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Seismic Structure of the Icelandic Crust Above Layer Three and the Relation Between Body Wave Velocity and the Alteration of the Basaltic Crust

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Abstract. Seismic refraction profiles from Iceland are studied with the aid of synthetic seismograms. The classical layered model of the Icelandic crust is shown to be an unacceptable interpretation of the available data. This is because the layered model does not satisfy the observed amplitude variation. On the other hand, a model which assumes continuously increasing velocity with depth does not contradict the observations and is therefore acceptable although it is not the only possible interpretation. The model represented here shows that the surface value of the *P*-velocity is variable from 2.0 km/s to 5.0 km/s, depending primarily on the degree of metamorphism. The *P*-velocity increases rapidly with depth in the velocity interval 2.0–3.5 km/s followed by an approximately constant gradient of about 0.57 s^{-1} This constant gradient continues down to the 6.5 km/s isovelocity surface below which the *P*-velocity becomes nearly constant. In view of this, it is more reasonable to divide the Icelandic crust into two parts: the upper crust with velocity continuously increasing with depth (corresponding to layers 0, 1, 2 in the layered model) and the lower crust with almost constant velocity (corresponding to layer 3 in the layered model). The depth to the lower crust is variable and depends on how deep the crust is eroded. A typical depth to the lower crust is 5–6 km for an uneroded basalt pile but can be considerable less where the basalt pile is deeply eroded, especially below extinct central volcanoes.

Key words: Seismic refraction – Synthetic seismograms – Poisson's ratio – Amygdale minerals – Crustal structure – Iceland.

Introduction

During the past few decades the classical method in interpreting seismic refraction data has been to assume layers of constant properties and to compute velocities and depth to the boundaries. By this method one has deduced the classical three-layer models of the oceanic and the Icelandic crust.

In the last few years several authors have pointed out the lack of uniqueness in this method. Because of measurement errors, it is not possible to decide whether the travel time curves are slightly curved or made up of straight line segments. It is therefore not possible, by the use of travel time diagrams for first arrivals only, to decide whether the velocity varies continuously with depth or in jumps.

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Table 1. Layered seismic structure of the Iceland crust after Pálmason (1963, 1971)

Layer No.	*P*-velocity km/s	*S*-velocity km/s	Poisson's ratio	Density g/cm^3
0	2.75			2.1–2.5
1	4.14	2.34	0.270	2.6
2	5.08	2.78	0.278	2.65
3	6.50	3.53	0.269	2.9
4 (Mantle)	7.20			3.1

Kennett and Orcutt (1976) have applied systematic inversion techniques to marine refraction profiles and they conclude that layer two is a region of strong velocity gradients, while layer three is relatively homogeneous. Lewis (1978) has concluded that the commonly assumed layerd model of the oceanic crust is an artifact of the method of data interpretation.

There is, however, more information than merely first arrivals to be had from a seismogram. By use of secondary arrivals such as wide-angle reflections and amplitude data it should be possible to discriminate to a certain extent between the various models. Use of synthetic seismograms is a powerful tool in such studies.

Båth (1960), Tryggvason and Båth (1961), and Pálmason (1963; 1971) have studied in Icelandic crust by refraction seismology and deduced the layering of the Icelandic crust. Pálmason's work includes studies of more than 80 refraction profiles distributed over Iceland. He concludes that the Icelandic crust consist of four seismic layers underlain by mantle with an anomalously low *P*-wave velocity of 7.2 km/s (Table 1).

Analysis of Seismic Refraction Profiles From Iceland

In order to determine whether or not the Icelandic crust is made up of homogeneous seismic layers, I have used some carefully chosen profiles from Pálmason's (1971) together with one new profile, and have made synthetic seismograms for various models. The profiles have been chosen so that they follow regional geological strike and are not interrupted by central volcanoes. This is done to minimize the possibility of lateral velocity variations in the direction of the profile. The field work and instrumentation involved in these measurements are described by Båth (1960) and Pálmason (1971). It is worth noting here that the recording was made on photographic paper and that the paper velocity was not the same for individual records within the same profile. I

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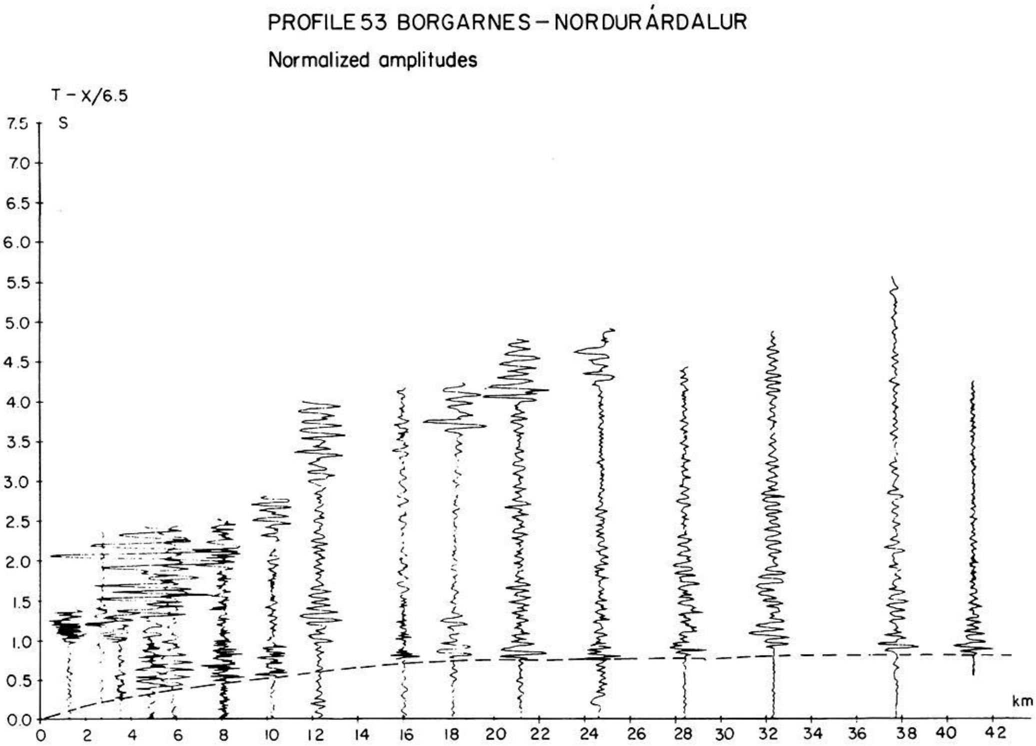

Fig. 1. The seismic record section for Profile 53

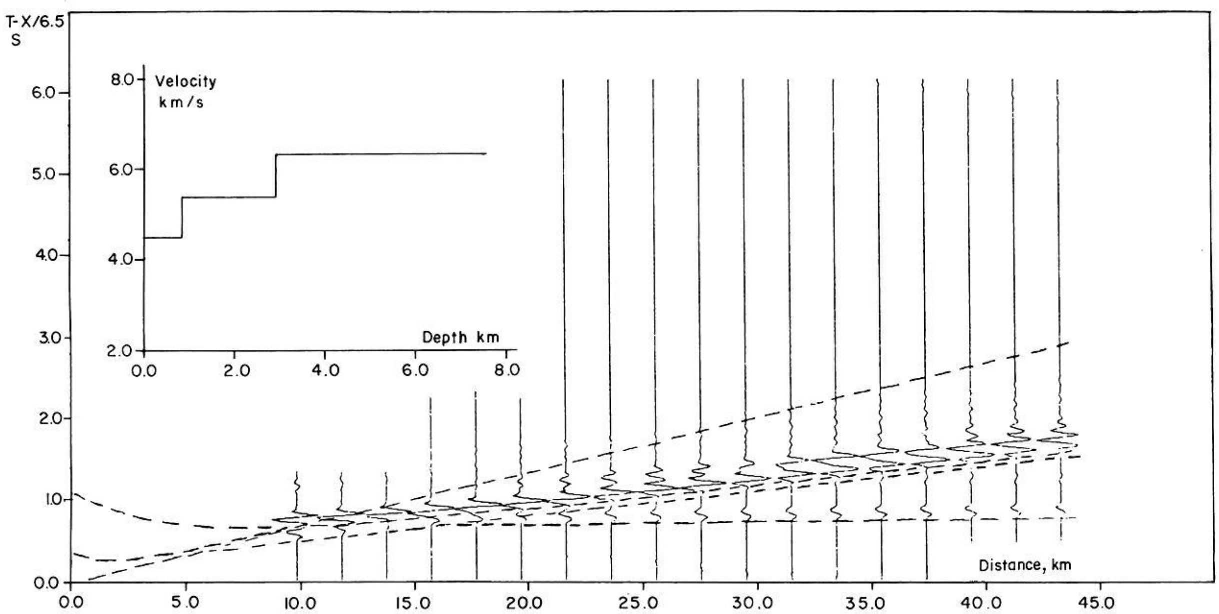

Fig. 2. Synthetic seismograms for the layered model of Profile 53

have therefore digitized the records from the photographic paper and plotted them with aid of a computer. In doing so the seismograms lose some of their characteristics.

For producing synthetic seismograms, I have used a computer program by Mykkeltveit (1978) based on a paper by Fuchs and Müller (1971). I have concentrated mainly on two possibilities: the layered model and a model based on continuously increasing velocity with depth. To deduce the latter model, I have used a computer program by Berge (1976). This program makes use of the Wiechert-Herglotz formula

$$R \cdot \ln\left(\frac{R}{R-z_p}\right) = \frac{1}{\pi} \int_0^{\Delta p} \cosh^{-1}\left(\frac{v_p}{v}\right) d\Delta$$

where R is the radius of the earth, Δp is the distance to the point on the travel-time curve where the apparent velocity is v_p; v_p is the apparent velocity for the ray for which one shall compute the greatest depth of penetration, v is the apparent velocity along the profile and z_p is the maximum depth of penetration of a ray which is recorded in the distance Δp from the shot point.

In computing the synthetic seismograms for a model with continuously increasing velocity with depth, I have assumed the den-

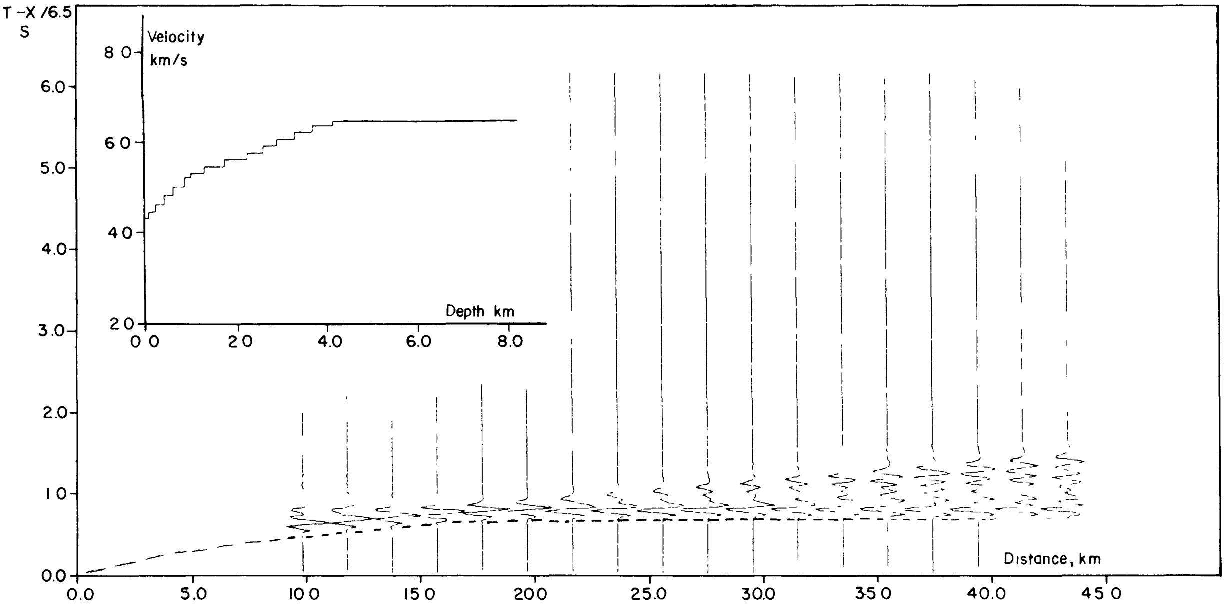

Fig. 3. Synthetic seismograms for a model of Profile 53, based on the Wiechert-Herglotz method

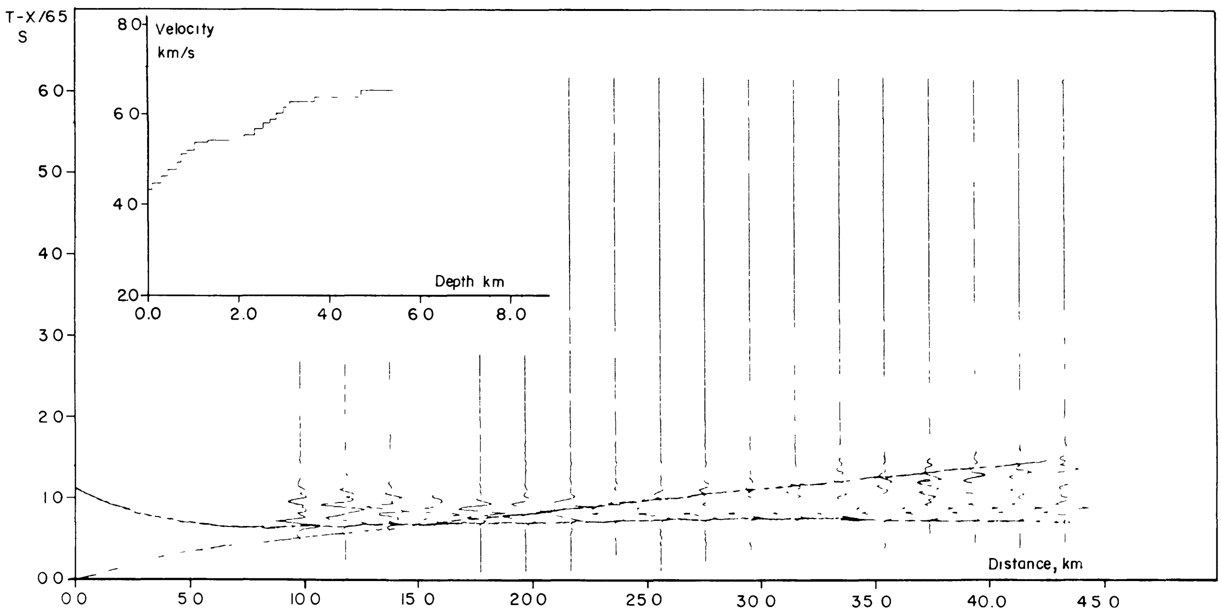

Fig. 4. The same as Fig. 3 except that there is a 0.8-km-thick constant-velocity layer within the upper crust

sity to be 2.60 g/cm^3 for rocks with P-velocities around 3.0 km/s, 2.65 g/cm^3 for rocks with velocity 5.0 km/s, 2.90 g/cm^3 for layer three (6.5 km/s) and interpolated linearly between them. These density below Moho is taken to be 3.1 g/cm^3 These values are based on Pálmason's (1971) estimate for the layered model and may be inaccurate. This will, however, not affect the synthetic seismograms seriously because they are not very sensitive to density.

Because of variations in charge size and magnification, and because the geophones were moved between each shot and the attenuation is unknown, I have used mainly synthetic seismograms with normalized amplitudes. It is possible, therefore, to compare the amplitudes within each trace but not to compare individual recordings. I have analysed six seismic refraction profiles from various geological provinces of Iceland. Two of them are described in this paper but the others are discussed in my thesis (Flóvenz, 1979). These two are profiles Nos. 53 and 1 of Pálmason (1971).

Profile 53, Borgarnes – Norðurárdalur, runs northeast from the village Borgarnes in the Tertiary basalt region of western Iceland. The age of the basalt is 6.2–7.0 Ma in Norðurárdalur (McDougall et al., 1977). The shot point is close to the coast. Figure 1 shows the seismic record section. The most characteristic feature of this profile is the relatively large amplitudes of the first arrivals compared with secondary amplitudes and it seems to be difficult to find systematic later arrivals. The travel-time curve for first arrivals can be considered to be composed of either three straight lines or of a curved line from the origin to approximately 6.5 km/s apparent velocity, from where it continues as a straight line. The first mentioned possibility leads to a layered

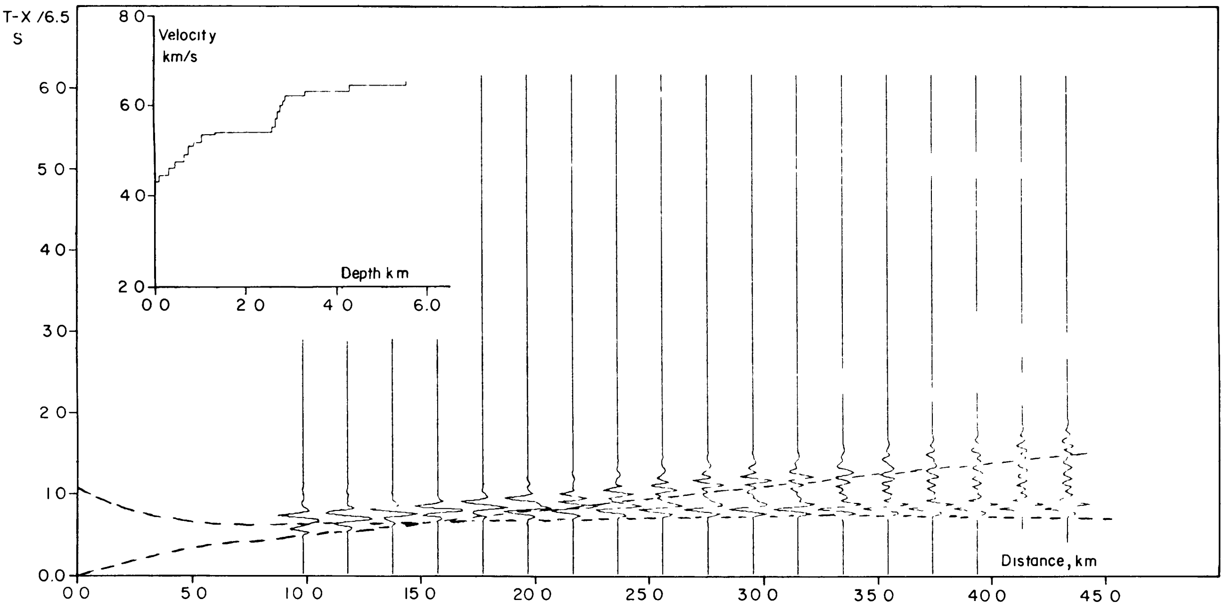

Fig. 5. The same as Fig. 4 but the constant velocity layer is now 1.3 km thick

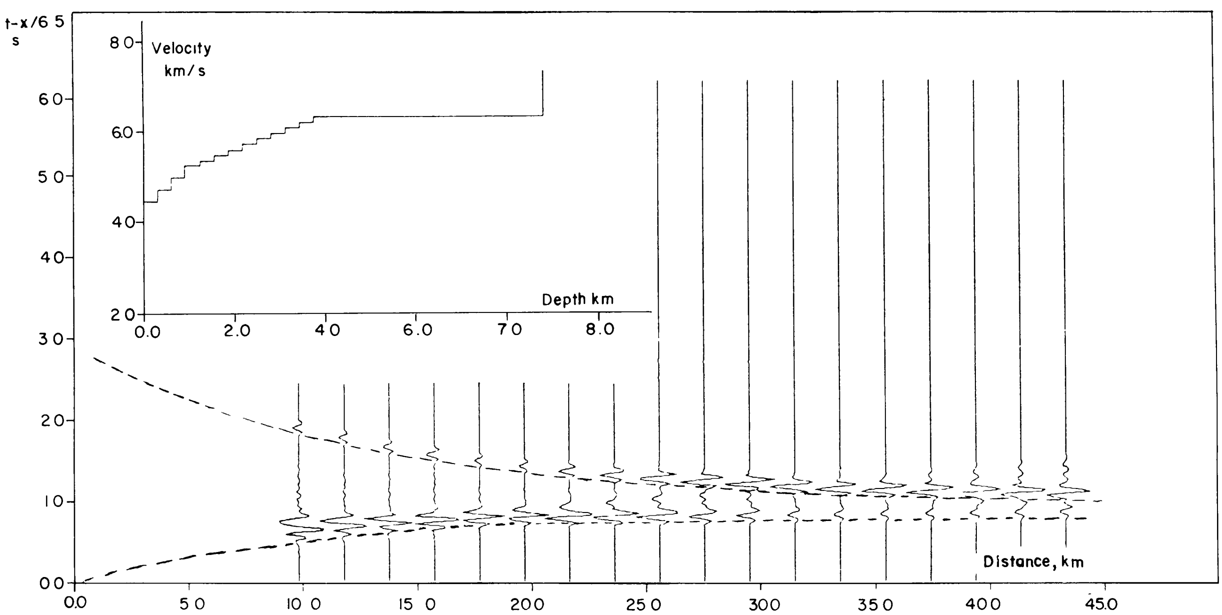

Fig. 6. The same as Fig. 3 except that the Moho is assumed to lie at 9 km depth

model. Figures 2 and 3 show synthetic seismograms for the layered model and a model which assumes continuously increasing velocity with depth.

It is immediately obvious from these figures that the layered model is not an acceptable interpretation of the data. In this model the amplitudes of the waves reflected at wide angles from the layer 2 – layer 3 boundary are much stronger than the refracted ones. These reflections are absent in the observations. On the contrary, the energy is concentrated in the first arrivals for the continuously-increasing velocity model; this bears much more resemblance to the observation.

The travel-time diagram in Fig. 1 is clearly curved at the beginning but over a short interval it approches a straight line with the apparent velocity of 5.4 km/s. It is possible that this part of the curve represents an interval of constant velocity within the crust. Figures 4 and 5 show synthetic seismograms for such models with a 0.8 km and 1.3-km-thick constant velocity layer, respectively. On these seismograms the energy is concentrated in the beginning of each signal just as on the observations. As the constant-velocity layer becomes thicker the velocity gradient between it and layer 3 becomes large enough to give triplication of the travel time curve, resulting in stronger secondary arrivals with an apparent velocity of around 5.5 km/s. As the velocity gradient increases further the amplitude of these secondary arrivals increases and should be detectable in the observations. But the observations do not show such secondary arrivals. It is possible, however that interference with possible bubble pulses can destroy these arrivals.

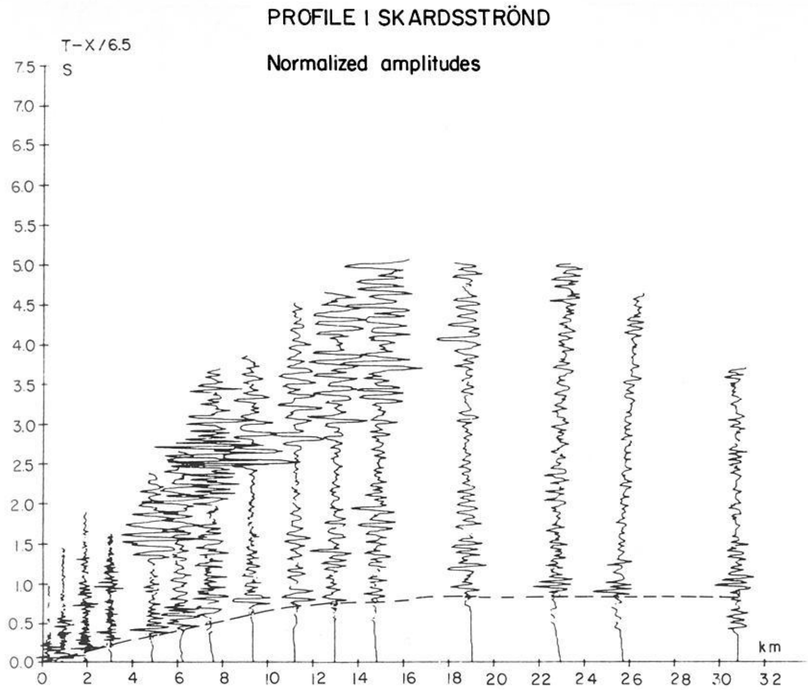

Fig. 7. The seismic record section for Profile 1

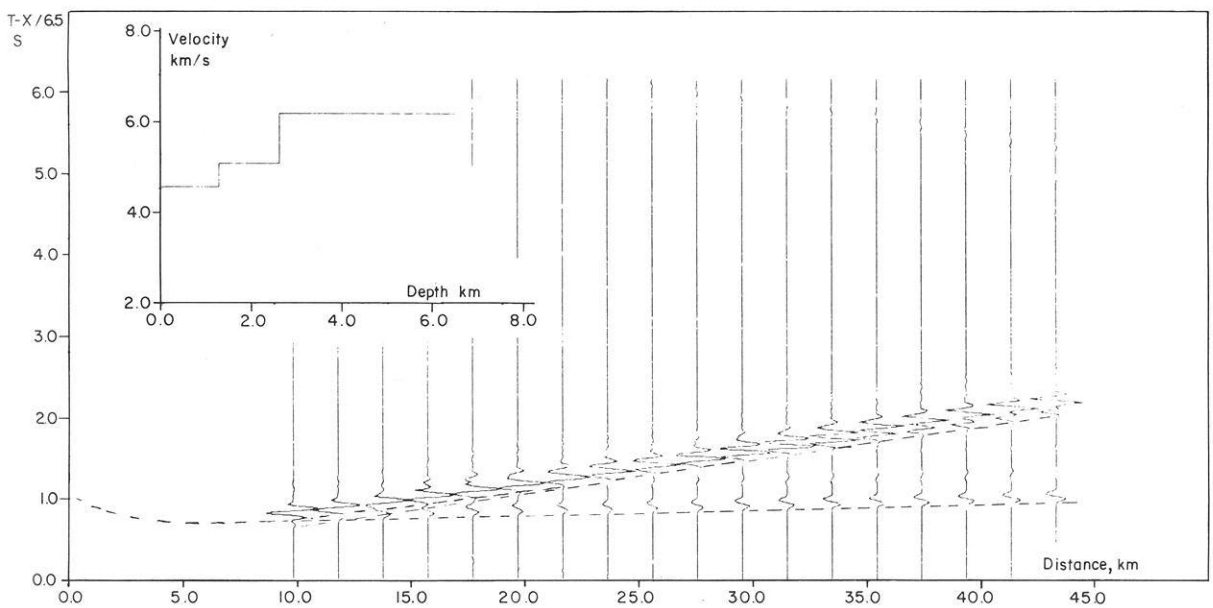

Fig. 8. Synthetic seismograms for the layered model of Profile 1

This profile is too short for recognizing critically refracted waves from the Moho as first arrivals. However, if the Moho discontinuity lies at about 9 km depth as Pálmason (1971) indicates, it is supposed to give rise to wide-angle reflections. Figure 6 shows synthetic seismograms for the model with continuously increasing velocity down to layer three and the Moho discontinuity at 9 km depth. These indicate that wide-angle Moho reflections should be recognizable from around 25 km shot point distance. The observations (Fig. 1) show no such reflections indicating that the Moho must be at a greater depth or be absent as a velocity discontinuity under the profile.

One can therefore conclude from this profile that the velocity increases continuously with depth at least down to the 5.4 km/s isovelocity surface. Below this surface the velocity either takes a constant value over a certain interval and increases after that rapidly to 6.5 km/s in layer 3, or the velocity increases continuously with depth down to layer 3 which seems more likely.

Profile 1 (Skardsströnd) runs northeastwards along the direction of strike on the south coast of Gilsfjörður in western Iceland, a region somewhat older (~ 10 Ma) than Borgarfjörður. Figure 7 shows the observations and Figs. 8 and 9 show the synthetic seismograms based on the layered and the continuous-velocity depth models, respectively. As for the previously discussed profile, the amplitudes of the first arrivals are usually the strongest except between 12 and 20 km shot point distance. There is however no regularity to be seen in the secondary arrivals. The synthetic seis-

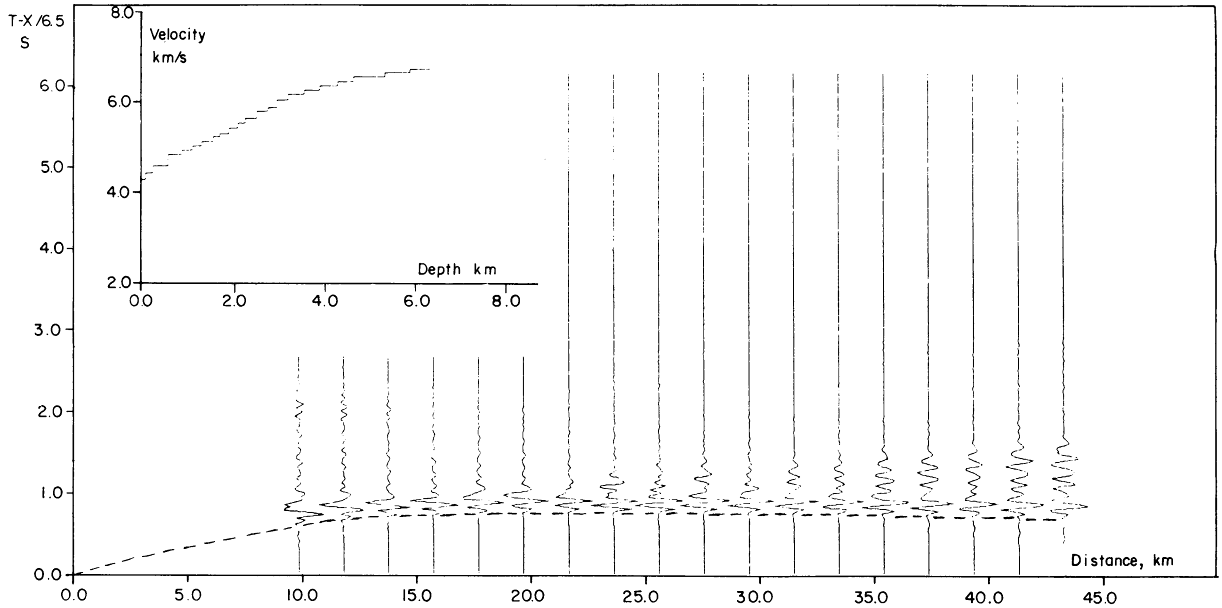

Fig. 9. Synthetic seismograms for a model of Profile 1 which assumes continuously increasing velocity with depth

Table 2. Thicknesses of Zeolite zones in Iceland After Walker (1974)

Types of zeolites	Thickness of each zeolite zone in m
No zeolites	150
Chabasite and thomsonite	450
Analsite	150
Mesolite and scolesite	900
Laumontite	1,400

mograms for the layered model show that a velocity discontinuity between layer two and three should result in much stronger amplitudes for the wide-angle reflections than for the first arrivals. This shows that the layered model is not consistent with the observations. The amplitude distribution for the model which assumes continuously increasing velocity with depth bears more similarity to the observations than the layered model and is therefore the preferable interpretation. It is possible, however, that the velocity can take a constant value over a certain depth interval. The travel-time curve has a fairly constant slope between 5 and 11 km shot point distance with an apparent velocity close to 4.7 km/s. On the basis of these observations alone, it is not possible to prove or disprove the existence of such a constant-velocity layer.

The conclusions that can be drawn from these profiles and the four others which are discussed in my theses (Flóvenz, 1979) are as follows: Both the layered model and the continuously-increasing velocity-depth model do satisfy the travel time curve for first arrivals but the latter better explains the energy distribution in the seismograms. This means that the layered model is generally not an acceptable interpretation of seismic refraction profiles in Iceland. This does not mean, however, that the continuously increasing velocity-depht model is correct, it is only a model that better fits the observations. It is, for instance, possible that the velocity takes a constant value over a limited range of depth. In the six profiles there are no signs of such a constant velocity zone but the available data are not good enough to disprove the existence of constant velocity layers.

In view of the conclusions drawn above, and the fact that these conclusions are based on profiles from all the main geological provinces of Iceland, it is reasonable to re-interpret all the seismic refraction profiles from Iceland by the use of the Wiechert-Herglotz method. This will give a more realistic picture of the Icelandic crust than the layered model. This has been done (Flóvenz, 1979) and the main results are outlined below.

Re-Interpretation of Published Data and the Relationship Between Body Wave Velocity and Alteration of the Basalt

I have re-interpreted about 70 of the 80 profiles described by Pálmason (1971). The position and direction of the profiles are shown on Figs. 1 and 2 of Pálmason (1971). For all these profiles the *P*-velocity distribution has been calculated, and where *S*-waves are available the *S*-wave velocity and Poisson's ratio has been calculated as a function of depth.

The body wave velocities at the surface vary greatly from one profile to another. The lowest *P*-wave velocity in the Icelandic basalt at the surface is slightly more than 2.0 km/s and the highest values are around 5.0 km/s. The surface velocity is lowest in the neovolcanic zone and becomes progressively greater away from it. In most of the profiles the *P*-wave velocity increases very rapidly with depth in the range 2.0 km/s to 3.5 km/s, followed by an approximately constant gradient of about 0.57 s^{-1}. This constant gradient continues down to the 6.5 km/s isovelocity surface below which the velocity becomes nearly constant. The *S*-wave velocity shows similar variations with depth.

The main geological and geophysical structure of Iceland is summarized by Pálmason and Sæmundsson (1974). Pálmason (1973) has calculated the trajectories of a lava element which reaches the surface in the neovolcanic zone. The resultant movement of such an element after it has been cooled on the surface

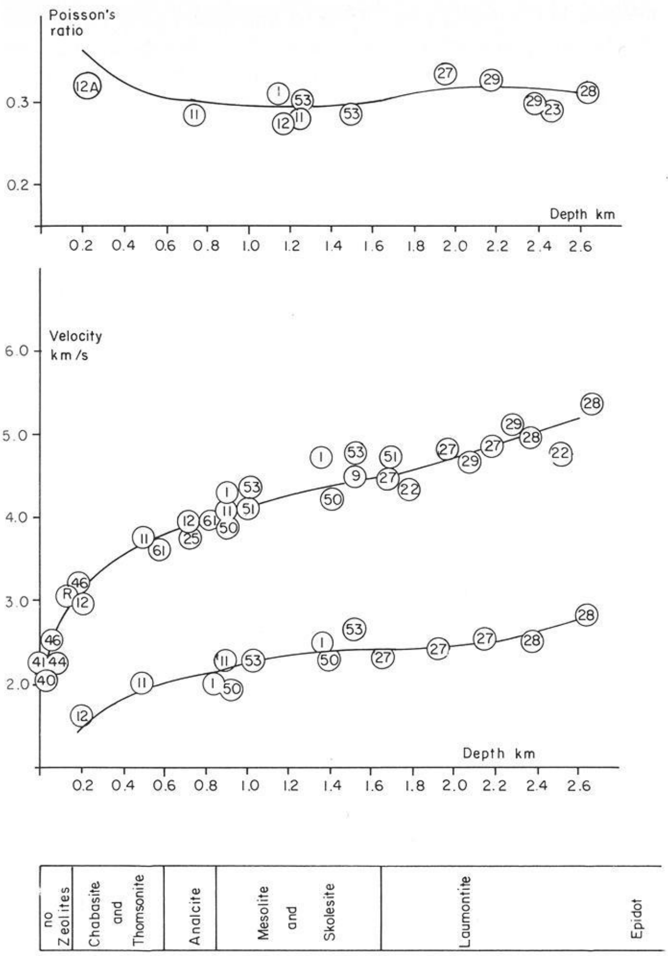

Fig. 10. The *P*- and *S*-wave velocity and Poisson's ratio as function of the alteration of the basalts based on surface values of these parameters. By assuming the thickness of the zeolite zones as given by Walker (1974) and Pálmason et al. (1978) this is equivalent to a plot of *P*- and *S*-wave velocities and Poisson's ratio versus depth in an uneroded basaltic crust, provided that the direct effect of themperature and pressure can be negledted

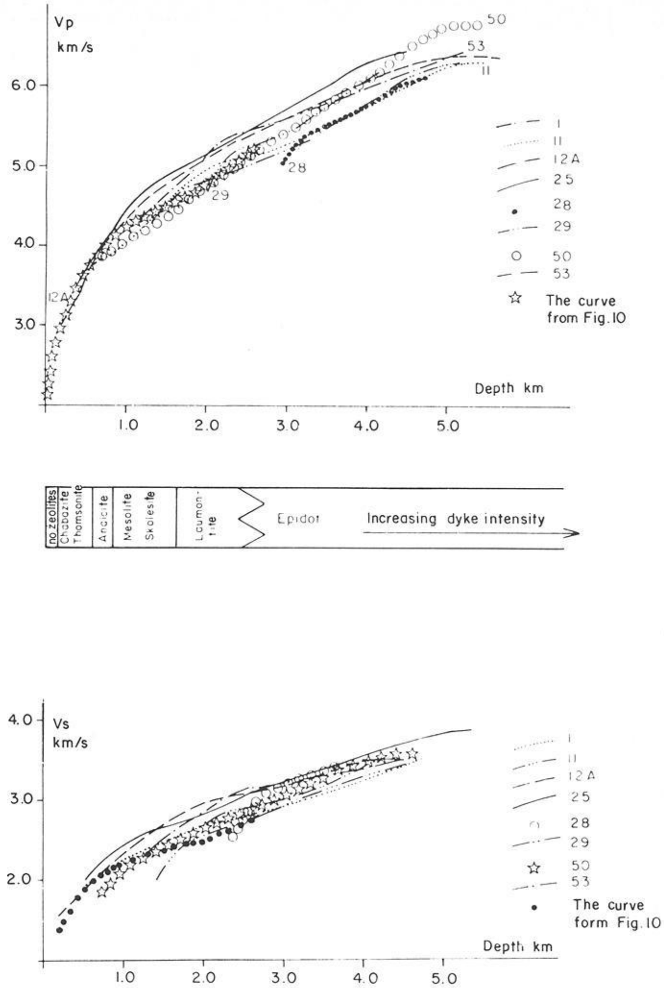

Fig. 11. The *P*- and *S*-wave velocity as function of depth in an uneroded basaltic crust. The figure is based on interpretation of profiles from different geological provinces in Iceland. Each profile is interpretated by the Wiechert-Herglotz method and brought to its relative depth in the uneroded crust as predicted by the zeolite zones. The curve on the *P*-wave figure (*asterixes*) and that one on the *S*-wave figure (*black circles*) are those of Fig. 10

is composed of a downward component of movement as a result of sagging and a lateral component prependicular to the rift zone as a result of ocean-floor spreading. As the lava element subsides it becomes reheated and some metamorphism occurs. As a result of Pleistocene glacial erosion and subsequent uplift due to isostatic adjustment, elements which have been buried down to a depth of 2 km are brought to the surface on both sides of the neovolcanic zone. Walker (1974) has studied the metamorphism produced in this manner in the basalt lavas in eastern Iceland. The low degree of metamorphism in the Icelandic crust is best described in terms of the formation of amygdale minerals. Walker (1974) has calculated the average thickness of the various zeolite zones in the plateau basalt of eastern Iceland (Table 2). Observations from Eyjafjörður in northern Iceland give similar results (Pálmason et al., 1978).

The various degrees of metamorphism near shot points on the seismic refraction profiles can be used to derive a relationship between the low-grade metamophism of the basalt and the body-wave velocity. These results are represented in Fig. 10. The surface is more or less an isotherm and an isobar. The alteration of the basalt should therefore be the only factor which influences the body-wave velocities there. As an example of how the curve is constructed, I take Profile 11. Geological evidence shows that the shot point for this profile is in the lowest part of the chabasite-thomsonite zeolite zone. This corresponds to about 0.5 km depth in the original crust (Table 2). The value of the apparent *P*-wave velocity for the direct wave to the nearest geophones is 3.7 km/s. This represents the *P*-velocity at 0.5 km depth in an uneroded basaltic crust, provided that direct effects of pressure and temperature on the body-wave velocity are small at shallow depts. By use of profiles with shot points in basalt at various stages of alteration it has been possible to construct a typical velocity-depth curve for body waves in the topmost 2.5 km in the Icelandic crust. In addition to the surface values, I have used the value of the velocity at 0.25 km depth from each profile to avoid the possibility of systematic error due to weathering. The use of these values gives the same results as the surface values. It is clear from Fig. 10 that there is no evidence for a discontinuity in the

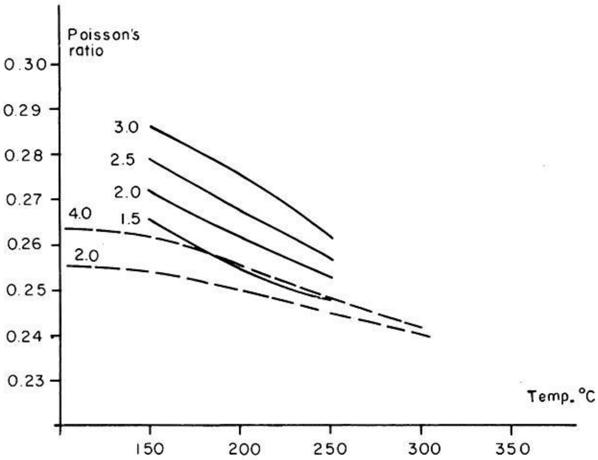

Fig. 12. The relation between Poisson's ratio and temperature for constant depth (pressure) in the Icelandic crust, based on seismic refraction measurements. The *broken lines* are the results of laboratory measurements by Hughes and Maurette (1957)

velocity-depth distribution of *P*- and *S*-waves down to the 5.0 km/s isovelocity surface for *P*-waves and 2.7 km/s for *S*-waves. This curve is constructed without any assumption as to how the velocity varies with depth. It lends firm support to the results abtained above from amplitude studies, namely that the velocity increases continuously with depth at least down to the 5.0 km/s isovelocity surface for *P*-waves.

It is possible to deduce the typical velocity-depth structure for body waves in an uneroded Icelandic basalt crust by taking the velocity-depth curves obtained by the Wiechert-Herglotz method for various profiles and displacing the individual curves to their original positions in the crust as predicted by the zeolites. This has been done in Fig. 11 for eight profiles from various geological provinces in Iceland. The curves overlap and make a narrow velocity-depth band. On the same figure the curve from Fig. 10, based on surface values only, has been plotted. This curve falls within the velocity-depth distribution obtained by use of the Wiechert-Herglotz method an gives almost the same structure. This proves the validity of using the Wiechert-Herglotz method in interpreting seismic refraction profiles from Iceland at least down to the 5.0 km/s surface for *P*-waves. What happens below this surface is not clear but there are no arguments against the idea that the velocity further increases continously with depth to layer 3. This cannot be proved, however, by the available data, and a detailed seismic profile would probably be necessary to demonstrate this.

The Relation Between Poisson's Ratio and Temperature

The body wave velocities in the Icelandic crust depend mostly on the low grade metamorphism but also on temperature and pressure. Varying dyke intensity is also likely to influence the velocities. It is, however, of no importance here because the increasing dyke intensity with depth is incorporated in the curves of body wave velocity with alteration. It is difficult to separate these factors. On the other hand, the Poisson's ratio measured at the surface shows small variations with the alteration in the depth-of-burial-interval 0.5–2.5 km of uneroded basaltic crust. By neglecting these variations it is possible to separate the effects of pressure and temperature by making use of refraction profiles positioned where the thermal gradient has been determined by boreholes. This permits the construction of a curve for Poisson's ratio as a function of temperature for constant pressure (depth). It is done by plotting Poisson's ratio versus depth as determined from the seismic profiles. By marking the temperature taken from extrapolated thermal gradients (from Pálmason et al., 1978) lines of constant temperature can be drawn and the plot can be inverted to give Fig. 12. The values on which Fig. 12 is based are fairly scattered but they do indicate the general trend. By a more accurate determination of this relationship from more detailed seismic measurements it appears to be possible to use seismic refraction measurements to obtain information about the absolute temperature in the Icelandic crust.

Central Volcanoes

The central volcanoes are an important exception to the typical velocity structure described above. As pointed out by Pálmason (1971) layer 3 lies at shallow depths under the roots of the central volcanoes. I have interpreted new seismic refraction data from the Stardalur central volcano near Reykjavík and, reinterpreted similar data from the Vatnsdalur central volcano in north Iceland (Flóvenz, 1979). The velocity seems to increase continuously with depth within the central volcanoes but at a much faster rate then outside, giving layer 3 at shallow dept (up to 1 km). After this velocity is reached, it takes on a constant value. Fan shooting over the Stardalur central volcano shows that the *P*-wave velocity in layer 3 (under the roots of the central volcanoes) is not significantly greater than elsewhere in layer 3. There appears to be a sudden change in velocity between the central volcanoes and their surroundings with much higher velocities inside the central volcanoes than at the same level outside. Thus the central volcanoes appear to act like chimneys through the Icelandic crust.

Division of the Icelandic Crust

In view of the results outlined above, it is reasonable to devide the Icelandic crust into two parts, namely the upper and the lower crust. The upper crust is characterized by continuously increasing velocity with depth from about 2.0 km/s to 6.5 km/s. In the range of 2.0 km/s–3.5 km/s the velocity increases very rapidly with depth but thereafter the velocity-depth gradient decreases and reaches an approximately constant value 0.57 s^{-1} at the 4.0 km/s surface. The lowest velocities correspond to fresh basalt lava at the surface and the rapidly increasing velocity reflects closing of fissures and pores as the basalt becomes buried. The constant velocity depth gradient corresponds to steadily increasing alteration of the basalt and the formation of secondary minerals which fill the pores. Increasing dyke density is also likely to contribute to the increasing velocity downwards. There is a possibility for the occurrence of a constant-velocity layer below the 5.0 km/s isovelocity surface. A possible interpretation of such a layer is that it represents a zone where the dyke intensity is almost constant and epidote is the dominant alteration mineral.

The lower crust (= layer 3) is characterized by a nearly constant body wave velocity and may be equated to layer 3 in the oceanic crust. A map of the depth to layer 3 according to the Wiechert-Herglotz interpretation is given in Fig. 13. The map is inaccurate in some important areas such as mid- and northwest Iceland. Reversal of some of the profiles would help to increase the accu-

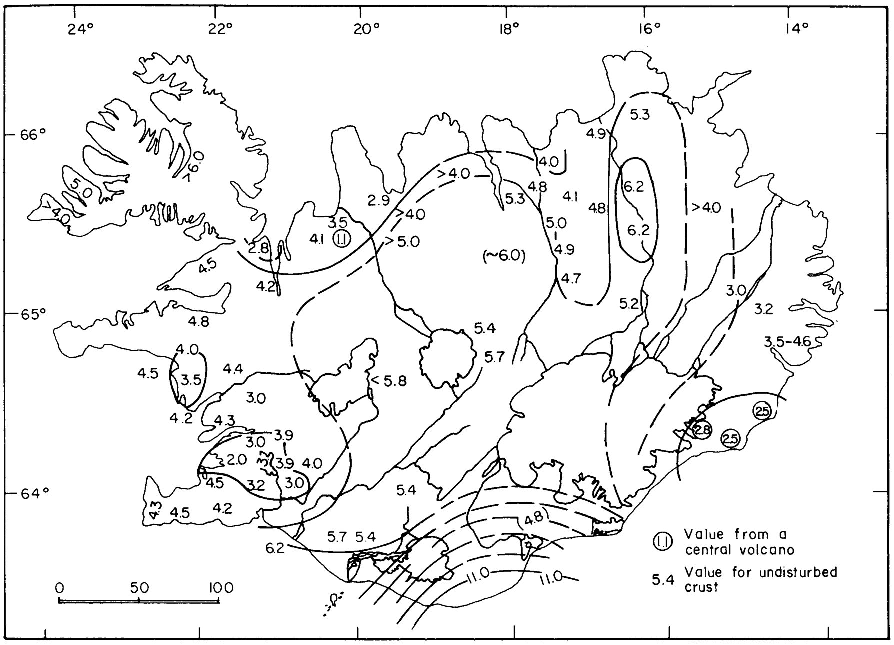

Fig. 13. Depth to the lower crust (layer 3) in Iceland. The velocity is assumed to increase continuously with depth

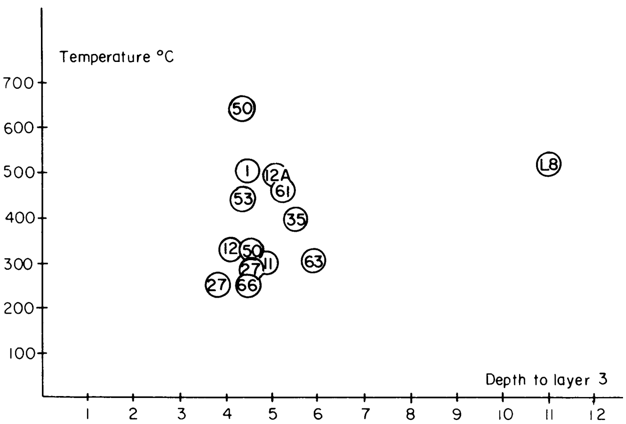

Fig. 14. Temperature at the top of the lower crust (layer 3) as function of depth for profiles outside the neovolcanic zone only. The temperature values are obtained by extrapolating the temperature gradient given by Pálmason et al. (1978) and the depth is taken from Fig. 13

racy of the map. From the map it seems not unlikely that the increasing depth to layer 3 towards the centre of Iceland can contribute to the negative Bouguer anomaly over Iceland. The anomalously great depth to layer 3 around southeast Iceland is not yet understood.

This map can be regarded as giving the maximum depth to the lower crust (correct if the velocity increases continuously with depth) and Pálmason's (1971) map as giving the minimum depth to layer three (correct if the layered model is valid).

The composition of the lower crust is not known, but seismic refraction measurements provide some constraints. The *P*- and *S*-velocities of the lower crust are 6.5 km/s and 3,6 km/s respectively, and Poisson's ratio is 0.28. Extrapolated temperature gradients from Iceland (Pálmason et al., 1978) and the map of the depth to the lower crust can be used to estimate the temperature in layer three. In Fig. 14 the temperature at the top of layer three is plotted versus the depth to the layer for profiles outside the neovolcanic zone. This shows that the top of the lower crust in Iceland lies at 500° C or lower except for one value which may be erroneous (Flóvenz, 1979). Inside the neovolcanic zone the temperature at the top of the lower crust is likly to be still higher. This means that the average temperature of the lower crust is more than 500° C and it is created at much higher temperatures than 500° C.

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