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BOREHOLE SEISMIC EXPERIMENTS AND THE STRUCTURE OF UPPER OCEANIC CRUST

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INTRODUCTION

Borehole Seismic Experiments

During the past five years it has been possible to carry out seismic experiments in the deep sea with clamped borehole geophones. This paper reviews three borehole seismic experiments in oceanic crust, discusses the implications of borehole seismology with respect to crustal structure and compares borehole results with more conventional ocean bottom receiver data.

A summary of the three borehole seismic experiments or oblique seismic experiments (OSE) is given in Table 6-3. All the experiments were carried out from the D/V GLOMAR CHALLENGER with the assistance of 'shooting ships' which fired small explosive charges out to ranges of 12 km (Figure 6-29). The first experiment was carried out in the Western Atlantic Ocean south of Bermuda (Figure 6-30) (Stephen, 1979; Stephen et al., 1979, 1980). Excellent quality three component data were collected at two depths in the hole from shots out to 12km range at four azimuths. Traveltime, synthetic seismogram and particle motion analyses were applied to the data. In the second experiment in the Gulf of California only vertical component data were obtained and traveltimes and amplitudes were analyzed (Stephen et al., 1983). The most recent experiment was carried out in the Costa Rica Rift area (DSDP Hole 504B) and only traveltime information was obtained for the borehole receiver (Stephen, 1983a). However, three ocean bottom hydrophones (Koelsch and Purdy, 1979) were also deployed. Detailed bathymetry within 15km of each site is given in Appendix A.

The original objectives of the borehole experiments were established on the assumption of two kilometers penetration into oceanic crust. The objectives were 1) to determine the lateral extent of the velocity structure intersected by the borehole, 2) to analyse the role of fissures and large cracks (greater than centimeter size) in the velocity structure of oceanic crust, 3) to look for seismic anisotropy in upper oceanic crust and 4) to obtain measurement of attenuation. The deepest geophone emplacement to date in oceanic basement has been only 542m, which is slightly greater than the seismic wavelengths involved in the experiment. Since techniques for studying wave propagation through laterally varying media are limited primarily to traveltime analysis the first objective has not been met. The accuracy of traveltime measurements using free falling explosive shots is not sufficient to resolve changes in wave velocity over very short distances. However, by comparing the refraction velocities from the borehole experiment to sonic log velocities and laboratory sonic measurements on recovered core material, an estimate of the large-scale porosity can be made (Salisbury et al., 1979). Seismic anisotropy can be studied effectively by particle motion analysis of three

Table 6-3

O.S.E. Operations Summary

Leg	52	65	70
Hole	417D	485A	504B
Date	March 1977	March 1979	December 1979
Co-ordinates	25 06.69'N	22 44.92'N	1 13.6'N
	68 02.82'W	107 54.23'W	83 43.8'W
Location	Western Atlantic	Gulf of California	Costa Rica Rift
Age of Crust	100 my	< 1 my	6 my
Water Depth	5479 m	2981 m	3463 m
Sediment Thickness	343 m	153 m	275 m
Geophone Depth in Basement (1)	8 m	71 m	52 m
Azimuths of Shot Lines (1)	N22.5 E	N30 E, N15 W N75 E, N60 W	N,E
Number of Shots (1)	45	220	96
Geophone Depth in Basement (2)	228 m		542 m
Azimuths of Shot Lines (2)	N22.5 E 67.5 W		N,E
Number of Shots (2)	120		79
Ocean Bottom Hydrophones	None	None	Three
Basement Topography	unknown	known	unknown
Spreading Direction	N22.5 E	N30 E	NO
Shear Wave Analysis	Yes	No	No
Amplitude Analysis	Yes	Yes	No
Shooting Ship	R/V Virginia Key	R/V Kana Keoki	R/V Gilliss

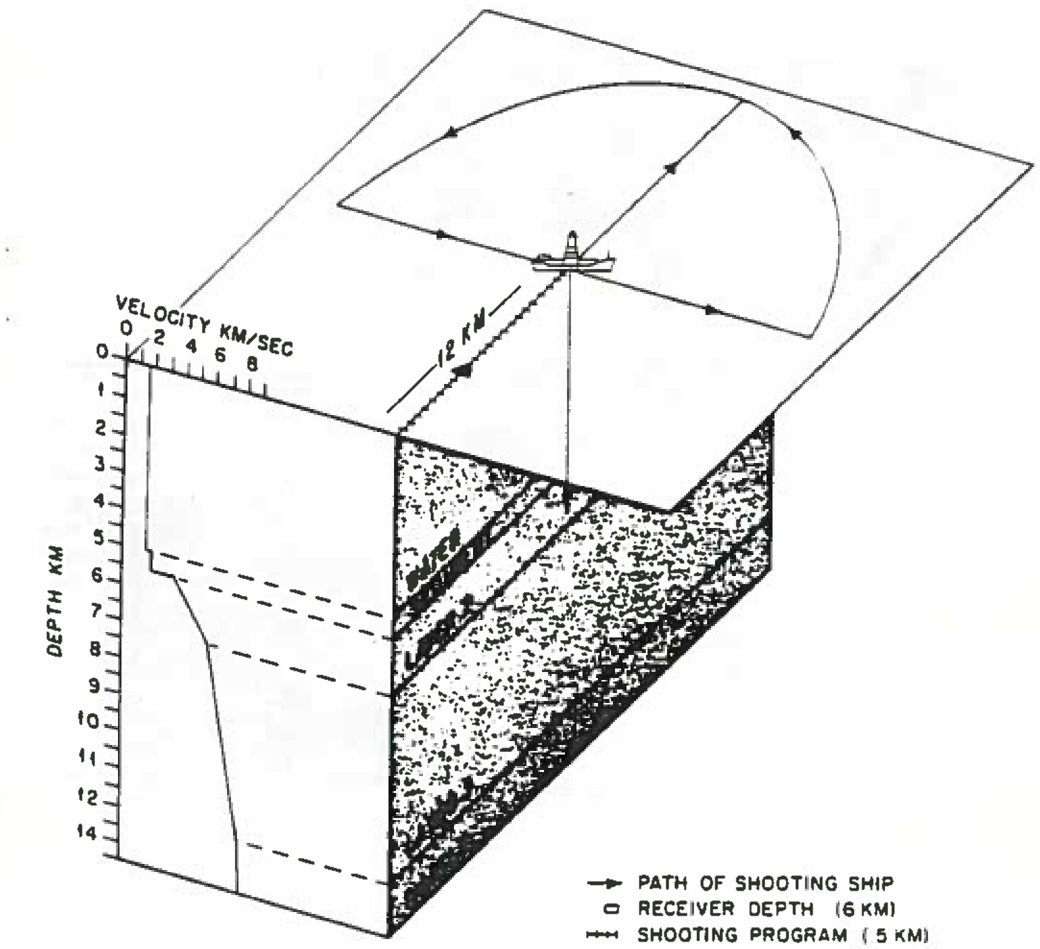

Figure 6-29 Schematic diagram of an oblique seismic experiment (OSE) over oceanic crust. In an OSE, a three-component borehole seismometer is clamped in the open hole at a fixed depth and shots are fired at the surface at a number of azimuths out to ranges of twelve kilometers or more. (Figure from JOIDES Journal).

component data as shown by Stephen (1981) for the Western Atlantic data. The observed anisotropy is attributed to the preferred orientation of large scale fissures and faults. Attenuation has remained an elusive objective because of the small penetration into basement material.

Although not originally an objective, the borehole seismic results have proved useful in determining shallow basement velocity structure. The structure of the upper 500m of basement is poorly resolved using ocean bottom or ocean surface receivers. Unfortunately, traveltime inversion schemes (e.g. Bessanova et al., 1974; Garmany et al., 1979; Dorman and Jacobson, 1981) require

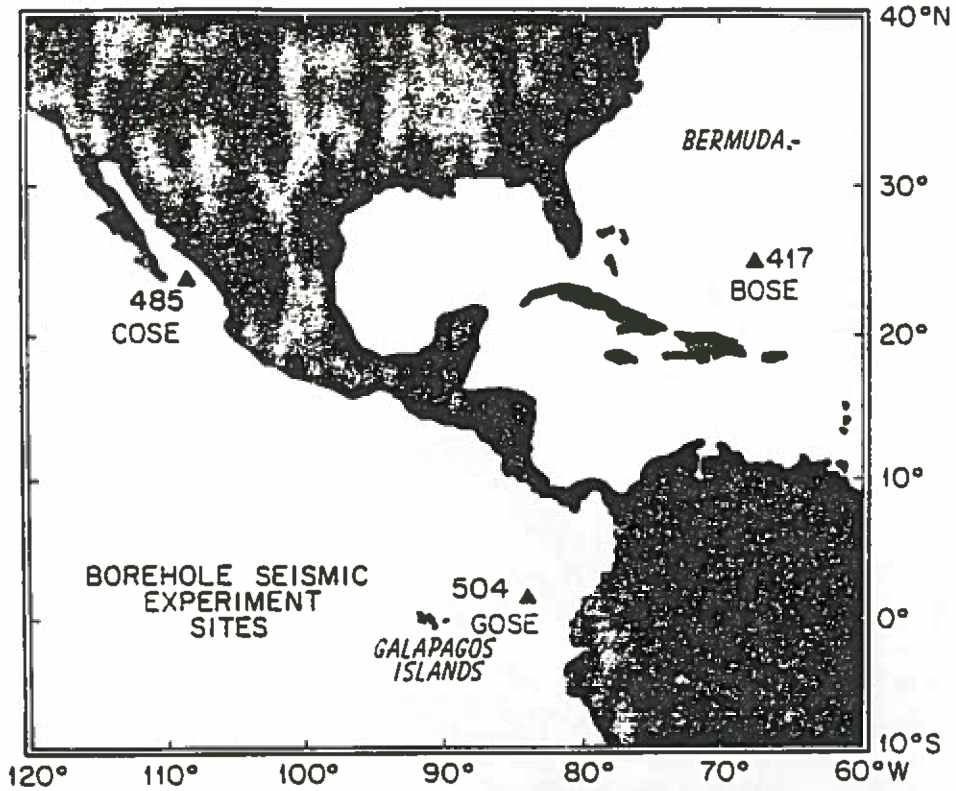

Figure 6-30 The locations of the three borehole seismic experiments discussed in the text.

an estimate of the uppermost velocity in a profile. This can be measured directly at *in situ* conditions and seismic frequencies from borehole receiver data and the velocity-depth profile in the uppermost crust is then considerably better resolved. Our traveltime data is inverted using the zeta tau inversion method (Dorman and Jacobson, 1981), modified for borehole receivers (see Stephen and Harding, 1983).

In this paper the traveltime results for the three borehole seismic experiments will be reviewed. The results of amplitude analysis by trial and error fitting of synthetic seismograms generated by the reflectivity method for borehole receivers (Stephen, 1977) and the evidence for seismic anisotropy based on particle motion analysis of the Western Atlantic data set will be presented. The geological significance of the borehole seismic experiment results will also be discussed.

Structure of Upper Oceanic Crust

Since the model of vertical velocity gradients replaced that of thick homogeneous layers for oceanic crustal structure, the trend in marine

seismology has been towards more detailed seismic analysis of fine-scale structure. This trend has been motivated by a desire to understand oceanic crustal processes such as crustal formation, hydrothermal circulation, and geochemical alteration, all of which require a detailed knowledge of structure, and by a desire to correlate field observations of ophiolites, deep-sea core and dredge sample analyses and well logging results with the gross *in situ* structure of oceanic crust. A number of intermediate issues have arisen. These include questions of scale, structural refinement, lateral variability, anisotropy and porosity. It is important to keep these issues in mind when addressing both the fundamental geological and geophysical processes and the seismic data interpretation.

The scales involved in the study of oceanic crustal structure range from detailed (even microscopic) analysis of hand sample specimens (less than 1.0 millimeter) to seismic refraction experiments (tens of kilometers). Scale must be considered when comparing results of different techniques and even in the same experiment, structures of different scales may affect different observations. For example, in the interpretation of a seismic refraction line, traveltimes reflect the structure along the whole ray path (about tens of kilometers) but the amplitude of a particular arrival is sensitive to finer scale topography and lateral variability in the region of the source and receiver (about hundreds of meters).

Purdy (1983) has pointed out the importance of structural refinement in interpreting oceanic crustal structure. Areas that would have been collectively referred to as 'normal' oceanic crust twenty years ago, are presently divided into ridges, trench, fracture zone and 'normal' oceanic crust. With the postulated 'zero offset' fracture zones (Schouten and White, 1980) 'anomalous' structure may occur every ten kilometers and the meaning of 'normal' oceanic crust in a seismic refraction context becomes nebulous.

Lateral variability is an issue on all length scales. It is frequently ignored because it is easier to conceive of purely depth dependent variations and in some techniques, such as seismology, the theoretical problem is considerably simplified by assuming lateral homogeneity. If lateral heterogeneity occurs over distances which are large compared to a seismic wavelength (~500m), the effect is to merely bend ray paths. However, if the heterogeneity occurs over distances on the order of a wavelength, anomalous amplitudes and attenuation due to scattering will occur. If the variation occurs over distances much less than a wavelength the medium can be represented by composite elastic properties. The most obvious violation of lateral homogeneity in the deep sea environment occurs at the top of igneous basement (the basalt-sediment or water interface) which is frequently assumed flat although it is notoriously rough on all length scales from millimeters to kilometers.

Anisotropy is another issue which is frequently avoided because of the difficulty and complexity of considering it. In a seismic context, general anisotropy requires twenty-one elastic constants to express the elastic behavior at every point in a medium (which is not homogenous) instead of only two elastic constants which are required to interpret an isotropic structure. It is not surprising that seismologists prefer to disregard anisotropy if possible, but structural geologists, sedimentologists, and petrologists regard anisotropy as a fundamental property of oceanic crustal rocks (cf. Karson, 1982).

Porosity of oceanic crust is the most important unknown parameter in oceanic crustal studies. (In this discussion porosity refers to all void space in a rock and includes fissures, cracks and inclusions). It affects all crustal measurements and processes. In correlating petrology to crustal structure by velocity analysis some assumption about porosity is essential. It is clearly important in upper oceanic crust (<500m) where porosity typically reaches 20 percent, but the depth at which it is negligible is The problem is further complicated by the unknown aspect ratio distribution of the voids which have a strong effect on both the permeability and seismic velocity.

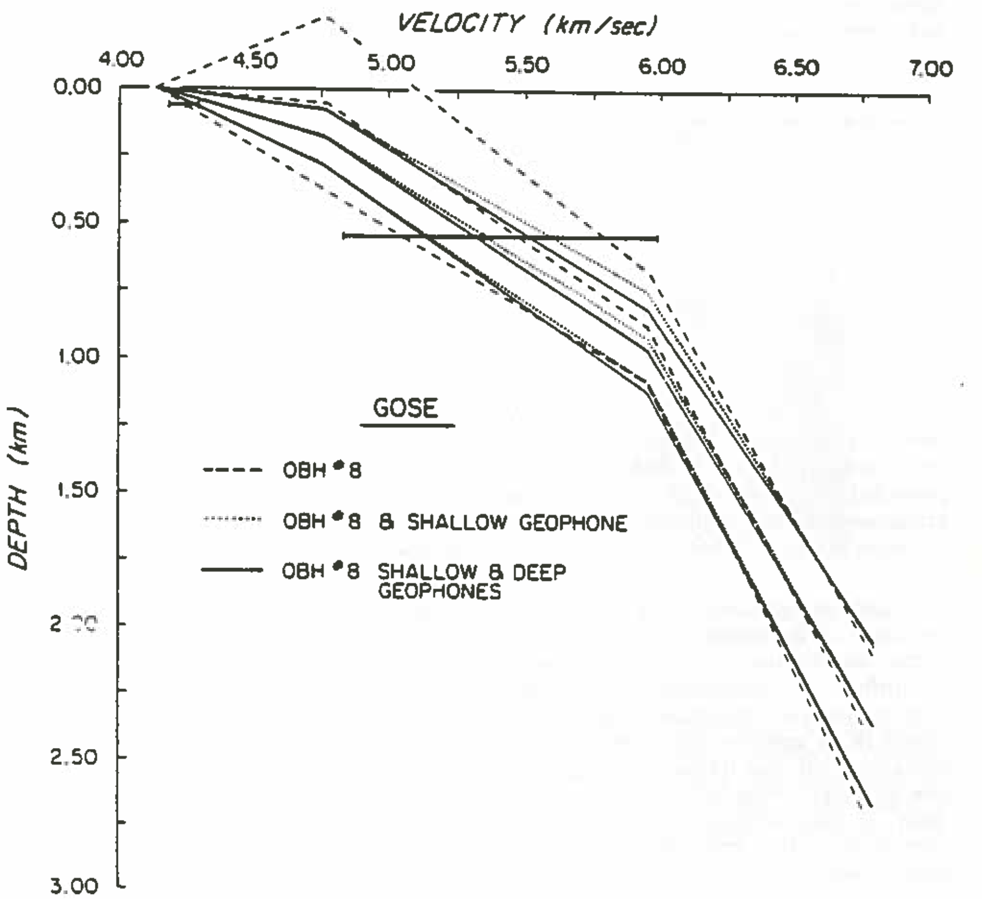

Figure 6-31: Velocity-depth profiles for real data at Site 504 (Costa Rica Rift Area - GOSE). The borehole data provides greater resolution at depths less than 1.0km. The independent velocity determination for the shallow geophone yields a value for the uppermost velocity (a necessary input to all traveltime inversion schemes) which is measured *in situ* at the same frequencies as the travel time data. For each data set the best velocity-depth function with 90% confidence limits is plotted. The heavy horizontal lines show the mean and one standard deviation bounds for the results of the windowing procedure for each geophone position (Figures from Stephen and Harding, 1983, *J. Geophys. Res.*).

These five issues are significant because each includes many unanswered questions, some of which can only be solved by innovative data collection, analysis, and interpretation techniques. Borehole seismic experiments provide some insights into these problems.

TRAVELTIME INVERSION AND INTERPRETATION

Comparison of field data sets

The Costa Rica Rift data set demonstrates clearly the advantage of borehole data in resolving shallow crustal velocity structure (Figure 6-31). The results from OBH data alone give higher shallow velocities and broader bounds

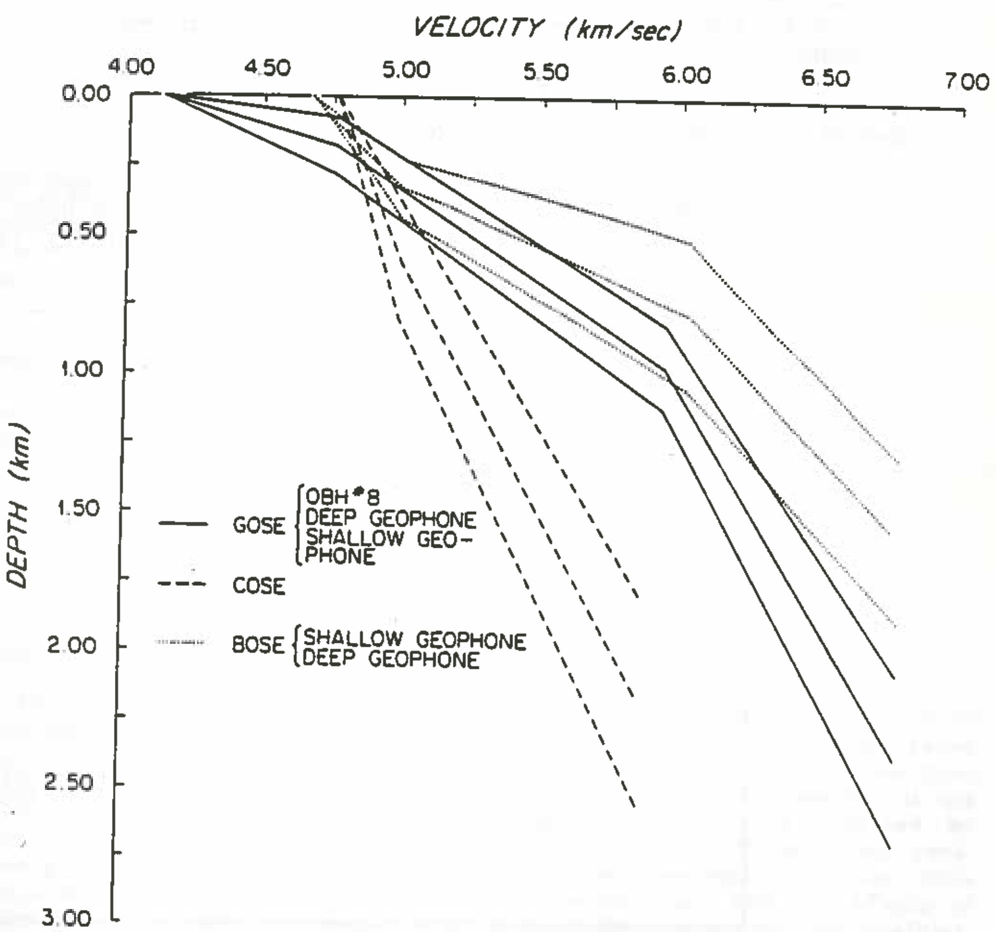

Figure 6-32 Velocity-depth profiles for the three data sets discussed in this paper. Despite the wide variation in crustal age (<1 my to 110 my) the velocity of the uppermost crust is remarkably similar. This contrasts sharply with the results of Houtz and Ewing (1976) based on sonobuoy data (Figure from Stephen and Harding, 1983, *J. Geophys. Res.*).

in the upper 500m of the .section than the results which include borehole receiver data. (In this figure the center line of each type is the 'best' determined velocity-depth profile and the lines to either side are the 90% confidence limits.)

The velocity-depth functions for all three sites are compared in Figure 6-32. The velocities of the upper 500m of crust are remarkably similar given the wide variation in ages: Costa Rica Rift (GOSE)-6my, Gulf of California (COSE) -1 my, Western Atlantic (BOSE)-110 my. The inversion procedure and the traveltime data themselves are given in Stephen and Harding (1983). This contrasts sharply with the results of Houtz and Ewing (1976) and Ewing and Purdy (1982) in which young crust has been observed to have upper velocities in the range of 2.5 to 4.0 km/sec and older crust to have upper velocities in the range 4.0 to 5.5 km/sec. A possible explanation for the discrepancy is that the borehole data have been obtained in areas with 100m or more of sediment. (See the section on 'Geological Implications' below.)

Delay Time Analysis of Gulf of California Data

As will be discussed later, it is not unreasonable to suspect that the velocity structure under ridges may differ from the structure under valleys because of systematic differences in the density of open cracks. This was

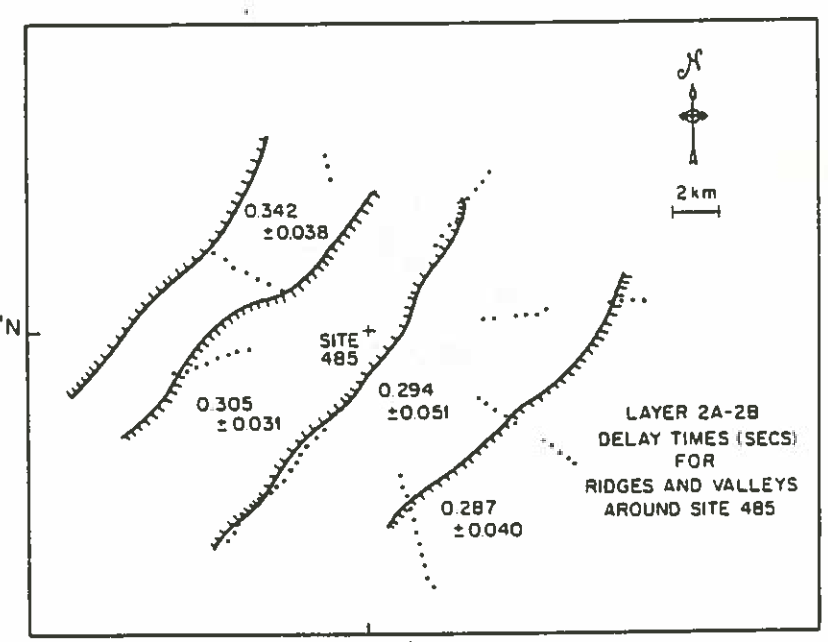

Figure 6-33 Summary of delay time analysis in the Gulf of California experiment. The objective was to check if the velocities of the hills differed from the velocities of the valleys. To the contrary, the velocities of upper oceanic crust appear to increase with increasing age (to the right) rather rapidly (Figure from Stephen *et al.* 1983).

studied using the Gulf of California data. Ridges and valleys in this context are defined by the height of the basement surface above a datum level (established at the basement depth at DSDP Site 485 (3134m Below Sea Level). This figure varies from 0 to 472m (see Figure 6-33). The correlation of velocity structure with topography was checked by two delay time analyses, one using shots over topography of less than 150m above the datum (valleys) and, the other, using shots over topography greater than 150m above the datum (ridges). Offsets in range were made to allow for the effect of non-vertical ray paths in the water column. Valleys are generally sediment filled and ridges are generally sediment free. For layer 2B, the velocity under the valleys is 4.64 ± 0.16 km/sec and under the ridges is 5.12 ± 0.16 km/sec. These were not considered to be the same refractor (a requirement for delay time analysis) and the delay time analysis for layer 2B arrivals was not carried further. For layer 2C, the velocity under the valleys is 5.62 ± 0.10 km/sec and under the ridges is 5.61 ± 0.18 km/sec. Since these do not differ significantly from each other or the average layer 2C velocity of 5.66 ±0.09 sec (Stephen *et al.* 1983) it is assumed that the layer 2C refractor is the same throughout the area.

The average layer 2A-2B delay times for two ridges and two valleys in the area is shown in Figure 6-33. The estimated standard measurement error in the delay times is (+-) 0.033 secs including the uncertainty in basement topography. The scatter in the delay times therefore falls mostly within the measurement error. However, the mean delay times differ as much between the two ridges as between the ridges and valleys so that we cannot infer different structure under ridges and valleys.

The layer 2A-2B delay times do decrease systematically with age. If one assumes a constant depth down to the layer 2C refractor the combined layer 2A-2B velocity increases from 4.1 km/sec under the western hill to 4.7 km/sec under the eastern valley. If one assumes a constant velocity for layer 2A-2B the depth to the layer 2C refractor decreases from 1.40 km under the western hill to 0.99 km under the eastern valley.

The Western Atlantic and Gulf of California data sets were not analysed in this fashion because the basement topography was not sufficiently well resolved.

AMPLITUDE ANALYSIS

The Western Atlantic data set is (at the time of writing) the only three component data set for a clamped receiver in oceanic crust. (Since this paper was prepared, a second three component data set for a clamped borehole receiver was obtained in the Western Pacific and a third was obtained at the Costa Rica Rift site.) It has many advantages over data collected by conventional techniques. The geophone is well coupled with the basement rock. Energy is only transmitted through the sediment-basement interface once because the receiver is located beneath the interface and the effects of basement topography are smaller than for sea bottom or sea surface receivers. The unique geometry allows direct waves to be received which have travelled horizontally in upper basement over ranges up to five kilometers. For all the shots fired at the deep geophone position (the four lines and the shots on the semi-circular arc), the geophone orientation was constant. The complete data set has been published by Stephen et al., (1979).

Figure 6-34 demonstrates a unique feature of borehole seismic experiments in oceanic crust. Because of the high velocity contrast between water or sediment and the top of igneous basement compressional wave arrivals travel almost horizontally to a receiver in the upper crust. Thus the dominant compressional wave particle motion (the largest P wave amplitude) occurs on the horizontal components rather than the vertical component as one would expect for a sea bottom seismometer. Similarly, the strongest shear wave motion (SV) occurs on the vertical component.

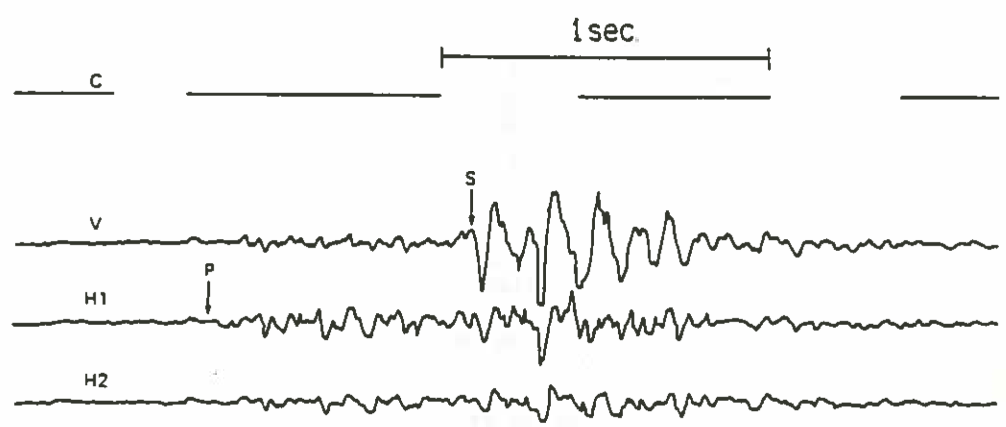

Figure 6-34 OSE seismograms from the Western Atlantic for the geophone at 228 m into basement for a shot on the north line at 6.58 km range. The three components are vertical (V), horizontal X (H1) and horizontal Y (H2). P and S refer to the compressional and shear first arrivals respectively. Note that the P wave arrivals are strongest on the horizontal components. This behaviour is expected for a P wave which has refracted at the strong velocity contrast of the sediment- basement interface (Figure modified from Stephen *et al.*, 1980).

Traditional synthetic seismogram analysis assumes lateral homogeneity and, before applying amplitude analysis to a data set, one must have some confidence that the actual structure satisfies this assumption. The strong effect of basement topography on seismic amplitudes is evident in both the Atlantic and Gulf of California data sets. Lines were chosen for interpretation which appeared to be least affected by topography. The main criteria here were gradual changes in amplitude with range (20db changes within 1 km were not considered gradual) and consistency of waveform.

With the limits of the lateral homogeneity assumption in mind, the goal of amplitude modelling is the determination of the variation of velocity with depth. The classic contribution of amplitude analysis has been that layers of constant velocity gradients rather than layers of constant velocity dominate oceanic crustal velocity structure. This picture is confirmed by the Atlantic data. Amplitude analysis of the vertical and horizontal components of the Western Atlantic data was carried out using the modified reflectivity method (Stephen, 1977). The Western Atlantic analysis is the first interpretation of horizontal component data in marine seismology (Stephen et al., 1980). Figures 6-35 and

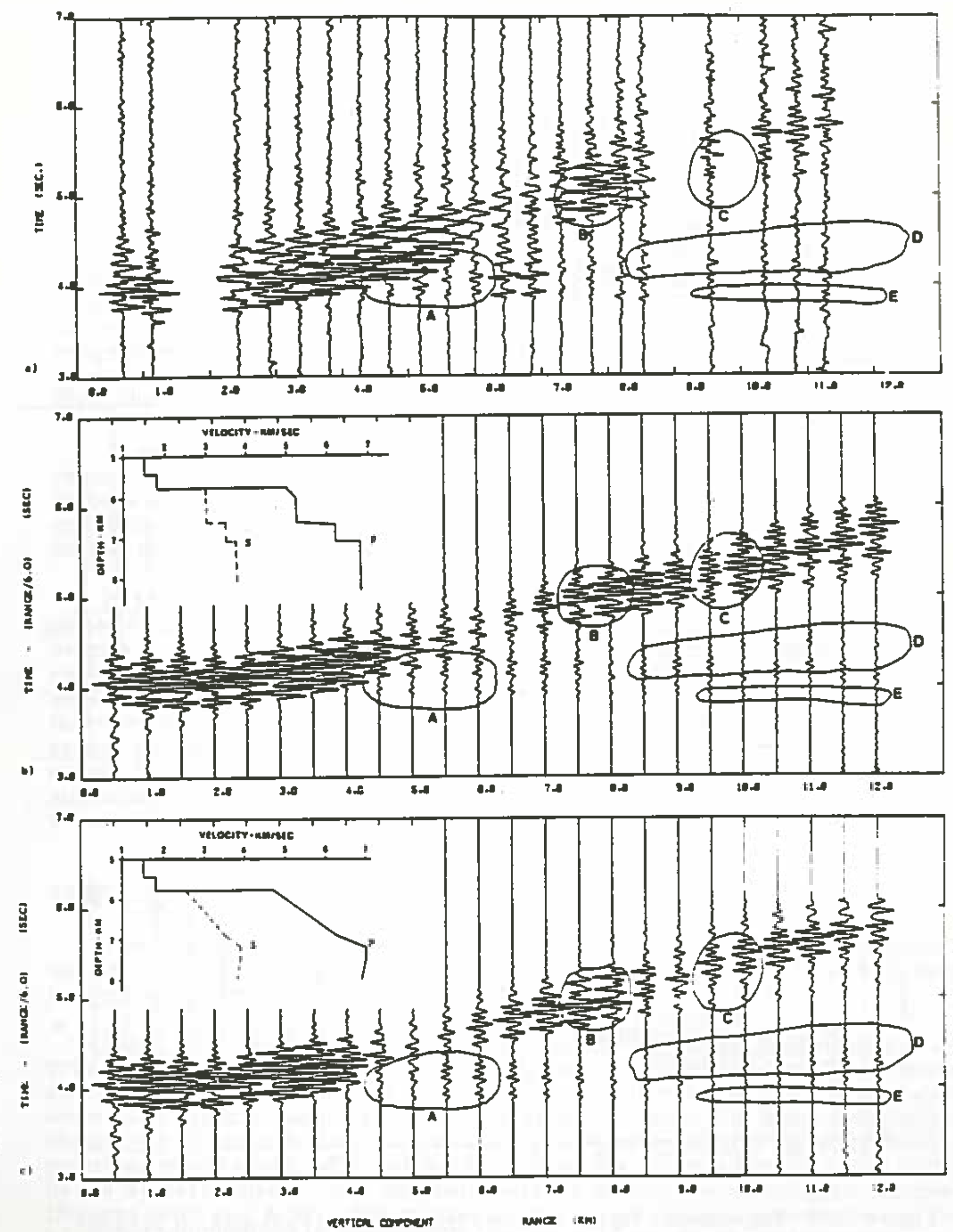

Figure 6-35 Synthetic seismogram interpretation from the Western Atlantic data of a vertical component record section for the south line fired to the deep geophone position (see Table 6-3). Figure 6-36(a) is the observed data and Figures (b) and (c) are two synthetic seismograms for the models shown. The regions A,B,C, D and E are described in the text. All the seismograms shown are displayed with an amplitude weighting of $(range/7.0)^{2.9}$ for ranges greater than 7.0km (Figure from Stephen *et al.*, 1980).

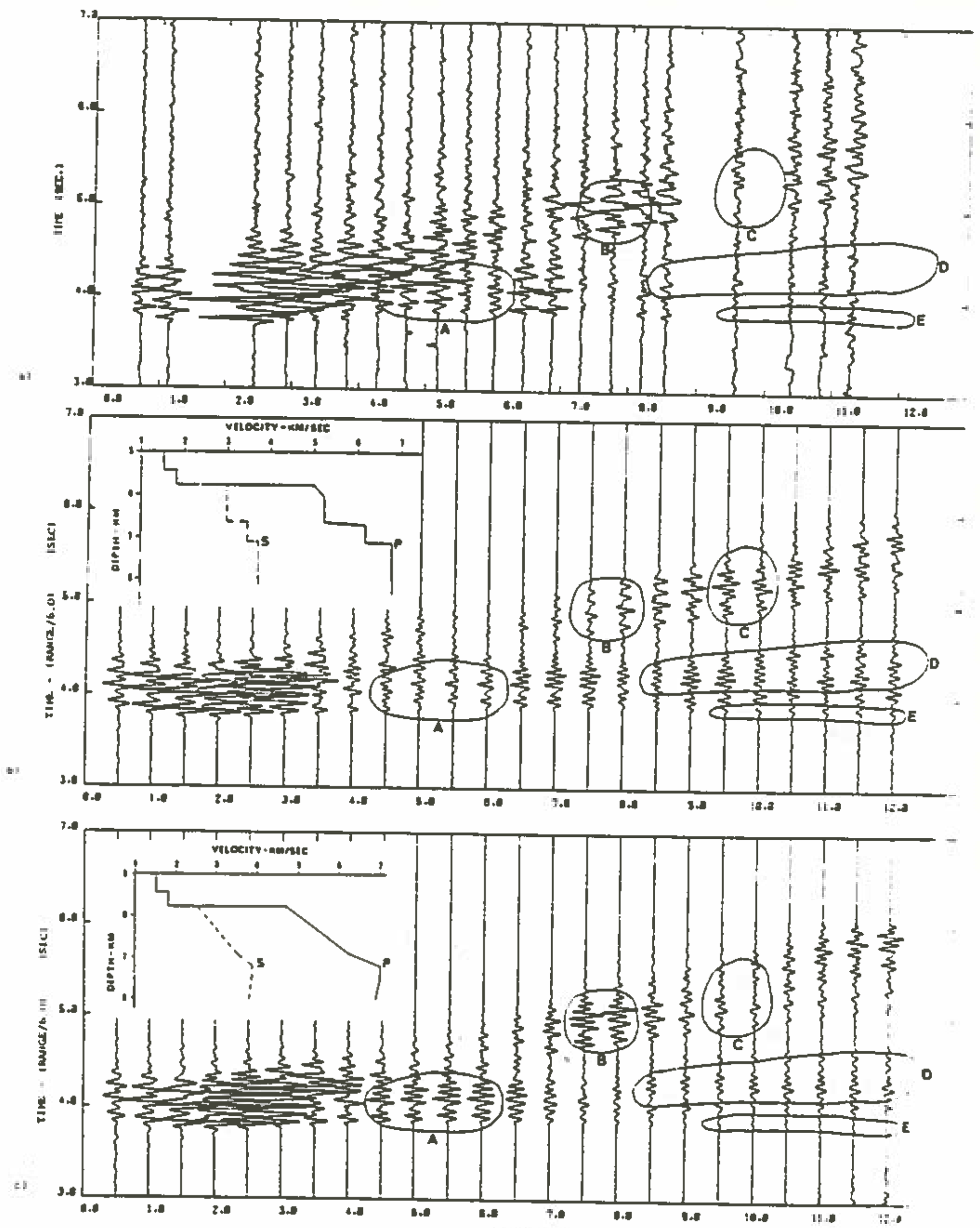

Figure 6-36 The same as Figure 6-35 for the horizontal component record section (Figure from Stephen *et al.*,1980).

6-36 show an interpretation of the vertical and horizontal component data respectively for one line. The results of a thick homogeneous layer model and a continuous gradient model are compared. The amplitude comparison of the vertical component data is the less diagnostic of the two models. Apparently in this application, the horizontal data are more sensitive to structure.

Rather than matching the exact amplitudes of an isolated phase, regions on the seismic section which seem to display characteristic behavior were considered. Regions A and B are the critical regions for P and S wave arrivals from layer 3. Amplitudes in these regions are large when compared to the amplitudes on the section in general. Regions C and D are areas of destructive interference in the S and P wave energy respectively. The observed section shows relatively low amplitudes in these regions. Region E includes the long range layer 3 refraction arrivals and is also an area of low amplitudes. The gradient model provides a more satisfactory fit, particularly to the horizontal component data, than the homogeneous layer model.

It was not possible to model the high amplitude arrivals to the right of region A using the reflectivity method and this is a good example of a seismic feature which may be attributed to lateral inhomogeneity. Finite difference synthetic seismogram methods which are suitable for studying lateral inhomogeneity in marine situations are being developed (Stephen, 1983b).

It should be noted that proper application of the reflectivity method to upper crustal gradient structure where significant compressional to shear wave energy occurs, requires the thicknesses of the homogeneous layers representing the gradient to be on the order of or less than a fifth of the shear wavelength, or approximately 7m (White and Stephen, 1980; Stephen, 1983). Interpretations which use coarser thicknesses are not implying gradients, but rather are implying the existence of thin homogenous layers in upper oceanic crust. This is significant when studying upper crustal shear wave velocity structure from transmitted P and S waves and reflected P waves (Spudich and Orcutt, 1980).

PARTICLE MOTION ANALYSIS

The evidence for seismic anisotropy in upper oceanic crust from particle motion analysis of three component borehole seismometer data has been presented by Stephen (1981). A short summary will be given here.

There are two particle motion anomalies associated with elastic wave propagation in an anisotropic medium. First, the primary (quasi-compressional) wave particle motion is not collinear with the propagation direction. Second, except along certain symmetry axes, the secondary (quasi-shear) wave consists of two distinctly polarized phases: quasi-vertically polarized shear waves, qSV, and quasi-horizontally polarized shear waves (qSH). In the Western Atlantic data set, both particle motion anomalies were observed (Figures 6-37 and 6-38). The scatter in the traveltime measurements was too large for the distances involved to resolve seismic anisotropy from velocity measurements.

The observed phenomena can be modeled quantitatively, assuming a flat, homogeneous, hexagonally anisotropic basement with all arrivals traveling

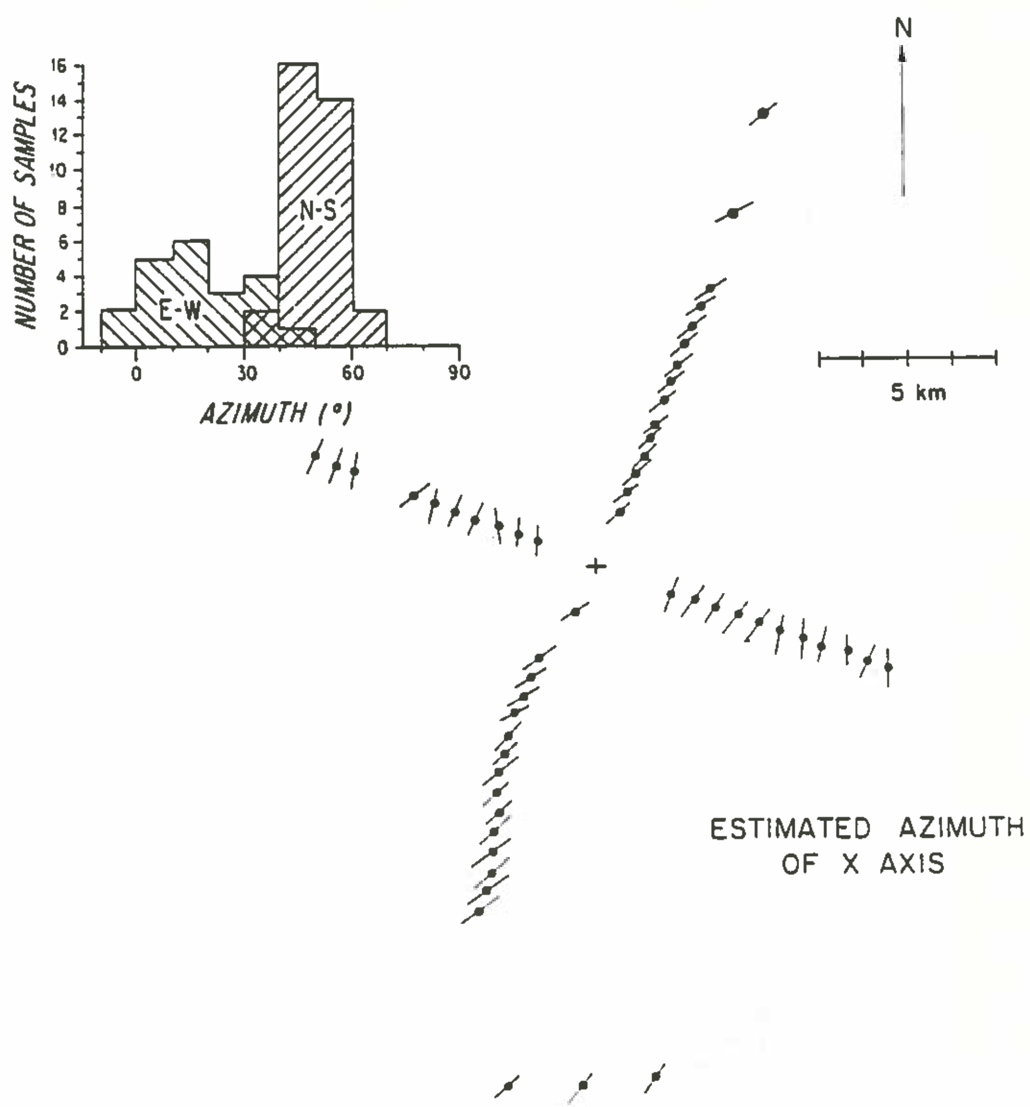

Figure 6-37 Estimated azimuth of the x-axis of the borehole geophone assuming the first compressional particle motion is in the direction of the shot. At each shot location ( ) the estimated azimuth of the x-axis at the geophone (+) is indicated. The direction of particle motion has been computed by polarization analysis (Kanasewich, 1973). All results are based on the first 0.1 sec of data which was bandpass filtered between 5 and 25Hz. Shots less than 1km in range have been omitted because they have near vertical ray paths which give inconsistent azimuths. Shots with power levels less than one standard deviation above the mean noise level were also disregarded. The distribution of the azimuths is summarized in the histogram (inset). In an isotropic, homogenous medium there should be no difference between azimuths estimated from north-south and east-west shots (Figure from Stephen, 1981).

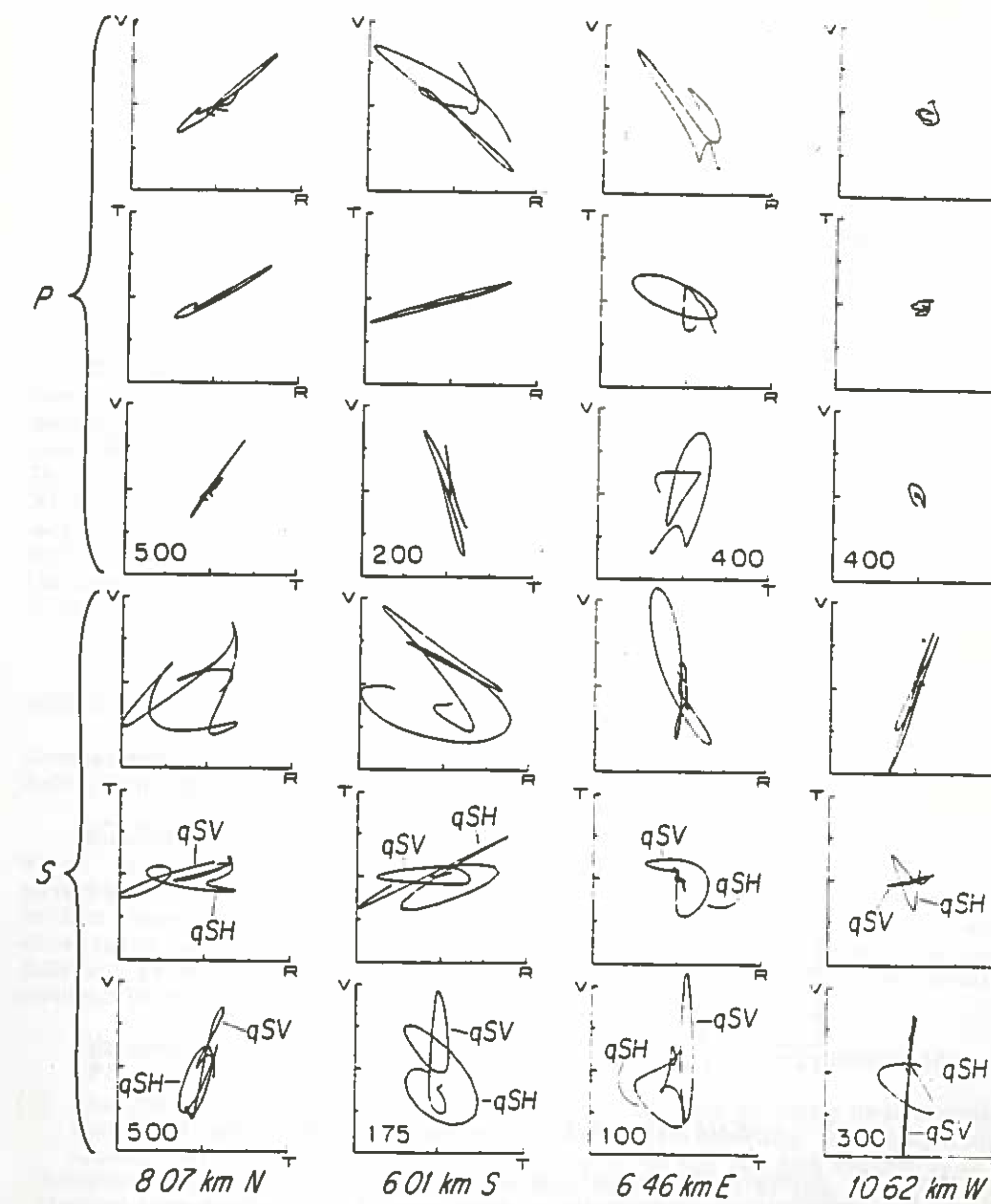

Figure 6-38 Particle motion diagrams for first compressional (P) and first shear (S) wave arrivals for four shots. The vertical (V), radial (R) and transverse (T) components are shown. A dot indicates the beginning of the arrival and 0.2 seconds have been plotted. The observation of quasi-vertically polarized shear waves (qSV) arriving prior to the quasi-horizontally polarized shear waves (qSH) indicates anisotropic elastic behaviour. The azimuth of the x-axis for all diagrams was assumed to be N34 ° E. The number in the lower left corner of the V-T plot is the amplification factor used in each case. The P wave arrival at 10.62km on the west line is below the noise level. The S wave arrivals for ranges less than 7.0km are from layer 2 and for ranges greater than 7.0km are from layer 3 (Figure from Stephen, 1981).

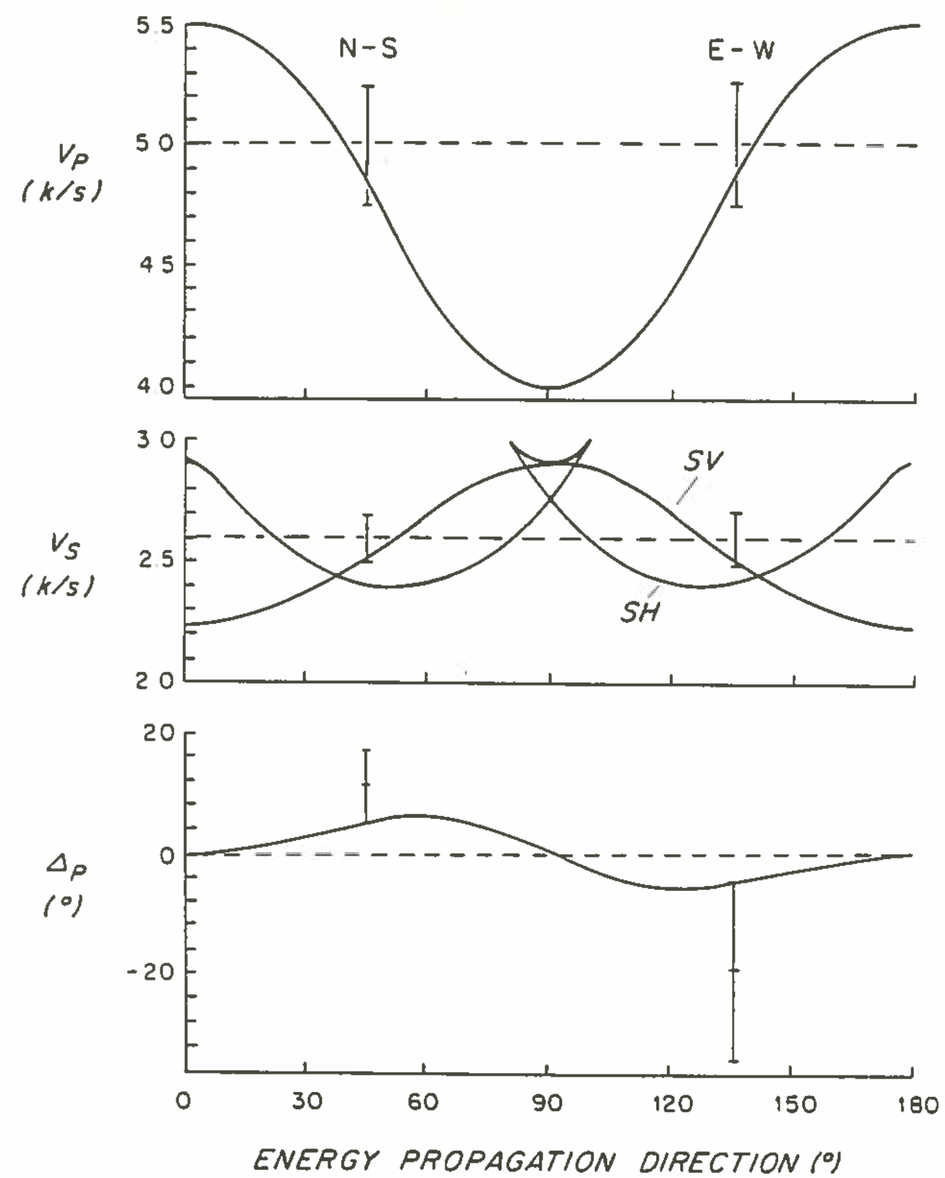

Figure 6-39 Hexagonally anisotropic model (——) which satisfactorily explains the observations. Compressional wave velocity (Vp), shear wave velocity (Vs) and compressional wave particle motion deviation (Δ p) are plotted against the energy propagation direction measured counter-clockwise from the 100 axis. The values with error bars are the observed values for north-south (N-S) and east-west (E-W) lines. An isotropic model (e.g. — — —) cannot explain the particle motion observations. The values of the tensor of elastic stiffness divided by density (Γijkl) used to model the anisotropy are
$\Gamma_{1111}=30.25$, $\Gamma_{2222}=16.00$, $\Gamma_{1212}=8.50$, $\Gamma_{1122}=10.00$, $\Gamma_{1313}=5.00 (km/sec)^2$
(Figure from Stephen, 1981).

horizontally just beneath the sediment-basement contact (Figure 6-39). (A hexagonally anisotropic medium is isotropic in planes perpendicular to a symmetry axis.) The symmetry axis of the medium is horizontal and its azimuth is one of the unknowns in the data interpretation procedure. The agreement of theory with observation is within the confidence limits of the data. The apparent compressional wave velocity anisotropy is large enough that one would expect to detect it from traveltime measurements if the shooting lines were in the proper azimuths. It should be noted, however, that the compressional and shear wave velocities are measurements over ranges of 4.0 to 12.0 km, and the P wave particle motion anomaly is caused by elastic properties over only about 300m. Lateral heterogeneity of the crust between these two scales could smear the results.

The shooting lines were originally chosen to coincide with the spreading direction and magnetic lineations in the area. It is disturbing that the maximum and minimum P wave velocities occur at 45 degrees to the shooting lines for the simple model shown. Arguments of scale can again be invoked. The spreading direction and magnetic lineations were obtained from data with 30 km spacing. Local spreading directions may not necessarily coincide with larger scale directions (eg. the case of enechelon spreading). It may also be significant that the bathymetry, which frequently mimics the spreading history (Andrews and Humphrey, 1980), trends at about 45 degrees to the shooting lines in this area (Figure 6-A1, Appendix A).

DISCUSSION

Comparison between Borehole Receiver Experiments and Conventional Refraction Experiments.

Borehole receiver experiments, (or oblique seismic experiments, OSE) which use clamped borehole geophones as receivers, have a number of advantages over conventional seismic refraction techniques, which use ocean bottom seismometers, ocean bottom hydrophones, or sonobuoys. These advantages can be attributed to theoretical factors, which are a result of the different geometries involved in the two experiments and operational factors involved in practical applications.

The advantages of the OSE which are a result of its unique geometry are:

(1) The OSE can give velocities of material near the hole by using near normal incidence shots. This is commonly called check shooting. By comparing traveltimes from surface shots to geophones at the top and bottom of the hole, the mean velocity of crust within about 500m of the hole can be obtained. The velocity measured by conventional refraction experiments is an average over the range of arrivals and the method requires ranges of 2 km or more to identify the refractor (Raitt, 1963 and Houtz, 1976). It is inherently impossible for conventional techniques to obtain the same degree of lateral resolution as the OSE or to obtain mean vertical velocities in layer 2.

(2) The OSE does not rely on the presence of reflecting or refracting horizons

in order to obtain a velocity determination. The mean velocity of the material above the receiver can be obtained from direct wave arrivals.

(3) Conventional refraction experiments cannot detect low velocity zones. The OSE can, of course, detect the presence of a low velocity zone above the receiver from the traveltimes of normal incidence direct wave arrivals to different depths in the hole. If the low velocity zone is caused by large-scale factors (e.g. large vertical fissures), even logging may not detect the zone.

(4) Much better layer 2 velocities can be obtained using the OSE with a receiver within layer 2 because waves which have traveled in layer 2 occur as first arrivals for offsets up to 10 km. This compares with distances of about 4.0 km for conventional refraction work. Also, in the presence of sediment thicker than about 100 meters, shallow layer 2 velocities are impossible to obtain from first arrival times using ocean bottom or surface receivers.

(5) Traveltimes in the OSE are less sensitive to the basement topography correction than traveltimes in conventional refraction experiments because the ray paths only pass through the basement sediment interface once. Because of the high contrast in velocity between sediment (~2.2 km/sec) and basement (~5.0 km/sec) and because this interface may not be well resolved by reflection profiling, the error introduced into traveltime analysis by this correction is significant. For example, if the basement topography is not known to within 50 m, an uncertainty of 0.02 sec is introduced into the traveltimes for each time the ray path crosses the sediment- basement boundary.

(6) Proper attenuation measurements in oceanic crust can be made with the OSE by comparing amplitudes of arrivals at deep and shallow positions in the hole. A correction for the effect of structure between the receivers can be made when the sonic log for the hole is available. It is impossible to obtain satisfactory attenuation measurements from conventional refraction data because the effect of scattering from small-scale structure cannot be adequately determined.

The operational advantages of the OSE over conventional refraction experiments are:

(1) The receiver is clamped to the rock wall of the borehole and hence has better coupling than an ocean bottom receiver which has fallen more or less randomly onto the sea bed and may, for example, be sitting on a rubble zone or poorly consolidated sediment. In an OSE, if a sonic log or core description is available, solid sections in the hole can be identified in advance. In addition, one would expect the background noise for a clamped borehole geophone to be less than for a receiver sitting on the sea bed which is exposed to bottom currents (Duennebier *et al.*, 1981). One would also expect that the amplitudes of both vertical and horizontal components would be reliable and directly correlatable.

(2) In the OSE case, the receiver is well located with respect to the borehole where in situ conditions have been measured and depths are known most accurately. For free falling bottom receivers the accuracy of emplacement is at best 200 m (Francis et al., 1977; Macdonald and Luyendyk, 1977; Creager and Dorman, 1982). Since changes on this scale in basement topography and in internal structure of layer 2 exist (Aumento et al., 1977), detailed studies, where one is interested in crust within a few hundred meters of the borehole, are best carried out with an OSE.

The major disadvantage of the OSE is that it is a difficult experiment to carry out. The drilling ship time is expensive compared to normal research vessel time and at its present stage of development the experiment requires two ships to rendezvous. The logistics of this are not simple. The OSE is dependent on a borehole and hence is not as flexible as conventional refraction work. At this stage, horizontal arrays of borehole seismometers are out of the question and even reversed profiles are unreasonable.

Geological Implications of Borehole Seismic Experiments

The geological impact of borehole seismic experiments results from a) the improved resolution of upper crustal velocity structure in the presence of sediments and b) the additional information provided by three component data from well-coupled receivers.

In the three borehole seismic experiments run to date, the upper crustal velocities have been consistently high (~4.5-5.0 km/sec). These high velocities persist even in very young crust where other investigators (e.g. Houtz and Ewing, 1976; Ewing and Purdy, 1982) have reported low velocities. There is a significant sampling bias in the borehole receiver results, however, because the drilling operation requires at least one hundred meters of sediment. All the borehole receiver results, regardless of age, have been obtained in regions of significant sediment cover (Table 6-3). The other observations made with radio sonobuoys and ocean bottom hydrophones were obtained in young crust with negligible sediment cover. These observations imply that sediment cover, not merely age, is a controlling factor for upper crustal velocity. Upper crustal velocity is high (~4.0-5.0 km/sec) in regions with significant sediment cover and low (~2.5-4.0 km/sec) in regions with insignificant sediment cover. A nominal value for "significant" sediment cover in the absence of better resolution data, can be assigned as 100m.

The data are consistent with the following geological hypothesis. The upper crust forms with relatively high porosity which is uniform with time. The high porosity provides the conduits for virtually unimpeded hydrothermal convection, which rapidly cools the crust. Subsequent sedimentation places an impervious 'blanket' on the crust. The hydrothermal circulation is reduced and temperatures in the crust increase, accelerating the formation of metamorphic minerals which seal the cracks, reduce the porosity and increase the velocity. The cementation of the cracks occurs on a time scale which is short (<1 my) relative to the time scale of spreading (Honnorez, 1981). (At a spreading half-rate of 1.25 cm/year, only 12.5 km of crust is generated in a million years). In this hypothesis the apparent increase of layer 2A velocity with age (Houtz and Ewing, 1976) is due to the average increase in sediment thickness with age. The apparent increase in upper crustal velocity in the Gulf of California may be a direct measurement of the rapid cementation of upper crust in a region of rapid sedimentation.

The hypothesis could be corroborated by a borehole seismic experiment in unsedimented crust. This would check for discrepancies which may be introduced by the different experiments (borehole vs. conventional receivers) and data reduction and interpretation techniques. The present drilling and re-entry techniques would have to be modified for operation in regions without sediment cover.

Particle motion analysis of three component data obtained with well-coupled receivers provides further information for the analysis of anisotropy in upper oceanic crust. Because the scatter in velocity measurements for upper crust, due in part to the medium scale heterogeneity (on the order of one hundred meters to a kilometer), seismic anisotropy is difficult to resolve from traveltime measurements. However, the observation of horizontally-polarized shear waves in oceanic crust from a compressional source strongly implies seismic anisotropy and simple anisotropic models can be constructed which predict all the observations. An alternative explanation for the presence of horizontally polarized shear waves is scattering from sharp lateral discontinuities. Since the basement interface is notoriously rough this cannot be ruled out, but at the present time, models which demonstrate the mode-conversion of SH waves at a rough interface have not been constructed and the computational effort required to do so using present theories is prohibitive. (This would require a three-dimensional, deterministic, scattering algorithm using either finite difference or finite element techniques.)

This alternative explanation could be ruled out by the observation of systematic changes in arrival time with range for horizontally and vertically-polarized shear waves. This method is plagued with the traditional problems of traveltime analysis mentioned above. In the Western Atlantic data set consistent differences in arrival times could not be distinguished from noise. A more intensive experiment, designed to look for this effect, is required.

The most obvious geological explanation for seismic anisotropy in upper oceanic crust is the presence of well-oriented, vertical fractures and fissures in the crust. In the pillow basalt sequence, the crust up to the meter scale is isotropic but the larger scale fractures and fissures give an anisotropic overprint to the gross structure which can be interpreted by anisotropic elastic constants since the scale of fracturing and fissuring is still small compared to seismic wavelengths (~500 m). Anisotropy should increase in the sheeted dyke complex where even the texture on the meter scale is anticipated to be anisotropic by comparison with ophiolite sequences. In fact, the measurement of anisotropy in the dyke sequence would provide important corroborative evidence for its existence. Deeper in the section the presence of listric faulting may be detectable by the change in axis of symmetry from horizontal to vertical. Particle motion analysis of borehole seismic data in a deep crustal hole would yield direct *in situ* evidence for geological features with length scales on the order of hundreds of meters which cannot be studied using the small sample size obtained from drill cores or seismic refraction experiments. Borehole seismic experiments provide an important link between the small-scale core and well log analyses and the large-scale seismic refraction observations which presently define oceanic crustal structure.

CONCLUSIONS

The large-scale simplicity of oceanic crustal structure models, mostly defined by seismic refraction experiments, led early investigators to sweeping, general models of plate tectonics, sea floor spreading, and internal composition and petrology. On the scale of tens to hundreds of kilometers, the approximations of lateral homogeneity and isotropy seemed perfectly valid in the sense that it was not necessary to invoke lateral heterogeneity or anisotropy to explain the seismic observations. Over the past decade there has been a trend, largely a result of the deep sea drilling capability and detailed studies of ophiolites, towards smaller scale observations and processes (e.g. hydrothermal circulation, geochemical processes, shallow crustal structure). Although there is some evidence that seismic structure remains simple down to the kilometer scale (e.g. Purdy, 1982) there is overwhelming evidence that, on the scale of hundreds of meters and less, the structure becomes significantly more complicated. In order to study this structure seismically , more innovative field techniques and data reduction and interpretation procedures must be developed. Direct observation by camera systems, diving submersibles and drilling reveals the basement interface as rough on length scales from millimeters to kilometers with significant lateral changes in structure and preferred orientations. At normal seismic frequencies the processes of mode-conversion and scattering from the lateral heterogeneities and the effect of anisotropy become significant.

Borehole seismic experiments have improved our knowledge of upper oceanic crust by providing accurate *in situ* velocities at seismic frequencies, measures of lateral heterogeneity on the scale of kilometers, and evidence for seismic anisotropy due to preferred fracture orientation.

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APPENDIX A: BATHYMETRY AROUND BOREHOLE SEISMIC EXPERIMENT SITES

The bathymetry within 15 km of each of the three sites discussed in this paper (DSDP Sites 417, 485 and 504) is shown in Figures 6-A1, 6-A2 and 6-A3.

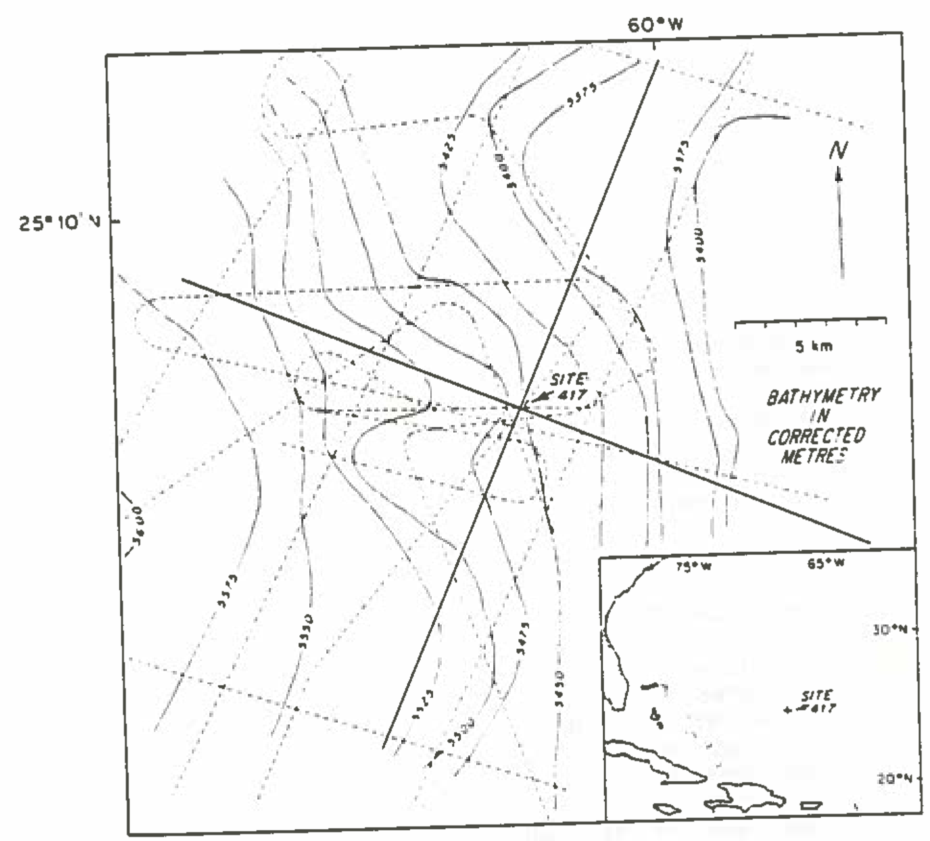

Figure 6-A1. Bathymetry at DSDP Site 417 (BOSE, Western Atlantic) (Figure from Stephen, 1981).

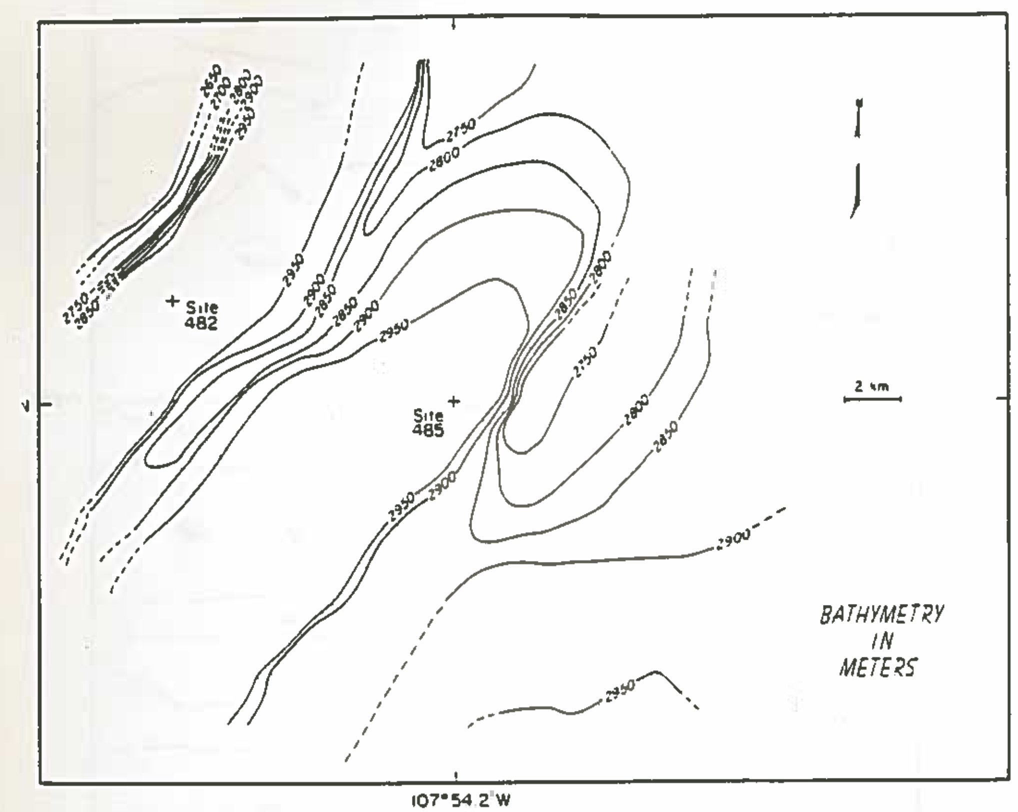

Figure 6-A2. Bathymetry at DSDP Site 485 (COSE, Gulf of California) (Figure from Stephen *et al.*, 1983).

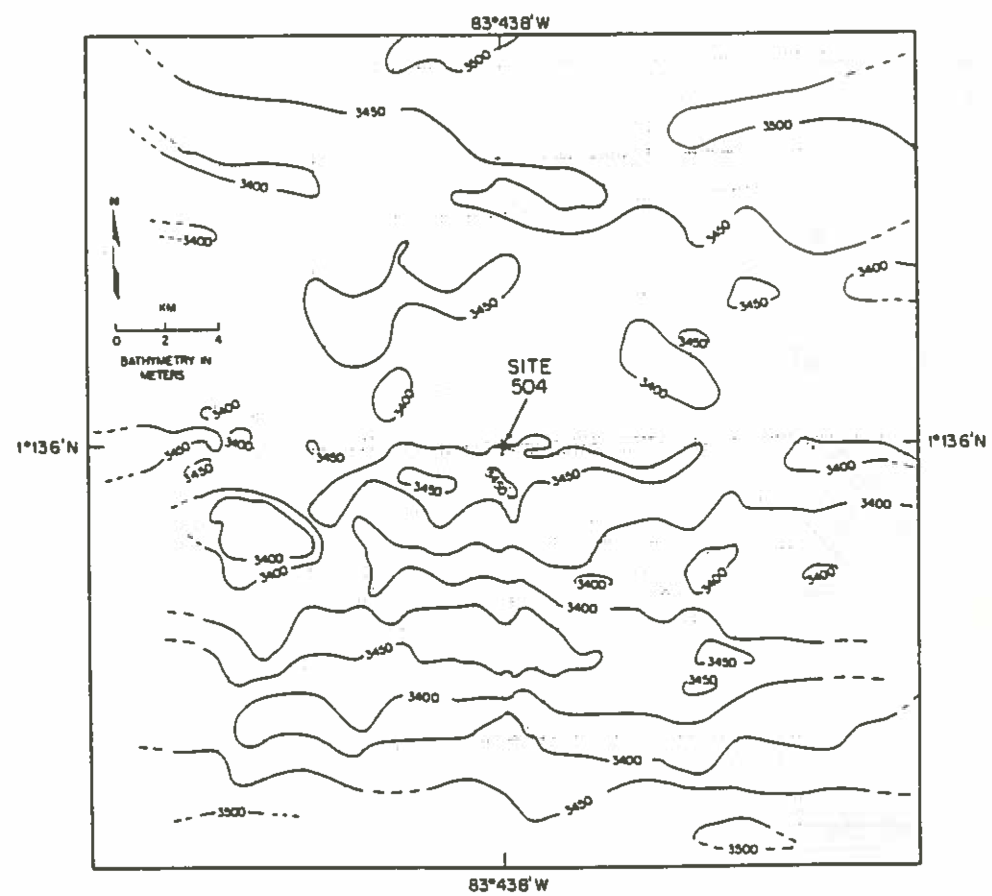

Figure 6-A3. Bathymetry at DSDP Site 504 (GOSE, Costa Rica Rift Area) (Figure from Stephen, 1983).