The crustal structure of the Reykjanes Ridge at 59° 30′ N

A. W. H. Bunch Department of Geodesy and Geophysics, University of Cambridge, Madingley Road, Cambridge CB3 0EZ
B. L. N. Kennett Department of Applied Mathematics and Theoretical Physics, University of Cambridge, Silver Street, Cambridge CB3 9EW

Received 1979 August 9; in original form 1979 May 23

Summary. In order to examine the development of the oceanic crust in the neighbourhood of a slowly spreading ridge, a seismic refraction experiment was carried out at 59° 30′ N on the Reykjanes Ridge. Three 120 km long overlapped split profiles were shot parallel to the trend of the ridge, on the eastern flank, and recorded on up to five recording sonobuoys. The profiles were at distances of 0, 30 and 90 km from the ridge axis, corresponding to approximate crustal ages of 0, 3 and 9 Myr. Data from the main profiles were supplemented by using a large chamber air gun during recovery of the buoys.

The analysis of the data combined standard travel-time interpretation, the ‘tau’ method of systematic travel-time inversion and detailed amplitude modelling using the Reflectivity Method to calculate synthetic seismograms. Detailed velocity–depth models were constructed for each of the profiles.

There is no indication of a significant magma chamber at the ridge crest, although a slight velocity inversion in layer 3 suggests a zone of elevated temperature. Away from the crest there was a slight positive velocity gradient in layer 3. Layer 2 was most effectively modelled by a region of varying velocity gradients, which thinned with age and the transition to layer 3 is marked by a sharp change in velocity gradient. The transition to mantle velocities is also best modelled by a high velocity gradient rather than an interface.

Although some lateral variation in properties is apparent along the profiles, the lateral velocity gradients were sufficiently weak to allow an effective analysis in terms of laterally uniform models.

1 Introduction

The crestal region of the Mid-Atlantic Ridge south of the Gibbs Fracture zone has been extensively studied by seismic refraction profiles during the last decade (Keen & Tramontini...
1970; Whitmarsh 1975; Fowler 1976, 1978; Fowler & Keen 1979). To the north of the Reykjanes Ridge there has been much less work, although a number of short (25–30 km) profiles were carried out by Talwani, Windisch & Langseth (1971).

In its southern portion the Reykjanes Ridge is dissected by fracture zones and shows the characteristic median valley of the Mid-Atlantic Ridge. From 57 to 60°N the Reykjanes Ridge has an orientation of 035° and shows a progressive change in morphology; at about 59°N the median valley is lost. The Ridge here acquires a triangular form of cross-section, and appears to break up into oblique en echelon spreading centres (Shih, Atwater & McNutt 1978) but without any obvious transform faults. Further north as the Reykjanes Ridge approaches Iceland the cross-section changes to a broader, shallower block.

The transition away from a median valley is also marked by a reduced level of seismicity relative to the southern portion of the Ridge (Francis 1973) and this could be associated with the existence of a ‘hot spot’ in Iceland.

The Reykjanes Ridge around 59°30′N is thus uncharacteristic of a slow spreading ridge (present spreading rate ~ 1.0 cm yr⁻¹) since, as is common with fast spreading ridges, it has no median valley.

Figure 1. Reference map of survey area and detailed layout of refraction lines. Lines X, Y, Z were shot as overlapped split profiles into recording sonobuoys marked by the dark stars. Line W was reversed between the OBS array, marked by hexagons, and a disposable sonobuoy (open star). The OBS array recorded arrivals from lines X, Y, W (see Fig. 20).
In 1977 July a seismic refraction experiment was conducted close to the crest of the ridge in this region by Cambridge University in association with the Institute of Oceanographic Sciences (Fig. 1). The object of the experiment was to find a detailed seismic structure for this portion of the ridge to enable comparisons to be made with other well studied regions further south on the Mid-Atlantic Ridge at 37° N (Fowler 1976) and 45° N (Fowler 1978; Fowler & Keen 1979).

Three major refraction lines, each 120 km long were shot into up to five free floating recording sonobuoys (lines X, Y, Z on Fig. 1) and in addition a short tie line W was shot into an array of OBS (Francis & Porter 1977) emplaced for seismicity studies. This array also recorded shots from the southern portions of lines X and Y. The refraction lines were oriented to be parallel to the trend of the Ridge, in an attempt to reduce the level of lateral variation in seismic properties along the profile and to follow an isochron. The approximate crustal ages for the lines are 0 Myr (line X), 3 Myr (line Y) and 9 Myr (line Z).

2 Conduct of experiment

On each of the three main lines we adopted the same basic shooting pattern along a 120 km line with two separated groups of receivers in the central portion of the line (Fig. 2) giving a maximum source–receiver offset of 65 km. This configuration gives a high data density with 24 shots recorded by four free floating sonobuoys. Thus we obtain a set of overlapped split profiles from the buoys with reversal on the upper crust in the central section, as can be seen from the pattern of arrivals shown in Fig. 2. This experimental design was intended to overcome the problem of charge size selection in single ship reversed profiles. The charge sizes employed were determined using experience gained on previous refraction profiles on the Mid-Atlantic Ridge (Fowler 1976) and to minimize the variability of source waveform along the profile only three charge sizes were used. In practice every shot fired produced a
readable and useable amplitude at each working receiver (see e.g. Figs 3–5). During the recovery of the buoys supplementary short range information was obtained using a 1000 in³ air gun.

In addition to try to link lines \(X\) and \(Y\) a reversed tie-line \(W\) was shot between a disposable sonobuoy and the OBS array at the southern end of line \(X\) with seven shots (Fig. 20).

On the two outer lines \((Y, Z)\) normal incidence reflection profiles were conducted along the refraction lines in order to provide information on sediment thicknesses along the lines; however poor weather prevented complete coverage. The limitation of using free floating sonobuoys is that it is difficult to determine the water depth beneath each buoy accurately. For lines \(X\) and \(Z\) there was no strong topography beneath the receivers but for line \(Y\) the profile was shot along the flank of a valley (Fig. 1) which gave rise to some difficulties with corrections.

Along the crestal line sediment cover is sparse and there are indications of structure oblique to the overall trend of the Reykjanes Ridge, but these features are not long enough to allow individual refraction lines. Since line \(X\) lies along the ridge trend our results will represent an average of the properties in the immediate crestal areas.

Navigation along the lines was primarily controlled by satellite fixes but frequent Loran C fixes were used to give relative positioning. At stations on lines \(X\) and \(Z\) sound velocimeter profiles were obtained to the sea floor and these were used to calculate the shot–receiver ranges by ray tracing sound through the water column. These ranges were compared with those derived from the navigation to give an estimate of sonobuoy drift. On all three lines only slight net drift occurred during the firing of the main charges. An incidental advantage of the shooting scheme we have used is that a visual check and accurate positioning is available for each sonobuoy as the ship passes through the central section of the line.

### 3 Data reduction

For each of the profiles digital record sections were prepared for every sonobuoy. These sections were corrected to a common water depth for each profile by compensating for the varying water delay times beneath shots and receivers. The correction factor used (Kennett & Orcutt 1976) assumes that all refractors mirror the sea floor topography. This is probably a reasonable hypothesis for lines \(X\) and \(Z\) where the topography was smooth, but for line \(Y\) rougher topography increases the difficulty of estimating corrections.

The record sections presented in this paper (e.g. Figs 3–5) have been filtered with a zero phase band pass filter with 3 dB points at 3 and 20 Hz, since analysis of averaged spectra of noise and signal samples indicated that the best discrimination against noise was in this band.

To compensate for variations in charge size and to make a partial allowance for geometrical spreading the amplitude of a seismogram at range \(R\) with charge weight \(W\) has been scaled by

\[
R/W^{0.64}
\]

using the empirical scaling factor given by O'Brien (1960). The factor \(R\) scales the arrivals as though they have been continuously refracted (Kennett 1977). This scaling factor has been applied to all the record sections and amplitudes presented in this paper.

Arrival times were picked from unfiltered records, but filtered sections provided a useful check on correlations between records. The times for lines \(Y\) and \(Z\) were also corrected for varying sediment thickness by subtracting an estimate of the sediment delay

\[
\Delta t = (T/\alpha_{x}) \left( \frac{\alpha_{s}^{-2}}{V_{app}^{-2}} \right)^{1/2}
\]
Figure 3. Composite record section of arrivals on line X. All good recordings on the inward (southern) leg are shown in the upper panel, and from the outward (northern) leg in the lower panel. No allowance has been made for the offset between receiver groups. Scaling for charge weight and range has been applied.

where $T$ is the vertical incidence in sediment delay, $v_s$ the sediment velocity and $V_{app}$ the apparent velocity of the arrival. A sediment velocity of 1.64 km s$^{-1}$ was adopted (see Fig. 5).

Composite record sections showing the majority of all recordings for each line are presented in Figs 3–5 (some omissions are made for clarity). The records from the inward leg at each buoy and the outward legs of the split profiles are separately presented.

Figure 4. Composite record section of arrivals on line Y, as in Fig. 3.
For each of the lines we note some gradation in behaviour along the profile with slightly larger amplitude on the outward (northern) end of each profile. On lines X and Z we note that we have a very consistent pattern of arrivals, and recordings at different buoys at the same range are almost identical. Indeed for these profiles the arrival times from all buoys as a function of range could be overlaid. Although the travel times can be fitted by a structure without lateral variations in elastic properties, the amplitude behaviour implies we have some lateral variation present. On line Y there is a more significant difference in behaviour between the inward and outward legs of the line. At the northern end of lines Y and Z we note that there is a very significant drop in amplitude for the shots at greatest range.

For line Y the difficulties of estimating water corrections lead to less consistent raw travel-times than for lines X and Z, but once systematic offsets between the travel-times to different buoys had been eliminated the data showed only a limited scatter.

4 Travel-time analysis

The travel-times for each of the three main lines were first interpreted using simple uniform layer techniques and then used to define bounds on the velocity-depth behaviour (Kennett & Orcutt 1976) to assist in subsequent amplitude and waveform modelling.

As far as possible all travel-time picks were of first arrivals. With the relatively shallow water over the ridge second arrival branches are difficult to trace because of interference between arrivals reflected at the sea surface and the seabed. In order to group arrivals so that they might be reasonably assigned to a single 'refractor' we have made use of amplitude criteria. Thus if there is a rapid change in the corrected amplitude distance curve at a particular range then this was taken as a likely point to change from one refractor to another. These ranges were then checked for consistency between the buoys.

Once first arrival groups were assigned, least-squares straight lines were fitted to the data and the residuals and variances determined. Where there was some ambiguity in cross-over
Crustal structure of the Reykjanes Ridge

Figure 6. Preliminary velocity depth models for lines X, Y and Z derived from first arrival refraction arrivals and wide angle reflections. Also indicated are velocity depth bounds derived from a $\tau(p)$ analysis of the travel-times.

range between refractors, this range was adjusted until the variance of the joint fit of the two crossing lines was minimized (Fowler 1976). This procedure generally allowed the cross-over points to be determined to better than 2 km.

Preliminary velocity—depth models determined from the least-squares analysis are shown in Fig. 6 with the sediment cover removed. These structures represent averages over the results from different sonobuoys and so represent an attempt to generate a laterally uniform approximation to the structure on each line. There were no indications of significant dip on any of the refractors on each of the three lines.

At the northern end of line Z we unfortunately had incomplete reflection profile coverage and thickish sediment (> 200 m) so that the errors are rather large and as a result the depth to the 8.2 km s$^{-1}$ refractor is not tightly constrained.

On line Y there is more variability along the length of the line, but the various corrections applied to the times make the results less reliable than for X and Z. The pattern of arrivals in the range 10–25 km suggested the existence of a velocity gradient and this has been modelled with three thin layers. At the southern end of line Y there is an indication of an arrival with a velocity of 7.8 km s$^{-1}$; whilst from the outer shots on the northern end there is possibly a weak arrival with an ill determined velocity greater than 8.2 km s$^{-1}$ whose intercept corresponds to a considerable depth, at least 14 km below sea-bed.

The first arrival times were also used in an adaption of the procedure of Bessonova et al. (1974), due to Kennett & Orcutt (1976), which allows the construction of bounds on possible velocity depth models via the intercept time $\tau$ as a function of ray parameter $p$.

With our high data density we have been able to use the data groupings established for the least-squares analysis to try to estimate the $\tau(p)$ segment corresponding to each branch of the travel-time curve. For each of these segments we used the graphical construction method of Bessonova et al. (1974) and found $\tau(p)$ from the maxima of reduced time plots. We assigned timing errors of not less than 0.03 s and used a smoothing spline technique (Kennett 1976) to compensate for the random errors and unevenness in data coverage. When excessive clustering of points occurred we used the mean over the cluster.

Since we only have first arrival information, the individual branches of the travel-time curve give only patchy coverage in ray parameter $p$. The gaps in the bounds on the $\tau(p)$ distribution were bridged using the method of parallelograms (Keilis-Borok 1971). This approach combined the monotonicity of $\tau(p)$ with estimates of the slope of $\tau(p) [-X(p)]$, from the minimum range for the faster arrival and the maximum range for the slower arrival, to extend the upper and lower bounds on $\tau(p)$. 
Velocity depth bounds were derived from the $\tau(p)$ bounds by a modification of the Wiechert–Herglotz technique (Bessonova et al. 1974; Kennett 1976) and are superimposed on the preliminary velocity depth models in Fig. 6. The bounds bracket these preliminary models, and for lines X and Z where we were able to split the data up into groups give strong indications of structures in possible velocity models. For line Y the errors were greater and so separation into travel-time branches was not so effective and relatively smooth bounds are obtained; as in the work of Kennett & Orcutt (1976) where the data density was insufficient to allow grouping.

These velocity depth bounds provide useful constraints on the possible velocity depth behaviour for subsequent detailed modelling of amplitudes and waveforms. Further by shrinking the bounds on the $\tau(p)$ distribution to zero we may find a smooth model satisfying the travel-time data, more conveniently than by linearized inversion (Kennett 1976).

With the two extreme classes of velocity–depth distribution (uniform layers and smooth models) available for each profile we can judge how to introduce a compromise velocity model which satisfies the amplitude information as well as the travel-times.

5 Supplemental information on near seabed structure

Additional short range information (< 18 km) was available from air gun data recorded on the sonobuoys during the process of recovery and from disposable sonobuoys launched during the reflection profiles.

5.1 Line X

A refracted arrival was visible over a limited range which, when tied to an intercept at the sea-bed reflection time (cf. Talwani et al. 1971), gave an apparent velocity of 2.23 km s$^{-1}$. At greater range a high amplitude wide angle reflection was visible from 6.7 to 10.2 km with peak amplitude around 8 km. This could be well fitted by a model with a layer 0.2–0.4 km of velocity 2.23 km s$^{-1}$ overlying about 1.6–2.0 km of 4.4 km s$^{-1}$, underlain in turn by the 5.5 km s$^{-1}$ layer seen on the longer range refraction work. The finite frequency of the source will displace the maximum reflected amplitude from the critical distance (~ 6.5 km) to around 8 km in good agreement with the observations.

The peak-to-peak amplitudes of the sea floor reflections were corrected for geometrical spreading through the water column and used to estimate reflection coefficients. Matching the behaviour with simple interface models suggested a thin veneer of sediment with velocity 1.62 km s$^{-1}$ overlying the 2.23 km s$^{-1}$ layer seen before.

This information was used to construct the top 2 km of the velocity model for line X shown in Fig. 6. The broad velocity bounds in this range reflect the lack of control from the refraction work.

5.2 Line Y

A strong reflected arrival was observed to fade in and out as the range from a disposable sonobuoy increased with sharp variations in amplitude with distance (Fig. 7), and this pattern was mirrored in the water multiple arrivals. The arrival times could not be fitted with a simple reflection from the base of a layer overlain only by the water column since this led to results which were incompatible with the refraction results.
Figure 7. An amplitude–distance plot of the multiple wide angle reflections seen on line Y. The amplitude of the last two peaks are contaminated by noise.

These arrivals can however be modelled as a multiple reflection in the upper part of the crust between the near seabed and the velocity transition at about 1.2 km. The underside of the seabed has high reflectivity at a wide range of angles of incidence whilst a large reflection from the transition will arise near the critical angle. A multiple of this type would have the regular repetition in time and range observed and the offsets match well with those calculated for the refraction velocity model. The sharpness of the amplitude peaks suggests that the reflection is not just a simple critical angle reflection but a caustic formed from a reflection off a velocity gradient (Červený & Zahradník 1972). The 5.2–6.2 km s\(^{-1}\) layers in the preliminary velocity depth model (Fig. 6) are an attempt to model this gradient which as we have noted also influences the first arrival times.

Strong second arrivals on this line were observed with an apparent velocity of 3.4 km s\(^{-1}\) and intercept 2.35 s and these may best be explained as converted shear waves refracted at the top of the transition zone we have just described.

5.3 LINE Z

Sediment velocities and thicknesses were here obtained from wide angle reflections and for the two buoys still recording during recovery we found 0.45 km of 1.62 km s\(^{-1}\) and 0.21 km of 1.52 km s\(^{-1}\) material. These are similar to the results of Talwani et al. (1971) from farther north.

Clear first arrival refractions showed apparent velocities of 3.8 and 4.7 km s\(^{-1}\) and tied in well with the sediment thickness. These results have been incorporated into the velocity model in Fig. 6. There were also some suggestions of a velocity gradient at the base of the 4.6 km s\(^{-1}\) layer from peaked reflections as on line Y.

6 Amplitude–distance behaviour

To aid the process of more detailed modelling we have constructed amplitude–distance curves for each of the lines. Consistent amplitude readings were obtained by taking the maximum peak to trough amplitudes in the first arriving energy on each seismogram. As previously noted we apply a charge weight correction and a linear spreading factor with range to all the amplitudes.

Although the travel-time data could be fitted by a laterally uniform velocity model, the variability in amplitudes from sonobuoy to sonobuoy visible in Figs 3–5 indicates the existence of significant lateral variation in structure, as indeed would be expected.
The varying density of shot-receiver pairs with range makes it difficult to construct an objective measure of amplitude variability. However we have found it convenient to consider, for each profile, an amplitude ‘snake’ (Fig. 8) whose character mirrors the amplitude behaviour and whose width gives a measure of the variability of the amplitude at the different receivers.

The measured amplitudes will include interference effects between different phases and as we see show rather different character for the three lines.

For line X consistently slightly higher amplitudes were recorded for the shots on the northern part of the line, but the character of the amplitude distribution matched that from the southern shots. On line Y there was a very sharp drop in amplitude at the northern end of the line. This low amplitude coincides with the recording of the 300 lb shots. We have therefore the possibility of strong lateral variations in structure leading to an amplitude shadow or of partial misfires. Inspection of hull geophone recordings on the shooting ship indicates that it is very unlikely that less than 150 lb fired. Even if we correct for the smaller charge size we are faced with a very rapid decrease in amplitude with range which will prove difficult to match.

A similar sharp amplitude drop occurs for the northernmost shots on line Z (cf. Figs 4 and 5). The location of the shots giving low amplitudes suggest a lateral change in structure correlating with a topographic depression just to the east of the northern end of line Y. This feature also seems to tie in with an offset in the oblique trends on the ridge crest.

On lines X and Z we note that we have definite amplitude minima between 20 and 40 km, more pronounced and narrower for line X than line Z; these features provide strong constraints on possible velocity models.

7 Synthetic seismogram modelling

The amplitude—distance ‘snakes’ provide very convenient discriminants on the behaviour of particular velocity—depth models, but the width of the ‘snakes’ makes it inappropriate to undertake detailed waveform modelling on any individual set of seismograms. We have, however, attempted to match the general character of the observed seismograms with synthetic seismograms calculated for proposed velocity models.

The synthetic seismograms were calculated using the Reflectivity Method (Fuchs & Muller 1971) modified to include attenuation (Kennett 1975) and to allow for a water
overburden (Fowler 1976; Orcutt, Kennett & Dorman 1976). This approach requires a model composed of homogeneous layers and it was found convenient to modify the velocity depth model by the addition of layers, adjusting the layer thickness to maintain the intercepts of first arrival refractions. As a guide to these modifications the position of the maximum amplitude corresponding to reflection from a particular interface was estimated to be 4/3 of the critical range, allowing for the finite frequency of the source (Červený & Zahradník 1972).

The restriction to uniform layers does not impose any problems when modelling gradients by a staircase of layers, provided that the steps are thinner than a quarter of the wavelength at the dominant frequency.

In all the calculations we have set the $P$-wave velocity ($\alpha$) to $S$-wave velocity ($\beta$) ratio to $\sqrt{3}$ and estimated the density from $\rho = 0.252 + 0.3788\alpha$; since the $P$-wave response is relatively insensitive to the form of these relations. For most synthetic seismograms a simple signal (dominant frequency 5 Hz) was assumed for all shots, but final comparisons were made including the effects of variation in charge weights on the source pulses (see e.g. Kennett 1977). The change in signal form improves the appearance of the seismogram but does not lead to any need to alter the velocity model.

7.1 LINE X

The amplitude—distance behaviour for this line (Fig. 8) has already been noted to show a consistent pronounced amplitude minimum between 20 and 40 km, but this feature does not appear on the seismograms calculated for the preliminary velocity model (Fig. 6). For comparison we show the seismograms recorded by one sonobuoy on the inward leg. The main areas of deficiency in the response of the preliminary model are: (i) too little amplitude at 15 km, (ii) too much amplitude at 30 km, this latter feature arises from the critical reflection from the top of the 7.1 km $s^{-1}$ layer.

![Figure 9. Experimental record section for one sonobuoy from line X and synthetic seismograms for the preliminary velocity model for line X. The observed seismograms are scaled for charge weight and both sets are scaled linearly with range.](https://academic.oup.com/gji/article-abstract/61/1/141/601877)
Modifications of the very near sea-bed structure improved the close range behaviour but had little influence on the seismograms beyond 12 km, as may also be seen in the work of Spudich & Helmberger (1979). However much more significant effects were achieved by changing the structure between 3 and 6 km below the sea floor by the introduction of a layer with velocity intermediate between 6.4 and 7.1 km s\(^{-1}\). An initial trial with a 6.7 km s\(^{-1}\) layer gave an amplitude low at 25 km and a peak at 45 km but with an insufficiently deep amplitude minimum. The minimum could be successfully matched with a 6.8 km s\(^{-1}\) layer but at the expense of displacing the amplitude peak to 50 km. This latter feature can be counteracted by introducing a very slight velocity reversal beneath a 6.8 km s\(^{-1}\) lid so that the average velocity is close to 6.7 km s\(^{-1}\). The solution adopted after a number of trials was to have 0.5 km underlain in turn by a further 0.85 km of 6.8 km s\(^{-1}\) material. In addition to sharpen the amplitude peak at 45 km the transition from 6.8—7.1 km s\(^{-1}\) was modified to a sharp gradient modelled by two 0.25 km thick layers.

The seismograms for this 'final' structure with due allowance for varying charge weight are compared with the experimental data in Fig. 10. The computed results agree well with this set of seismograms and with the general characteristics of the recordings for line X (Figs 3 and 10). After about a 1.5 s delay water multiples occur in the experimental data but they have not been modelled in the synthesis.

The final velocity—depth model and the preliminary model are compared in Fig. 11. It has still not been possible to introduce much detail into the shallowest structure but the transition from velocities around 5—6.4 km s\(^{-1}\) appears to represent a gradual change of properties. The velocities beneath this zone are typical of those associated with 'layer 3', even including the velocity reversal. The data provide very little constraint on attenuation within the reversal and \(\text{Q}\_a\) can be made quite low before any significant effects are introduced. \(\text{Q}\_a\) in this zone is probably greater than 50 and no attenuative contrast across the reversal is required by the data. The highest velocity seen is 7.1 km s\(^{-1}\) and we have no evidence for higher velocities at greater depth from later arrivals over the range span of our data.

Figure 10. Synthetic seismograms for proposed model for line X, with allowance for varying source waveform with range.
Crustal structure of the Reykjanes Ridge

In Fig. 11(a) we have also indicated the smooth velocity model obtained from inverting the $\tau(p)$ estimates with zero assigned errors. The final velocity model has a character which has a greater resemblance to the smooth model than the preliminary model, and the resulting amplitude distance curve fits well into the amplitude distance 'snake' for line X as may be seen in Fig. 11(b), except at the largest ranges. An improved fit would require increased attenuation in the 7.1 km s$^{-1}$ layer or a negative velocity gradient beneath a 7.1 km s$^{-1}$ lid, or some combination of these effects.

It is difficult to provide any objective measure of the uncertainty in the velocity model. We have to recognize that as a consequence of the finite frequency of the source it would be difficult to resolve features much less than 0.15 km in extent. The data constraints are also uneven; we have for example less control on upper crustal structure for this line than is desirable. Small velocity perturbations could certainly be made to the model without affecting the fit to the data but the gross features in terms of velocity gradients must remain (at least within our assumptions of a laterally uniform model).

7.2 LINE Z

A similar approach was adopted to try to improve the velocity model for this line, since the synthetic seismograms for the preliminary model (Fig. 12) do not show the sharp fall in first arrival amplitude noted on the experimental data. Once again we present a set of observed seismograms from a single sonobuoy.

The structure was varied to try to produce the amplitude contrast between 15 and 25 km and in Fig. 13 we show the amplitude distance curves from these trials superimposed on our average amplitude 'snake'. The models were:

(a) The 4.6—6.6 km s$^{-1}$ material was replaced by two velocity gradients chosen to preserve the intercepts corresponding to apparent velocities of 6.0 and 6.6 km s$^{-1}$. It was hoped that this change would increase the contrast beyond 15 km, but an amplitude minimum occurred at too short a range corresponding to screening of supercritical reflection amplitude by the gradients (Cervený & Ravindra 1971).

(b) The 6.6 km s$^{-1}$ layer was made attenuative ($Q_\alpha = 100$) to damp the critical reflection from the top of the 7.06 km s$^{-1}$ layer at a range of 14.2 km. The amplitude—distance curve

\begin{figure}[h]
\centering
\includegraphics[width=\textwidth]{figure11}
\caption{
(a) Comparison of velocity models for line X. --- Preliminary model, --- proposed model, --- smooth model from $\tau$ analysis. (b) Comparison of amplitude—distance behaviour for proposed model with 'snake'.
\end{figure}
Figure 12. Experimental record section for one sonobuoy from line Z and synthetic seismograms for the preliminary velocity model for line Z. The observed seismograms are scaled for charge weight and both sets are scaled linearly with range.

has the right character but does not decay quite fast enough. A good fit could only be produced by a very attenuative zone which would greatly damp arrivals from deeper layers.

(c) The 6.6–7.06 km s$^{-1}$ interface was replaced by a velocity gradient intended to focus energy at less than 20 km and thus increase the amplitude contrast between 15 and 25 km range. An amplitude peak was formed as expected but it occurred at too great a range producing amplitude minima at 14 and 26 km range.

(d) A 6.8 km s$^{-1}$ layer was introduced above the 7.1 km s$^{-1}$ medium with consequent thinning of the 6.6 km s$^{-1}$ layer as in line X. The layer was within the $\tau$–$p$ velocity–depth bounds and gave critical ranges of 11.5 km for reflection from the top of the layer and 19.4 km from the base. This trial produced a sharp amplitude drop from 15 to 20 km of the right order. A similar but improved solution was to replace the step at the top of the 6.8 km s$^{-1}$ layer by a staircase with velocities 6.66, 6.76 and 6.86 km s$^{-1}$ all designed to have a critical range of 13 km.

The model (d) provides an acceptable representation of the short range seismograms and there remains the problem of producing the amplitude peak at 40 km. This could be

Figure 13. Comparison of amplitude–distance behaviour of models a, b, c, d (discussed in text) with amplitude distance 'snake' for line Z.
achieved by replacing the interface between 7.1 and 8.2 km s\(^{-1}\) by a velocity gradient of 0.66 km s\(^{-1}\) leading to a reduction in reflected amplitude on either side of the cusp at 40 km (Braile & Smith 1975; Červený & Zahradník 1972). A single intermediate layer with velocity between 7.1 and 8.2 km s\(^{-1}\) does not give such a good match to the amplitude distribution beyond 55 km, but a combination of an intermediate layer and a gradient would be a suitable alternative to the simple gradient.

The synthetic seismograms for the model combining the upper structure from (d) with the gradient into the 8.2 km s\(^{-1}\) layer are compared with some of the experimental data for line Z in Fig. 14. The agreement is quite good, but it must be recalled that there was greater amplitude variability on line Z than on line X so that the discrepancies between the synthetics and the data are only of the same order as between those at different receivers. A good fit of the amplitude behaviour to the amplitude distance ‘snake’ for this line has been achieved as can be seen from Fig. 15(b). The amplitudes at large ranges are a little high and this suggests that a slight negative velocity gradient or increased attenuation underlies the mantle transition.

The velocity model used to generate the synthetics in Fig. 14 and the preliminary model are compared in Fig. 15(a), where the smooth model from tau inversion is also indicated. For this line we do not have such a close fit to the smooth model for the upper crustal structure which was in part constrained by air gun data which was not used in the \(\tau(p)^\) analysis. The main changes from the preliminary model occur in the more gradual transition into ‘layer 3’ and in the Moho transition; although as we have noted the details of this transition could be modified and still give a suitable response, we would not have a simple interface.

7.3 LINE Y

The analysis of this line presented different problems. For the previous two lines the structure had been quite well defined by the first arrivals, giving good control to the synthetic seismogram analysis.
For line Y the timing control imposed by the first arrivals was weaker because of the difficulties with corrections. There is also a significant variation in seismogram character between the inward (S) and outward (N) legs of the line (Fig. 4). We have therefore used the amplitude distance curves as a guide to the general behaviour of the model and then look for laterally uniform models which have the appropriate character for the two legs.

Both inward and outward legs show an amplitude dip around 18 km with some recovery to 30 km. On the inward leg the amplitude then decays but on the outward leg increases to a maximum at 40 km but then diminishes extremely rapidly.

Figure 16. Experimental record section for one sonobuoy for outward leg on line Y and synthetic seismograms for preliminary velocity model for line Y. The observed seismograms are scaled for charge weight and both sets are scaled linearly with range.
In Fig. 16 we compare synthetic seismograms for the preliminary model with a set of seismograms for the outward leg. There is too little amplitude around 10 km and a large undesired amplitude peak at 22 km. Also compared with the outward leg there is insufficient amplitude at 40 km.

The amplitude peak at 22 km arises from constructive interference between the critical reflections from the top of the 6.8 and 7.4 km s$^{-1}$ layers. To try to break up this interference and to increase the energy before 16 km the 5.8 and 6.2 km s$^{-1}$ layers were replaced by a single 6.0 km s$^{-1}$ layer. This had the desired effect of increasing the short range amplitude but led to a dearth of amplitude from 30 to 50 km. Although not appropriate to the outward leg of line $Y$, this behaviour matches quite well the data for the inward leg as may be seen in Fig. 17.

From the experience gained on the previous lines intermediate structure was introduced between the 6.2 and 7.4 km s$^{-1}$ layer to try to improve the fit to the outward leg. The general behaviour could be quite well matched out to 42 km by a shallow velocity staircase resembling a weak velocity gradient as can be seen from the comparison of the synthetics with outward leg seismograms in Fig. 18. Beyond 45 km there is far too much energy in the synthetics. This could be reduced by making the lower portion of the model extremely attenuative. However as discussed above (Section 6) it is likely that the very low observed amplitudes correspond to transmission across a lateral change of structure.

In Fig. 19(a) we compare the model corresponding to the synthetics in Fig. 18 with the preliminary model and the smooth model from tau inversion. The errors in the travel-times reduce the utility of the tau model for this line but we see it reflects the general character of the distribution quite well. The main change introduced in the synthetic modelling for the outward leg is to make "layer 3" a weak gradient zone.

The models proposed for the inward and outward legs are compared in Fig. 19(b). We see that on the southern inward leg there does not seem to be a noticeable change of velocity gradient around 6 km s$^{-1}$ and we get a shallower transition to both the 6.8 and 7.4 km s$^{-1}$ layers. The southern part of line $Y$ did however show rather more rugged relief which reduces the reliability of this model.
To put these two slightly different structures for the northern and southern parts of the line into context, we must recall that neither represents a vertical depth section beneath the receiver point, and that both have been derived subject to assumptions of lateral homogeneity. Approximately, at least, these models are sections along the locus of seismic turning points within the structures. Thus the deeper parts of the models correspond to larger offsets from the receiver of roughly half the range at which the arrival is seen. For a 7.4 km $s^{-1}$ arrival we therefore have an offset of around 20 km from the source.

This simple picture suggests that we may have lateral velocity gradients of up to 0.03 km $s^{-1}$/km for velocities around 6.0 km $s^{-1}$, which compares with vertical gradients of about 0.5 km $s^{-1}$/km in this velocity range and at least 0.1 km $s^{-1}$/km at greater depth. The main features of our velocity structures should therefore be reasonably well determined by an analysis in terms of laterally homogeneous structures.
Since line Y is displaced south eastwards by about 10 km compared with the configurations of lines X and Z we have used the velocity model proposed for the northern (outward) portion of line Y for comparison purposes.

8 OBS results

The pattern of shots recorded at the OBS array is shown in Fig. 20, the line W shots were also recorded at the disposable sonobuoy.

On line X arrivals were obtained out to 62 km range and from least-squares fitting we have an apparent velocity of 6.5 km s$^{-1}$ out to 35 km and 7.0 km s$^{-1}$ beyond this range. These velocities agree quite well with those obtained from the sonobuoy work on line X, but the intercept times at both the OBS are 0.4 s greater than those expected from the sonobuoy results.

The first arrivals from line W were analysed at the disposable sonobuoy and for both the working OBS. At the sonobuoy we have a refractor with velocity of 6.3 km s$^{-1}$ and an intercept of 2.62 s. At the OBS we have a poorly determined refractor with velocity around 6.3 km s$^{-1}$ (certainly less than the 6.5 km s$^{-1}$ seen from line X) and an apparent velocity at greater ranges of 7.6 km s$^{-1}$. For both refractors there is an offset in intercept between OBS III and OBS IV, on the western flank of the ridge, close to the 0.4 s already noted for line X.

The results from line Y did not warrant a separate interpretation but were used to extend the results for line W. The arrivals at OBS IV were of poor quality, due to high background noise, but were offset from those at OBS III, whilst both sets of observations were delayed relative to line W. Our topographic corrections assume that refractors mirror the sea floor but the southern end of line Y lies above a steep sided valley with sea floor relief of up to 500 m within 1 km of most shots, along the ray paths joining the shots and the OBS. The observed delays for line Y, relative to line W, may be accounted for by assuming that all the relief is due to thickened upper crust.

![Figure 20. Arrivals recorded at OBS array. ▲, ▲ Delayed arrivals at OBS III, IV, v delayed arrivals at OBS IV, △ additional crustal delays.](https://academic.oup.com/gji/article-abstract/61/1/141/601877)
If we take the sonobuoy results for lines X and W as a reference then the pattern of delayed arrivals is as indicated on Fig. 20: OBS III — all shots from line X and shot 7 of line W, OBS IV — all shots from lines X and W. The delay is of the order of 0.4 s and would be consistent with a region with markedly reduced crustal velocities in the speckled area of Fig. 20. Since we have noted a delay on the refractors with velocities between 6.3 and 6.5 km s\(^{-1}\) it would appear that the delays arise in the uppermost crust perhaps due to shattering of the rocks. In this context we may note that the OBS array was dropped at the most northerly point of the active teleseismic zone on the Reykjanes Ridge. Also near OBS IV there is a topographic low through which water rushes across the ridge crest giving high noise levels at this OBS.

The line W results as seen at OBS III and the disposable sonobuoy could be fitted by a linear transition from a structure appropriate to line X to that for line Y with a thinning of the uppermost crust, as the distance from the ridge crest increases. This gives a slight dip (< 2\(^{\circ}\)) on the 6.3 km s\(^{-1}\) refractor towards the axis and the apparent velocity of 7.6 km s\(^{-1}\) can be explained by a return from a refractor with velocity 7.1 km s\(^{-1}\) and a dip of about 3\(^{\circ}\) as given by the progressive transition model.

9 Comparison of velocity models

The velocity depth models suggested from the combination of travel-time and synthetic seismogram analysis described above are compared in Fig. 21. If the spreading rate in this survey area (Fig. 1) has been maintained for the last 10 Myr then line X is approximately 0 Myr, line Y is 3 Myr and line Z is 9 Myr old.

The set of velocity models enables us to make some characterizations of the process of crustal ageing on this portion of the Reykjanes Ridge. It is convenient to divide the models up into velocity bands and these are indicated on Fig. 21.

The trends with increasing age are:

(a) The velocity of the sea-floor refractor increases. At the Ridge axis the surface refractor has a velocity of 2.2 km s\(^{-1}\) which rapidly rises to 3.8 km s\(^{-1}\) for crust 9 Myr old. The layer has a relatively constant thickness of 400 m.

(b) The 4.6—4.7 km s\(^{-1}\) layer becomes thinner but its velocity remains relatively constant. The layer thins from 1.3 km thick at the Ridge axis to 0.8 km thick at 9 Myr old.

(c) The 5.4—6.2 km s\(^{-1}\) layers become shallower but their combined thickness and mean velocity remain relatively constant. The velocity structure of these layers is not well defined.

Figure 21. Comparison of proposed velocity models for lines X, Y and Z. The letters refer to the text discussion.
Crustal structure of the Reykjanes Ridge

as neither the first arrivals nor the synthetic seismogram analysis are well controlled for this velocity or depth. The main velocity and thickness of this interval are 6.0 km s\(^{-1}\) and 0.9 km respectively. The sharpness of the growth and decay of reflections from this interval suggest that the 5.4–6.2 km s\(^{-1}\) velocity range may be a region of velocity gradient, rather than a constant velocity layer. The velocity gradient would be approximately 2 km s\(^{-1}\)/km.

(d) The 6.4–6.6 km s\(^{-1}\) layer also becomes shallower, again maintaining its velocity and thickness, 0.8 km. This layer’s velocity is more accurately defined than that of the 5.4–6.2 km s\(^{-1}\) layer. The layer appears to be a transition from the rapid increase in velocity of the 5.4–6.2 km s\(^{-1}\) layering to the relatively constant velocity layer below.

(e) The 6.6–7.2 km s\(^{-1}\) layer increases its thickness and mean velocity with age. The top of this layer has a well defined velocity of 6.8 km s\(^{-1}\) for all three lines. The top interface may be a region of a strong velocity gradient. This interpretation proved best for the 9 Myr line. At the Ridge axis it appears that the 6.8 km s\(^{-1}\) material is underlain by a small velocity contrast low-velocity zone, 6.6 km s\(^{-1}\), the synthetic seismogram analysis suggests a lid thickness of 0.5 km. There is no low velocity layer below the 9 Myr line but rather a slight positive velocity gradient. The layer thickens from above and below with age, the main thickening coming from below. The synthetic analysis is particularly sensitive to the velocity structure within this layer.

(f) The maximum velocity measured (Moho?) increased with age. The transition to this velocity is a gradient rather than a first order discontinuity. The maximum velocity measured at the Ridge axis was 7.1 km s\(^{-1}\). This value is at the lower limit of anomalous mantle velocities recorded. The 9 Myr line gives a typical Moho velocity of 8.2 km s\(^{-1}\). The synthetic seismograms suggest that this interface is a gradient varying from 0.60–0.66 km s\(^{-1}\)/km for the 0 and 9 Myr lines respectively.

10 Discussion

In a compilation of seismic refraction results of oceanic crust Raitt (1963) defined three major layers of the crust, layer 1 – sediments, layer 2 – ‘basement’ and layer 3 – the ‘oceanic layer’. These were underlain by mantle, layer 4. Three general properties were found:

(i) layer 2 exhibited a wide range of velocities, 5.0 ± 0.63 km s\(^{-1}\),
(ii) layer 3 had a well defined velocity, 6.69 ± 0.26 km s\(^{-1}\) and
(iii) the mantle velocity was variable, being low near the Mid-Atlantic Ridge.

Further work in terms of this layered structure suggested other relationships. Le Pichon et al. (1965) found that layer 2 thinned and layer 3 thickened away from the Mid-Atlantic Ridge, and Christensen & Salisbury (1975) found that layer 2 was thicker for more slowly spreading ridges. The crustal structure thus depends both on the rate of extrusion and the crustal age, and its appears that the most rapid changes occur in the first few million years after crustal formation (Orcutt et al. 1976; Kennett et al. 1977).

More recent analysis has modified the layered model of the crust to multiple layer models (Houtz & Ewing 1976) and to velocity gradient models (Helmberger & Morris 1970; Orcutt et al. 1976; Spudich, Salisbury & Orcutt 1978; Helmberger 1977; Whitmarsh 1978). The velocity–depth models we have found in this study are also relatively smooth. A suitable seismic model for the oceanic crust is thus a generally smooth velocity–depth profile with variations in velocity gradient rather than a model of thick constant velocity layers.

Although no thick layering is now visible there is a pronounced change in velocity gradient at around 6.5–6.8 km s\(^{-1}\). The material above the depth of this change of gradient
roughly corresponds to 'layer 2' of Raitt (1963), under which title it is discussed. Similarly the material with only a mild velocity gradient and velocities between 6.8 and 7.3 km s\(^{-1}\) is discussed under 'layer 3'.

10.1 LAYER 2

The seismic layering in layer 2 has been ascribed to cracked rocks with healing of the cracks with depth (Matthews et al. 1971; Miyashiro, Shido & Ewing 1970) and this explanation for the variation of velocities within layer 2 is supported by many authors (Schrieber & Fox 1976; Hyndman & Drury 1976; Whitemarsh 1978).

Three general subdivisions of layer 2 were found in the Reykjanes Ridge results, their velocities being 2.2–3.8 km s\(^{-1}\), 4.6–4.7 km s\(^{-1}\) and 5.4–6.5 km s\(^{-1}\). The first two subdivisions correspond to layer 2A and the last division to layers 2B and 2C of the classification of Houtz & Ewing (1976) in a general study of velocity variation in layer 2.

The rapid rise in the velocity of the top 400 m of the oceanic crust, 2.2–3.8 km s\(^{-1}\) in 9 Myr, for the Reykjanes Ridge is comparable to the difference in surface refractor velocity between leg 37 DSDP, 2.8 km s\(^{-1}\) at 3.5 Myr (Robinson et al. 1976) and legs 51–53 DSDP, 4.8 km s\(^{-1}\) at 106 Myr (Salisbury et al. 1979). Both layers are reported as fractured basalt but the level of infilling of voids with smectite and calcite is high for legs 51–53 (Salisbury et al. 1979). Thus the rise in velocity of this layer on the Reykjanes Ridge is taken as being due to the infilling of cracks in the basalt pillow layer.

The 4.6–4.7 km s\(^{-1}\) layer can either be explained as a low density basalt or a highly fractured basalt. The lower velocity limit for basalt samples measured by Christensen & Salisbury (1972) was 4.53 km s\(^{-1}\). Using the basalt velocities measured for legs 417A, 417D and 418 DSDP and the formula of Wyllie et al. (1958) to calculate rock velocity for a porous rock, Stephen (1978) found that a basalt with 11 per cent porosity would have a velocity of 4.8 km s\(^{-1}\).

The final subdivision, velocities 5.4–6.6 km s\(^{-1}\), is taken to correspond to the gradual closing of cracks and pores in the rock with depth. This subdivision shallows with age probably due to the infilling of voids by hydrothermal mineralization, a process also favoured by Houtz & Ewing (1976). Why this process should affect the lower levels of layer 2 more than the 4.6–4.7 km s\(^{-1}\) subdivision is unclear (Stephen 1978). The range of basal velocities measured for samples from leg 37 was 5.4–6.6 km s\(^{-1}\) with a mean of 5.94 km s\(^{-1}\) (Hyndman & Drury 1976). The highest velocities of the subdivision are therefore high for basalts, but it is unlikely that there is a sharp boundary between the basaltic layer 2 and higher velocity layer 3 rocks. This is suggested by the differences between holes 332A and 332B of leg 37 drilled only 100 m apart (Robinson et al. 1976). The higher velocities of this subdivision may thus be due to an average of the basaltic velocities and layer 3 rocks (Matthews 1978, private communication).

10.2 LAYER 3

In contrast to 'layer 2', this layer has a very uniform velocity in all the oceans. The main rock types which are recovered by dredging are gabbros, metagabbros and metabasalts (Christensen & Salisbury 1975). For a sample from DSDP leg 37, Hyndman & Drury (1976) report a velocity of 7.15 ± 0.15 km s\(^{-1}\) which could be reduced to the mean velocity for the layer 6.7 km s\(^{-1}\) by only 2 per cent of water-filled voids.

Layer 3 is taken by some petrologists (Cann 1968, 1974; Bryan & Moore 1977) to correspond to the site of a large magma chamber at the ridge crest from which the crust is formed.
Such a picture may well be appropriate for a fast spreading ridge, since seismic refraction experiments on the crest of the East Pacific Rise have indicated the presence of a distinct seismic low velocity zone (Orcutt et al. 1976; Rosendahl et al. 1976) which has been interpreted as a magma chamber (Rosendahl 1976). Seismic reflection work on the East Pacific Rise in the same area has also indicated a reflector which may correspond to the top of the low velocity zone (Herron et al. 1978).

For the slowly spreading Mid-Atlantic Ridge, seismic refraction experiments designed specifically to find a magma chamber (Fowler 1976) showed that any such chamber must be very narrow. Thermal calculations by Sleep (1975) suggest that the magma chamber on a ridge with spreading rate 1 cm yr⁻¹ would only be 1–2 km wide. Nisbet & Fowler (1978) have recently proposed the ‘infinite leek’ theory of crustal formation for slowly spreading ridges by which oceanic crust can be formed without the presence of a large magma chamber.

The data from the Reykjanes Ridge indicate the presence of a small contrast velocity inversion at the Ridge axis, the velocity passing from 6.8 to 6.6 km s⁻¹. This low velocity zone was needed to explain the amplitude–distance behaviour of the seismograms. Using the velocity–temperature coefficient of Birch (1958), \(-6 \times 10^{-5} ^\circ\text{C}^{-1}\), the velocity contrast is equivalent to a temperature difference of 500°C between the 6.8 and 6.6 km s⁻¹ layers. This would imply a vertical temperature gradient of approximately 500°C km⁻¹ at the ridge axis. These conditions could be met by Sleep’s (1975) thermal model of oceanic crust formation at 1 cm yr⁻¹ if the 100°C isotherm corresponds to the base of layer 2. Thus, unlike the low velocity zone of the East Pacific Rise where the velocity contrast is 6.7–4.8 km s⁻¹ (Orcutt et al. 1976) the velocity inversion found on the Reykjanes Ridge may be explained by the effects of temperature. Since, even allowing for possible oblique spreading, line X passed within a kilometre of the centre of spreading there is no indication of a large magma chamber beneath this portion of the Reykjanes Ridge. The seismic velocity of the material within the inversion would increase slightly away from the ridge as the zone cooled.

From the three main refraction profiles we find that layer 3 appears to thicken with age both from the top and base of the layer. The thickening from below may be due to off-axis intrusion from the mantle (Christensen & Salisbury 1975; Nisbet & Fowler 1978).

10.3 GENERAL COMPARISON WITH OTHER CONSTRUCTIVE OCEANIC RIDGES

As we have mentioned before the form of the Reykjanes Ridge in this area is atypical of slow spreading ridges. Nevertheless the velocity models obtained for the main refraction profiles are quite similar to those proposed by Fowler (1976) for the Mid-Atlantic Ridge at 37°N, but there are some points of difference:

(i) The upper part of layer 2 (2.2–5.2 km s⁻¹) is slightly thicker on the Reykjanes Ridge and has a lower mean velocity (4.6 km s⁻¹) than for the Mid-Atlantic Ridge (4.9 km s⁻¹). The lower velocity may be due to the difference in depth of burial as Palmason (1970) noted that the surface lava velocity on the Reykjanes peninsular was highly pressure dependent.

(ii) In contrast to previous work on the Mid-Atlantic Ridge a thin zone of material with velocities in the range 6.5–6.8 km s⁻¹ is present over the axis of the Reykjanes Ridge, overlying a small contrast velocity inversion (6.6 km s⁻¹). This picture is somewhat similar to results from the East Pacific Rise (Orcutt et al. 1976) but there the velocity contrast proposed for the low velocity zone 6.7–4.8 km s⁻¹ is much greater. If an attempt is made to force the same contrast in the Reykjanes Ridge model we get a 0.26 km thick velocity
inversion and the reverberations from the low velocity zone give an unsatisfactory behaviour to the synthetic seismograms. There is thus a greater difference between the Reykjanes Ridge and the East Pacific Rise, than for the Mid-Atlantic Ridge.

(iii) The highest velocity found at the axis of the Reykjanes Ridge was 7.1 km s\(^{-1}\), much lower than the velocity at corresponding depth at 37° N, 7.6 km s\(^{-1}\). The differences between the lower part of the crust were less marked but layer 3 was about 0.5 km thicker on the Reykjanes Ridge. The transition to mantle velocities was modelled as a zone of velocity gradient which may correspond to a thicker cumulate layer (Nisbet & Fowler 1978) on the Reykjanes Ridge compared to the Mid-Atlantic Ridge at 37° N.

10.4 MANTLE TRANSITION

If, indeed, the highest velocity seen on each refraction line corresponds to mantle arrivals, the mantle velocity increases from 7.1 – 8.2 km s\(^{-1}\) in 9 Myr. These results are in agreement with the work of Ewing & Ewing (1959). A 60 km reversed profile E-3 along the ridge crest just south of the OBS array gave a velocity of 7.2 km s\(^{-1}\). Farther north profile E-4 parallel to the ridge about 50 km (5 Myr) from the crest gave a higher velocity of 7.6 km s\(^{-1}\). The short (30 km) sonobuoy profiles of Talwani et al. (1971) to the north of this region gave a well determined highest velocity of 7.40 km s\(^{-1}\) about 1 Myr away from the axis. At 37° N on the Mid-Atlantic Ridge, Fowler (1976) found normal mantle velocities only 10 km (1 Myr) from the ridge axis. Here we have weak indications of a 7.75 km s\(^{-1}\) arrival of line Y from the southermost portion of the line (Fig. 4), closest to the development of a median valley (Fig. 1) which would correspond to a rather deep refraactor.

The mantle transition was generally best modelled by a velocity gradient zone rather than abrupt transition. Such behaviour has been noted previously in the Bering Sea (Helmberger 1968, 1977) and near Guadalupe Island (Spudich et al. 1978).

The velocity gradient zone may correspond to the cumulate gabbro–peridotite layer formed at the base of layer 3 in the model of Nisbet & Fowler (1978). Such a mix of gabbro and peridotite would certainly give velocities intermediate to layer 3 and mantle.

The rate of decay of energy at long ranges on the two closest profiles to the ridge crest suggests that we have strong attenuation or a negative velocity gradient beneath the mantle transition and this effect is even seen weakly on line Z (Section 7.2). This would correspond quite well to the thermal model of Sleep (1975) or Parker & Oldenburg (1973). Thus although we have no significant crustal magma chamber at the crest, underlying the crust we have a zone of elevated temperature with incipient or actual partial melting giving rise to reduction in seismic velocities and increased attenuation. The temperature at the mantle transition will diminish as the distance from the axis increases, and so we would expect more efficient propagation of the mantle arrival for older crust, as is indeed seen.

Acknowledgments

We would like to thank D. H. Matthews and C. M. R. Fowler for advice and encouragement and the Master and crew of RRS Shackleton for their help at sea. We are grateful to T. J. G. Francis (principal scientist for the leg) and R. Lilwall for making available their OBS data. This work would not have been possible without the efforts of T. Claydon, M. Mason and K. Louden at sea and on land.

This work was supported, in part, by NERC grant GR3/1651 for Marine Geophysics and Geodynamics.
References


Salisbury, M. H., Stephen, R., Christensen, N. I., Donnelly, T. W., Francheteau, J., Hamano, Y., Hobart, M. & Johnson, D., 1979. The physical state of the upper levels of Cretaceous Basement from the results of logging, laboratory studies and the oblique seismic experiment at DSDP sites 417 and 418, in press.


