earthquake data from the 1991 Hengill array are used to construct compressional and shear wave record sections, which show unambiguous P_mP and S_mS Moho reflections, as well as P, P_n and S. The appearance of the S_mS phase on the Hengill array is qualitatively similar to the one on profile E. Differential attenuation measurements support Menke et al.'s (1995) estimate of lower crustal shear wave quality factors in the 800 range.

The RRISP-77 shear wave data have been shown to be consistent with the idea that the crust of central Iceland is both thick (30-35 km) and cool (i.e. subsolidus).

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Figure Captions

Fig. 1. Map of Iceland, showing location of RRISP-77 shots E and G (triangles), RRISP-77 Profiles I and II (dotten lines), SIST profile (solid line), Hengill array (circle), and two earthquakes recorded by the Hengill array (numbered squares). The RRISP-77 E profile passes close to the Askja central volcano, and the G profile passes close to the Katla and Oraefajokull central volcanoes. Map after Angenheister et al. (1980).

Fig. 2. RRISP-77 shear wave data, after Gebrande et al. (1980). (Top) E shot observed on Profile I (Bottom) G shot observed on Profile II. The dotted line is the S wave arrival time predicted from the P wave traveltime and V_p/V_s=1.76. Note the delay in the actual S wave for ranges >140 km on the E profile and for the entire G profile.

Fig. 3. SIST shear wave data (courtesy IRIS Data Management Center). The solid line is the S wave arrival time predicted from the P wave traveltime and V_p/V_s=1.79. The position of the Western Volcanic Zone (a branch of the mid-Atlantic plate boundary in southwestern Iceland) is shown with a bar. Note the close agreement of actual and predicted S wave arrival times.

Fig. 4. RRISP-77 G profile compressional wave record section, after Gebrande et al. (1980). Note offset between neighboring seismograms, which are from two different shots at the same shotpoint. Circles show arrival times at neighboring Icelandic permanent stations (Einarsson 1979), which are systematically advanced with respect to RRISP-77 arrival times.

Fig. 5. Vertical (bottom) and horizontal (top) component record sections from

earthquake 1 (Archival file E220/185318) observed on the Hengill array (data courtesy IRIS Data Management Center). Traveltimes and distances are based on the SIL catalog location. Note the clear P, P_mP, S and S_mS arrivals. The phase velocity of the P wave is 6.75 +/- 0.2 km/s. Apparent velocities for all phases are consistent with V_p/V_s=1.76-1.77.

Fig. 6. Vertical (bottom) and horizontal (top) component record sections from earthquake 2 (Archival file E251/015238) observed on the Hengill array (data courtesy IRIS Data Management Center). Traveltimes and distances are based on the SIL catalog location. Note the clear P_n, P, and S. The S_n phase is not detected, possibly because of the large amplitude P coda. The phase velocity of the P_n phase is 8.01 +/- 0.15 km/s.

Fig. 7. (Solid curve) Compressional wave traveltime computed for a vertically stratified earth with the upper crustal structure of Bjarnason et al. (1993) and a 30 km thick crust. Vertical bar. Differential arrival times for the data in Figs. 5 and 6. Absolute traveltimes are not used, because the earthquake origin times are not known to sufficient precision.

Fig. 8. (Top) RRISP-77 E profile, after Gebrande et al. (1980). The dotted line is the S wave arrival time predicted from the P wave traveltime and V_p/V_s=1.76. The solid line is the S wave traveltime predicted for a 35 km thick crust. Note that the late arriving phase in the 150-200 km range has a traveltime matching the late arriving phase of the triplication (i.e. S_mS). (Bottom) Trace-normalized plot of radial-horizontal component synthetic seismogram, computed using a full-waveform code. Compare with observed seismogram.

Fig. 9. S_mS:S spectral ratio for the three clearest seismograms in Fig. 5. The least-squares line (solid), with a slope of -0.06 +/- 0.02, indicates that the S_mS phase is slightly more attenuated than S.