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Seismic Structure of Iceland Along RRISP-Profile I

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Abstract. As part of the RRISP 77 combined land-sea refraction seismic experiment, observations were carried out on Iceland itself with special emphasis on resolving the deep structure beneath Iceland and its transition towards the eastern flank of Reykjanes Ridge. The data, interpretational procedures, and results for the land part are described in this paper. A structural model of Iceland is presented which is characterized by a generalized two-layered crust of variable thickness underlain by anomalous mantle with *P*-wave velocities of 7.0 km/s at the base of the crust increasing to 7.4 km/s at 30 km depth. Two regions of relatively low velocity have been identified in the lower crust, possibly indicating zones of high melt concentration. A normal *P*- to *S*-wave velocity ratio of 1.76 is found within the crust, whereas this ratio reaches unusually high values of up to 2.2 in the anomalous mantle. From this and the *P*-wave velocity distribution the amount of partial melt is calculated. The melt content is highest (17%–23%) at the top of the mantle and decreases with increasing depth indicating differentiation processes in the upper mantle. The anomalous mantle is confined to Iceland and a sharp transition exists in the area of the shelf edge where normal oceanic lithosphere replaces the updoming asthenosphere.

Key words: Iceland – Reykjanes Ridge – Deep seismic sounding – Crust – Lithosphere – Asthenosphere – Anomalous mantle – Partial fusion.

Introduction

This is paper 2 of a set of three papers on the Reykjanes Ridge Iceland Seismic Project 1977 (RRISP 77) the general objectives and execution of which have been discussed in paper 1 (RRISP Working Group 1980). The present paper focusses on the deep structure of Iceland and its transition towards the flank of Reykjanes Ridge, while paper 3 (Goldflam et al., 1980) is concerned with the latter itself.

The crustal structure of Iceland has been investigated in considerable detail by a large number of seismic refraction lines which were summarized and homogeneously interpreted by Pálmason (1971, see also Pálmason and Saemundsson, 1974, for a summary). A characteristic layering has been found, resembling the oceanic crust in velocity values, but with greater thickness of individual layers. A surface layer of variable velocity and thickness has been subdivided by Pálmason (1971) into three sub-layers 0, 1, 2, but can also be interpreted as a single layer with *P*-wave velocities changing continuously with depth (Flóvenz, 1980). It is underlain everywhere beneath Iceland by a rather homoseneous layer 3 with a mean *P*-wave velocity of about 6.5 km/s which may be equated with the typical oceanic layer. The depth to layer 3 has been found to be quite variable, usually in the range of 1–5 km but up to 10 km in the southeastern part of Iceland (Pálmason, 1971). Its thickness, as far as known, is usually 4 to 5 km. A simple relationship to the gross geolocical structures (see Fig. 1 of paper 1) is not evident.

The base of layer 3 has been reached only by few of Pálmason's profiles, by one profile of Båth (1960) in the western part of Iceland, and by the NASP observations in northeastern Iceland (Zverev et al., 1976; Bott and Gunnarsson, 1980). Velocities from 7.2 to 7.4 km/s have been attributed to an anomalous mantle by Pálmason (1971) but to the lower crust by Zverev et al. (1976). Combining gravity and refraction seismic data, Zverev et al. (1976) concluded that Iceland may be underlain by a very thick crust of continental affinity and normal upper mantle (with *P*-wave velocities of 8 km/s) at about 50 km depth. A sialic crust underlying Iceland has also been postulated by van Bemmelen (1972) because of the abundance of acidic volcanism and the elevation relative to the Mid-Atlantic Ridge. Beloussov and Milanovsky (1976) have used these and other arguments as evidence against any significant amount of sea-floor spreading in the North Atlantic at the latitude of Iceland, but this reasoning seems not very convincing. Nevertheless, the possible existence of continental fragments beneath Iceland would imply serious complications for the kinematics of sea-floor spreading in the North-Atlantic and earlier paleogeographic reconstructions (Bullard, 1965; Laughton, 1971) would at least have to be modified.

A normal mantle at 50 km depth below Iceland as suggested by Zverev et al. (1976) is in contrast to some seismological observations. Teleseismic travel-time residuals (Tryggvasson, 1964; Long and Mitchell, 1970) as well as apparent velocities from Mid-Atlantic Ridge earthquakes across Iceland (Francis, 1969) have been interpreted as evidence for an anomalous mantle extending to some 240 km depth. On the other hand, Stefánsson (1966) claims that travel-time delays from a large earthquake in Iceland can be better explained with a rather shallow 7.4 km/s layer and a 8.0 to 8.2 km/s layer underneath.

These rather inconsistent results were the incentive for the RRISP experiment. Since the crust was studied in some detail before (Pálmason, 1971), the experiment was mainly designed for the investigation of the so-called anomalous mantle. Large penetration of seismic rays and therefore a long range seismic profile was necessary for this purpose.

As is to be seen from Fig. 1 of paper 1, the line follows for the greater part the trend of the eastern zone of active rifting

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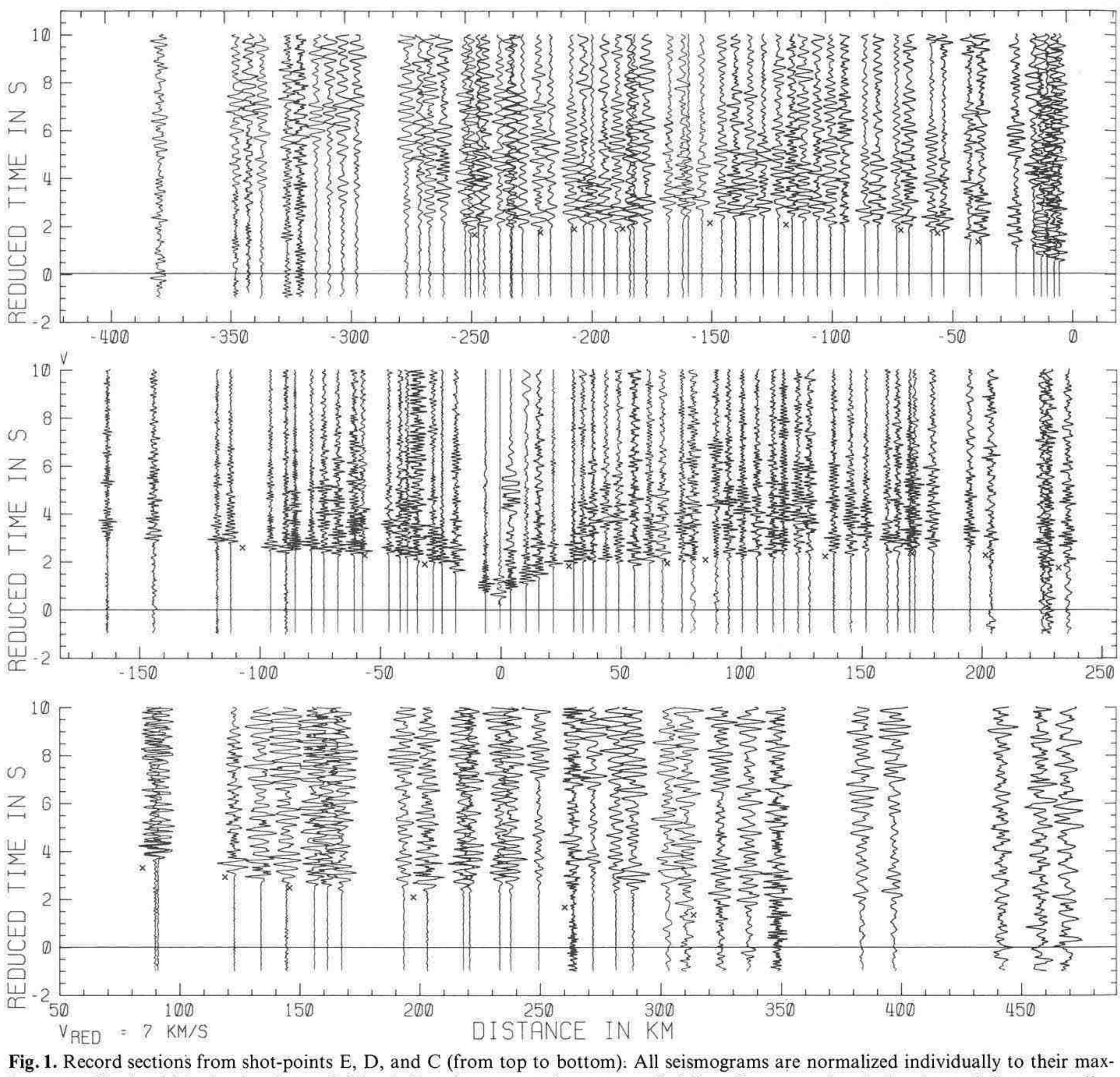

Fig. 1. Record sections from shot-points E, D, and C (from top to bottom). All seismograms are normalized individually to their maximum amplitude within the time interval. Identical stations are on the same vertical line. *Crosses* mark calculated travel-times according to the ray tracing calculations of Fig. 6. Note the delay of the first arrival of the leftmost record of shot C, which is related to a region of very low velocity within the lower crust beneath Heimaey

and volcanism, which for simplicity is called thc neovolcanic zone. The reasons for which we chose to observe along this line were the following:

1. This zone is the continuation of the Mid-Atlantic plate boundary, which thus can be studied by land meassurements.
2. Refraction measurements along the strike of geologic units generally provide more reliable velocity information than observations in other directions.
3. Since the eastern neovolcanic zone is offset from the mid-Atlantic spreading axis by the Reykjanes transform fault zone, the combined land-sea experiment enabled us to shoot from the normal oceanic realm into the active spreading zone along its trend. It was felt that by this scheme of observations the effects of absorption could be kept at a minimum because only the land part of the line would lie above possible anomalous mantle. Ray tracing calculations had also shown that the different structural models discussed above would yield significant differences in the characteristics of travel -time curves.

Observations and Interpretation

As the technical part of the experiment has been described to some extent in paper 1 and in RRISP Working Group/Angenhei-

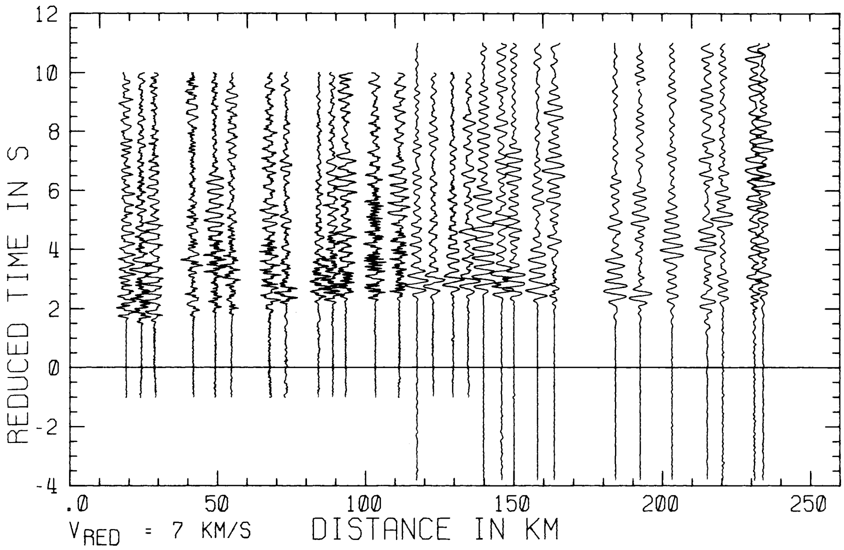

Fig. 2. Record section from shot-point F towards the southeast into the Tertiary basalts. Records starting at a reduced time of −1 s were obtained by the Soviet group, the others by the German group

ster et al. (1979), we can confine ourselves to details of observational results and their evaluation. The record section for shots F (Axarfjördur) through the eastern neovolcanic zone has already been shown in Fig. 4 of paper 1 as a typical example of RRISP data. Figure 1 shows the record section of shot E (Vopnafjördur), D (Thveraldavatn), and C (SW of Surtsey). Reduction velocity is 7 km/s and the sections are arranged in such a way, that seismograms obtained at the same location appear one above the other. Slight deviations from this are caused by the observation taken on a not strictly straight line. One notes that the apparent velocities of the first arrivals are close to and slightly over 7.0 km/s in all sections at distances beyond some 120 km. Whereas for shots D and E, apparent velocities generally increase with distance, they decrease with distance for shot C, which is of particular interest as it suggests the existence of lateral velocity variations. In Fig. 2, the record section for the line from shot-point F towards the southeast coast of Iceland is given. It is very similar in appearance to the record sections of shot-points E and F along the main line, even though the second half lies fully within the Quaternary and Tertiary basalts of East-Iceland, i.e., outside the neovolcanic zone.

In all record sections significant differences can be seen in the first 50 to 100 km, which must be attributed to variations in the crustal layers. The subcrustal structure seems to have no regional differences along the land lines as indicated by the similar apparent velocities observed at distances greater than 100 km. The seismogram sections from Iceland are characterized by the lack of distinct later arrivals. This is in contrast to continental areas, where frequently observed clear later arrivals can be interpreted as overcritical reflections from discontinuities or as diving waves bottoming in zones of strong velocity gradient. A more or less continuously varying velocity distribution is thereby manifested for the Icelandic structure. On the other hand the first onsets, especially at greater observational distance are not impulsive, but rather emergent and there is considerable energy in the later parts of the seismograms. This may be caused by scattered waves from small-scale heterogeneities. Therefore, the small-scale structure may be highly heterogeneous, whereas the deep structure can be characterized by a continuous velocity function when averaged over some wavelengths. With the mean station spacing of

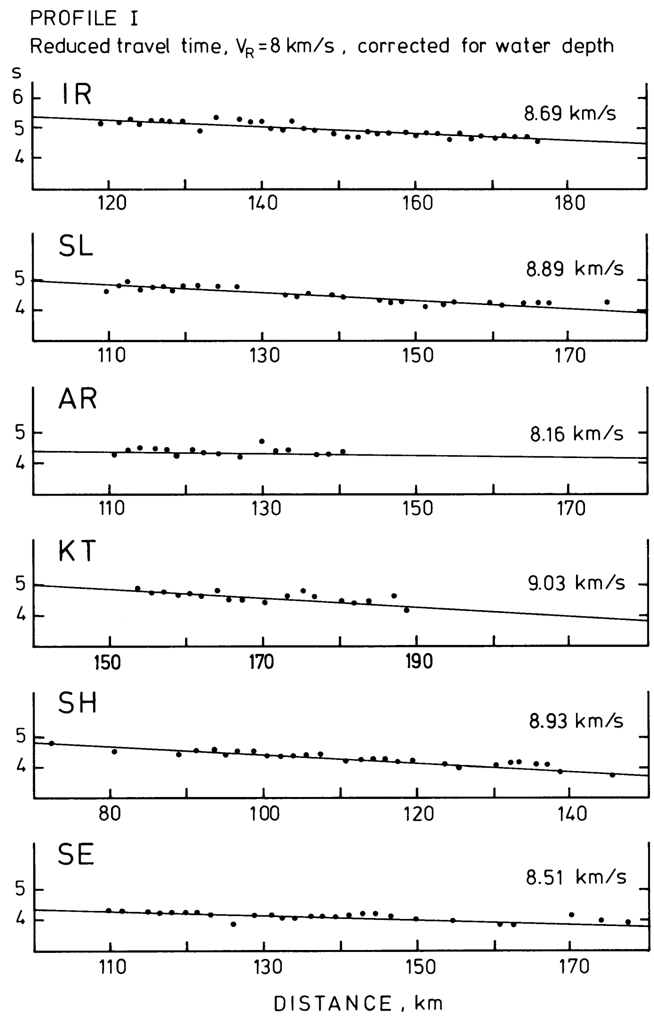

Fig. 3. Travel times from shots on Profile I to stations of the Icelandic seismological network. Travel times are corrected to the ocean bottom. Lines are least-squares fit and numbers give the apparent velocity in km/s

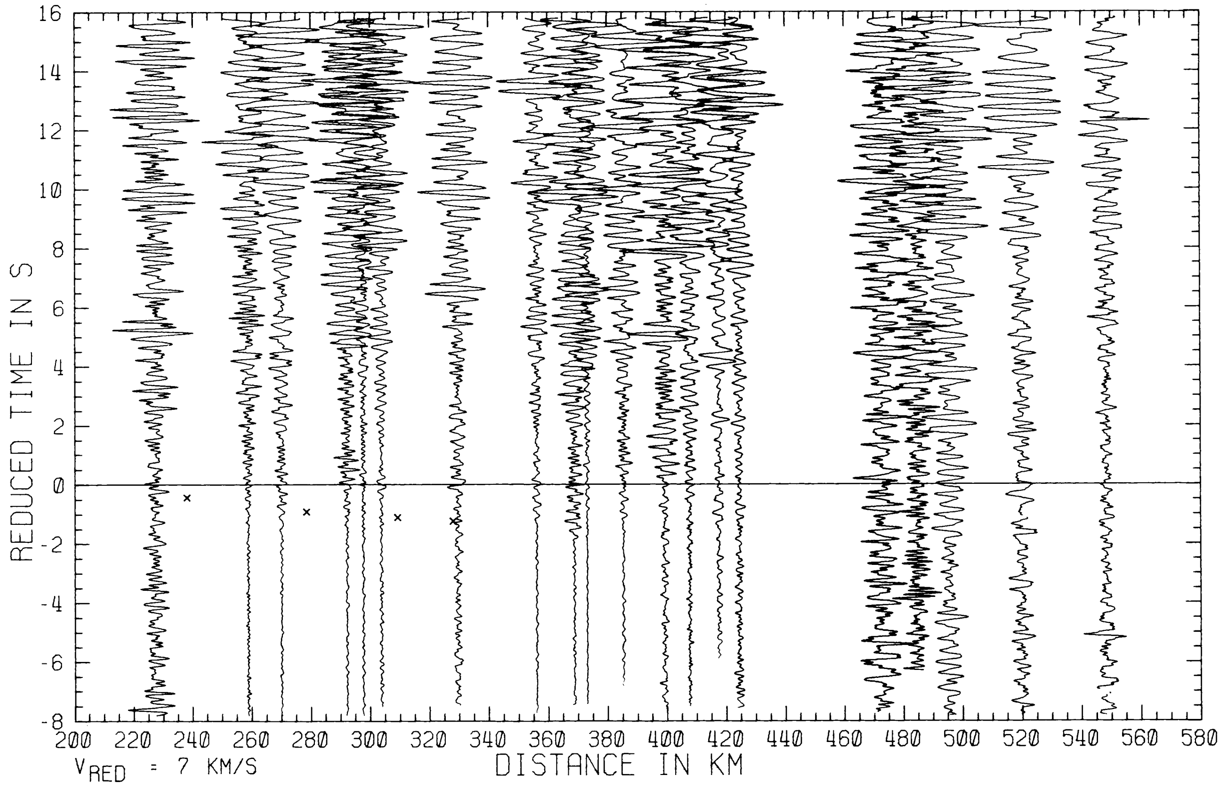

Fig. 4. Record section from shot B, filtered and normalized. The emergent character of the first onset is evident. Main energy within each seismogram begins to arrive at reduced times of some 4 to 5 s. *Crosses* mark calculated travel-times according to the model of Fig. 5

7 km chosen for the RRISP observations it is evident, that we cannot resolve small-scale heterogeneities from the analysis of travel-time data but have to be content with the evaluation of the main features of the deep structure. The lack of overcritical reflections which would give useful information on average velocity, puts further limits on the resolving power of the seimic refraction method.

The only later arrivals which might possibly be correlated over some distance, are found in the record section from shots D towards the northeast (Fig. 1) in the distance range between 90 and 170 km. Reduced travel times are between 4 and 2.5 s and apparent velocity approximately 7.8 km/s. Similar later arrivals are not to be found in the opposite direction from shot D and also not on the reversed line from shotpoint E. This feature may be related to the different frequency content of the D and E signals and may indicate the presence of thin layers of only local extent beneath central Iceland at about 30 km depth. The higher signal frequency content of the D-shots comes from the shooting technique with dispersed charges in shallow water. The dominant frequency of the other shots agrees well with theoretical predictions (Wielandt, 1972) for single charges at optimum depth (see paper 1, Table 1 for further information regarding the shots).

Figure 3 shows travel-time data obtained at some stations of the Icelandic seismological network for the series of small shots at sea between B and C. With the exception of the top two stations IR (IR-Skáli) and SL (Selfoss) all stations shown here lie very close to or on the main line (for location see Einarsson 1979). The travel times have been corrected for water depth at each shot position as follows:

$$t_w = \frac{d}{V_w}(1 - V_w^2/V_n^2)^{\frac{1}{2}} \tag{1}$$

where d is the water depth, V_w the velocity in water and V_n the velocity in the deepest layer of penetration. Surprisingly high apparent velocities between 8.2 and 9 km/s are calculated from these data. Absolute travel times are greater by 2.0 to 3.0 s as compared to travel times (corrected for water depth) at OBS BI02 in the same distance range (see paper 3). The apparent velocities for individual shots of the same shot series at sea, measured between different stations in Iceland, are much smaller and close to 7.4 km/s. Similar apparent velocities have also been observed along the main line up to distances of about 500 km for the more distant shot-point B. The pertinent record section is shown in Fig. 4. It is an extreme example for the emergent character of the onsets at greater distances and the shift of wave energy to greater travel times, mentioned above. At distances beyond approximately 500 km no onsets can be detected in spite of a rather low ground noise level (on the average 0.3 μm/s in the passband 1 to 10 Hz). The record section of the 4-ton shot at A, which covers the observation range from 420 to 800 km, is not shown here, because no recognizable onsets are to be seen.

The data have been evaluated as follows: First, direct inversion methods were used to find preliminary models, which satisfied the observational data in parts. Thereafter, numerous model calculations with a ray-tracing program by Gebrande (1976) were carried out in order to derive a model for crustal and upper-mantle structure along the main line, that satisfies travel-time data for all shots observed, and includes earlier results (Pálmason, 1971). The model building technique with the necessity of working with a manageable number of model parameters required a certain generalization of the available information for the topmost part of the crust. This was attained by choosing velocities, gradients, and thicknesses in such a way that they give the same transit times as through the more detailed structures given in paper 3

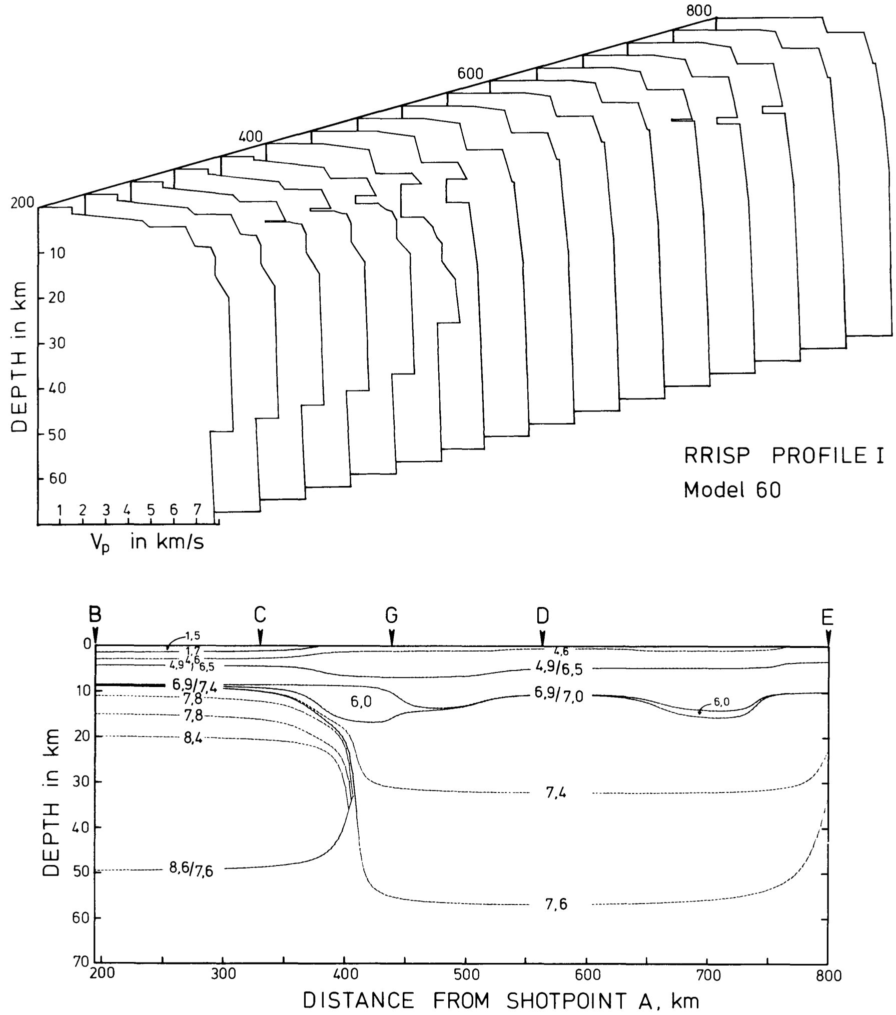

Fig. 5. Crustal and upper-mantle model beneath Iceland and the east flank of Reykjanes Ridge between shot-points B and C. Distances are given from shotpoint A. *Top:* Velocity-depth distributions along RRISP profile I. *Bottom:* The model is defined by the velocity isolines and linear vertical interpolation in between. *Numbers* give the velocity at each isoline. The velocity of 6.0 km/s in the two blisters at the base of the crust must be less than 5 km/s to explain observed local travel-time delays

for the marine part and for the land part by Pálmason (1971). His constant-velocity layers 0, 1, 2 have been modelled by two gradient layers. In parts this may even be a better approximation to the real structure.

The lower crust was modelled as a layer with a discontinuous velocity increase at its top to 6.5 km/s and a continuous increase to 6.9 km/s at its base along the whole length of the profile.

The best fitting model derived in this way is shown in Figs. 5 and 6. In the top part of Fig. 5 the velocity-depth structure along the RRISP main line is given in a semi-perspective view and in the lower part it is shown in the form of velocity contours. The model is defined by the velocity contours and linear vertical interpolation in between. As an example of the ray tracing calculations, Fig. 6 shows calculated seismic rays for shotpoints C, D, and E within part of the complete model in order to convey an impression of the sampling of the structures by the seismic rays. Only a few selected rays are shown to keep the figure legible. The calculated travel times are marked as crosses in the record sections of Fig. 1 to give an idea on the degree of fit. Mean deviation is generally less than 0.1 s. In view of the obvious small-scale heterogeneity a better fit would not be very meaningful.

An exceptional large discrepancy between observed and calculated travel-time is found on Heimaey, where a delay of 0.3 s is observed with respect to our model. This discrepancy could be removed by decreasing the velocity in the blister below Heimaey from 6 km/s to less than 5 km/s. The source of this delay cannot lie within the upper crust, since the records of the nearby shots G1, 2 show normal travel-times. If this low velocity is accepted, it may indicate the existence of a zone of relative high degree of partial fusion in the lower crust beneath Heimaey. The horizontal dimensions and the depth of the low velocity body are not well constrained by our data. A similar low-velocity body at the base of the crust is inferred in the northern part of the neovolcanic zone (Askja-Herdubreid).

The base of layer 3 has somewhat arbitrarily been identified with the 6.9 km/s isoline. Contrary to Båth (1960) and Pálmason (1971) we find no pronounced discontinuity to velocities of 7.2 or 7.4 km at that depth range. The small discontinuity from 6.9 to 7.0 km/s present in our model is consistent with the data, but is not an inevitable consequence. We could just as well have a continuous increase in velocity. A more significant change in this depth-range seems to be a reduction of the *P*-wave velocity gradient from about 0.07 to 0.02 s^{-1}. At the moment we will call this zone of low-gradient layer 4 and will show later that it should be identified with anomalous upper mantle.

The crustal layers are continuous across the transition from Iceland to the eastern flank of Reykjanes Ridge and vary only in thickness. Layer 4, on the other hand, is present only beneath Iceland. Beneath Reykjanes Ridge a normal and well developed oceanic crust (paper 3) rests on a layer with a *P*-wave velocity

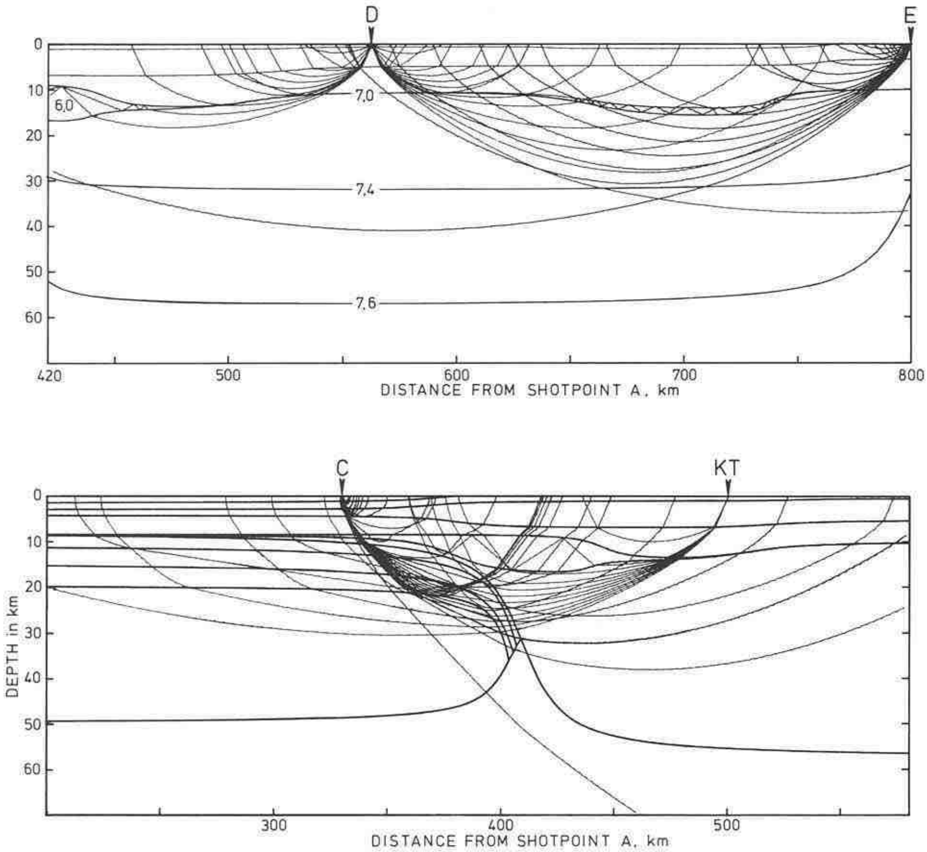

Fig. 6. Some calculated rays through parts of the model of Fig. 5 to give an idea on the sampling of the structures. The calculated travel-times have been plotted into the record sections of Fig. 1.
Top part: Shot-points D and E; *Lower part:* Shot-point C and station KT

of 7.8 km/s followed by a high velocity layer (up to 8.6 km/s) at greater depth. The velocities and the geometry of the transitional part of the model were derived from a combination of the land and OBS data sets by ray tracing. For this purpose the data set recorded at the stations of the Icelandic seismological network was especially useful as it provided reversed information to the OBS data on the marine part.

Looking at the rays for shot-point C (Fig. 6), one may see how the sudden termination of the oceanic lithosphere affects the observed travel times. Particularly evident is the decrease of apparent velocity with increasing distance. Rays bottoming in the high velocity layer of the oceanic lithosphere emerge at distances up to 170 km giving rise to apparent velocities of about 8 km/s, whereas the travel-time segment at greater distance with an apparent velocity of about 7.2 km/s is produced by rays bottoming in layer 4 beneath Iceland. The model structure also explains the observed travel times for shot B as well as for the small shots between B and C recorded on land (see Figs. 4 and 3).

As mentioned above, almost no signal energy could be recorded from the most distant shot A. In our model this is explained by the small thickness of the oceanic lithosphere. The rays sampling the lithosphere emerge south of Iceland, while Iceland itself, with respect to shotpoint A, is situated in a shadow zone caused by the lower velocities in the asthenosphere. Even though lack of signals is a poor means of structural determination, we are quite certain that the lithosphere thickness must be limited, since from the other shot-points we find no evidence of excessively high absorption necessary to reduce signal strength in the manner observed. Because of the negative evidence, the lower boundary of the occeanic lithosphere as shown in the model is by no means certain and was put somewhat arbitrarily at a depth of 50 km.

Although the recorded waves, according to our ray tracing calculations, did not penetrate deeper than some 30 km below Iceland it can be stated, that a continuous layer with *P*-wave velocities around 8 km/s, such as beneath the marine part of the profile cannot exist in the upper 60 km beneath Iceland. This is the result of several ray tracing calculations for models modified accordingly. If an 8 km/s layer or half-space with its top above a depth of 60 km is incorporated in the models, travel-time segments with apparent velocities around 8 km/s arise for which no evidence can be found in the record sections.

It is rather likely that the model presented in Fig. 5 will be subjected to improvements in the future, but we do not expect that the general features of the model will have to be changed. It should be mentioned in passing that the results from Profile II along the southeast coast of Iceland not presented here, corroborate our model and indicate that the structure derived is not only representative for the neovolcanic zone, but for all of Iceland. The details of the model may not be equally accurate in different parts due to different fit of calculated and observed travel times, different coverage of the lines, and different quality of the data, but also due to the principal resolving properties of the refraction seismic method. The reader will judge the accuracy of the model realistically if he keeps in mind, that the refraction seismic method is a good diagnostic for strong positive velocity contrasts. Its resolving power however, is rather weak if, as is the case below Iceland, the velocity gradients are small.

Discussion

The model presented has certain implications for the physical state and the petrology of the material at depth, which we would

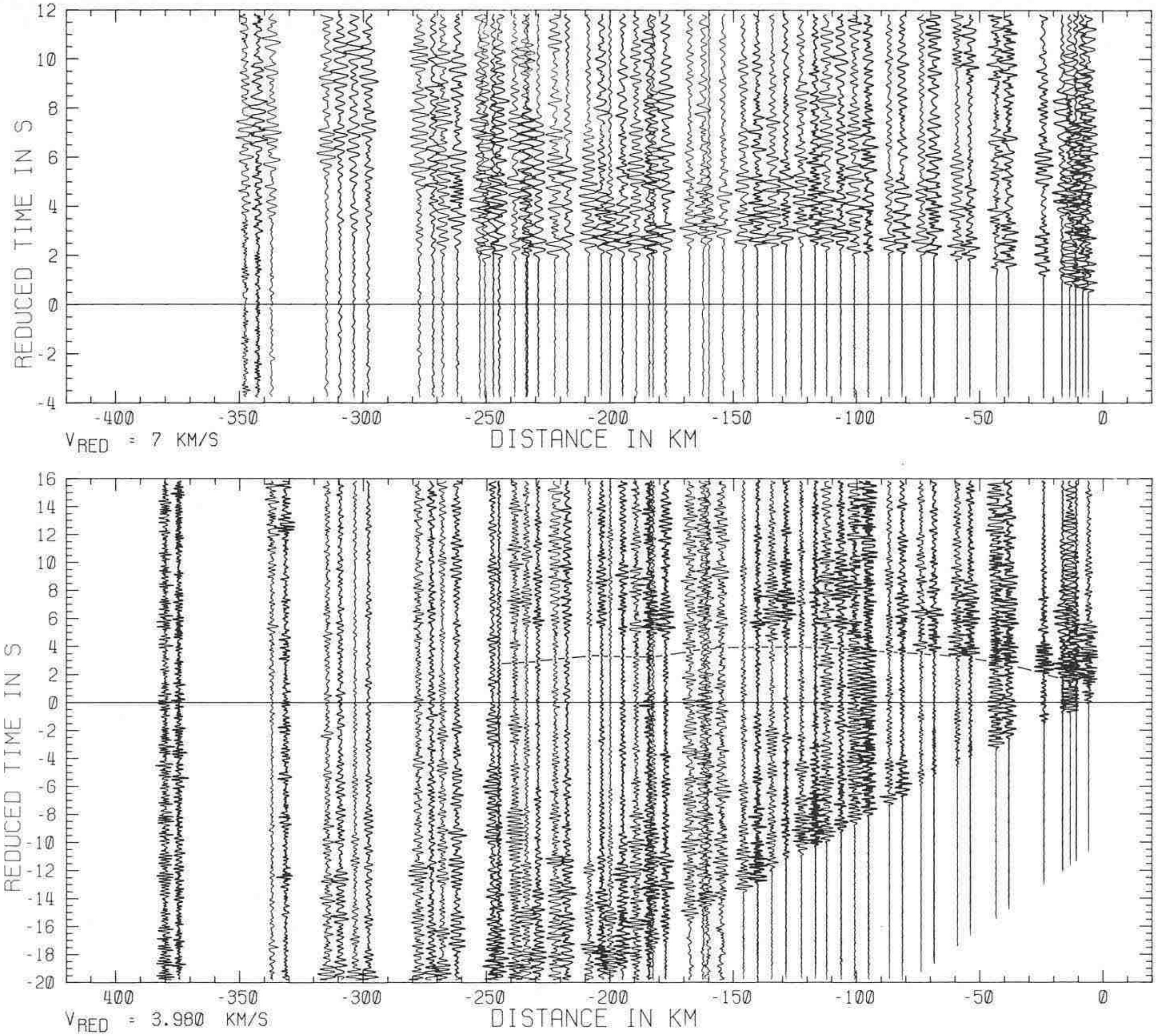

Fig. 7. *Top:* Record section from shotpoint E reduced by 7 km/s. Vertical component seismograms. *Bottom:* Record section from shotpoint E reduced by 3.98 km/s. Horizontal components were used mostly for this section, which shows *P*- and *S*-wave arrivals. Reduction velocity and time axis were chosen such, that *S*-arrivals should be congruent to *P*-arrivals in the top section, if the ratio of *P*- to *S*-wave velocity is 1.76. The *dashed line* gives *S*-arrivals as calculated from the *P*-arrivals with this ratio. The fit is good up to distances of 140 km, beyond which *S*-arrivals become progressively late

like to discuss with special emphasis on layer 4 beneath Iceland. Some conclusions concerning the oceanic lithosphere have already been presented in paper 1.

Additional to the structural information is the *P*- to *S*-wave velocity ratio beneath Iceland determined for shots E and G, which all generated considerable *S*-wave energy. This was not the case for the large shots at sea, which were suspended and not fired on the sea floor. Figures 7 and 8 show the records used. In the upper part of the figures the normal vertical component record sections of shots E and G are shown, reduced by 7 km/s. In the lower part the horizontal component record sections from the same shots are shown but reduced by 3.98 km/s, corresponding to a *P*- to *S*-wave velocity ratio of 1.76. The time axis has been compressed by the same factor. This representation of the data proves particularly useful when one section is laid on top of the other, because deviations from the chosen *P*- to *S*-wave velocity ratio can readily be recognized since *P*- and *S*-arrivals will not be congruent any longer. Since the reader cannot easily do this, for comparison the *P*-wave correlation for shots E has been marked in the corresponding *S*-wave section.

It is quite clear, that beyond distances of 140 km the *P*- to *S*-wave velocity ratio changes to higher values as *S*-arrivals are becoming progressively 'late' with repect to the *P*-arrivals. At the same range, *S*-wave amplitudes decrease in comparison with

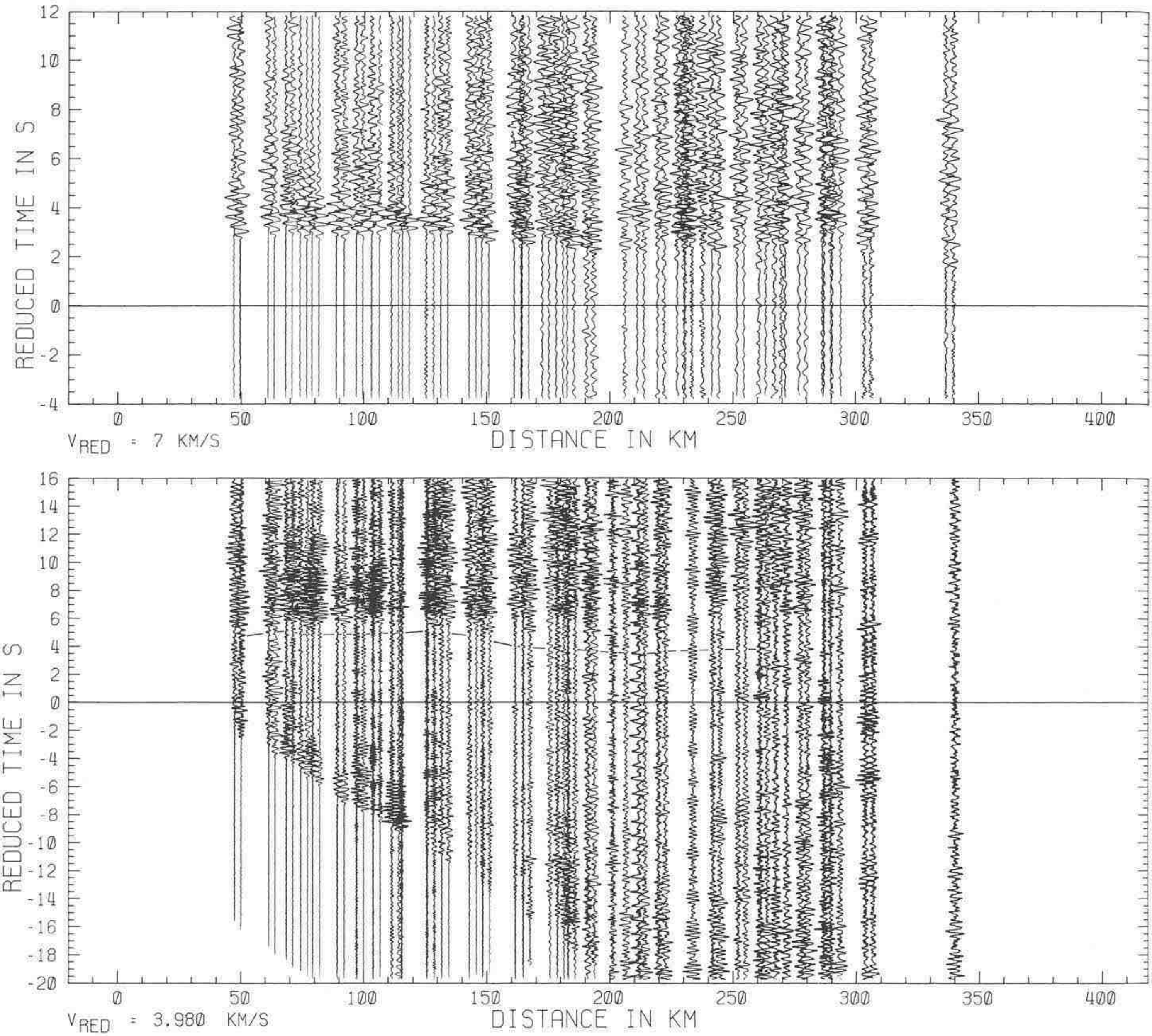

Fig. 8. Same as Fig. 7 but for shots G1 and G2 recorded along the line at the southeast coast

the *P*-wave amplitudes and *S*-wave signals are lost at distances greater than 250 km. It is a rather dramatic amplitude attenuation considering that, e.g., for shots E, the *S*-wave amplitudes are larger than the *P*-amplitudes by at least a factor of 2 at distances between 20 and 130 km (some seismograms are distorted by amplifier saturation).

When looking at the rays from shotpoint E drawn in Fig. 6, one notices that beyond distances of 140 km emergent rays have penetrated the low velocity body mentioned before and have sampled layer 4. We must therefore conclude that at the base of layer 3 a more fundamental change of physical properties takes place than is indicated by the almost negligible change in *P*-wave velocity.

A quantitative evaluation of the *P*- to *S*-wave velocity ratio is given in Fig. 9 where a Wadati diagram for shots E has been plotted. Up to a *P* travel time of 21 s, corresponding to a distance of approx. 140 km, the *P*- to *S*-wave velocity ratio is constant at 1.76. With greater *P*-wave travel-times the slope of the curve reaches values as high as 2.2 with a mean of 1.96.

The knowledge of the change in *P*- to *S*-wave velocity ratio with depth as well as the absolute values help us answering the question how layer 4 should be interpreted. The question, whether the velocity values between 7.0 and 7.4 km/s should be assigned to the crust or upper mantle has been debated extensively in the literature (Båth, 1960; Tryggvason, 1962, 1964; Bott, 1965, 1974; Francis, 1969; Pálmason, 1971; Pálmason and Saemundsson, 1974; Zverev et al., 1976).

In the last decade velocities in the range from 7.0 to 7.7 km/s have more widely been found than assumed previously and are generally assigned to layer 3b of the oceanic crust (e.g., Peterson et al., 1974). Comparison with laboratory measurements of seismic velocities on samples dredged from the ocean floor and taken

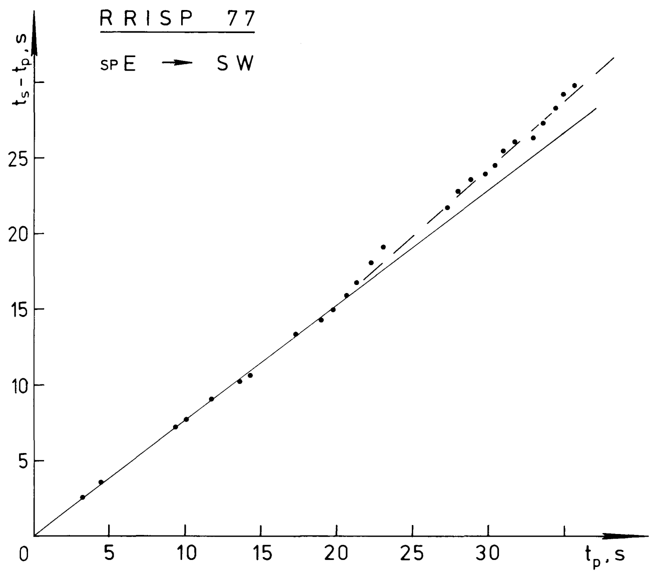

Fig. 9. Wadati plot for shots E1 and E2 along the main line. For P travel-times greater than 21 s the P- to S-wave velocity ratio changes from 1.76 to a mean of 1.96

from ophiolites of presumably oceanic origin shows, that gabbros and metagabbros have about the right velocities at appropriate pressures and moderate temperatures. At the same time, gabbros also exhibit the correct P- to S-wave velocity ratio of 1.9 corresponding to a Poisson's ratio $\sigma=0.31$ (Christensen and Salisbury, 1975; Kroenke et al., 1976; Christensen, 1978). At temperatures of 1,000° to 1,100° C, such as are indicated beneath Iceland at 10 km to 20 km depth by the geothermal gradient (Pálmason and Saemundsson, 1974) and by magneto-telluric data (Hermance and Grillot, 1970; Beblo and Bjönsson, 1978, 1980), the combined influence of pressure and temperature would produce a decrease in the P-wave velocity of gabbroic material of some 1.2 km/s (Kroenke et al., 1976) and this disagrees with the in situ velocities. An additional decrease would result from incipient melting at 1,000° to 1,100° C in the presence of small amounts of water. Therefore gabbroic or other basic material must be discounted as the only or at least principal constituent of layer 4; predominantly ultramafic material must be assumed. It is therefore most natural to attribute layer 4 to the upper mantle. But this is more or less a question of definition. It seems more important to us that, whatever the exact petrological composition of the upper mantle is, a rather large density contrast is to be expected with respect to the normal lower lithosphere beneath Reykjanes Ridge, as long as the material is in the perfect solid state. Taking 7.2 km/s and 8.4 km/s as mean P-wave velocity values for the depth range from 20 to 50 km beneath Iceland and the Reykjanes Ridge (Fig. 5) and a Birch relationship

$$V_P=3.31\,\rho-2.55\,(\text{km/s}); \qquad (\rho \text{ in g/cm}^3) \tag{2}$$

as confirmed by Kroenke et al. (1976) for possible lower crust and upper mantle rocks, the density contrast should be some 0.35 g/cm^3. This is reduced to 0.3 g/cm^3 if the relationship given by Christensen and Salisbury (1975) is used. In any case, taking only the depth range of 20 to 50 km into account, this would imply a Bouguer anomaly over Iceland of some −350 mgal, which is about three times the observed value (Einarsson, 1954). This disagreement can be overcome as mentioned by Bott (1965) in a similar context, if the low seismic velocities are mainly attributed to partial fusion. In this case the seimic velocities are relatively much more affected than the density and the usual velocity-density relationship breaks down.

One may now ask, whether quantitiative information on the degree of partial fusion can be extracted from the seismic velocities. This is possible for a simplified model of partially molten material. We assume that it consists of only two phases, a solid phase 1 with properties k_1, μ_1, ρ_1 (bulk modulus, shear modulus, and density) and a liquid phase 2 with properties k_2, $\mu_2=0$, ρ_2. With Green and Ringwood (1963) we may identify the solid phase with peridotite and the liquid phase with basaltic melt. We are interested in how the seismic velocities of the inhomogeneous composite material depend upon the properties of the homogeneous phases and the fractional volume of the melt. A solution to this problem should then allow us to determine the melt concentration from the measured seismic velocities, if the properties of the homogeneous phases are known. Unfortunately a unique solution is not possible since the properties of the two-phase material depend in general on the entire 'phase geometry', i.e., the geometry of the phase interfaces. This ambiguity is not removed, if statistical homogeneity is assumed.

Walsh (1968, 1969) has investigated theoretically the special case of isolated melt inclusions in the form of randomly oriented oblate spheroids with minor axes much smaller than major axes. Since melting in polycrystalline material starts at grain boundaries as thin films, this seems to be a good model for incipient melting. It turns out that the elastic moduli of the partially molten material depend not only on the volume concentration c_2 of melt, but also on the aspect ratio α of the inclusions, i.e., the ratio of minor to major diameter d of the oblate spheroids. A smaller aspect ratio of the inclusions requires a much smaller melt content than a larger aspect ratio does, to give same velocity decrease. The velocities therefore cannot be interpreted unambiguously without making assumptions about the aspect ratio. An $\alpha=0.01$ has been used by several authors. It has however, often been neglected, that the melt content c_2 and the aspect ratio are not independent variables. At a constant aspect ratio, an increase of melt content necessarily requires an increase of the inclusion diameter, which consequently leads to a coalescence of previously isolated inclusions. If this process proceeds too far, the Walsh theory can no longer be applied. It can be shown (Gebrande, in preparation) that with randomly distributed inclusions the fraction of isolated inclusions is less than 90% if $c_2 \leqq \alpha/10$. For $c_2=\alpha/2$ only 60% of the inclusions can be expected to be isolated. Since α has to be small anyway, the Walsh theory is valid only for rather minimal melt content. It is therefore not possible to apply this theory to the anomalous mantle without violating its inherent assumptions.

For our purpose extremal bounds for bulk and shear moduli k, μ as derived by Hashin and Shtrikman (1963; ref. Hashin 1966) from some elasticity extremum principles are more useful. These bounds can be transformed into bounds for the seismic velocities V_P, V_S and the seismic parameter ϕ. These bounds are the best possible in terms of k_i, μ_i $(i=1,2)$, and c_2; the velocities of any two-phase material independent of its phase geometry must lie within these bounds. Unfortunately in the case of a solid-fluid mix the bounds for V_P and V_S are rather far apart. This is due to the fact that the lower bound for the shear modulus vanishes. This is physically plausible, since the rigidity of the composite is zero if all solid particles are surrounded by melt. The bounds for the seismic parameter ϕ, and the 'hydrodynamic wave velocity' $\phi^{1/2}$ (Birch 1969) are reasonably close, however. If the P- and S-wave velocities are known, an estimate of the melt content can be derived from these bounds. They are given by:

$$\phi \leqq \frac{1}{\rho}\left\{k_1 + c_2\left(\frac{1}{k_2-k_1} + 3\,\frac{1-c_2}{3k_1+4\mu_1}\right)^{-1}\right\} \tag{3}$$

$$\phi \geqq \frac{1}{\rho}\left\{k_2 + (1-c_2)\left(\frac{1}{k_1-k_2} + \frac{3c_2}{3k_2+4\mu_2}\right)^{-1}\right\} \tag{4}$$

$$\rho = \rho_1 + c_2(\rho_2 - \rho_1). \tag{5}$$

If we assumę that the anomalous mantle beneath Iceland is a mixture of the material of the high-velocity layer observed beneath the Reykjanes-Ridge and basaltic melt, we can calculate the extremal bounds according to these formulae. From our model we obtain a P-wave velocity of 8.47 km/s in the lower lithosphere at a depth of of 30 km. Correcting for a possible temperature difference of some 200 K between the 10 Ma old Reykjanes Ridge and Iceland at this depth by using a temperature coefficient of

$$(\partial V_p/\partial T)_p = -4\cdot 10^{-4}\ \mathrm{km/s\cdot K}$$

(Anderson et al. 1972) we obtain a P-wave velocity of 8.39 km/s for the solid component. For basaltic melt a P-wave velocity of 4.1 km/s is given by Röber and Thyssen (1978). Taking these values and appropriate densities

$$\rho_1 = 3.3\ \mathrm{g/cm^3} \quad \text{and} \quad \rho_2 = 2.76\ \mathrm{g/cm^3} \quad \text{(basalt glass)}$$

the extremal bounds given by the heavy solid lines in Fig. 10 were calculated. The dashed lines are based on values of Birch (1969), who has used the bulk modulus of basalt glass for the melt component. On the other hand, $\phi^{1/2}$ can be calculated from the observed P-wave velocity and P- to S-wave velocity ratio according to

$$\phi^{\frac{1}{2}} = V_p[1 - \tfrac{4}{3}(V_s/V_p)^2]^{\frac{1}{2}}. \tag{6}$$

The corresponding values for the anomalous mantle beneath Iceland have been marked by horizontal lines in Fig. 10. If follows from the theoretical bounds, that the melt content must be between 10% and 16.7% (or 15.5% to 23% for the values of Birch) to explain a P-wave velocity of 7.3 km/s and higher values of 17% to 27% are obtained for a P-wave velocity of 7.0 km/s. These results seem to reflect a differentiation process in the upper mantle and an enrichment of basaltic melt at the base of the crust. The possible existence of local and rather thin reflecting elements at a depth of approximatly 30 km may well be correlated with the decreasing melt content. Taking a melt content of 13% as the average for the depths from 20 to 50 km the mean density difference between Reykjanes-Ridge and Iceland within this depth range becomes 0.07 g/cm^3 according to Eq. (5). This difference is much smaller than expected from a Birch relation for solid rocks; in order to explain the observed gravity anomaly this density anomaly beneath Iceland most likely must extend to depths greater than 50 km. The same conclusion is reached by studying the teleseismic travel-time residuals. The upper 50 km of the model presented in Fig. 5 account only for a travel-time delay of 0.2 s and therefore a much deeper extent of the low-velocity body beneath Iceland must be assumed to explain the 1.4 s delay as observed by Long and Mitchell (1970).

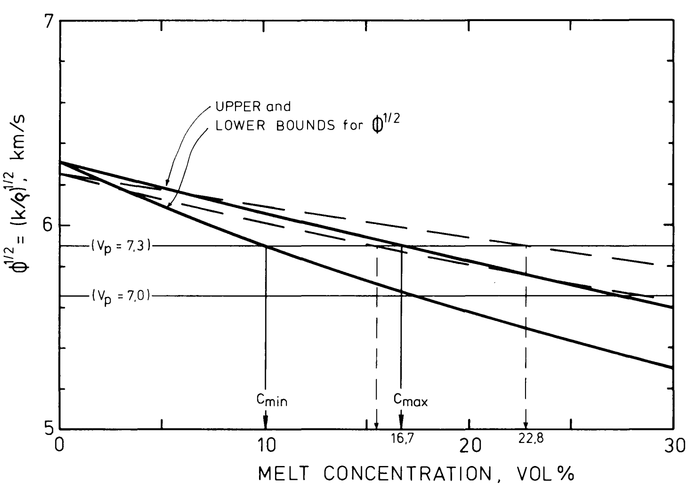

Fig. 10. Upper and lower bounds on the hydrodynamic wave velocity $\Phi^{1/2}$ against melt concentration in a solid-fluid two phase system. The two different bounds (*solid and dashed lines*) are based on different values of elastic parameters

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