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Seismic Structure Along RRISP – Profile I on the Southeast Flank of the Reykjanes Ridge

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Abstract. During the first leg of the 'METEOR-Expedition 45', July 1977, crustal seismic refraction measurements were obtained in the vicinity of the Reykjanes Ridge, south of Iceland. Profile I was located approximately along magnetic lineation anomaly 5 (8.34–9.74 Ma) and was a part of an 800-km-long land-sea seismic experiment. The purpose of the overall experiment was to study the changes in crustal structure of the Ridge near Iceland and to resolve the seismic structure at greater depth than was previously possible by extending the seismic line south of Iceland. At the Mohorovičić discontinuity the velocity increases to 7.7 km/s, a 'typical' low mantle velocity observed frequently in oceanic refraction profiles near ridge crests. A normal upper mantle velocity of 8.2 km/s is observed at a depth greater than 16 km. Reflection profiles show a rough basement topography in the south, becoming smooth towards the north. The results indicate: (i) that a normal oceanic crust is in place within 100 km of an active ridge crest; (ii) that the presence of Iceland has only a second-order effect on the oceanic crust to the south; (iii) that a normal upper-mantle velocity is present underneath a low (7.7 km/s) velocity at the Mohorovičić transition zone; and (iv) that velocity gradients in the lower crust and the upper mantle are consistent with the results of the inversion of time-distance data, but should be confirmed by synthetic seismogram modelling.

Key words: Reykjanes Ridge – Iceland – Seismic structure – Hot spots – Explosion seismology – Anisotropy – Anomalous mantle – Upper mantle – Asthenosphere flow – Mantle plume – Extremal inversion.

1. Introduction

Iceland has fascinated explorers and travellers since the first Norsemen settled there, more than 1,000 years ago. Its rugged volcanic terrain, majestic glaciers and waterfalls, mysterious hot springs and often violent erruptions of lava, all seemed to hold a key to the better understanding of the Earth as a dynamic engine. It is not surprising that the modern research into the nature and structure of the outer rock layers of the Earth, often returned to Iceland and the surrounding ocean for more insight and understanding. The evidence from many directions is presented in this issue. Our paper describes the contribution of the marine program to the Reykjanes Ridge Iceland Seismic Project (RRISP).

Iceland is located astride the Mid-Atlantic Ridge, between latitudes 63° N and 67° N. The edge of Eurasian and North American lithospheric plates (LePichon et al., 1973) is exposed in the neovolcanic zone of Iceland and offers an unparalelled opportunity for study. As a geological feature, Iceland is young: the outpouring of some half a million cubic kilometers of lava occurred within the last 18 Ma (Jakobsson, 1972; Moorbath et al., 1968; Everts et al., 1972). Although Iceland is a part of the Mid-Ocean Ridge system, it differs from a typical cross-section in many important ways: (i) the extensive volcanic activity has created an excess of mass so that it outcrops above the sea surface and forms a 'blister' on the crest of the Mid-Atlantic Ridge; (ii) it is an area of active seismicity and tectonic development; (iii) it is a centre of convective heat flow from the mantel as a postulated 'hot spot'; (iv) the seismic structure indicates a depression of the Moho discontinuity and/or unusually low upper mantle *P*-wave velocity; (v) geochemistry of Icelandic basalts differs from the average composition of mid-ocean ridge basalts (Brooks and Jakobsson 1974). For these and other reasons, Iceland is considered a 'hot spot' (Morgan, 1971) and complicated convection patterns have been postulated to explain its origin (Vogt, 1974).

It is thus accepted that Iceland is an 'anomalous' section of the Mid-Atlantic Ridge but it is not known how far this anomaly extends along the length of the ridge crest. Two transform faults, Tjörness Fracture Zone to the North and Reykjanes Fracture Zone to the south displace the ridge crest through Iceland eastwards. These zones seem to confine the Iceland eruptives but the hot-spot may affect the ridge beyond the fracture zones. The southward extension of the mid-ocean ridge is called Reykjanes Ridge; it is an atypical ridge, probably because of its slow spreading rate (Talwani et al., 1971) and possibly because of the proximity of the hot-spot to the north. Based on an analysis of V-shaped magnetic anomalies (Vogt and Avery 1974) and the distribution of trace elements revealed by geochemical analysis (Schilling, 1973), it has been proposed that some of the lava brought up by the mantle plume under Iceland, finds its way southwards along the ridge axis through hydraulic channeling (Vogt, 1974). In studying the deeper structure of Iceland, it is therefore of great interest to investigate the seismic structure of the surrounding sub-oceanic crust (Bott et al., 1971). The marine seismic experiment, described here, was designed to supplement the land investigations and additionally, to study the uppermost structure of the oceanic lithosphere by extending the length of the seismic refraction Profile I. Data collected along other profiles shown in Fig. 1 will be reported elsewhere.

Long seismic lines at sea are difficult to carry out and only four had been performed prior to the RRISP experiment· two in the Pacific (Asada and Shimamura, 1976, 1979; Orcutt and Dorman, 1977), one in the Atlantic (Steinmetz et al., 1977), and one in the Mediterranean (Hirn et al., 1977). Among the questions

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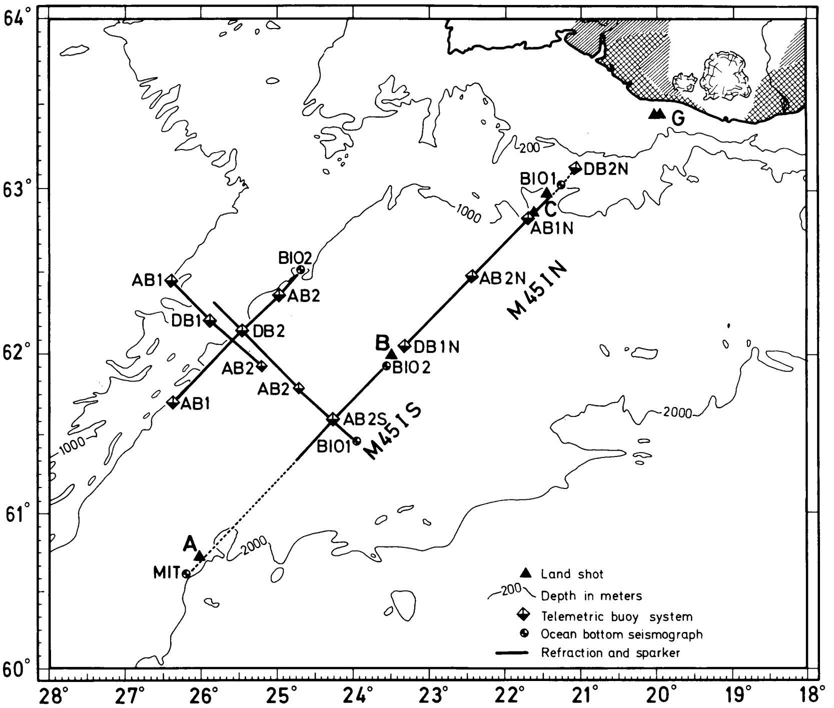

Fig. 1. Geophysical investigations during the METEOR Cruise 45, south of Iceland

posed by these experiments and RRISP were: is there a high-velocity layer 3 B at the base of the crust as shown by Sutton in the Pacific (Sutton et al., 1971) or is there a low velocity zone near the crust-mantel interface (Lewis and Snydsman, 1977); is P-wave velocity in the mantle 'normal', i.e., in the range 8.2 ± 0.2 km/s or is there any anomalously low-velocity mantle away from the axial spreading zone? Is there a velocity gradient in the crustal layers or is a horizontally layered crust an adequate approximation?

The topography of the Reykjanes Ridge is rough in the crestal region and becomes smoother towards the flanks and southwards. The considerable thickness of sediments near the bottom of the continental slope and away from the ridge crest represent an accumulation of erosional material transported from the Iceland Plateau, perhaps by turbidity currents (Fleischer, 1974). The volcanic basement is hidden by a blanket of these sediments and it is not known whether the basement under the flanks is block-faulted and broken up as much as it is near the crest. The structure of the ridge is nearly symmetrical with respect to its axis, and so are the magnetic lineations. The clear correlation of the magnetic stripes in this area confirmed the hypothesis of sea-floor spreading (Vine and Matthews, 1963; Heirtzler et al., 1968). Because of the nature of development of plate boundaries in the North Atlantic, Iceland and Reykjanes Ridge have been the subject of extensive exploration. Gravity and magnetic measurements represent a dense net of observations (Heirtzler et al., 1966; Talwani et al., 1971; Talwani and Eldholm, 1972; Fleischer et al., 1973; Fleischer 1974; Grønlie and Talwani 1978) and reveal a noticeable Bouguer gravity minimum and a high axial magnetic anomaly of over 1,000 *nT*. There is also a substantial body of information regarding the heat flow in the vicinity of the ridge (Talwani and Eldholm, 1977; Grønlie and Talwani, 1978; Bram, 1980; Sclater and Crow, 1979).

In contrast to these extensive geophysical observations, seismic information about the deeper structure of the crust is sparse. The results of early refraction seismic experiments were presented in a compilation by Ewing and Ewing (1959). In this early work usual procedures involved two surface ships. The signal-to-noise ratio was relatively poor and the seismic lines were often too short to detect deeper refractors. About 60 km west of the axis, these authors found a consolidated layer with a compressional wave velocity of 5.7 km/s under local sedimentary troughs of small thickness, and below 6 km depth an anomalous mantle *P*-velocity of about 7.4 km/s. These results only roughly agree with those of Talwani et al. (1971) farther north who found a rise of the material with a similar *P*-velocity (7.4 km/s) from the flank towards the ridge axis. For the crestal area these results correspond to those obtained from earthquake surface wave dispersion observations (Trygvason, 1962). Ariç (1972) presented a crustal section of the west flank of the Reykjanes Ridge from deep reflection seismic results. His depth calculation is based mainly on the *P*-wave velocity structure from the mid-Atlantic Ridge between 0° and 30° N given by Ewing and Ewing (1959) and LePichon et al. (1968). The model calculated shows an extensive body with *P*-velocities varying from 7.3 to 7.7 km/s, underlain by a layer with a velocity of 8.1 km/s. This low-velocity body reaches a depth of 40 km under the crest, and extends about 200 km from the axis. At a distance of about 600 km, this anomalous mantle zone melds into a normal oceanic crust. This interpretation is analogous to the results obtained under Iceland by Zverev et al. (1976). As part of the IPOD/DSDP site surveys (Leg 49, site 409), Snoek and Goldflam (1978) found, at a distance of 18 km from the axis, a high *P*-velocity of 7.9 km/s at a depth of 7 km, a slightly higher velocity than found by Talwani and Eldholm (1977), though still classified an 'anomalous' mantle.

The concept of a mantle plug of lower density and lower *P*-velocity ('anomalous mantle') was introduced by Talwani et al. (1965) as a possible model to explain the observed gravity anomalies across the Mid-Atlantic Ridge. This body extended about 400 km on either side of the median valley and the model was based on the earlier compilation of seismic observations by Ewing and Ewing (1959). From the study of surface wave propagation

Table 1. Type and locations of seismic systems used

System	Latitude (Degree) N	Longitude (Degree) W	Depth (km)	Sensor
M45/IN				
DB2N	63° 7.8′	21° 3.0′	0.366	GHG
BIO1	63° 1.5′	21° 14.1′	0.819	GHG
AB1N	62° 49.8′	21° 40.2′	1.124	GH
AB2N	62° 28.2′	22° 25.2′	1.385	GH
DB1N	62° 2.8′	23° 18.7′	1.542	GH
BIO2	61° 55.0′	23° 34.2′	1.544	GHG
M45/IS				
BIO1	63° 1.5′	21° 14.1′	0.819	GHG
DB1S	62° 2.8′	23° 33.5′	1.439	GH
BIO2	61° 55.0′	23° 34.2′	1.544	GHG
AB2S	61° 35.3′	24° 14.3′	1.605	GH

DB,AB = Digital and analog buoy system, BIO = Ocean bottom seismographs, GH = Ground hydrophones, GHG = Ground hydrophones and geophones

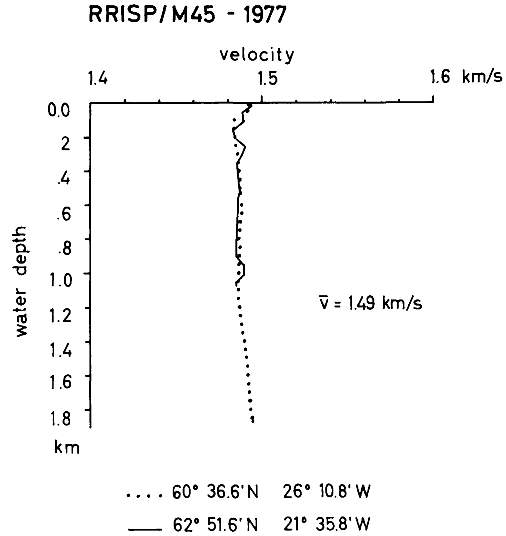

Fig. 2. Sonic log measurements in water at two positions on profile *M45/I*

and teleseimic *P*-delay times (Tryggvason, 1962, 1964; Francis, 1969a, Long and Mitchell, 1970), it was suggested that low-velocity mantle in the vicinity of Iceland may extend to a depth of 150 to 250 km. An examination of the relationship of terrain-corrected Bouguer anomaly to bathymetry within the detailed survey area of the Mid-Atlantic Ridge near 45° N, led Woodside (1972) to suggest that a density deficiency *or* buoyant forces in the upper mantle are responsible for the overall elevation of the crestal mountain region and that the topography of the high-fractured plateau may be partially compensated by undulations of the crust-mantle interface. The search for the low-density, low *P*-velocity mantle under the crestal region and the delineation of the extent of this 'anomalous mantle' away from the ridge has been an objective of much research during the last decade.

Working on Mid-Atlantic Ridge near 45° N, Keen and Tramontini (1970) found the Mohorovičić discontinuity at a mean depth of 7.5 km with a mean velocity of 7.9 km/s for the underlying material. No evidence was found for anomalous mantle material except within the immediate vicinity of the median valley and low *P*-velocities were interpreted as a result of anisotropy. Later work of Fowler (1978) within the same area, re-interpretation of Keen and Tramontini (1970) data by Fowler and Keen (1979), work of Whitmarsh (1975) and Fowler (1976) on the Mid-Atlantic Ridge near 37° N (FAMOUS area), and work of Whitmarsh (1978) and others on the ridge flanks north of the Azores, all confirmed that the crustal structure is more complicated than that described by the standard oceanic model (Raitt, 1963), that a low-density, low *P*-velocity mantle is confined to a narrow axial zone, perhaps not more than a few kilometres wide, and that away from the axis a 'normal' oceanic crust is formed within a few million years.

On the basis of all these results it appeared that Reykjanes Ridge may be different compared to the rest of the Mid-Atlantic Ridge. The question of how it is related to Iceland could only be answered by the knowledge of the deep crustal structure of both Iceland and Reykjanes Ridge. This was the reason for carrying out an 800-km-long refraction profile with 53 mobile and 37 permanent stations on Iceland and 8 stations at sea (consult RRISP Working Group 1980, Gebrande et al., 1980). This combined land/sea profile runs along the eastern flank of the Reykjanes Ridge parallel to the bathymetry contours and somewhat obliquely to magnetic anomaly 5 (8.34–9.74 Ma; LaBrecque et al., 1977), crosses Surtsey and Vestmannaeyjar and continues into the young volcanic zone of Iceland. We present in this paper the results from the sea end of this profile.

2. *Description of the Experiment*

The seismic refraction experiment along the seaward extension of the land profile, called profile M45/I (Fig. 1) was carried out in two parts. Along the 187 km northern section, Profile M45/IN, six seismic receiving systems were launched (Table 1). Four of these were anchored telemetering buoy systems of the Institut für Geophysik, Hamburg (Kebe, 1971, Weigel et al., 1978) and the other two were ocean bottom seismometers (BOBS) of the Bedford Institute of Oceanography, Dartmouth, N.S. (Heffler and Barrett, 1979). All systems included a hydrophone near the sea floor. In addition, BOBS had two geophones (one vertical, one horizontal) but the records from these were too noisy for detailed analysis. The two ocean bottom seismometers, BIO 1 and BIO 2 (Table 1), stayed on the bottom for 14 days and were also used for measurements along the 80-km-long southern section – Profile M45/IS. Along this section, two additional telemetering buoys were anchored, again with sea-floor hydrophones. During the whole seismic experiment, all systems worked satisfactorily except for the buoy DB1S. This buoy drifted 8 km due to unknown reasons, so the results can be used only for qualitative purposes. For the combined land/sea experiment, charges of high explosive GEOSIT II from 25 to 4,000 kg were detonated electrically. In total 141 shots, about 1.8 km apart were fired.

For the determination of the shot and seismic-system coordinates, Loran C and integrated satellite navigation were used, and the raw data were corrected by the method described by Goldflam and Goldflam (1979). Sound velocity in water measurements were carried out at two stations (Fig. 2) and were used to calculate shot to receiver distances. A very weak low-velocity channel appears within the first 100 to 200 m. The mean sound velocity of the sea water remains constant at 1.486 km/s and this value

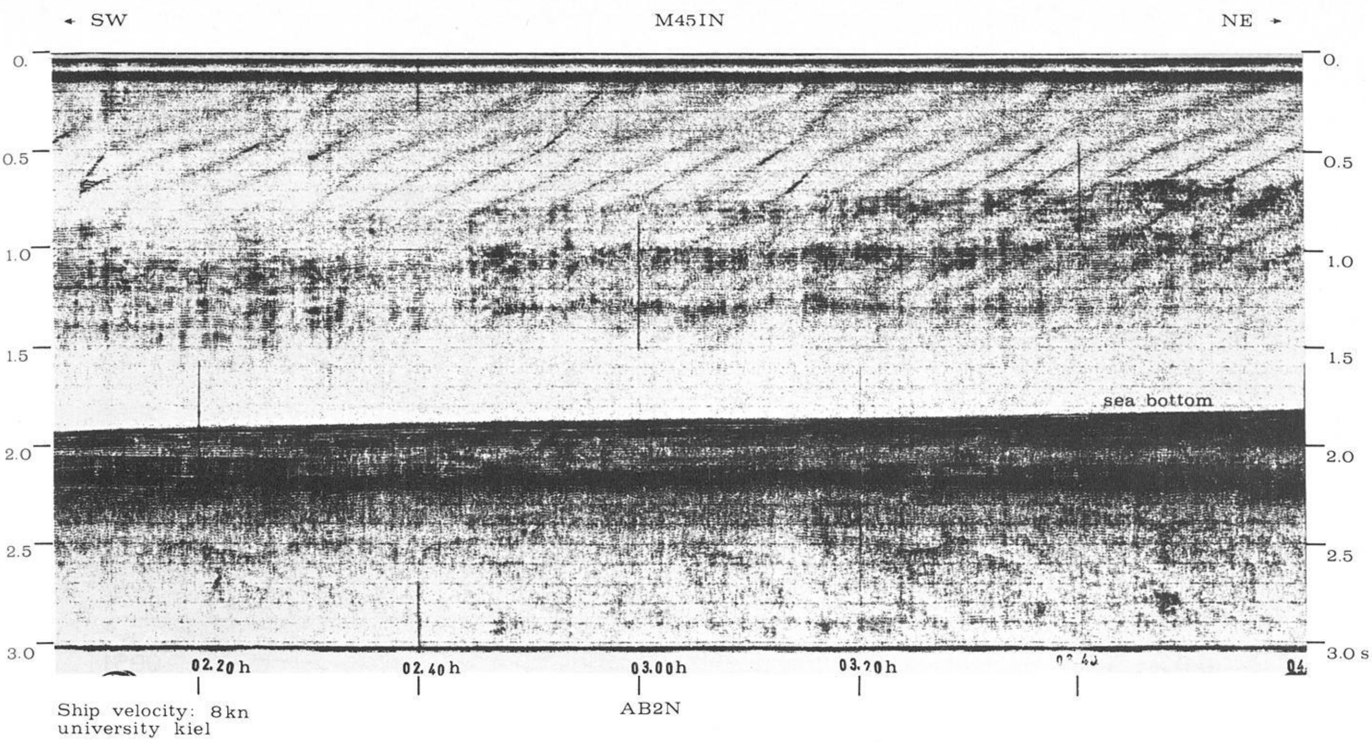

Fig. 3. Sparker results in the area of the buoy AB2N characterized by smooth acoustical basement

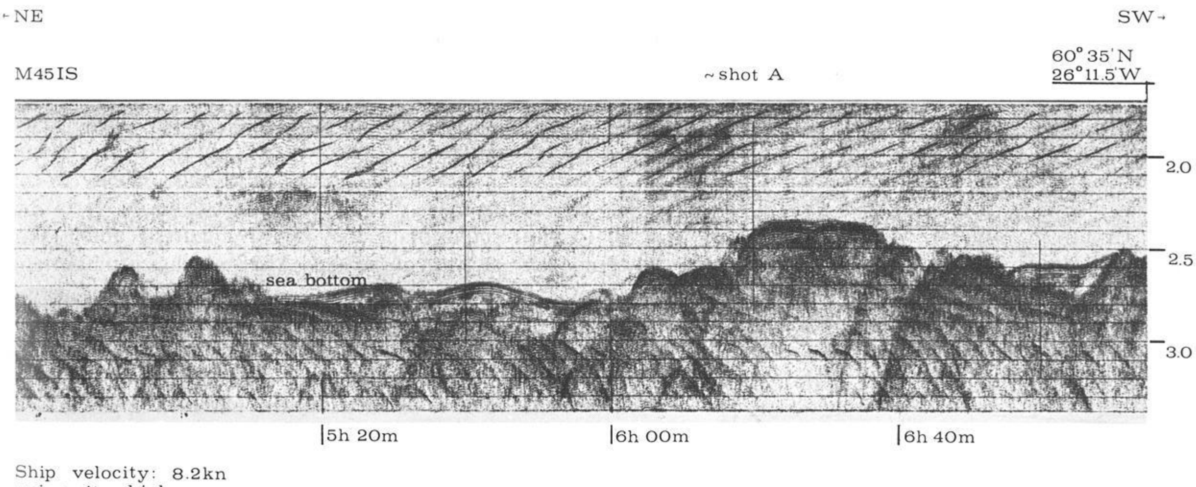

Fig. 4. Sparker results NE and SW of shot A. Sediments are concentrated only in local depressions of the roughly shaped outcropping basement

was used for all calculations. Comparisons of geodetic distances with distances calculated from water-wave arrivals indicate an accuracy of geographical position determination of 200 to 300 m.

To obtain the fine structure of the upper sediments, seismic reflection data using a sparker system were collected along the profile. Over most of the profile we observed smooth acoustic basement and sediment thicknesses up to about 400 m which thin from north to south. In the southern part of the profile, close to shot A, the basement becomes rough (Fig. 4) and the sediments are concentrated only in local depressions between the outcropping basement highs. These features of the basement topography are important for tectonic interpretation of the refraction data (see also RRISP Working Group 1980).

3. The Results

Out of a total of 15 seismogram sections compiled for the interpretation of the refraction observations on Profile M45/I, only four representative examples are presented in this paper in order to save space. The ranges of observations of each system are discussed in RRISP Working Group 1980, (see Fig. 2). All Hamburg seismograms were filtered with a band-pass of 2 to 20 Hz which roughly corresponds to the frequency response of BOBS. The sections are not corrected nor are the arrivals normalized for variations in shot charges. Only the BIO 1 record section is corrected for the attenuation with distance. According to the theory of propagation of head waves (Červeny and Ravindra, 1971), at long ranges

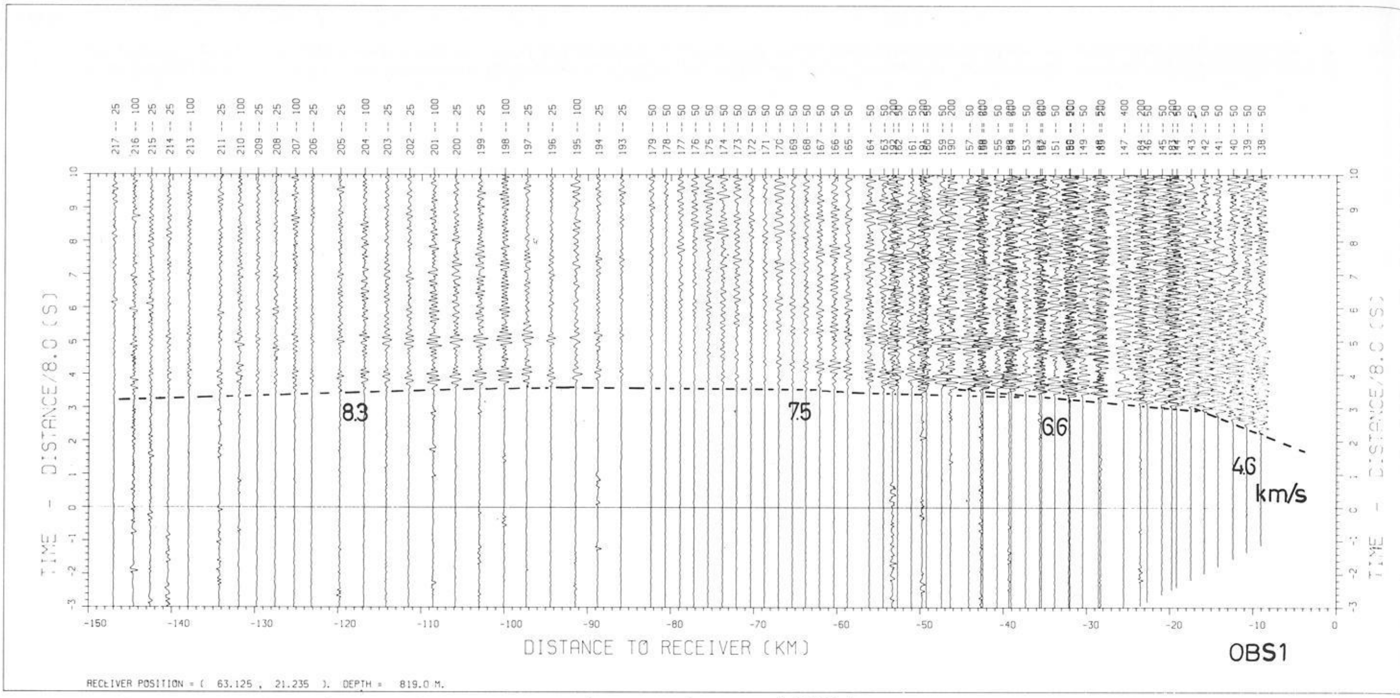

Fig. 5. Seismogram section of the northernmost ocean bottom seismograph BIO 1

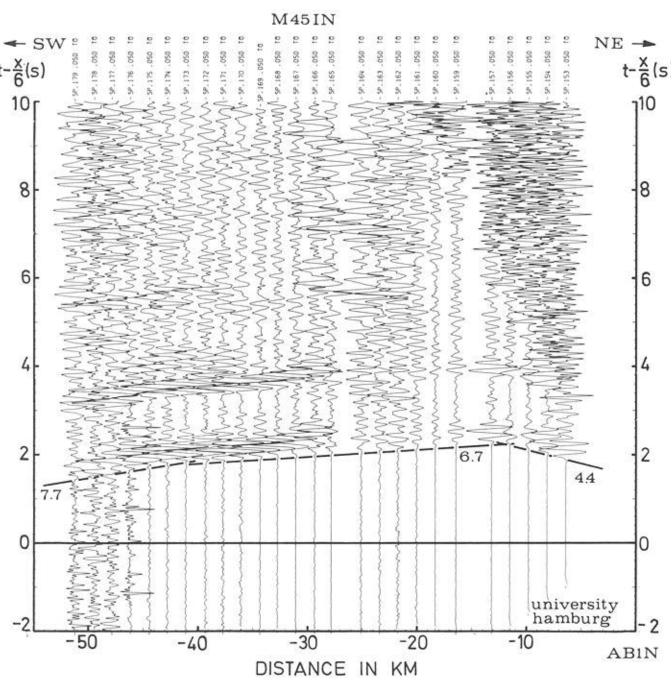

Fig. 6. AB1N seismogram section. Very good phase correlation (supported by clear multiple refraction arrivals)

the first arrival separates from the interference packet by more then the pulse length and so the amplitude decays as r^{-1}. This relationship was used and the constants of proportionality so adjusted that a signal at 120 km range is corrected by a factor of two (i.w., an attenuation of 50% is assumed at that range).

In the illustrations given, the time axis corresponds to the reduced time with a reduction velocity of 6 km/s except for BIO 1 (Fig. 5). Because of long range of observations – almost the whole length of profile M45 I – a reduction velocity of 8 km/s was used for this section. In this seismogram section we can recognize the main horizons which are here characterized by the apparent velocities: $V_{4-}=4.6$; $V_{5-}=6.6$; $V_{6-}=7.5$; and $V_{7-}=8.3$ km/s. Similar results were obtained in the remaining sections, as for example in the reversal part between the buoys AB1N and AB2N (Fig. 6) except that no upper-mantle velocity was observed. The absence of any arrivals indicating the presence of this velocity (V_7) is due to the shorter range of observations on these buoys as compared to the results obtained by BIO 1. Figures 7 and 8 show the seismogram sections of the buoy AB2S which is of importance bacause the apparent velocities V_{4+} and V_{4-} indicate an increase of the true *P*-velocity for this particular layer southwards. This indicates a different crustal constitution in the south as already suggested by the changes in basement morphology.

In Table 2 we have summarized the observed apparent velocities for comparision of the results of all seismic systems. We designate with V_{i+} the apparent velocities of the arrivals originating from shots northeast of the seismic receivers and with V_{i-} those fired southeast of the receiver (Fig. 1). With regard to the horizontal nature of the sediment layer (Fig. 3) and taking into account all the calculated apparent velocities (Table 2), a model was developed to fit the observed travel times on Profile M45/IN by ray-tracing. The velocities at the upper boundaries of the model are as follows: $V_1=1.49$, $V_2=1.6$, $V_3=2.2$, $V_4=4.4$, $V_5=6.7$, $V_6=7.7$, and $V_7=8.2$ km/s. The algorithm used in calculating the ray tracing model required a small velocity gradient in the layers (Gebrande 1976). Similar model calculations along the southern section of this profile (M45/IS) – northeast and southwest of the position of AB2S – show slightly different V_p values for layers 4 to 6 namely: $V_4=4.9$, $V_5=6.5$, and $V_6=7.7$ km.

The extremal inversion of travel time data by the tau-p method (Bessonova et al. 1974; Bessonova et al. 1976; Kennett, 1976) was applied to the data recorded at BIO 1. We find that the model velocities calculated by conventional methods are within the extremal bounds on possible velocity-depth distributions obtained by

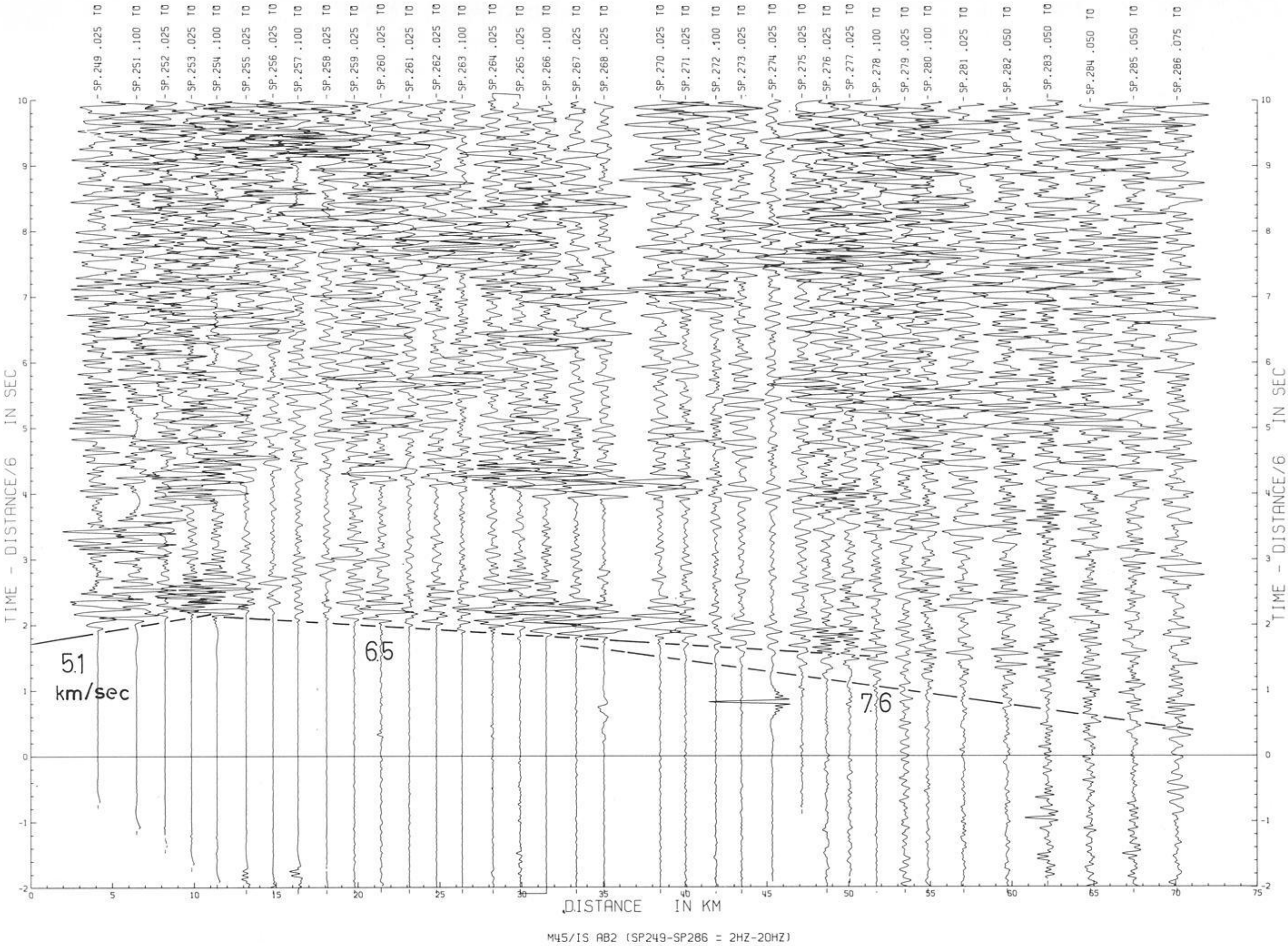

Fig. 7. Results of the observation of the Buoy AB2S on the southern section M45/IS (NNE direction)

the tau-p inversion (Fig. 9). The upper boundary of possible V(z) functions indicates that the transition from 6.7 to 7.7 km/s and from 7.7 to 8.2 km/s layers may be step like. The absence of velocity gradients (i.e., a layered solution) and a presumed absence of 'interference head waves' (Kennett 1977) may account for the low amplitudes of first arrivals at BIO 1 (Fig. 5). The larger amplitudes between 90 and 120 km may be due to a slight positive velocity gradient in the upper mantle. Another explanation for the increased amplitude at greater range is the possible arrival of wide-angle reflections from the mantle boundary *(PnP)*, as observed by Lewis and Snydsman (1977) in the Pacific. For strong reflections, the boundary would have to be sharp and this has implications for petrologic models which may explain our observations as discussed later. These speculations are quite qualitative, however, and should be verified by calculation of synthetic seismograms.

4. Interpretation and Discussion

The main part of the RRISP experiment (see RRISP Working Group 1980; Gebrande et al. 1980) has confirmed the findings of many earlier investigators that the crust under Iceland is 'anomalous'. In contrast, we find under the flanks of the Reykjanes Ridge south of Iceland nearly normal crust, though of greater thickness than elsewhere on mid-ocean ridges. Two aspects of our results deserve further discussion: (i) the absence of strong along-strike changes in crustal thickness as Iceland is approached; and (ii) the nature of 7.7 km/s layer and the transition from 7.7 to 8.2 km/s at a depth of 16 km.

The uppermost (unconsolidated sediment) layer decreases in thickness with distance from the Iceland plateau as is to be expected. This is a surficial phenomenon due to erosion and redistribution of sediments and is not important for understanding deeper structure and tectonic development of the region. Of more interest is the slight increase in the velocity of layer 2 from 4.4 to 4.9 km/s (north to south) and perhaps some thickening southwards. This result needs confirmation from the other seismic lines in the area. These velocities are identical to those of wide-spread basaltic nappes on and around Iceland. Because of the oblique crossing of anomaly 5 by our profile the rocks under the southern portion of Profile M45/I are older and get progressively younger towards the north. Our results would thus suggest an increase in velocity of layer 2 with age. The change (of about 10%) is too large to be explained by compression of basalts and closure of cracks and, if real, may be due to a slight compositional variations in the chemistry of the rocks away from the Iceland plume. Of greater interest, however, is the almost horizontal layering of deeper crustal layers.

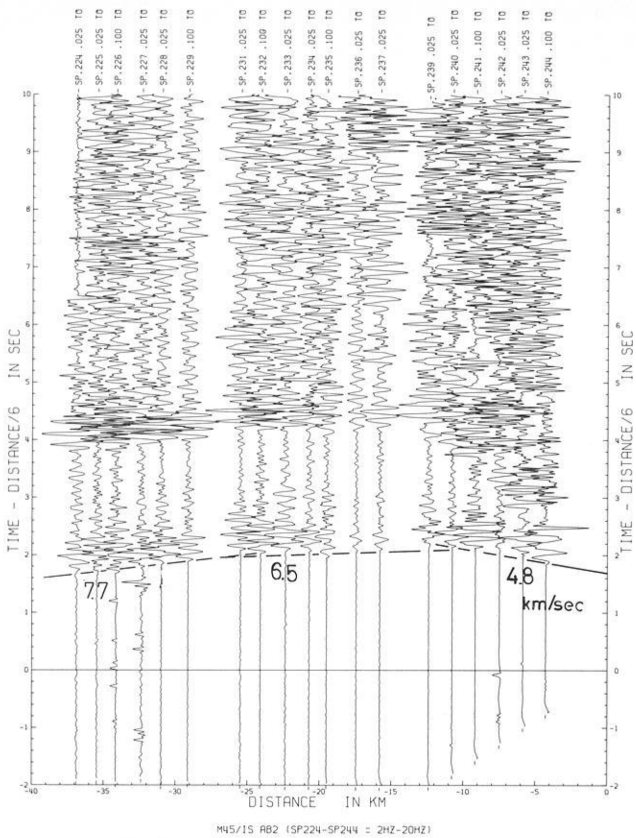

Fig. 8. Results from AB2S to SSW

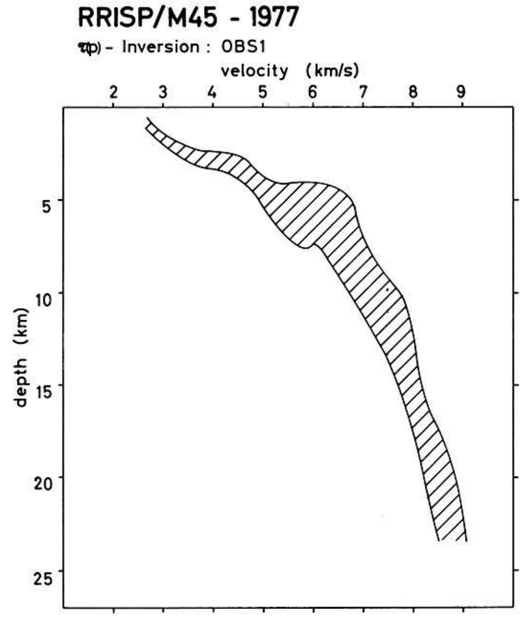

Fig. 9. Extremal bounds for the possible velocity-depth function V(z) obtained by the tau-P inversion applied to the travel time data of the BIO 1

The main crustal layer under Iceland of velocity 6.35 km/s (Pálmason, 1971) to 6.8 km/s (RRISP Working Group 1980) corresponds to the 6.7 km/s layer observed under Profile M45/1 (Fig. 9). The base of this layer under Iceland is at a depth of 8 to 18 km and is underlain by a layer of velocity 7.0 km/s which has a slight positive velocity gradient and may extend to a depth of 150 to 200 km (Tryggvason, 196', 1964; Long and Mitchell, 1970). This is at complete variance with sea observations. The oceanic structure determined under M45/I must come to an abrupt termination just south of the mainland of Iceland, perhaps along the walls of the Reykjanes Fracture Zone. This result places a constraint on the depth of asthenosphere flow outwards from a mantle plume as proposed by Vogt (1974; Vogt and Avery, 1974). This has further implications for search for 'archeo-plumes'. If the anomalous crustal structure is confined to the immediate vicinity of the mantle plume, then traces of ancient plumes will be found only by studying anomalous crustal structures over small areas. Systematic search for ancient or exhausted mantle plumes by seismic refraction techniques is not a practical project at the present time.

A *P*-velocity of 7.7 to 7.9 km/s has been measured frequently at the base of the oceanic crust and interpreted as the crust-mantle or Mohorovičić discontinuity (Wyllie 1971). For example, in an experiment carried out about 200 km southeast from the position of BIO 1, Whitmarsh (1971) found a deep layer of velocity 7.84 km/s at a depth of about 9.5 km. A layer of similar velocity (7.74 km/s, depth 7 km) was detected by Steinmetz et al. (1977) at the eastern flank of the Mid-Atlantic Ridge north of the Azores. Measurments on the younger crust of faster spreading ridges in the Pacific have also found similar velocities: 7.3 to 7.9 km/s near the Explorer Ridge (Malecek and Clowes 1978) and 7.5 to 8.2 km/s (age dependent) on the Cocos Plate (Lewis and Snydsman 1979). None of these experiments reported a further sharp increase in *P*-velocity at a greater depth as we observed under M45/I. This is significant and can be explained in several ways. (i) The 7.7 to 8.2 km/s transition is widespread but has not been observed

Table 2. Observed apparent *P*-velocities in km/s

Observed to SW				Shot point	Buoy	Shot point	Observed to NE				
$\bar{v}_{4-}$	$\bar{v}_{5-}$	$\bar{v}_{6-}$	$\bar{v}_{7-}$				$\bar{v}_{3+}$	$\bar{v}_{4+}$	$\bar{v}_{5+}$	$\bar{v}_{6+}$	$\bar{v}_{7+}$
		7.7		138–146	DB2 N						
4.6	6.6	7.5	8.3	138–217	BIO 1						
4.4	6.7	7.7		153–179	AB1 N	138–148		4.6	6.8		
	6.7	7.7		199–220	AB2 N	147–179		4.5	6.8		
					DB1 N	169–219		5.0	6.9	7.4	
	6.7			252–283	DB1 S						
4.9		7.6		224–277	BIO 2	147–278	2.1	4.9	6.7	7.5	8.3
4.8	6.7	7.7		224–244	AB2 S	249–286		5.1	6.5	7.6	

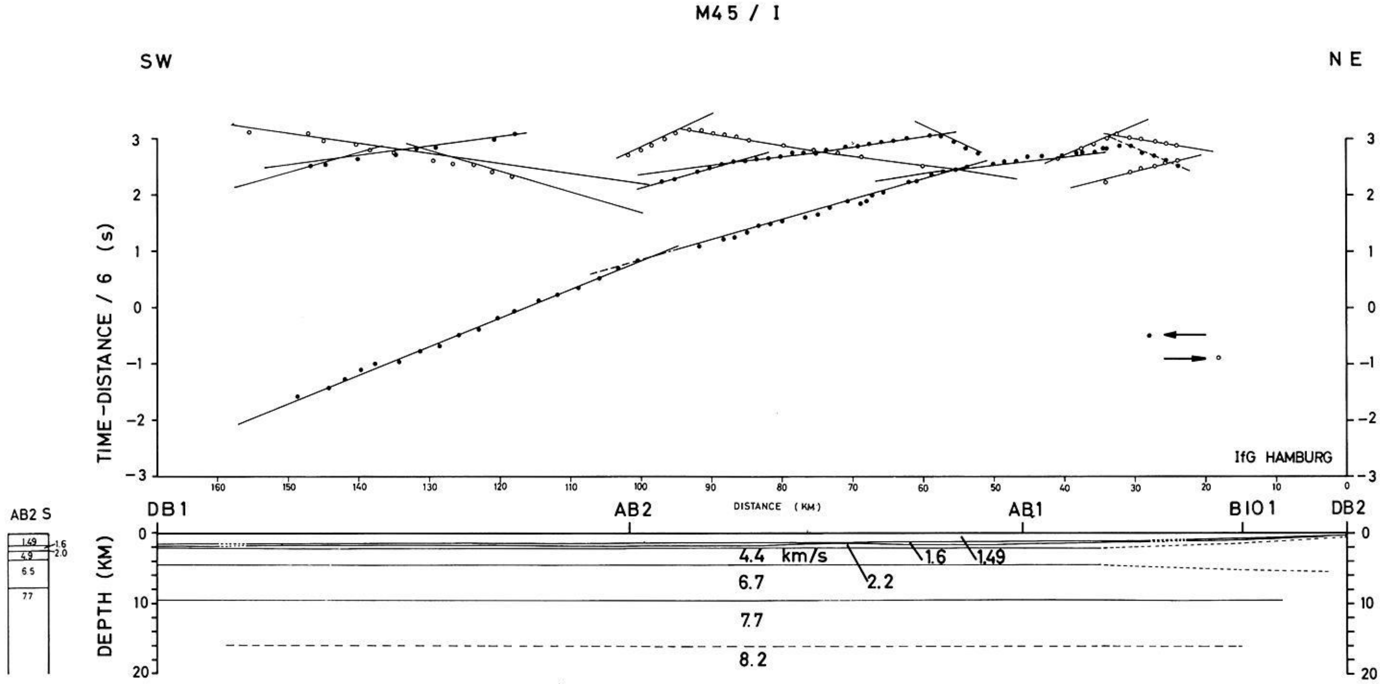

Fig. 10. Crustal model of the eastern flank of the Reykjanes Ridges calculated by means of the ray tracing theory *(bottom)*. Upper part: calculated *(solid lines)* and observed *(symbols)* travel time

elsewhere because of insufficient length of profiles (typically less than 100 km); (ii) this transition is widespread and normally at a greater depth; peculiar conditions under Reykjanes Ridge (elevated temperatures? asthenosphere flow?) make it possible to observe the transition there; and (iii) the transition is a special feature of Reykjanes Ridge. As will be seen from the subsequent discussion, we favour explanation (iii) allowing that (ii) also may be possible. The correct choice depends on a plausible explanation of the 7.7 to 8.2 km/s transition. We offer here one explanation based on anisotropy. We also note that there are other possible explanations based on phase transitions in different petrologic models proposed for the composition of the lower crust and the upper mantle.

The anisotropy in seismic velocity measurements is a measure of dependence of elastic parameters on the direction of propagation of *P*-waves. As was originally shown by Birch (1960, 1961), the strong variation of compressional wave velocity with propagation direction in ultramafic rocks is related to prefered orientation of olivine. At temperatures less then 1,000° C an low strain rates over a geological time scale, olivine crystals may be aligned in the direction of maximum compressional velocity (a crystallographic axis) parallel to the principal glide plane (Francis 1969 b). Measurements on single crystals of olivine by Verma (1960) detected velocities of 9.87, 7.73, and 8.65 km/s in the *a*, *b*, and *c* crystallographic directions, with intermediate velocities possible in intermediate directions. Recent detailed measurements on field samples from the ophiolite complex of western Newfoundland by Christensen and Salisbury (1979) have confirmed this range of variations on a suit of rocks presumed to represent an upthrust and exposed section of the oceanic mantle.

In several marine seismic refraction experiments where anisotropy was detected, the variation in velocity is of the order of 5 to 8% with the direction of minimum usually perpendicular to the direction of spreading (Raitt et al. 1969, 1971; Keen and Tramontini 1970, Keen and Barrett 1971, Lewis and Snydsman 1979). All of these experiments were performed relatively close to the spreading axis, on crust less than 12 Ma old. On the flanks of Reykjanes ridge, on crust 35 Ma old, careful experiment by Whitmarsh (1971) failed to measure anisotropy. Lewis and Snydsman (1979) have shown that on a fast-spreading ridge (4.4 cm/a half spreading rate on Cocos plate) the anisotropy decreases from 0.6 km/s to 0.3 km/s within the first 10 Ma of lithospheric plate development. It is thus possible that Whitmarsh's experiment was too far from the spreading axis and his results do not preclude the possibility that our observation of 7.7 km/s layer represents mantle with low velocity due to anisotropy. [Our additional lines in the area (Fig. 1) may resolve this question. Unfortunately there are problems with the data reduction and we cannot report the results from these lines at the present time.]

In the deepest layer observed in this experiment the velocity increases from 7.7 km/s to 8.2 km/s. This range of the *P*-velocity change is of the same magnitude as the change that could be caused by anisotropy due to a prefered orientation of olivine crystals. We therefore wish to consider a model in which the uppermost part of the mantle is composed of petrologically uniform material but with a change in *P*-velocity with depth due to rheological causes.

We could envisage a couple of forces acting on the uppermost lithosphere, perhaps over different time scales. The first, and perhaps dominant, force is the driving force of the ocean-floor spreading. This causes strain and the minimum of *P*-velocity would be parallel to the Reykjanes Ridge axis. This would be the 7.7 km/s velocity observed along our profile M45/I. The orientation of the crystals which gives this low velocity could be 'frozen-in' early in the cooling stage of the lithosphere development.

At depth, another force may be acting according to the hypothesis of outward flow from a mantle plume as developed by Vogt and Avery (1974). If there is a significant flow at depth away from Iceland parallel to the direction of, and under, the Reykjanes Ridge, then the temperature and viscous stress may be sufficient to re-orient the olivine crystals along their gliding planes. The maximum P-velocity in this depth range would be in the direction of flow, i.e., parallel to the ridge crest. This would be the observed 8.2 km/s velocity.

The transition between 7.7 km/s and 8.2 km/s could be gradual or sharp, depending on the actual mechanism of olivine crystal re-orientation. If the temperature gradient and stress field are uniform, the crystals may be reoriented gradually and all the intermediate velocities between 7.7 and 8.2 km/s would be present along our profile. On the other hand, if there is a 'stickiness threshold' which must be reached before the plastic deformation begins and the crystals start to re-orient, and if, once this threshold is reached, the reorientation proceeds until the alignment in the new direction is completed, then the change from 7.7 to 8.2 km/s could be quite sharp. Synthetic seismograms may discriminate between the models with sharp and gradual transitions between 7.7 and 8.2 km/s and we intend to make these calculations in the near future. Incidentally, the couple produced by the spreading in the southeasterly direction and the asthenosphere flow in the southwesterly direction, may explain the highly developed en echelon fracturing of the ridge crest (Laughton et al., 1979).

Finally, we want to comment on the possibility that the decrease of velocity from 8.2 to 7.7 km/s is due to petrological changes in the upper mantle, the most likely candidate being serpentinization of the ultramafics at the base of the crust. Serpentinization was recognized as an important process in the development of the lower crust by Hess (1955). Geophysical consequences of serpentinization are a decrease in seismic velocity (Christensen 1966) and an increase in volume. To explain the observed change in P-velocity the degree of serpentinization would not have to be great, perhaps 10 to 15%. This is consistent with observations that in most ophiolite suites, the lowermost section of ultramafics is serpentinized (Clague and Straley, 1977). The volume expansion could provide the upward thrust to maintain the crest of the Reykjanes Ridge at a high elevation and thus explain the remarkable absence of an isostatic gravity anomaly compared to other cross-sections of the Mid-Atlantic Ridge (Cochran, 1979). Serpentinization, however, requires water, and it is difficult to envisage hydrothermal circulation to the depth of 10 to 16 km (Lister 1974; Fehn and Cathles 1979). Elevated temperatures at the bottom of the crust in the neighbourhood of the hot spot would prevent the serpentinization reaction above 450° C even if water was present. For these reasons we reject the serpentinization hypothesis and suggest vertical anisotropy as an explanation of our observations.

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