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Upper Mantle Structure Along the Axis of the Mid-Atlantic Ridge near Iceland

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Summary

Apparent velocity measurements of P arrivals from Mid-Atlantic Ridge earthquakes across pairs of stations in Iceland and Greenland have been made using bulletin data. They define both the vertical and horizontal extent of the low velocity anomalous mantle zone between 50° and 70° N.

Introduction

Several geophysical techniques have been employed in the study of upper mantle structure beneath the mid-oceanic ridges, though no single technique can provide the complete picture. Seismic refraction shooting has defined a low velocity anomalous mantle a few hundred kilometres wide along the axes of the ridges (Raitt 1956; Ewing & Ewing 1959; Le Pichon *et al.* 1965), but is unable to throw any light on the vertical extent of this anomalous mantle beneath the ridges. The dispersion of long period surface waves from earthquakes has been used to study the deeper structure (Kovach & Press 1961), but the great path-lengths involved have inevitably crossed ridge flank and ocean basin as well as the ridge axis (since seldom does a mid-oceanic ridge extend more than a thousand kilometres without being offset by a fracture zone) and the structure obtained is probably intermediate between that of ocean basin and that of ridge axis. Dispersion studies of propagation along short straight sections of ridge (Tryggvason 1962) have concerned periods too short to have been much influenced by the structure of the mantle, except near its very surface. Gravity measurements (Talwani *et al.* 1965) show that the mid-oceanic ridges are in approximate isostatic equilibrium, and have been interpreted with various assumptions of density to indicate anomalous mantle extending, in the case of the Mid-Atlantic Ridge, to a depth of 30 to 40 km below sea level. Magnetometry, bathymetry and sediment profiling provide detailed information about the crust and the history of the ridge system and tell us much about the processes involved in the mantle beneath, but say little about its actual structure.

The application of earthquake body waves to the elucidation of ridge structure has been mainly the work of Tryggvason in the region of Iceland. In one study (1961) he used close range observations ($\Delta = 3°$ to 25°) of four earthquakes in the Greenland and Norwegian Seas, and concluded that the 7·4 km s^{-1} anomalous mantle of the Mid-Atlantic Ridge in this region was 1000 km wide and 140 km deep. In a later paper (1964), Tryggvason compared the P wave residuals of teleseisms at Reykjavik, Iceland with those at Kiruna, Sweden, and explained the delays at Reykjavik with a 240 km thick layer of 7·4 km s^{-1} velocity.

This paper presents the results of a study of body waves originating along the Mid-Atlantic Ridge north and south of Iceland and recorded in Iceland and Greenland. The data used were obtained from the bulletins of the B.C.I.S., I.S.C. and I.S.S., from the earthquake data reports of the U.S.C.G.S. from Tryggvason (1962), Sykes (1965) and from Iceland and Greenland preliminary bulletins. To avoid any preconceptions of ridge structure the approach has always been to work in time differences between observing stations. Ideally only *iP* arrivals would have been used, but such a restriction limited too much the data available. In effect only four stations have been used: Reykjavik (REY), Akureyri (AKU) and Sida (SID), all on Iceland, and Scoresby sund (SCO) in Greenland. From 1964 onwards the latter is replaced by Kap Tobin (KTG) but since the two sites are only a few kilometres apart this has no effect on regional interpretations. Fig. 1 is a map of the region on which the stations and epicentres of earthquakes used are marked. Table 1 lists the parameters of these earthquakes and the phases at each station which have been employed.

Table 1

Earthquakes used in this study

No	Date	Origin time (G.M.T.)			Epicentre		Observing stations							
							REY		AKU		SID		SCO/ KTG	
		h	m	s	°N	°W	*P*	*S*	*P*	*S*	*P*	*S*	*P*	*S*
1	14 Dec. 1933	07	42	14	56·1	33·9	✓						✓	
2	19 Oct. 1954	17	48	15	58·0	32·6	✓		✓				✓	
3	11 Dec. 1954	12	57	08	52·8	31·7	✓		✓					
4	6 Feb. 1955	02	27	52	70·7	14·4	✓	✓	✓	✓				
5	1 Apr. 1955	18	41	27	64·0	21·3	✓		✓				✓	
6	30 Oct. 1956	00	11	04	66·5	17·7	✓	✓	✓	✓				
7	5 June 1957	07	16	17	52·5	35·0	✓						✓	
8	18 June 1958	01	15	01	68·8	16·5	✓	✓	✓	✓				
9	18 June 1958	02	23	26	68·9	16·3	✓	✓	✓	✓				
10	18 June 1958	04	34	00	68·8	16·6	✓	✓	✓	✓				
11	25 June 1959	06	46	52	61·8	27·3	✓	✓	✓	✓	✓		✓	
12	28 June 1959	04	23	29	64·0	19·3	✓		✓				✓	
13	8 Dec. 1959	08	08	20	66·9	18·8	✓	✓	✓	✓	✓			
14	30 Mar. 1960	12	58	55	69·1	17·0	✓	✓	✓	✓	✓			
15	29 June 1960	10	23	00	47·2	27·3	✓						✓	
16	6 Oct. 1960	19	55	40	58·4	31·9	✓						✓	
17	14 Oct. 1960	22	55	44	55·5	35·0	✓						✓	
18	30 Apr. 1961	07	33	51	52·5	31·9	✓						✓	
19	14 May 1961	15	08	01	67·8	18·8		✓		✓				
20	14 May 1961	15	38	07	67·8	18·4	✓	✓	✓	✓	✓			
21	7 Mar. 1962	01	42	49	61·9	26·6	✓	✓	✓	✓	✓			
22	7 Mar. 1962	02	07	11	62·1	26·5	✓		✓		✓			
23	15 Jan. 1963	01	32	20	68·9	17·1	✓		✓		✓			
24	15 Jan. 1963	05	23	10	69·0	16·6	✓		✓		✓			
25	28 Mar. 1963	00	15	48	66·3	19·6	✓		✓		✓			
26	27 Apr. 1963	03	42	34	66·7	19·2	✓	✓	✓	✓	✓			
27	23 Aug. 1964	02	56	13	59·4	30·3	✓		✓					
28	23 Aug. 1964	04	47	47	59·5	30·2	✓		✓					
29	17 Sep. 1964	15	02	01	44·6	31·3	✓		✓		✓			
30	5 May 1966	15	16	32	61·4	27·5	✓	✓	✓	✓	✓			
31	5 May 1966	15	25	12	61·5	27·4	✓	✓	✓	✓	✓			
32	5 May 1966	15	52	41	61·5	27·5	✓	✓	✓	✓	✓			
33	16 May 1966	20	02	25	61·9	26·8	✓		✓		✓			
34	22 Aug. 1966	21	49	17	71·9	11·4	✓		✓					
35	1 Sep. 1966	21	27	39	58·3	32·6	✓		✓					
36	6 Nov. 1966	08	29	14	59·8	30·0	✓		✓					
37	17 Dec. 1966	05	59	10	70·7	14·0	✓		✓		✓			
38	13 Feb. 1967	23	14	20	52·7	34·1	✓	✓	✓	✓	✓		✓	

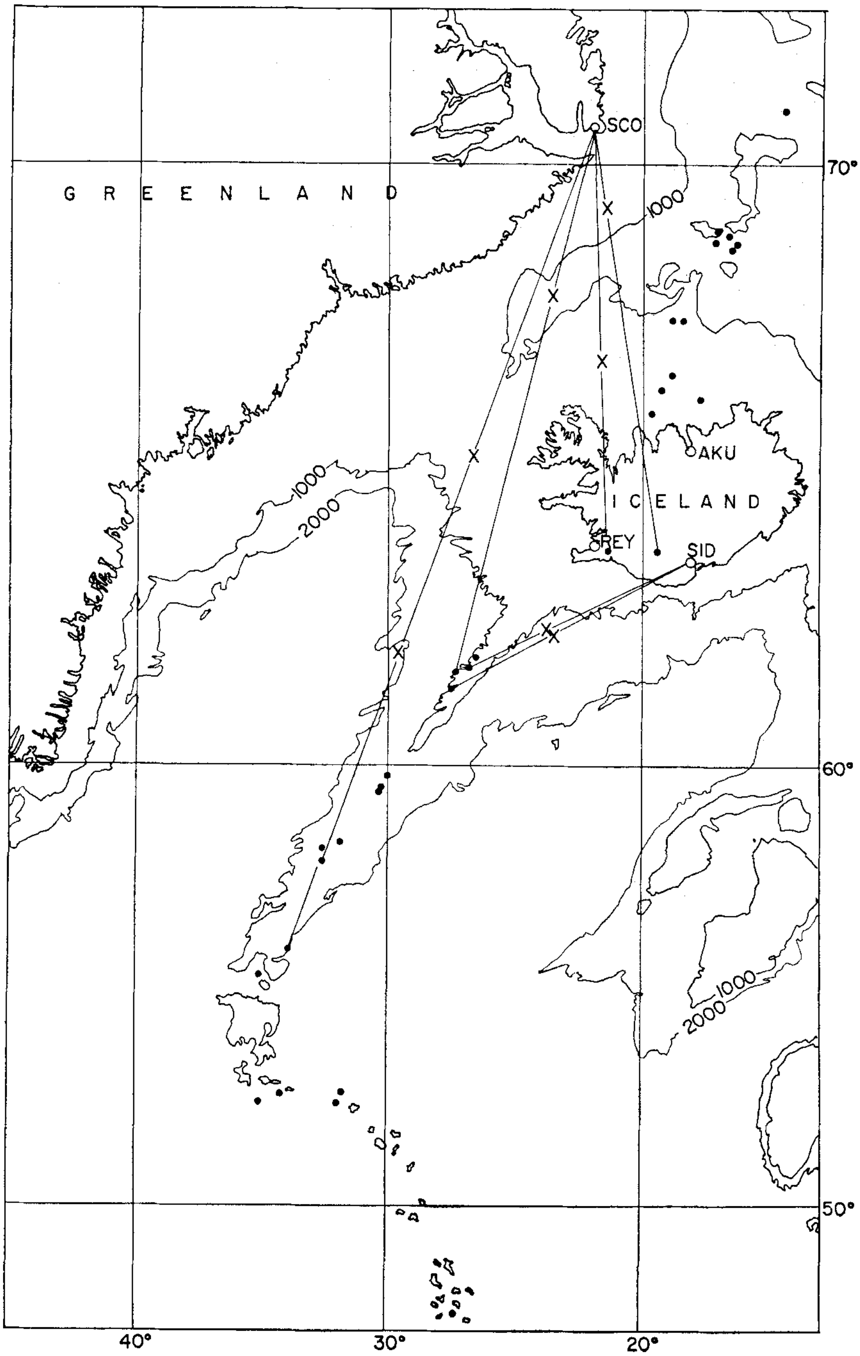

FIG. 1. Bathymetric sketch map of the North Atlantic Ocean near Iceland showing seismographic stations (open circles) and epicentres (solid circles) of earthquakes used in this study. A few epicentres have been omitted for clarity. 1000-m and 2000-m contours and coastlines from the International Hydrographic Bureau General Bathymetric Map of the Oceans sheet B1 (1966). Crosses mark the boundary of the anomalous low-velocity mantle (see text).

Apparent velocities across the REY, AKU array

As it passes through Iceland the Mid-Atlantic Ridge changes its direction from approximately NE–SW along the Reykjanes Ridge to NNE–SSW north of Iceland. This reorientation is clear from plots of earthquake epicentres (Sykes 1965; Stefansson 1967), although in Iceland itself the earthquake distribution is complicated by small east–west fracture zones offsetting the pattern near the north and south coasts of the island. South of Iceland the ridge is rectilinear until offset by a large fracture zone near 53° N (Johnson 1967). North of Iceland it remains approximately straight until offset by the Jan Mayen Fracture Zone (Johnson & Heezen 1967) near 72° N. Rays travelling to the pair of stations REY and AKU from epicentres along these sections of the ridge are confined wholly to the ridge axis.

By taking differences in the P wave arrival time and in the epicentral distance for the two stations, $dT/d\Delta$ has been found as a function of Δ for the axial structure. dT is simply the difference in arrival times given in the bulletins. $d\Delta$ was generally taken as $D\cos\theta$ where D is the distance between the two stations and θ the angle the ray direction makes with the line joining the two stations at its mid-point. D was computed using Richter's method (Richter 1958, p. 701) and θ by spherical trigonometry. In the few cases where Δ was given to 0·01° (in the I.S.C. bulletins) it was sufficient merely to take the difference. Since all the events used lay outside of and approximately along the line of the REY, AKU array, $d\Delta$ is

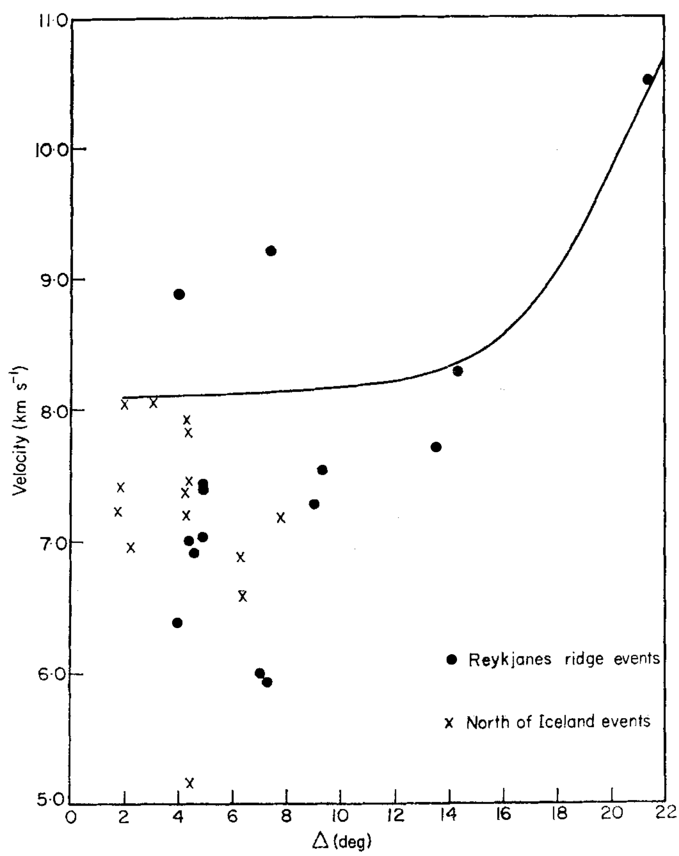

FIG. 2. Apparent velocities of P waves from events north and south of Iceland observed across the AKU, REY array. The solid line represents average mantle and is the inverse of Herrin's (1968) $dT/d\Delta$.

virtually independent of error in the positions of the epicentres. The Δ corresponding to the $dT/d\Delta$ measurement was taken as the average for the two stations. The results of using this method with earthquakes north and south of Iceland are given in Table 2 and plotted in Fig. 2. $dT/d\Delta$ has been expressed in reciprocal form as velocity for easier comparison with the results of other types of seismic measurement. In the measurement of distance 1° has been taken as 111·5 km throughout.

It is clear that the velocities are considerably lower than the average for the uppermost part of the Earth's mantle, represented in Fig. 2 by the inverse of Herrin's $dT/d\Delta$ (Herrin *et al.* 1968), but the measurements show considerable scatter and some discussion of errors is desirable. It might be thought that using a two-station array with such a wide spacing ($D = 248{\cdot}9$ km), i.e. approximating dT by δT and $d\Delta$ by $\delta\Delta$, introduces an appreciable error. This is not so if $d^2T/d\Delta^2$ is small, which is reasonable to assume. Complications, such as cusps, of the travel time curve are likely to be missed, however. Errors in Δ will be much the same as those in the determination of the epicentres, say not more than 0·3°. These are clearly too small to contribute much to the scatter in Fig. 2. As has already been mentioned above, $d\Delta$ is little affected by epicentral error and has been computed to be accurate to the nearest kilometre. Thus the errors in the velocities are essentially those contributed by the travel times. Measurements of dT ranged from 22 to 42 s, averaging about 30 s. If dT is the difference of two *iP* arrivals, each with a estimated error of ± 1 s, its error will be $\pm\sqrt{2}$ s or about 5 per cent. This is the most favourable situation and accounts for half the points on Fig. 2. Where dT is the difference of two *eP* arrivals it may be in error by as much as 6 s or 20 per cent.

The method might be criticized further for two possible sources of systematic error in the velocities. Firstly, it may generate lower velocities than in fact exist Since the distance across the array is about 2°, comparable with mean Δ for the nearest events, the effect of attenuation might be to produce a systematically later measurement of arrival time at the farther station and hence a systematically low velocity. This might be expected to show with $iP - iP$ measurements of dT producing higher velocities than other measurements. No such tendency was found, however, and the scatter in the velocity measurements is therefore assumed to be random. The other source of systematic error in velocity could arise from differences in the station residuals associated with REY and AKU. Station residuals range from about -2 to $+2$ s for the most extreme differences of geological structure, the vast majority lying between -1 and $+2$ s (Cleary & Hales 1966). Even though we are concerned here with fairly local events, it is probably safe to assume that within Iceland itself the residuals for the three stations involved in this study are the same to within a few tenths of a second. This assumption appears to be supported by the results of Cleary & Hales, who found the following for two Iceland stations:

REY	$0{\cdot}3 \pm 0{\cdot}2$ s
SID	$0{\cdot}2 \pm 0{\cdot}6$ s

An appreciable difference in the residuals of AKU and REY would show itself as a difference in velocities measured with events from north of Iceland compared with those from south of Iceland events. No significant difference was detected. For $\Delta < 10°$, the geometric mean of velocities for north of Iceland earthquakes is $7{\cdot}15\ \mathrm{km\,s^{-1}}$, that for south of Iceland events is $7{\cdot}14\ \mathrm{km\,s^{-1}}$.

Thus the measurements of $dT/d\Delta$ as a function of Δ, though subject to large random errors, appear to be free from any systematic trends. It is reasonable, therefore, to average the results and interpret them in terms of a velocity structure for the uppermost mantle along the axis of the ridge. Averaging was performed by taking geometric means of velocity measurements in 1° intervals. The values obtained are plotted in Fig. 3. Again, the velocity curve corresponding to Herrin's $dT/d\Delta$

Table 2

Apparent velocities across the REY, AKU array

Reykjanes ridge events					Events north of Iceland				
Event	Mean (Δ)	Apparent velocity, P (km s^{-1})	Apparent velocity, S (km s^{-1})	V_p/V_s	Event	Mean (Δ)	Apparent velocity, P (km s^{-1})	Apparent velocity, S (km s^{-1})	V_p/V_s
2	9·2°	7·55			4	6·25°	6·88	3·95	1·74
3	13·45	7·72			6	1·85	7·41	4·08	1·81
11	4·5	6·92	4·79	1·44	8	4·25	7·91	3·49	2·27
21	3·95	6·39	3·00	2·13	9	4·3	7·45	3·60	2·07
22	3·95	8·90			10	4·25	7·37	3·55	2·07
27	7·3	9·23			13	2·25	6·95	3·72	1·86
28	7·2	5·93			14	4·4	5·15	3·17	1·62
29	21·3	10·50			19	3·0		3·75	
30	4·8	7·03	3·07	2·29	20	3·0	8·05	4·19	1·92
31	4·8	7·43	4·61	1·61	23	4·2	7·20		
32	4·8	7·41	3·61	2·05	24	4·35	7·83		
33	4·3	7·01			25	1·7	7·23		
35	8·9	7·29			26	1·9	8·05	4·87	1·65
36	6·9	6·00			34	7·7	7·18		
38	14·2	8·29	3·24	2·56	37	6·3	6·58		

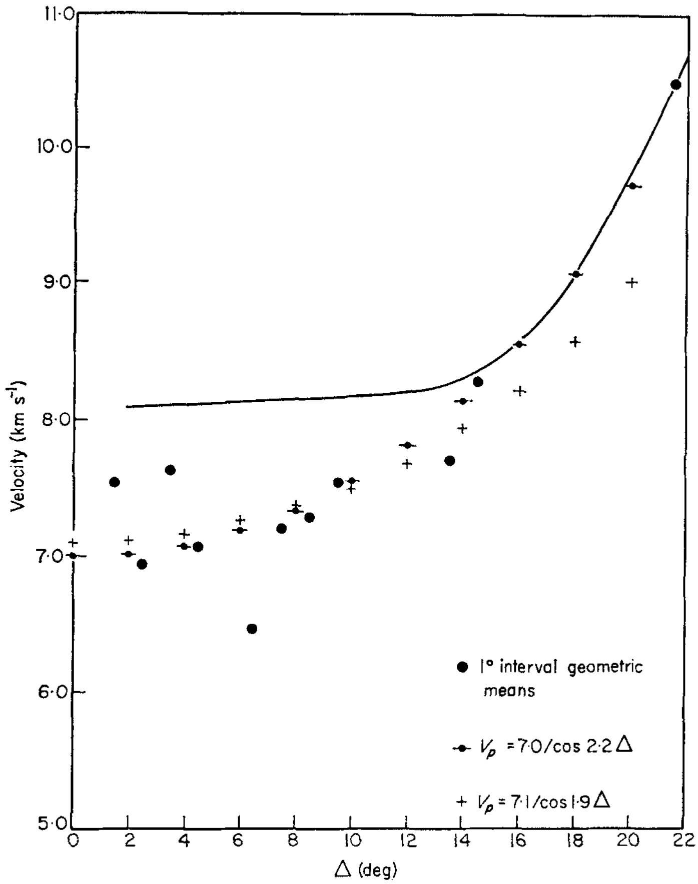

FIG. 3. 1° interval geometric means of the apparent velocity data plotted in Fig. 2.

curve is shown for comparison. The ridge axis velocity falls well below that for the average earth until about $\Delta = 15°$. The single point on the graph beyond this range is from an earthquake south of the 53° N fracture zone, whose ray paths to REY and AKU do not closely follow the ridge axis. Its close agreement with Herrin's curve, however, is taken as further evidence that beyond about $\Delta = 15°$, $dT/d\Delta$ for the ridge axis does not appreciably differ from that for the average mantle.

Inversion of apparent velocity data

Before inverting the data to obtain a velocity structure for the mantle beneath the ridge it is necessary to make some assumptions about the overlying structure. Beneath the submerged axis of the ridge the top of the anomalous 7·3 km s^{-1} mantle lies at about 6 km below sea level (Talwani *et al.* 1965). Beneath Iceland the corresponding depth is about 15 km below sea level (Båth 1960; Tryggvason 1962). Given the quality of the data, this overlying crust is thin enough that it may be neglected without altering appreciably $dT/d\Delta$ as a function of Δ. In other words the top of the mantle is taken as the surface of our model. The fact that the ridge earthquakes must occur somewhere near the boundary of the crust with the anomalous mantle makes this simplification more acceptable. The velocity–distance function can then be inverted into one of velocity–depth such that the velocity for $\Delta = 0$ is equal to the velocity at the surface of our model.

A straight forward analytic approach has been used. A function of the form $V_p = V_0/\cos a\Delta$ has been fitted to the apparent velocity data. This corresponds to a velocity distribution of the form $V = V_0\,(r_0/r)^b$ where V_0 is the velocity of the Earth (mantle) at its surface, r_0 is the radius of the Earth, V is the velocity at radius r and $b = 2a-1$ (Bullen 1963, Chap. 7). If $d = r_0 - r$ is the depth at which the velocity is V, then when $d \ll r$ it is a good approximation to write $V = V_0(1+(bd/r_0))$. Thus we are essentially finding the linear velocity gradient in the top few hundred kilometres of mantle which best matches our observations. The quality of the data hardly merits a more sophisticated approach.

Two velocity functions are shown with the data on Fig. 3, the best fit being given by $V_p = 7{\cdot}0/\cos 2{\cdot}2\,\Delta$. The corresponding velocity–depth functions are plotted in Fig. 4 together with two velocity distributions deduced from the dispersion of surface waves and Herrin's model based on an analysis of P waves. Model 1588 (Kovach & Press 1961) represents the structure of the East Pacific Rise between Easter Island and Pasadena, model 6EGHP1′ (Aki & Press 1961) is an average structure for the Atlantic and Indian Oceans. Herrin's velocity distribution (Herrin

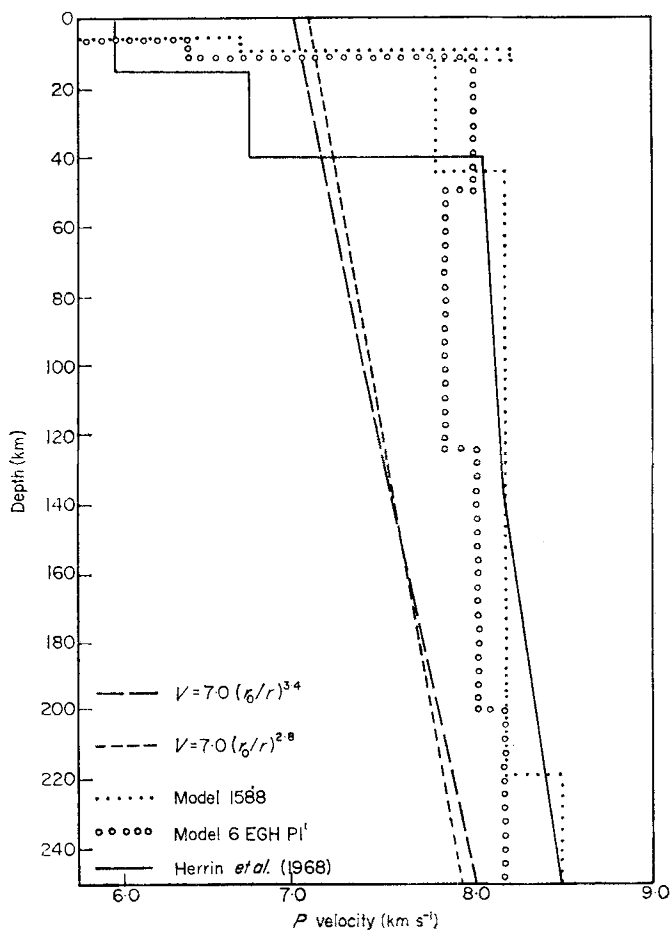

FIG. 4. P velocity distributions in the upper mantle. The two analytic functions correspond to the (V_p, Δ) functions in Fig. 3. Radius of the Earth $r_0 = 6370$ km, depth is $(r_0 - r)$. Models 1588 (Kovach & Press 1961) and 6EGHP1′ (Aki & Press 1961) are based on surface wave dispersion measurements and apply to the East Pacific Rise and to the Atlantic–Indian Oceans respectively. Herrin's distribution is an average Earth based on P wave observations of earthquakes and explosions.

et al. 1968), though capped by a continental structure, does not differ widely from the surface wave models below 40 km depth. The velocity distribution which fits the structure of the ridge axis near Iceland is $V = 7{\cdot}0(r_0/r)^{3{\cdot}4}$ (model A). Correcting this structure for the assumptions made about the overlying crust would involve shifting the curve perhaps 10 km down the depth axis, hardly affecting the broad picture.

Between about 250 and 300 km depth model A coincides with the average Atlantic Ocean structure represented by model 6EGHP1′. This depth of penetration corresponds to $\Delta \simeq 15°$. At shallower depths model A differs more and more from the average structure, and may be regarded as an accentuated low-velocity zone reaching to the base of the crust. It must be added, however, that neither the data nor the method used here are sufficient to detect velocity reversals in the mantle beneath the ridge axis.

From the systematically late arrivals of *P* waves at Reykjavik, Tryggvason deduced that the upper 240 km of mantle beneath the station had a velocity of 7·4 km s^{-1}. The geometric mean velocity of the top 240 km of model A is 7·46 km s^{-1}. The figures to 200 km and 300 km are 7·38 km s^{-1} and 7·58 km s^{-1} respectively. Thus model A agrees closely with Tryggvason's result.

The velocity at the surface of the mantle according to model A is rather lower than 7·3 km s^{-1} found by seismic refraction measurements over the Mid-Atlantic Ridge axial zone (Ewing & Ewing 1959; Le Pichon *et al.* 1965). The difference may be due to the simplifications of the model, for whilst model A fits the data reasonably over the whole range of Δ, for $\Delta < 6°$ the mean velocity observed is 7·3 km s^{-1}.

S velocity data

S arrivals at both REY and AKU were reported for only about half the events for which *P* velocities have been calculated. *S* velocities were calculated in the same manner as the *P* velocities and are given in Table 2. However, the *S* measurements show somewhat greater scatter than the *P*, and of 16 measurements all but two were for $\Delta < 5°$. A detailed analysis as with *P* was not considered worthwhile and it was thought sufficient merely to compare the *S* and *P* velocities. Values of the ratio V_p/V_s are also given in Table 2. At small Δ this ratio is effectively identical to the ratio of compressional to shear velocity (α/β) in the uppermost mantle, hence a measure of Poisson's ratio. Neglecting the single measurement for $\Delta = 14{\cdot}2°$, 14 measurements of V_p/V_s for small Δ give a mean of 1·89. This value is greater than normal for the top 100 km of the mantle, but typical of values for the low-velocity zone. For example, in the detailed model of the oceanic upper mantle, CIT11A, (Anderson & Toksoz 1963; Kovach & Anderson 1964) the ratio α/β of the top 100 km of mantle averages about 1·76, but deep down in the low-velocity zone, between 200 and 400 km, it averages about 1·9. So the factors which combine to produce soft rock and a high ratio of compressional to shear velocity in the low-velocity zone beneath ocean basins appear to operate in the very uppermost mantle beneath the Mid-Atlantic Ridge. However, the value 1·89 differs from 1·76 at only the 90 per cent confidence level, and more measurements are necessary to substantiate this conclusion.

Apparent velocities across the AKU, SID array

Apparent velocities were measured across the AKU, SID array using the method previously described, are listed in Table 3 and plotted in Fig. 5. The paucity of observations at SID has limited the list to eight measurements from North of Iceland events and nine from Reykjanes Ridge earthquakes. The former group straddle the apparent velocity function of model A, as one would expect. The latter group, however, fall well below the ridge axis curve. This must mean that the average velocity along a ray from the Reykjanes Ridge to SID is greater than along the axis of the ridge to AKU. The average velocity to SID can be calculated if we make some simple assumptions about the velocity to AKU and the time residuals at each station.

Table 3

Apparent velocities across the SID, AKU array

Reykjanes ridge events			Events north of Iceland		
Event	Mean (Δ)	Apparent velocity, P (km s^{-1})	Event	Mean (Δ)	Apparent velocity, P (km s^{-1})
11	5·2°	5·62	13	2·2°	7·50
21	4·6	4·82	14	4·4	4·91
22	4·6	9·33	20	3·0	6·84
29	21·5	9·90	23	4·2	6·81
30	5·5	5·24	24	4·3	7·50
31	5·4	6·19	25	1·8	7·37
32	5·4	5·55	26	1·6	7·29
33	4·9	5·55	37	6·2	6·89
38	14·7	7·48			

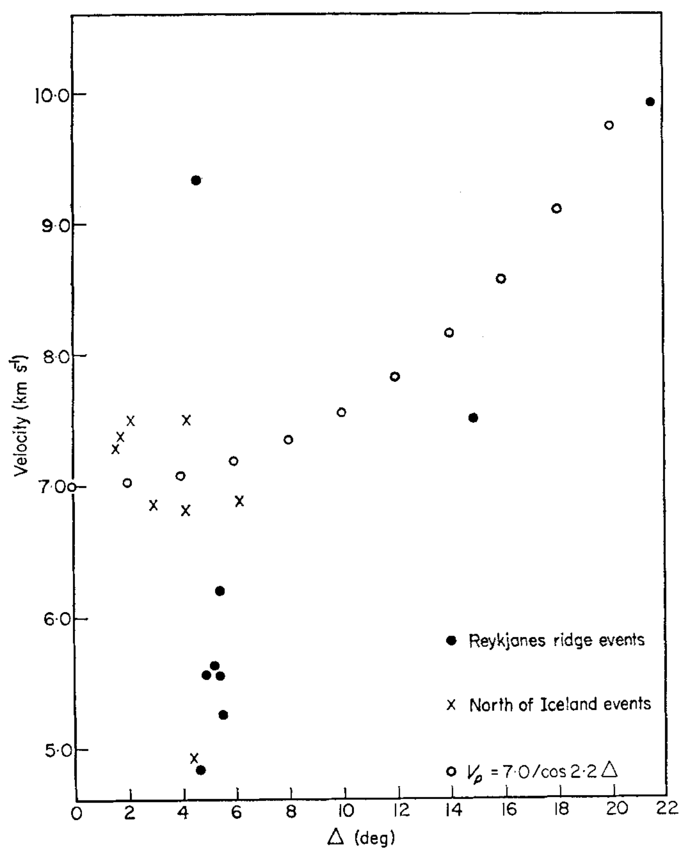

FIG. 5. Apparent Velocities of P waves from events north and south of Iceland observed across the AKU, SID array.

For small Δ the time–distance graph can be approximated by a straight line and we can write, using subscript 1 for the ridge axis station, 2 for the other

$$T_1 = \alpha+\beta_1+\frac{\Delta_1}{V_1}$$

$$T_2 = \alpha+\beta_2+\frac{\Delta_2}{V_2}$$

where α is some unknown source term, β_1 is the time residual of the axial station, V_1 is the average velocity along the ridge axis, etc . . . Making the assumption that $\beta_1 = \beta_2$ we can write

$$V \text{ array} = \frac{\Delta_2-\Delta_1}{T_2-T_1} = \frac{\Delta_2-\Delta_1}{(\Delta_2/V_2)-(\Delta_1/V_1)}.$$

If we now assume that the average velocity along the ridge to AKU is $V_1 = 7{\cdot}3 \text{ km s}^{-1}$, we find that the average velocity to SID, in the case of event 33 for example, is $V_2 = 7{\cdot}86 \text{ km s}^{-1}$. Two interpretations of this striking velocity difference can be proposed: (a) anisotropy of the upper mantle; (b) the low velocity axial zone is quite narrow so that rays travelling to SID leave it and traverse more normal, higher velocity mantle. A velocity difference of over 7 per cent for two paths differing in azimuth by less than 30° implies much too great an anisotropy to be plausible, particularly as detailed seismic refraction measurements of anisotropy have found a maximum range of velocity of less than 4 per cent (Raitt *et al.* 1968). The second interpretation is not only plausible, however, but can be carried further and used to estimate the fraction of the path to SID which lies in the low velocity axial zone. If this fraction is f, the velocity in the axial zone is V_a and in the normal mantle V_b, we can write

$$\frac{\Delta}{V_2} = \frac{f\Delta}{V_a}+\frac{(1-f)\Delta}{V_b}.$$

Putting

$$V_a = 7{\cdot}3 \text{ km s}^{-1}$$

$$V_b = 8{\cdot}2 \text{ km s}^{-1} \text{ gives } f = \frac{66{\cdot}51}{V_2}-8{\cdot}11.$$

This formula has been used to deduce the positions of the boundary of the low velocity axial zone along the rays from events 32 and 33 to SID. These positions are marked by crosses on Fig. 1.

Apparent velocities across Iceland/Greenland station pairs

Apparent velocities were measured across pairs of Iceland, Greenland stations of P arrivals from Reykjanes Ridge and Southern Iceland earthquakes, are listed in Table 4 and plotted in Fig. 6. In contrast to the AKU, SID values all these measurements fall above the apparent velocity curve (model A) fitted to the ridge axis data. This means that the average velocities to Greenland are higher than along the ridge axis, a difference which is interpreted in exactly the same way as the SID result. Rays travelling to SCO or KTG leave the low-velocity axial zone and travel in higher velocity, normal mantle. Making the same assumptions as previously, the average velocities to Greenland were computed and hence estimates of the position of the boundary of the axial zone. Some of these are marked with crosses on Fig. 1, the remainder being left out for the sake of clarity.

It should be emphasised that the formula computing f, being the difference of two similar large quantities, is very sensitive to changes in V_2. A 1 per cent error in V_2 gives an error of 14 per cent in f. Nevertheless, the boundary estimates give a consistent picture of the low-velocity axial zone, extending more or less throughout Iceland and about 150 km to each side of the axis of the submerged part of the ridge.

Table 4

Apparent velocities across Iceland, Greenland array

Reykjanes ridge events				South Iceland events			
Event	Mean (Δ)	Apparent velocity, P (km s^{-1})	Stations	Event	Mean (Δ)	Apparent velocity, P (km s^{-1})	Stations
1	12·6°	8·96	REY, SCO	5	3·5°	7·72	REY, SCO
2	10·7	8·28	REY, SCO		4·4	8·04	AKU, SCO
	11·8	9·04	AKU, SCO	12	3·8	7·75	REY, SCO
7	16·2	8·76	REY, SCO		4·2	7·23	AKU, SCO
11	6·2	7·57	REY, SCO				
	7·3	8·13	AKU, SCO				
15	20·3	10·91	REY, SCO				
16	10·1	7·73	REY, SCO				
17	13·4	8·45	REY, SCO				
18	15·7	10·02	REY, SCO				
38	15·9	8·94	REY, KTG				
	17·0	9·27	AKU, KTG				

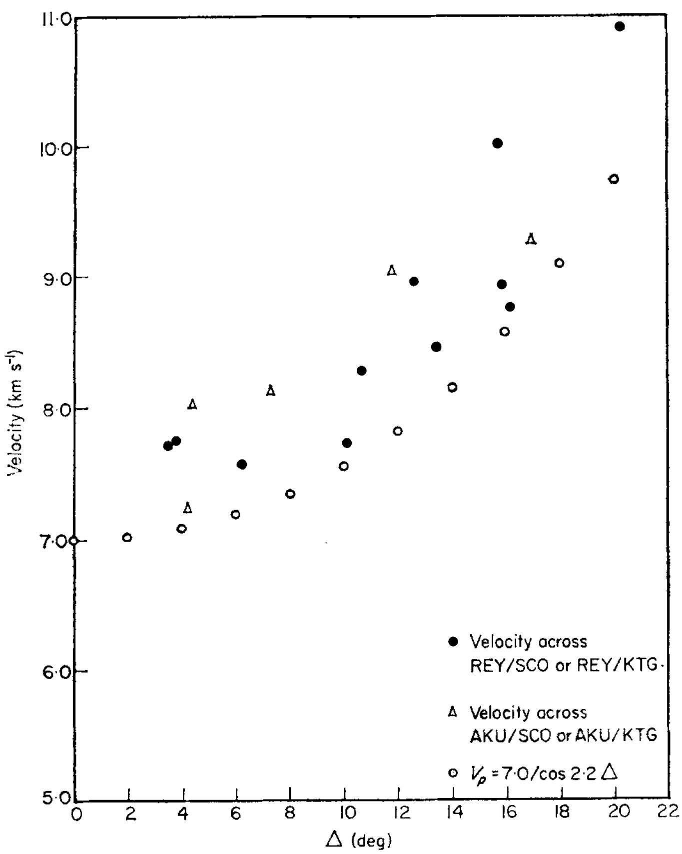

FIG. 6. Apparent velocities of P waves from Reykjanes Ridge and South Iceland events across Iceland/Greenland pairs of stations.

Conclusions

Apparent velocities of P arrivals from Mid-Atlantic Ridge earthquakes measured across the REY, AKU arrays satisfy the velocity–distance function $V_p = 7{\cdot}0/\cos 2{\cdot}2\Delta$ for $\Delta < 15°$. This corresponds to the velocity–depth distribution $V = 7{\cdot}0(r_0/r)^{3{\cdot}4}$, which gives a lower velocity than that typical of the whole ocean down to a depth of about 250 km. This low velocity structure is confined to a belt about 300 km wide between 50° and 70° N. The high ratio of P to S velocity in the upper part of this belt suggests that the low-velocity zone lying some hundred kilometres deep beneath the ocean basins commences at the base of the crust beneath the axis of the Mid-Atlantic Ridge.

The isostatic equilibrium of the Mid-Atlantic Ridge has been explained by a large density contrast ($\sim 0{\cdot}2\,\mathrm{g\,cm^{-3}}$) extending to depths of 30 or 40 km (Talwani *et al.* 1965). This implies the presence of normal mantle beneath the ridge axis at depths greater than 40 km, and is not compatible with the velocity structure deduced here. The alternative is for a small density contrast ($\sim 0{\cdot}03\,\mathrm{g\,cm^{-3}}$) extending some 200 km deep, which requires partial fusion to produce the low P velocities (Bott 1965). This view is compatible with the velocity structure deduced here, and receives further support from the high ratios of P to S velocities observed.

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