Summary

Two ocean bottom seismographs have been operated on the Mid-Atlantic Ridge near 45° N. Earthquakes up to $M_L = 3.8$ were observed in a total of eight days recording and the seismicity was found to be largely confined to the Median Valley. The observations are related to other geological and geophysical work in the area and to the teleseismically reported activity there during the last 10 years. A number of models of the structure and seismicity of the ridge axis are discussed. The most plausible model which can account for the observations indicates a magma chamber at a depth of about 3 km beneath the Median Valley floor.

Introduction

Between 45° and 46° N the Mid-Atlantic Ridge is at present better known and understood than any other part of the ridge. Detailed bathymetric, magnetic and gravity surveys carried out between these latitudes from 27° to 30° W together with seismic refraction and reflection experiments, heat flow measurements and dredging have combined to give the most comprehensive picture yet of the Mid-Atlantic Ridge crest (Loncarevic, Mason & Matthews 1966; Barrett & Aumento 1970; Keen & Tramontini 1970; Keen & Manchester 1970; Aumento, Loncarevic & Ross 1971; Bhattacharyya & Ross 1972; Hyndman & Rankin 1972; Woodside 1972). This made the area the obvious choice for the first operational deployments of ocean bottom seismographs developed over the past few years under the joint auspices of the Institute of Geological Sciences and the United Kingdom Atomic Energy Authority. A preliminary report of this work has already been published (Francis & Porter 1972). This paper presents the results of a detailed study of the records obtained.

Two instruments were available for the work, carried out in May and June 1972 from R.R.S. Shackleton. Since this was their first operational use only one instrument was deployed at a time. The first seismograph (OBS 1) was dropped into a small sedimented valley some 30 km east of the Median Valley, the second (OBS 2) into the Median Valley itself. OBS 2 was deliberately dropped over the deepest part of the Median Valley between 45° and 46° N, finishing up less than one nautical mile from the deepest sounding observed. The positions of the seismographs in relation to the bathymetry of the area are shown in Fig. 1; the geographical co-ordinates, water depths and operating times of the two instruments are given in Table 1.

The seismographs themselves will be the subject of a separate paper and a brief description of their operation was given in the first paper (Francis & Porter 1972). Each instrument contains a three-component short period seismometer (natural frequency 2 Hz) and a low frequency hydrophone. The amplified outputs of these
Fig. 1. Bathymetric map of the Mid-Atlantic Ridge showing positions of seismographs. Bathymetry after Bhattacharyya & Ross (1972). Water depths greater than 2600 m, hatched, less than 1800 m, stippled. Circles mark earthquake epicentres reported from 1963 to September 1972. The three events marked with open circles occurred on 1972 May 13, only 16 days before OBS 1 started operating.

Table 1

Parameters of ocean bottom seismograph stations

<table>
<thead>
<tr>
<th>Instrument</th>
<th>OBS 1</th>
<th>OBS 2</th>
</tr>
</thead>
<tbody>
<tr>
<td>Latitude</td>
<td>45° 30.4' N</td>
<td>45° 40.3' N</td>
</tr>
<tr>
<td>Longitude</td>
<td>27° 27.4' W</td>
<td>27° 47.4' W</td>
</tr>
<tr>
<td>Water depth (m)</td>
<td>2772</td>
<td>3592</td>
</tr>
<tr>
<td>Recording ON</td>
<td>1900Z, 1972 May 29</td>
<td>2300Z, 1972 June 2</td>
</tr>
<tr>
<td>Recording OFF</td>
<td>1500Z, 1972 June 2</td>
<td>1900Z, 1972 June 6</td>
</tr>
</tbody>
</table>
Median Valley seismology

Fig. 2. Response curves for seismometer and hydrophone channels of the ocean bottom seismograph. (1) gives relative magnification when no filtering is used on playback, (2) when low-pass filtering at 10 Hz is employed, (3) gives the relative gain of the hydrophone system with no filtering on playback, (4) when band pass filtering from 3 to 10 Hz is employed. The relative level of the various curves is arbitrary.

Seismic context

Earthquake epicentres reported for the area by the USCGS and its successors for the period 1963 to September 1972, recomputed where available by the ISC, are also plotted in Fig. 1. The same events are presented in the form of a Benioff strain release plot in Fig. 3. It is apparent that the ocean bottom seismographs were recording only 16 to 24 days after three major events were detected teleseismically from this
section of ridge, yet for more than two years prior to 1972 May 13 the area was tele-
seismically quiet. This, of course, was not known at the time the seismographs were
deployed, but may have had a pronounced effect on the level of microearthquake
seismicity observed. A tendency for earthquakes to occur in swarms along mid-
oceanic ridge crests has been noticed (Sykes 1970; Francis & Porter 1971) and it is
possible that the activity observed by the ocean bottom seismographs was higher than
the usual level. Statistical predictions of the level of seismic activity must therefore
be treated with caution. Nevertheless it is useful to make such an estimate if only
to provide a yardstick against which the observations of the seismographs can be
compared.

The number of events observed in the area of Fig. 1 between 1963 and September
1972 are too few to determine a reliable magnitude–frequency relation, but previous
seismicity studies of the Mid-Atlantic Ridge have established that, in the case of body
wave magnitudes, the slope \( b \) of this relation has the value 1.72 for rift zone regions
(Francis 1968a, b). Furthermore, it has been shown that since 1963 the efficiency of
the network has been sufficient to detect all events from the Mid-Atlantic Ridge for
which \( m_b \geq 5.0 \) (Francis & Porter 1971). Two such events occurred between 44°
and 46°N from 1963 to September 1972. The magnitude–frequency relation for
this region and this period of time may therefore be written as

\[
\log N = 8.90 - 1.72 m_b
\]

where \( N \) is the number of events of magnitude \( m_b \) and greater. This expression can
be used to estimate the level of microearthquake activity to be expected in the vicinity
of the seismographs. Expressed in more suitable form for application to the ocean
bottom data, the expression indicates that for \( m_b \geq 1 \) the number of events expected
is 19 per kilometre of ridge axis per day. For \( m_b \geq 3 \) the number falls to 0.0068 per
kilometre per day. But the body wave magnitude scale is inappropriate for locally
recorded earthquakes and the local magnitude scale must be used. The relationship between these two scales, based on observations of major earthquakes, has been given as (Gutenberg & Richter 1956)

\[ m_b = 1.7 + 0.8 \times M_L - 0.01 \times M_L^2. \]

If the same relation holds true for microearthquakes, then \( m_b = 3 \) is roughly equivalent to \( M_L = 2 \), \( m_b = 1 \) to \( M_L = -1 \). Thus, on the basis of teleseismically observed earthquakes, one may conclude that if OBS 2 were ‘sampling’ between 20 and 50 km of ridge axis during its 4-day recording period, it might observe one earthquake of \( M_L \approx 2 \), but larger events than this would be unlikely. The seismic energy released by an earthquake of \( M_L = 2 \) is approximately \( 10^{14} \text{ erg} \) (Richter 1958; Duda 1965).

Description of observations

Although the noise levels at the two sites were similar, there was a marked difference in their seismicities. An average of 30 events per day were recorded by OBS 2 in the Median Valley, but of only six per day by OBS 1 30 km off the axis. Clear P and S phases could be distinguished for the larger events. In addition compressional waves reflected from the sea surface were observed, being labelled \( R_1, R_2 \ldots \) according to the number of sea surface reflections. In general the amplitude of the S-wave was an order of magnitude greater than that of the P, and on the hydrophone trace the amplitude of \( R_1 \) was approximately twice that of P. The S-waves were best detected by the horizontal components of the seismometer and \( R_1 \) by the hydrophone. The hydrophone also proved superior to the vertical component of the seismometer in detecting P. This was probably because the seismometer, located inside the same pressure vessel as the tape recorder, was detecting some vibration from the latter, whereas the hydrophone, located outside, was relatively immune to this source of noise.

A sequence of earthquakes of increasing size recorded by OBS 2 is shown in Fig. 4. Only the shear waves can be identified for the smallest event, but as the size of the events increases first \( R_1 \), then P and multiply reflected phases in the water layer emerge above the noise. An event recorded by OBS 1 is shown in Fig. 5. P, \( R_1 \) and S phases are again clearly identifiable, but the record differs from those recorded by OBS 2 in having a much larger \( (S-P) \) time and consequently \( R_1 \) precedes S.

A remarkable feature of the records obtained by OBS 2 was the occurrence of identical sets of records differing only in amplitude and hence the ratio of signal to noise. In general these occurred as pairs of events spaced a few minutes apart, but two such identical pairs occurring on consecutive days were also identical. Three different sets of identical records were observed. Their relevant characteristics are given in Table 2. Three records from the largest set are shown in Fig. 6. When measured independently the \( (S-P) \) or \( (R_1-P) \) times of each set agreed to within a few hundredths of a second, but when laid over each other it was apparent that these differences were the result of the varying noise background and that no measurable time differences actually existed, i.e. the \( (S-P) \), \( (R_1-P) \) times agreed to within 0.01s. The only possible explanation for a matching set of records is that all emanated from a common source. The agreement of \( (S-P) \) times within a set indicates (assuming an \( (S-P) \) velocity of 7 km s\(^{-1}\)) that the individual foci were within 70 m of each other. This is of the same order as the focal dimensions of the events being observed. The limited bandwidth of the recording implies that some differences between the source functions of events within a set could have existed, but their polarity was always the same and the only essential difference observed was in their amplitudes. The most plausible interpretation is that each identical set of records was the result of repeated movement on an active fault.
Fig. 4. Set of four earthquakes of increasing size recorded by OBS 2. Equivalent ground displacement is given on the seismometer traces for the peak magnification at about 6 Hz. The pressure calibration of the hydrophone channel is approximate. The orientation of the horizontal components (X, Y) is unknown. The times of these events to the nearest minute were (a) 1452 6 June, (b) 1258 3 June, (c) 1137 5 June, (d) 1502 6 June.
Fig. 4(b)
Fig. 4(c)
Fig. 4(d)
Fig. 5. Earthquake recorded by OBS 1 at 0233 30 May. The S-P time is 4.51 s.
Fig. 6. Identical set of records, differing only in amplitude, recorded by OBS 2. The times of these events to the nearest minute were (a) 1111 3 June, (b) 1508 4 June, (c) 1513 4 June.
Fig. 6(c)
Characteristics of identical sets of events

<table>
<thead>
<tr>
<th>Time of arrival</th>
<th>S-wave amplitude relative to smallest event</th>
<th>S-P time measured on largest event</th>
</tr>
</thead>
<tbody>
<tr>
<td>0910/3 June</td>
<td>2.1</td>
<td>1.09 s</td>
</tr>
<tr>
<td>0912/3 June</td>
<td>1.0</td>
<td></td>
</tr>
<tr>
<td>1111/3 June</td>
<td>1.7</td>
<td></td>
</tr>
<tr>
<td>1113/3 June</td>
<td>1.0</td>
<td></td>
</tr>
<tr>
<td>1508/4 June</td>
<td>2.5</td>
<td></td>
</tr>
<tr>
<td>1513/4 June</td>
<td>35.0</td>
<td></td>
</tr>
<tr>
<td>0524/4 June</td>
<td>1.3</td>
<td>1.31 s</td>
</tr>
<tr>
<td>0532/4 June</td>
<td>1.0</td>
<td>0.81 s</td>
</tr>
</tbody>
</table>

The records shown in Fig. 6 indicate a very simple source function. Many records were considerably more complex, making it difficult sometimes to identify the various water borne phases on the hydrophone channel. Some of these may be similar to the identical sets but with the individual events occurring so close to each other that their records overlap and make separation and matching either difficult or impossible. One of the more complex records obtained by OBS is shown in Fig. 7.

Many of the records shown in Figs 4–7 have a noticeable 3 Hz component in their coda, especially prominent on the X and Y channels. A damped oscillation of the same frequency was produced when the seismograph was tapped by a diver during underwater tests in a tank, and is generated by a resonance of the instrument structure. On the ocean floor it seems to be most excited by water borne energy propagating downwards from the sea surface. It may be possible to suppress this resonance by redesign of the package. But to some extent the situation is analogous to operating a seismometer on the second storey of a five-storey building and it is unlikely that we shall ever be able to occupy the basement. A few events recorded by both instruments consisted almost entirely of this frequency and were also devoid of phases reflected from the sea surface. These have been interpreted as fish bumps.

A cumulative plot of the distribution of amplitudes of earthquakes is shown in Fig. 8(a). Here the logarithm of the number of events greater than or equal to a given amplitude is plotted against the logarithm of that amplitude. The amplitude used is the maximum peak to peak horizontal deflection, \( A = (X^2 + Y^2)^{1/2} \), of each event, which invariably occurred in the S-wave train and refers to the pen deflection at a particular gain setting on playing back the magnetic tape onto paper. The straight line relationship for the logarithmic plot

\[
\log N = C - b \log A
\]

does not fit all the point shown on the plot particularly well, but gives a better fit when confined to amplitudes above about 2 cm. The value of \( b \) then obtained is 0.84, in close agreement with that for microearthquakes recorded in Iceland (Ward, Palmason & Drake 1969). Similar values of \( b \) have been obtained in other microearthquake studies and are typical of tectonic earthquakes. But if only one class of earthquake were being observed, as the amplitude decreased from 2 cm the points should eventually fall away below the line of slope \( b = 0.84 \) as events became undetectable in the noise. In fact they do the opposite. This may indicate the presence of another class of earthquake whose cumulative distribution relation has a steeper slope. An analogous situation has been found in seismic observations on the Moon, where the cumulative plots of moonquakes have a steeper slope than those of meteoroid impacts (Latham et al. 1972). If two classes of earthquake were detected by OBS 2, then a better fit to the points of Fig. 8(a) is given by the curve in Fig. 8(b), the sum of two straight lines with slopes \( b = 0.80 \) and \( b = 2.29 \), respectively, than by a single straight
Fig. 7. More complex record obtained by OBS 2 at 1433 6 June. The $S$-$P$ time is 1.54 s.
Fig. 8(a). Cumulative plot of earthquakes recorded by OBS 2. The maximum peak to peak amplitude, $A = (X^2 + Y^2)^{1/2}$, always occurred in the S-wave train and refers to the pen deflection at a particular gain setting when playing back the magnetic tape onto paper. At this magnification the measurement of amplitude in centimetres is roughly the same as that of horizontal ground motion in microns.

line. This curve is not claimed to be the best fitting combination of two straight lines, but it fits the data better over a wider range of amplitude than the single straight line of Fig. 8(a) with the added advantage that the data points now ‘roll-off’ normally below the line from about 0.7 cm. An interesting consequence of this interpretation is that whilst retaining the slope $b = 0.8$, there is now also the slope $b = 2.29$. And whereas low $b$-values are associated with tectonic earthquakes, high $b$-values are associated with volcanic earthquakes, probably generated by subsurface movement of magma. Both types have been recorded in the vicinity of active volcanoes (Minakami 1960). However the data are too few and restricted to too small a range of amplitude for the two-population interpretation to be more than tentative. No evidence could be found from the characteristics of individual records for dividing them into two classes. Furthermore it is difficult to understand at this stage what is the relationship between these locally determined $b$-values and the high $b$-values found for teleseismically reported earthquakes from the rift zones of the Mid-Atlantic Ridge.

The restricted range over which earthquakes were recorded by OBS 2 allows the amplitude scale of the cumulative plot to be calibrated in terms of $M_L$ (Richter 1958). An amplitude of 1.13 cm in Fig. 8(a) is approximately equivalent to $M_L = 2$ and 11.3 cm to $M_L = 3$. The largest recorded event (Fig. 6(c)) has a magnitude of 3.8 on the local scale. The event recorded by OBS 1 shown in Fig. 5 has a magnitude $M_L = 3.4$. The number of events with $M_L \geq 2$ is 35, considerably more than pre-
dicted from the teleseismic body-wave magnitude-frequency relation. This might indicate an unusually high level of activity after the major earthquakes reported from the area on 1972 May 13, following the pattern of swarming which has already been noted for the Mid-Atlantic Ridge. An alternative interpretation is that the local magnitudes of these events have been overestimated.

Duda (1965) has studied seismic wave propagation in the region of the Tonto Forest Seismic Array over the distance range $14 \text{ km} \leqslant \Delta \leqslant 640 \text{ km}$, paying particular attention to the $P_g$ and $S_g$ phases. These are the equivalent phases in continental structure to the phases identified as $P$ and $S$ here. At the closest range he found the ratio of $S$ to $P$ wave energy to be the order of 100 to 1, similar to that observed here. But between 14 and 100 km from the source this ratio decreased by an order of magnitude because of the greater attenuation of $S_g$ than $P_g$. It is quite likely that $S$-wave attenuation is much greater than that for the $P$ wave in the vicinity of OBS 1 and OBS 2. If so, local magnitudes based on $S$-waves observed very close to the source are overestimated. Put another way, the log $A_0$ term of the $M_L$ formula (Richter 1958), which corrects for the attenuation of seismic waves with distance, may not be the same for the Mid-Atlantic Ridge as found in Southern California. This would not be surprising.

Further doubt can be cast on the comparability of the $M_L$ measurements obtained for OBS 2 from another aspect of Duda's work. He found that the duration of the

![Graph](https://academic.oup.com/gji/article-abstract/34/3/279/630851)

FIG. 8(b). Two straight lines which account for the cumulative plot better than the single straight line in Fig. 8(a). The curve is the sum of the two straight lines and fits the cumulative plot down to 0·7 cm where the normal roll-off starts.
Sg phase was independent of distance in the range investigated but was related to magnitude according to the relation

$$\log t(S_g) = 0.254 + 0.391 M_L$$

where $t(S_g)$ is the duration of the $S_g$ phase in seconds. This relationship suggests that most of the earthquakes recorded by OBS 2 were in the region of $M_L = 0$. On the other hand, the fact that an identical set of earthquake records (Fig. 6, Table 2) spanned a magnitude range of 1.5 with no appreciable difference of waveform suggests that Duda's formula may not be applicable.

**Simple models**

The simplest model which can account for the various phases observed on the ocean bottom seismograph records consists of two homogeneous layers of water and rock, the thickness of the water layer being equal to the water depth at the OBS site (Fig. 9). The P and S waves follow straight paths from earthquake to seismograph, and $R_1$, after propagation in the rock as a compressional wave, is refracted at the sea bed according to Snell's law and totally reflected at the sea surface. A sea surface reflected phase which propagated as $SV$ in the sea bed may also be present but will always follow $R_1$, and hence is difficult to observe. A flat sea bed is not of course a very good approximation to the rugged terrain prevalent near the axis of the Mid-Atlantic Ridge. On the other hand the lack of sedimentary cover in the Median Valley implies that the velocity contrast $\alpha_1 : \alpha_2$ is great and hence the horizontal distance, $x$, between the seismograph and the point where $R_1$ emerges from the sea bed is small. For the same reason ($R_1 - P), (R_2 - R_1), \ldots$ will not differ greatly from the sounding time at the OBS site.

![Fig. 9. Simplest model for explaining phases observed on the ocean bottom seismograph records: two homogeneous layers of water and rock separated by a horizontal boundary. P and S follow the straight line from earthquake to seismograph, $R_1$ is refracted at the sea bed according to Snell's law and totally reflected at the sea surface.](https://academic.oup.com/gji/article-abstract/34/3/279/630851)

$(S - P)$ measurements could be made on about a quarter of the total number of events recorded by OBS 2 and on three of those recorded by OBS 1. Histograms of these measurements are shown in Figs 10 and 11. Since $R_1$ was almost invariably more energetic than $P$, it was possible on a number of records to distinguish $S$ and $R_1$ but not $P$. The cross-hatching shows $(S - P)$ times from these records assuming $(R_1 - P) = 4.68$ s, its average value, and setting $(S - P) = 4.68 - (R_1 - S)$. The error involved in this assumption is unlikely to exceed 0.2.
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![Fig. 10. Histogram of (S-P) observations made by OBS 2 in the Median Valley. The single-hatched part of the histogram represents those measurements made directly from the records, the cross-hatched portion those obtained from (R1-S) assuming the relationship (S-P) = 4.68-(R-S).](image)

![Fig. 11. Histogram of (S-P) observations made by OBS 1.](image)

Although only a small proportion of the total number of events recorded by each seismograph gave sufficiently clear records to yield (S-P) times, there is no reason for supposing that these measurements are not representative of the whole samples. Thus the ranges at which earthquakes were detected differ strikingly between OBS 1 and 2. Assuming a (S-P) velocity of 7 km s\(^{-1}\), earthquakes recorded in the Median Valley ranged from 4 to 12 km, whereas those recorded outside ranged from 24 to 55 km. The most plausible interpretation of this observation is that all the earthquakes originated in the Median Valley (Francis & Porter 1972). The much higher seismicity of the OBS 2 site supports this conclusion. The remarkably limited range of (S-P) times at OBS 2 must reflect in some way either the propagation or the earthquake distribution within the Median Valley. This problem is discussed in detail later in the paper.

In addition to (S-P) times, OBS 2 yielded a number of (R1-P) times. A histogram of these measurements is shown in Fig. 12. As already indicated (R1-P) should approximately equal the sounding time of 4.79 s. This is observed to be the case. But if the flat bottom model strictly applied no (R1-P) time should exceed
4.79 s. That a few values do is an indication of the ruggedness of the terrain. Less than a nautical mile from OBS 2 the deepest point of all the Median Valley shown in Fig. 1, 1939 uncorrected fathoms, was observed, equivalent to a sounding time of 4.85 s. With a flat bottom the maximum $(R_1 - P)$ time occurs when the $R_1$ phase emerges vertically from the sea bed. As $\theta_2$ increases $(R_1 - P)$ decreases and acquires its minimum value when $\theta_2 = 90^\circ$. In this situation $\theta_1$ is the critical angle. Thus the minimum $(R_1 - P)$ value observed is a measure of $\alpha_2$, for

$$(R_1 - P)_{\text{min}} = \frac{2d}{\alpha_1} \cos \theta_1$$

and

$$\alpha_2 = \frac{\alpha_1}{\sin \theta_1}.$$

The minimum value of $(R_1 - P)$ observed was 4.51 s which corresponds to $\alpha_2 = 4.48 \text{ km s}^{-1}$. This agrees remarkably well with the velocity, 4.58 km s$^{-1}$, obtained for the uppermost layer of the crestal mountains by seismic refraction shooting less than 50 km to the west (Keen & Tramontini 1970). It also agrees well with laboratory determined compressional wave velocities of basalts dredged from the Median Valley in the immediate vicinity of the OBS 2 site (dredge stations 173 and 197, Barrett & Anmento 1970). The velocities of these rocks, measured at their in situ hydrostatic pressures, ranged from 4.44 to 4.75 km s$^{-1}$.

With $\alpha_2$ known all the parameters necessary to specify $P$ and $R_1$ in the model of Fig. 9 are known and it is possible to compute the value of $(R_1 - P)$ as a function of $\theta_2$. This is plotted in Fig. 13 for three different values of $r$. It is clear that $(R_1 - P)$ is a good measure of $\theta_2$ over the whole range of $r$, i.e. the whole range of $(S - P)$ times, observed by OBS 2. Thus for each earthquake for which $(R_1 - P)$ and $(S - P)$ have been measured it is possible to deduce a focal depth. The validity of these estimates
of focal depth will depend on how well the simple model of Fig. 9 fits the real situation. To test this it is worthwhile examining the ratio of the amplitudes of $R_1$ and $P$ as measured on the hydrophone trace, since this also contains information on the parameter $\theta_2$.

Consider the ray paths shown in Fig. 9. The direct $P$ wave to the hydrophone is attenuated by (1) geometrical spreading in the rock, (2) the anelasticity of the rock, and (3) transmission across the rock/water boundary. The sea surface reflected wave $R_1$ is similarly attenuated but in addition suffers (4) geometrical spreading in the water, (5) total reflection at the sea surface and (6) reinforcement at the seismograph by its own sea bed reflection (since the hydrophone is much closer to the sea bed than the wavelengths involved). The $Q$ of seawater is so great that absorption of $R_1$ in the water layer is negligible. Furthermore the ranges $r$ and $y$ in the rock are short and do not differ greatly so that absorption in the rock even for low values of $Q$ is also negligible. If the assumption is now made that the transmission coefficients across the rock/water boundary are the same for both $P$ and $R_1$, the amplitude ratio can be written down as follows:

$$\frac{A_{R_1}}{A_P} = (1 + R) \cdot \left( \frac{y'}{2d} \right) \cdot \left( \frac{r}{y + \cos \theta_1} \right).$$

reinforcement
of $R_1$ by spreading
seabed
reflection
geometrical
gerating
spreading in $R_1$ in water rock

\[ d = 3.592 \text{ km} \]
\[ v_1 = 1.501 \text{ km/sec} \]
\[ v_2 = 4.5 \text{ km/sec} \]
where $R$ is the reflection coefficient of the seabed. It will be sufficient for this discussion to take $R$ as constant and equal to the normal reflection coefficient, and

$$y' = \frac{\cos^2 \theta_1}{\cos^2 \theta_2} \frac{\alpha_2}{\alpha_1} y$$

is the apparent range of the earthquake from the point where $R_1$ emerges into the

![Amplitude ratio AR1/AR as a function of $\theta_2$ computed for various values of $r$ according to the model shown in Fig. 9. $R$ is the normal reflection coefficient of the sea bed, taking $\rho_1 = 1.0$ and $\rho_2 = 2.8$ g cm$^{-3}$.](https://academic.oup.com/gji/article-abstract/34/3/279/630851)
water. In the limiting case of vertical and horizontal rays the expression is exact and reduces to

$$\frac{A_{R_1}}{A_P} = \frac{(1+R)}{\left(1 + \frac{2\alpha_1}{\alpha_2 h}\right)}$$ when $\theta_2 = 0$

$$\frac{A_{R_1}}{A_P} = \frac{(1+R)r}{(r-x)}$$ where $x = \frac{2\alpha_1}{\sqrt{(\alpha_2^2 - \alpha_1^2)}}$ when $\theta_2 = 90^\circ$.

In general the expression for $A_{R_1}/A_P$ is a good approximation when $r > x$. For OBS 2 $\alpha_2 = 4.5 \text{ km s}^{-1}$ and $x$ cannot exceed 2.6 km, so the approximation should be good for all the events recorded. $A_{R_1}/A_P$ has been computed as a function of $\theta_2$ for three values of $r$ and is plotted in Fig. 14. The amplitude ratio increases with $\theta_2$, the increase being much more marked for small $r$. For any event for which $A_{R_1}/A_P$ and $(S-P)$ have been measured it is possible to deduce a focal depth. As before, the validity of this approach depends on how well the simple model of Fig. 9 fits the real situation.

$A_{R_1}/A_P$ measurements were made on a dozen records with clear $P$ and $R_1$ phases. Many more records showed these phases, but the signal/noise ratio was too small to permit a reliable measurement of amplitude ratio even though one could be quite confident of the $(R_1-P)$ measurements. The $A_{R_1}/A_P$ ratios are plotted against $(R_1-P)$ times in Fig. 15. If the two alternative methods of obtaining $\theta_2$ were consistent there should be an inverse correlation between $A_{R_1}/A_P$ and $(R_1-P)$. In fact the points are very scattered and no noticeable correlation exists. The main conclusion to be drawn from this is that the estimates of $\theta_2$ and hence focal depth should not be taken very seriously. This does not necessarily mean that the model applied is too
simple, for the large errors involved, especially in the measurement of amplitude ratio, may be confusing the issue. That large errors are present became apparent when measurements were made on identical sets of records. Thus the three records shown in Fig. 6 gave \((R_1 - P)\) values from 4.65 to 4.68 s and \(A_{R1}/A_P\) ratios from 2.22 to 2.82 when measured independently, and yet when laid over each other were clearly identical waveforms differing only in amplitude and noise background.

Geologically plausible models for the Median Valley floor

Before discussing the implications of more complex models of the Median Valley floor it is worth reviewing the geological constraints which every model should satisfy. Extensive dredging has been carried out in the vicinity of the ridge axis at 45° N and has established to a good degree of certainty both the prevalent rock types and their distribution (Barrett & Aumento 1970; Aumento, Loncarevic & Ross 1971). It appears that the crust as originally formed in the Median Valley has a layered structure and consists of a sequence of pillow basalts, often vesicular, grading downwards into more massive basalts, meta basalts and ultimately gabbro. As the crust has moved away from the Median Valley it has been dislocated by block faulting and the occasional intrusion of serpentinite diapirs. The grades of metamorphic rock observed imply a steep temperature gradient beneath the Median Valley, in excess of 200 °C per kilometre. Thus plausible models of the ridge postulate the presence of a magma chamber at a few kilometres depth beneath the Median Valley floor (Cann 1970).

The existence of a temperature gradient of an order of magnitude greater than that found generally in the oceanic crust is likely to have a profound effect on the seismic velocity structure. In general velocity increases with pressure and decreases with temperature, but very few measurements of the temperature coefficients of seismic velocity have been made (Press 1966). However a good rule of thumb is that at constant pressure the velocity decreases by 1 per cent for every 100 °C rise in temperature. Thus, if the rock between the magma chamber and the Median Valley floor were massive and homogeneous, temperature effects would be likely to dominate pressure effects and the seismic velocity would decrease with depth. Such a velocity structure and the ray paths it would produce are shown in Fig. 16. An interesting consequence of this structure is that, because of the curvature of the ray paths, the maximum range at which an earthquake can be observed by an ocean bottom seismograph is a function of its focal depth. If the gradient is constant with value \(g\) and the maximum depth at which an earthquake can occur is \(h_{\text{max}}\), then it is easy to show from ray theory that the maximum horizontal range at which an earthquake can be detected is given by

\[
l_{\text{max}} = \sqrt{\left(\frac{2a/z h_{\text{max}}}{g} - h_{\text{max}}^2\right)}.
\]

![Fig. 16. Negative velocity gradient model for Median Valley floor. The deepest earthquake gives the maximum possible range to an ocean bottom seismograph.](https://academic.oup.com/gji/article-abstract/34/3/279/630851)
But the velocity structure shown in Fig. 16 is not in keeping with the geology to be expected beneath the Median Valley floor. Even in the presence of a strong temperature gradient a layered structure of vesicular basalts giving way to more massive basalts, various metamorphic grades of basalt and dolerite and ultimately to gabбро will display a marked increase of velocity with depth. From the gabбро downwards to the magma chamber, however, changes in lithology will be slight and in this region temperature will control the seismic velocity. Thus the geologically reasonable velocity structure will have a positive gradient for the first kilometre or so beneath the Median Valley floor, becoming negative at some intermediate point between sea floor and magma chamber. At the magma chamber itself there will be an abrupt drop in the compressional velocity as the rock loses its rigidity, provided of course that there is an abrupt boundary to the chamber. Since coarse-grained rocks will only crystallize where the rate of cooling is slow, i.e. the ambient temperature is high, one can make an informed guess at the extent of the positive velocity gradient: if the temperature of the magma chamber is 1200°C, one would not expect gabbro to crystallize much above the 900°C isotherm. So the positive velocity gradient will extend perhaps three-quarters of the way to the magma chamber before reversing. This geologically reasonable velocity structure and the ray paths it would produce are shown in Fig. 17. One can argue further that above about 900°C the rock will be approaching a ductile condition so that earthquakes are unlikely to occur at depths greater than \( h_{\text{max}} \), the depth to which the positive velocity gradient penetrates. Thus the maximum range at which an earthquake can be recorded by an ocean bottom seismograph in this velocity structure is limited by the maximum depth to which rays can penetrate and still return to the sea bed. The shallowest earthquake will be detectable at the greatest range and once again it is easy to show from ray theory that this is given by

\[
I_{\text{max}} = 2 \sqrt{\left( h_{\text{max}}^2 + \frac{2\alpha_2 h_{\text{max}}}{g} \right)}.
\]

As far as an individual seismograph is concerned it is more useful to know the maximum range along the ray, \( r_{\text{max}} \), since a measure of this is given by the maximum \((S-P)\) time. This is given by

\[
r_{\text{max}} = (\pi - 2\theta_2) \frac{\alpha_{\text{max}}}{g} \quad \text{where} \quad \sin \theta_2 = \frac{\alpha_2}{\alpha_{\text{max}}}
\]

and \( \alpha_{\text{max}} = \alpha_2 + gh_{\text{max}} \)

\( \alpha_{\text{max}} \) is the maximum velocity the ray experiences, i.e. the velocity at its greatest depth of penetration.

Before discussing the application of these ideas to the observations made by OBS 2 it is important to consider what effect absorption might have on compressional and shear waves travelling between the Median Valley floor and a magma chamber. It might be thought that absorption would be high in the presence of a high ambient temperature. If fact measurements on basalts and gabбро suggest that \( Q \) is fairly constant for temperatures < 900°C and only as the melting temperature is approached does it fall sharply (Bradley & Fort 1966). Thus it is unlikely that rays propagating back to the sea bed in the velocity structure shown in Fig. 17 experience low \( Q \) over a significant portion of their path. It is also unlikely that the ratio \( \alpha/\beta \) will differ much from \( \sqrt{3} \) in the zone of positive velocity gradient, and only as the magma chamber is closely approached will this ratio show a marked increase. So the ray paths followed by \( P \) and \( S \) waves will not differ greatly.
We now have a model which is not only reasonable on geological grounds but which can explain the limited range to which \((S - P)\) times are observed. Furthermore, from our knowledge of the geology it is possible to make a good estimate of the maximum compressional velocity \(\alpha_{\text{max}}\) between Median Valley floor and magma chamber, since this is likely to be the velocity of gabbro at a temperature of about 900°C and a pressure in the region of 1 kilobar. At 0°C, 1 kbar gabbro has a compressional velocity of about 6.9 km s\(^{-1}\) (Press 1966). Applying the rule of thumb mentioned above gives \(\alpha_{\text{max}} = 6.3\) km s\(^{-1}\). There is now sufficient information to make an estimate of \(g\) and \(h_{\text{max}}\). In Fig. 18 the relationship between \(r_{\text{max}}\) and \(\alpha_{\text{max}}\) is plotted for various values of velocity gradient \(g\), the velocity at the sea bed \(\alpha_2\) being 4.5 km s\(^{-1}\) in every case. With \(\alpha_2\), \(\alpha_{\text{max}}\) and \(g\) known the value of \(h\) is fixed, so lines of constant \(h_{\text{max}}\) are drawn on the same diagram.

The maximum value of \((S - P)\) observed by OBS 2 was 1.64 s. This must be multiplied by an effective \((S - P)\) velocity to convert it to \(r_{\text{max}}\). 7 km s\(^{-1}\) is probably an underestimate when \(\alpha\) ranges from 4.5 to 6.3 km s\(^{-1}\), and 7.5 km s\(^{-1}\) is more likely. Thus \(r_{\text{max}} \approx 12\) km. Entering the diagram of Fig. 18 for \(\alpha_{\text{max}}\) and \(r_{\text{max}}\) we find that the velocity gradient \(g\) is about 0.75 s\(^{-1}\) and its depth of penetration \(h_{\text{max}}\) just over 2 km. It is also interesting to note that these values do not depend greatly on an accurate estimate of \(\alpha_{\text{max}}\).

If this is the correct interpretation for the limited range of \((S - P)\) values then not only has a measurement been made of the depth to the magma chamber but also, since the temperature of the magma chamber and the thermal conductivity of its rock cap can be estimated quite precisely, of the heat flow through the Median Valley floor. With \(h_{\text{max}}\) just over 2 km, the depth to the magma chamber is about 3 km. Assuming a magma chamber temperature of 1200°C and a thermal conductivity of \(5 \times 10^{-3}\) cal cm\(^{-1}\) s\(^{-1}\) °C\(^{-1}\) for its rock cap, the temperature gradient between magma chamber and sea floor is approximately 400°C per kilometre and the heat flow is 20 μcal cm\(^{-2}\) s\(^{-1}\).

The full ray diagram for the positive velocity gradient model is shown in Fig. 19. Also shown in this model is the method used for constructing ray paths in the region of the velocity gradient (Officer 1958, p. 60). In the case where \(g = 0\) the ray paths become straight lines and Fig. 19 becomes identical to Fig. 9. With the values for \(\alpha_2\) and \(g\) deduced above \((R_1 - P)\) has been computed as a function of \(r\) for a range of values of \(h\) and is shown in Fig. 20. In Fig. 21 the observations of \((R_1 - P)\) and \((S - P)\) made by OBS 2 are plotted. Multiplying the abscissae by 7 km s\(^{-1}\) transforms...
Median Valley seismology

Fig. 18. Diagram for deducing $g$ and $h_{\text{max}}$ for the model shown in Fig. 17, given $r$ and $a_{\text{max}}$. $a_2 = 4.5 \text{ km s}^{-1}$. Maximum range in positive velocity gradient.

Alternative interpretations of the limited range of OBS 2 observations

Two other interpretations might be proposed for the restricted range over which OBS 2 was able to observe earthquakes in comparison with OBS 1: (1) the value of $Q$ beneath the Median Valley might be much lower than that outside it; (2) OBS 2 was fortuitously dropped some $4 \text{ km}$ away from an active fault of approximately $10 \text{ km}$ extent, possibly a small transform fault.
FIG. 19. Ray path diagram for model with positive velocity gradient beneath sea bed. Rays below the sea bed are arcs of circles whose centres lie on the plane $\alpha_2/g$ above the sea bed.

FIG. 20. $(R_1 - P)$ as a function of $r$ computed for various values of $h$ according to the model shown in Fig. 19. The parameters specifying the model are shown on the figure.
Low $Q$ Median Valley

The amplitude of a wave propagating away from an earthquake through a medium with specific dissipation constant $Q$ can, at ranges larger than the focal dimensions, be expected to decrease according to the relation

$$A = \frac{A_0}{r} \exp \left( -\frac{\pi fr}{QV} \right)$$

where $r$ is the range, $f$ the predominant frequency and $V$ the velocity of propagation of the wave. This expression can be used to estimate how low $Q$ must be to have the desired effect.

Consider the largest earthquake recorded by OBS 2, shown in Fig. 6. This event had an $(S-P)$ time of 1.31 s, equivalent to a range of about 9 km. Assuming $Q = 100$ and using the expression above it is easy to show that doubling the range of this earthquake would only attenuate the record by about a factor of 3. Thus it should still produce a record with a good signal/noise ratio on which $P$, $S$ and $R_1$ phases could easily be read at $r = 18$ km. A similar case can be made for many of the larger events recorded by OBS 2. So if the seismic activity was fairly continuous along the Median Valley but its observation by OBS 2 were limited by a low $Q$, the value of $Q$ must be much less than 100. Such a low value might be expected to reveal itself in the character
of the records observed. But no loss of high frequency content with increasing \( r \) was apparent. Moreover microearthquake studies in Iceland, the most comparable study available from work on land, indicate that in the vicinity of highly fractured geothermal areas \( Q = 150 \) (Ward & Bjornsson 1971). It has already been pointed out that at temperatures \( < 900^\circ C \) the rock floor of the Median Valley is unlikely to possess low \( Q \) as a result of temperature alone. One can conclude that a general level of \( Q \) much less than 100 is not a plausible explanation for the range of observations made by OBS 2.

However, absorption could be high because of the presence of highly faulted zones of rock across which propagation might be difficult. Propagation over short paths which did not encounter such obstacles could be good, but over longer paths might be impossible. A drawback to this type of explanation is that such fault zones are just as, if not more, likely to exist in the block faulted crestal mountains as in the Median Valley. The long range observations of OBS 1 would then be anomalous.

A variation of the low \( Q \) model would be one in which the rock cap of the magma chamber is in places much thinner than elsewhere in the Median Valley. If cupolae or dykes project upwards from the magma chamber there might be zones in which temperatures \( > 900^\circ C \) come very close to the surface. The presence of low \( Q \) in these zones could then make propagation through them as difficult as across the fault zones cited above.

A criticism of any explanation which zones the Median Valley is that it violates the principle of Occam’s razor. Between 45° and 46° N the Mid-Atlantic Ridge is essentially a two-dimensional structure whose characteristics vary only with distance from the axis. Any explanation which can explain the seismic observations in these terms is preferable to one which invokes variation along the ridge axis as well.
Vicinity of an active fault

Transform faults offsetting the ridge axis are the principle feature which disturb the two-dimensional symmetry of the ridge. If such a fault were active some 4 km away from the OBS 2 site and it was approximately 10 km in extent, this could explain the range of observations observed by OBS 2. The topographic expression of a fault of this scale might be lost in the rugged terrain of the ridge axis.

There is some rather sketchy evidence for a small transform fault in this region. Woodside (1972) has interpreted a gravity low over the western crestal mountains to indicate a short ‘leaky’ transform fault at about 45° 45’ N. The magnetic anomaly map (Bhattacharyya & Ross 1972) could also be interpreted to indicate a small transform fault at about 45° 42’ N, offsetting the ridge axis north of this latitude to the east, but large changes in the topography of the eastern crestal mountains occur at about the same place. Unfortunately these features are too close to the northern limits of the survey area for any definite conclusions to be drawn. Within the limits of the bathymetric map (Bhattacharyya & Ross 1972) there is no evidence for an offset of the Median Valley and if the axis of the valley determined between 45° and 46° N is projected northwards it coincides with that determined by a survey near 47° N (Hill 1960). However, a piece of bathymetric evidence which might support the presence of a small transform fault near OBS 2 is that it was sited close to the deepest part of the Median Valley between 45° and 46° N. Sleep & Biehler (1970) have pointed out that the sea floor is often deeper at the intersection of rift and fracture zones than along the rifts and fractures themselves.

But the main objection to the hypothesis of a single active fault close to OBS 2 comes from the seismological evidence. The observation by OBS 1 of events between 24 and 55 km is not compatible with the activity being confined to a limited zone close to OBS 2 and indicate a wider distribution of activity along the Median Valley. Similarly the teleseismically reported events show no concentration of epicentres near 45° 40’ N, but suggest a fairly uniform distribution between 44° and 46° N (Fig. 1). If there is any concentration of teleseismic epicentres, it is close to 45° N. So even if OBS 2 were sited close to an active fault, there must be some other factor which prevented it from observing activity at greater range elsewhere along the Median Valley.

Conclusions

Two ocean bottom seismograph drops have been made onto the Mid-Atlantic Ridge near 45° N, one into the Median Valley and the other 30 km off the axis. Thirty events per day were recorded in the Median Valley at ranges between 4 and 12 km. Six events per day were recorded off the axis at much greater ranges—from 24 to 55 km. It is likely that all events originated in the Median Valley itself.

Many of the earthquakes observed were large by microearthquake standards, ranging up to $M_L = 3.8$ in magnitude. But the relevance of the local magnitude scale as developed in Southern California to this work is questionable. The cumulative plot of amplitudes indicates the presence of two earthquake populations, one with a low $b$-value 0.8, typical of tectonic earthquakes, the other with a high $b$-value 2.3, typical of certain types of earthquake found in the vicinity of active volcanoes. Only the low $b$-value is well established.

The presence of an overlying layer of water of known velocity and thickness makes it possible for the velocity of the uppermost layer of the sea floor to be determined by a single instrument. The velocity so obtained, 4.48 km s$^{-1}$, is in good agreement with neighbouring seismic refraction results and with laboratory measurements on dredged rocks.
The limited range of observations made by the seismograph in the Median Valley has been interpreted as indicating a positive velocity gradient of limited extent which is reversed in the vicinity of a magma chamber. Hence the depth of the magma chamber beneath the sea floor is found to be about 3 km.

Methods have been developed for the determination of the focal depth of individual events from the \((S - P)\), \((R_1 - P)\) times and the amplitude ratio \(A_{R1}/A_P\) recorded at a single seismograph. But the errors involved in the determination of \((R_1 - P)\) and \(A_{R1}/A_P\) are too great for them to be useful. It is unlikely, however, that any of the events observed were deeper than about 2 km beneath the sea floor.

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