CRUSTAL STRUCTURE FROM
AN OCEAN BOTTOM SEISMOMETER SURVEY OF
THE NOOTKA FAULT ZONE

by
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A THESIS SUBMITTED IN PARTIAL FULFILMENT OF
THE REQUIREMENTS FOR THE DEGREE OF
DOCTOR OF PHILOSOPHY

in
THE FACULTY OF GRADUATE STUDIES
(Department of Geophysics and Astronomy)

We accept this thesis as conforming to the required standard

The University of British Columbia
June, 1981
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ABSTRACT

The Nootka fault zone is the boundary between the small Explorer and Juan de Fuca plates which lie off western Canada between the America and Pacific plates. To investigate the crustal structure in the region, three refraction lines were shot with explosives and a large airgun into three ocean bottom seismometers (OBS's) each equipped with three-component geophone assemblies. The P and S wave profile data are analysed primarily by synthetic seismogram modelling using the WKBJ algorithm. The interpretation gives relatively consistent results for the upper crust. Below the 1 km thick sediment layer, the P wave velocity ranges from 3.7 to 4.7 km/s and increases with depth at a moderate velocity gradient of 0.5 km/s/km to a depth of 1.9 km. A zone of high velocity gradient marks the transition from layer 2A to 2B, below which the velocity again increases moderately with depth to a range of 6.0 to 6.4 km/s at the base of layer 2B. The S wave velocity also increases with depth in the upper crust but no detail is available for the $v_s$ structure in the shallow region of layer 2A. Within layer 2B, the P and S wave velocity values give a Poisson's ratio with an unusually low value of 0.24. This may be related to the presence of quartz-rich trondhjemites. Both $v_p$ and $v_s$ are relatively uniform at the top of layer 3A. Good agreement is found between the seismic velocities of layer 3A and the seismic velocities of rock samples from the corresponding depths of ophiolite complexes measured in the laboratory, resulting in consistent values for Poisson's ratio.

Sub-bottom crustal thickness varies from 6.4 to 11.2 km among the
various profiles. Some aspects of this variation can be explained by consideration of a recent tectonic model for the development of the fault zone. This requires, within the past one million years, variation in the process of crustal formation at the ridge, crustal 'maturing', or both. The abnormally thick crust may result in part from the complex interaction of the Juan de Fuca and Explorer plates with the larger and older America and Pacific plates.

Apparent velocities of compressional waves in the upper mantle, calculated for areally distributed ray paths, show significant anisotropy. A maximum velocity of 8.3 km/s is found in the inferred direction of plate motion and a minimum velocity of 7.5 km/s is found parallel to the spreading ridge. This type of velocity variation can be approximated by a mixture of 42% transversely isotropic olivine and an isotropic material with a constant velocity of 7.0 km/s. Velocity of shear waves in the upper mantle, varying from 4.5 to 4.6 km/s, is isotropic within the resolution of the interpretation. This causes a prominent anisotropy in the values of Poisson's ratio.

The variation in velocity with depth in the crust, values of Poisson's ratio, the P wave velocity anisotropy, and the approximately isotropic S wave velocity are consistent with laboratory measurements on rock samples from ophiolite complexes.
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ACKNOWLEDGEMENTS

I wish to express my sincere gratitude to Dr. R. M. Clowes, my thesis supervisor, for his helpful guidance and encouragement during the period of this study. His constant support and understanding are deeply appreciated.

I acknowledge the cooperation and assistance of the officers and crew of CFAV Endeavour during the field survey. Collaboration with Dr. R. D. Hyndman of the Pacific Geoscience Centre, EMR, Sidney, B. C., whose ocean bottom seismometers made the refraction program possible is greatly appreciated. L. Haynes and W. Nehring of the Fleet Diving Unit, Pacific Maritime Command, CFB Esquimalt, B. C. prepared and detonated the explosive charges. I thank C. H. Chapman for providing a copy of his WKBJ synthetic seismogram routine and G. A. McMechan for his ray tracing/synthetic seismogram algorithm.

Financial assistance for this project was provided by a Research Contract and Research Agreement Nos. 167-3-78, 110-3-79 and 172-3-80 from Energy, Mines and Resources (Earth Physics Branch) Canada and by operating grant A7707 from the Natural Sciences and Engineering Research Council Canada. Additional financial support was provided by Shell Canada Resources Limited and Mobil Oil Canada Limited. I acknowledge receipt of an H. R. MacMillan Family Fellowship from the University of British Columbia during the period of this study.
1.1 Tectonic Setting of Study Area

The northeast Pacific Ocean off the coast of western Canada has played a significant role in the development of theories related to plate tectonics. Since the original proposals of Wilson (1965) and Vine and Wilson (1965) based on the magnetic anomaly map of Raff and Mason (1961), the tectonics of the region have been studied by many scientists. The general configuration and approximate relative motions of the plates have been established for some time (e.g. Atwater, 1970). However, certain important and critical details have been worked out only recently. These are discussed in an extensive review by Keen and Hyndman (1979).

Figure 1.1 outlines the plate configuration in the region of the continental margin of western Canada. In the northwest, the Queen Charlotte transform fault zone, which separates the Pacific and America plates, has right-lateral motion of about 5.5 cm/yr. The small Explorer and Juan de Fuca plates are separated from the Pacific plate by an en echelon series of spreading centres and transform fault zones, from the Tuzo Wilson seamounts in the north to the Juan de Fuca ridge in the south. The pattern is probably formed by a complex process of ridge jumping, reorientation of spreading axes and initiation of new spreading centres (Keen and Hyndman, 1979; Hyndman et al., 1979). Full spreading rates vary from 4 to 6 cm/yr. The convergent boundaries between the two small plates and the America plate are approximately coincident with the
Figure 1.1. Major tectonic plate boundaries off Canada's west coast. Arrows indicate the direction and magnitude of the relative plate motions with respect to the America plate. The magnitude scale is shown in the lower left corner. Area enclosed by dashed lines defines the coverage of the location map in figure 2.1.
base of the continental slope. Subduction rates for these young plates are low, lying between 1 and 3 cm/yr (Riddihough, 1977).

The different directions and rates of motion of the Explorer and Juan de Fuca plates require differential motion across some boundary between them. Hyndman et al. (1979) established the existence of this boundary, naming it the Nootka fault zone. Motion across it is left lateral at a rate of about 2 cm/yr. The Nootka fault zone forms the core of the area of interest in the seismic refraction project discussed in this thesis.

1.2 Outline of Study

The objective of this study is to obtain detailed seismic information on the crustal structure of the Juan de Fuca and Explorer plates in the region of the Nootka fault zone, and to relate the seismic results to the petrology of the oceanic crust and upper mantle. Insight concerning various tectonic processes may be gained from the knowledge drawn from this study.

The experimental set up and preliminary data processing relevant to all data in this study are described in Chapter Two. The main body of the study is divided into three major sections.

In the first section, Chapter Three, the various aspects of interpretation of P wave profiles are discussed. Current techniques of refraction seismology used include the linearized tau-p inversion of travel times, ray tracing, and synthetic seismogram modelling of the
amplitude data. To relate the seismic results to the lithology of the oceanic crust and upper mantle, a detailed comparison of the results with the observed properties of ophiolite complexes is given.

For the purpose of reducing the ambiguities in the petrological interpretation of seismic results, shear wave information is extremely important; therefore, considerable effort has been made in this study to obtain high quality shear wave data. The second section, Chapter Four, deals with the problems of shear wave observation and the interpretation of two shear wave profiles.

The third section is concerned with velocity anisotropy, which is shown by the results of the profile interpretations to be very prominent in this study site. The configuration of shots and receivers in this experiment provided some areally distributed travel times which allowed study of azimuthally varying velocities in the upper mantle. The interpretation of this data set and the implications of anisotropy for tectonic processes are discussed in Chapter Five.

Finally, Chapter Six gives a summary of the conclusions drawn from the results of this study.
2. ACQUISITION AND PROCESSING OF DATA

2.1 The Experiment

In 1977, four ocean bottom seismometers were deployed near the Nootka fault zone for a study of seismic refraction and seismicity. The seismicity was analysed by Hyndman et al. (1979). Analysis of the refraction experiment constitutes the present study.

Three lines were shot using a total of 150 explosive charges ranging in size from 5 to 270 kg. The location and orientation of the shot lines and the positions of the OBS's are shown in figure 2.1: EX1 crosses the Nootka fault zone in an east-west direction; EX2 crosses the zone in a south-north direction; and EX3 runs parallel to the general trend of the zone. Reversals were achieved by deploying OBS's on both ends of the profiles; however, data from the OBS that was located at the western end of EX1 were lost due to instrument problems, so EX1 was unreversed. Maximum shot-receiver range was approximately 80 km. The configuration of the OBS array was a compromise between effective earthquake location and convenient layout of seismic refraction lines.

In addition to the explosive shots, a 16 litre airgun provided the energy source for a number of short range refraction lines. Shot spacing of 300 m for the airgun profiles provided detailed information for the upper crust at each of the receiver sites.
Figure 2.1. Location map for the three ocean bottom seismometers (solid circles) and the refraction shot lines (heavy solid lines). Arrows indicate the forward direction of the profiles. Bathymetric contours are in meters (from Tiffin and Seeman, 1975). The approximate extent of the Nootka fault zone, indicated by the stippled area, is from the results of Hyndman et al. (1979).
2.2 Instrumentation and Data Digitization

The ocean bottom seismometers are of the free fall pop-up type similar to those described by Lister and Lewis (1976) and Johnson, Lister and Lewis (1977). The same instruments were used in an earlier refraction experiment carried out near the Explorer ridge region (Cheung and Clowes, 1981).

The OBS's record in continuous direct AM mode using a 4 channel tape head at a tape speed of 1 mm/s. The overall frequency response is bandlimited between 2 and 100 Hz with a dynamic range of 80 db. The rather wide dynamic range is achieved by a bipolar square-rooting signal compression scheme described by Lister and Lewis (1976). However, such signal compression scheme does introduce some distortions in the waveforms recorded (for example, see figures 4.3 and 4.4). A 20 Hz time code from an internal clock is recorded on the first channel while the outputs from one vertical and two horizontal 4.5 Hz geophones are recorded on the other three channels.

In converting the original analog data into digital form, a two stage process is necessary due to the limited capacity of the available instrumentation for playback and digitization. The procedures involved have been described in detail by Cheung (1978) and are outlined below for completeness.

First, the data is transcribed from AM to FM at a playback speed of 15/16 inch per second (ips) and a recording speed of 15 ips. The FM tape is then played back at a speed of 15/16 ips (approximately 1.5 times
real time) for digitization at a sampling rate of 312.5 samples per second (sps). The overall effect of this process is to give a real-time data sampling rate in the neighbourhood of 200 sps. The actual sampling rate varies with time and differs between the OBS's due to minor variations in tape drive motor speed and a small degree of tape stretching. However, accurate estimates of the data sampling rate are obtained by decoding the digitized time code on the first channel. It is found that for a given record section, the sampling rate can be considered as constant; but between different record sections, variations in the order of 5 digital samples per second are found.

2.3 Range and Travel Time Corrections

All shot locations and receiver positions are determined from Loran C navigation supplemented by satellite checks. Relative position accuracy was ±200 m with greater uncertainties in absolute position (Hyndman et al. 1979). Shot-receiver distances determined by this method have accuracies similar to or better than those determined from direct water wave arrivals.

To bring the source and receiver to the same depth level, all shots are corrected to an equivalent source depth of 2.5 km by ray tracing through the water column. The sound velocity profile of the water column (figure 2.2) is based upon the compilation of the National Oceanographic Data Center in Washington, D.C. for the study area. The corrections are phase velocity dependent. For a phase velocity of 6.0 km/s, the offset in the shot range due to the travel path through water is 0.64 km.
Figure 2.2. Sound velocity profile of the water column based upon the compilation of the National Oceanographic Data Center in Washington D. C. for the study area.
while the travel time correction is 1.74 s.

The rapid sedimentation rate in this region has resulted in a very subdued seafloor topography which requires only minor corrections; however, substantial corrections are required for the variations in basement topography. Since no reflection profiles along the shot lines are available, the basement topography over most of the study area is interpreted from the reflection profiles of Davis and Lister (1977). Their results give detailed information on basement depths over a grid with spacing of approximately 10 km in the region of Juan de Fuca ridge and vicinity. Additional information on basement topography is provided by Hyndman et al. (1979) in the area not covered by Davis and Lister (1977). The seafloor and basement topography for the three shot lines are plotted in figures 3.2c, 3.3c, and 3.5c. Major geological features can be delineated from the basement topography. At distances between 15 and 30 km along profile EX1 (figure 3.2c) where the Nootka fault zone is crossed, the basement shows a subdued expression. The prominent basement relief between 40 and 50 km is Middle Ridge of the Juan de Fuca ridge system. A basement ridge associated with the Sovanco fracture zone emerges beyond 70 km. Along profile EX3 (figure 3.3c), the major features of the basement are East Ridge at 40 km and Middle Ridge beyond 50 km. For profile EX2 (figure 3.5c), the basement variation is less abrupt: East Ridge is located between 5 and 30 km and the Nootka fault zone extends beyond this to 40 km.

To calculate the topographical corrections for a given interface, a
datum level is first chosen. Corrections for the variation in the depth of this interface are obtained simply by replacing material above or below this reference level with the appropriate material. This procedure is carried out for the water-sediment interface and for the sediment-basement interface. Figure 2.3 illustrates the situation for an exaggerated topographical bump on the sediment-basement boundary. $\Delta t$, the correction to be added to the travel time, is given by

$$\Delta t = \Delta h \cdot q(p)$$  \hspace{1cm} (2.1)$$

where

$$q(p) = v_{\text{SED}}^{-1} \left[ 1 - (pv_{\text{SED}})^2 \right]^{-1/2} - v_{\text{BASE}}^{-1} \left[ 1 - (pv_{\text{BASE}})^2 \right]^{-1/2}$$ \hspace{1cm} (2.2)$$

The ray parameter $p$ is determined from the slope of the uncorrected travel time curve while $v_{\text{SED}}$ and $v_{\text{BASE}}$ are the seismic velocity of the sediment and the basement respectively. This approach assumes that all structures at lower levels are horizontal (Whitmarsh, 1975). While the magnitude of this correction is insensitive to the value of $p$, it is very sensitive to the values chosen for the velocities (Detrick and Purdy, 1980).

Another approach for correcting topographical variations is to remove the travel time from the shot to the entry point at the interface concerned (A to B in figure 2.3). In this approach, it is assumed that all structures at lower levels are parallel to the topography of this boundary. This calculation does not involve the velocity immediately below the interface, but it is very sensitive to the value of $p$. 
Figure 2.3. Schematic diagram illustrating the various parameters used in calculating topographical corrections to the travel time data. The depths are in units of km and the velocities are in units of km/s.
There are two main reasons why the first approach is favoured in the present study. First, due to large scatters in the travel time data points (see for example figure 2.4), the estimate of the ray parameter is not very reliable, rendering the second method unstable. Second, the velocities of the sediment and the basement are known sufficiently well in the study area for the first method to give reasonable corrections. Although no refracted arrivals are observed from within the sediment, information for the sediment velocity is taken from Davis et al. (1976) who obtained a number of velocities for the sediment at the northern end of the Juan de Fuca ridge by using bottom sources and ocean bottom seismometers. The basement P velocity of 4.5 km/s is obtained from the average of the apparent velocities of the first branch of the refracted arrivals on the airgun profiles. S velocity for the basement is obtained from the P velocity by assuming a Poisson's ratio of 0.34 (Spudich and Orcutt, 1980a). Values of the various parameters used in making the topographical corrections are shown in figure 2.3.

In making corrections for the different receiver depths, a vertical travel path is assumed. This is acceptable because both the water and basement depths at the OBS sites differed only slightly from the reference levels and the actual angles of emergence are estimated to be near vertical.

A comparison of the raw travel time data and the corrected data is shown in figure 2.4 for profile EX3. The seafloor and basement topography are also plotted. The obvious travel time advance due to the basement ridge at 40 km is much reduced in the corrected travel time.
Figure 2.4. An example showing the travel time data before and after applying topographical corrections for profile EX3. The sea floor and basement reliefs are plotted at 6 times vertical exaggeration.
curve; however, some degree of scatter still exists in the data. Maximum deviation of the basement depth from the datum level is approximately 1.0 km which corresponds to a correction of 0.31 s for P arrivals and 0.10 s for S arrivals. For variations in the water-sediment interface, the typical travel time correction is much smaller in magnitude, in the order of a few milliseconds.

As with other marine refraction studies in areas of considerable topography (Spudich and Orcutt, 1980a; Detrick and Purdy, 1980), the topographical correction is the largest uncertainty in the travel time data due to possible errors in the velocity estimates of the upper layers and uncertainties about the ray entry points. Errors associated with the topographical corrections are estimated to be in the order of ±0.03 s.

Further uncertainties in timing arise from factors such as shot origin time determination, shot depth estimate, and OBS clock drifts. The contribution to the timing error from these factors are estimated to be ±0.03 s, giving a maximum total error of ±0.06 s.
3. P WAVE PROFILE INTERPRETATION

3.1 Record Sections

Vertical component record sections, corrected for topography and the travel path through water, are shown in figures 3.2a to 3.6a for the explosion profiles and figures 3.7a to 3.9a for the airgun profiles. Trace amplitudes on all sections are scaled with an $r^{-2}$ spreading factor to enhance the weak arrivals at greater distances. To compensate for the different charge sizes in the explosion profiles, an amplitude normalization factor of weight$^{2/3}$ is applied. This normalization is based on empirical results relating the size of the charge with the amplitude of seismic energy it generates (O'Brien, 1960; Kaneström and Øvrebø, 1978).

Figure 3.1 shows the Fourier power spectra for a refracted P arrival from an explosion and a segment of the noise in the data prior to any seismic arrivals. Both spectra are calculated from a signal duration of one second. It is apparent that much of the noise is concentrated in the frequency range of 23 to 27 Hz and at 2 Hz while the P arrival has a well defined peak at 6.5 Hz and a small secondary peak at 19 Hz. This illustrates that the signal and noise are quite distinct in frequency content and improvement in data quality may be gained by digital filtering.

The seismic traces in the record sections have all been bandpassed with a 5 to 20 Hz eight pole zero phase Butterworth filter. Filtering of the data does improve the appearance of the seismic traces thus
Figure 3.1. Fourier power spectra, normalized to the maximum power peak, for a segment of noise in the data prior to any seismic arrivals and for a refracted P phase (see trace (a) of figure 4.3). The signal duration is one second for both spectra. Note the low power of the noise peaks relative to signal in the P wave spectrum.
Figure 3.2. (a) Record section along profile EX1 corrected for topographical variations of the seafloor and the basement. Amplitudes of the traces have been corrected for charge size and normalized by a spherical spreading factor of $r^2$. Travel times and ranges have been corrected for the one-way travel path through water. First arrival picks are noted by horizontal tick marks. The travel time curves are transferred from the synthetic seismogram section below. (b) Best fit synthetic seismogram section for profile EX1 calculated by the WKBJ approach of Chapman (1978) and using the velocity-depth model of figure 3.11a. An $r^2$ spreading factor is included. (c) Topography of the seafloor and the basement along profile EX1. Vertical exaggeration is 3X. The profile runs approximately westward from the origin.
Figure 3.3. (a) Record section, (b) synthetic seismogram section and (c) topography along profile EX3. Description is the same as that in figure 3.2. Synthetic seismograms are calculated using the velocity-depth model of figure 3.1lc. Profile direction is from NE to SW.
Figure 3.4. (a) Record section and (b) synthetic seismogram section for profile EX3R. Synthetic seismograms are calculated using the velocity-depth model of figure 3.11c. Origin is located at the range of 45 km on profile EX3. Profile direction is from SW to NE.
Figure 3.5. (a) Record section, (b) synthetic seismogram section and (c) topography along profile EX2. Description is the same as that in figure 3.2. Synthetic seismograms are calculated using the velocity-depth model of figure 3.11b. Profile direction is from south to north.
Figure 3.6. (a) Record section and (b) synthetic seismogram section for profile EX2R. Synthetic seismograms are calculated using the velocity-depth model of figure 3.11b. Origin is located at the range of 53 km on profile EX2. Profile direction is from north to south.
Figure 3.7. (a) Record section and (b) synthetic seismogram section of the airgun profile AG2N shot along the same direction as EX2. The synthetic seismograms are calculated to a distance of 18 km using the same velocity-depth model as EX2 but with a denser trace spacing. Some subcritical reflection travel time branches are included, though the amplitudes of these arrivals are small.
Figure 3.8. (a) Record section and (b) synthetic seismogram section for airgun profile AG3S shot along the same direction as profile EX2R. The extended coda on traces between 12 and 14 km is well modelled by the synthetic seismograms. The large amplitude secondary arrivals between 7 and 8 km, probably due to multiply reflected phases, are not modelled. The velocity-depth model used in the calculation of the synthetic seismograms is the same as for EX2R.
Figure 3.9. (a) Record section and (b) synthetic seismogram section for airgun profile AG1S, shot south of the OBS that recorded profile EX1. The velocity-depth model used in the calculation of the synthetic seismograms is the same as for EX1. The secondary arrivals between 8 and 10 km are not modelled.
facilitating comparison with synthetic seismograms for the purpose of amplitude interpretation; however, it also smooths out the first breaks of the P arrivals which is undesirable. Usually the impulsive first breaks of crustal arrivals can be picked more accurately from the unfiltered data while weak Pn arrivals contaminated by severe noises would benefit from filtering. The arrival picks are indicated by horizontal tick marks on the data traces in the explosion record sections. First arrival breaks are not marked individually on the airgun sections since the breaks are obvious in most cases. Superimposed on the data are theoretical travel time curves transferred from the best fit synthetic record sections which are plotted immediately below the data (see figures 3.2b through 3.9b).

There are a number of common features in the explosion record sections (figures 3.2a to 3.6a): (1) large amplitude arrivals over a short distance interval associated with the triplication caused by a velocity gradient immediately above the crust-mantle boundary; (2) few prominent secondary arrivals; and (3) small amplitude Pn phases.

There are also a number of important differences among the record sections. The distance range over which the large amplitude triplication occurs differs from profile to profile and even between forward and reverse profiles. For example, the triplication is near 25 km in EX2 but is near 35 km in EX2R (figures 6a and 7a). This indicates that there are some lateral variations in the structures at depth and that the crustal thickness varies between the different locations. The travel time curves of EX2 and EX2R also show noticeable differences from those of the other
profiles. The slopes of the T-X curves of EX2 and EX2R reach an apparent velocity of 7.5 km/s at a much closer distance than the other profiles and this slope is maintained without change to beyond 75 km. Also note the sudden decay of amplitude along EX2 and EX2R at distances greater than 45 km. The observed amplitudes are approximately four times smaller than the headwave amplitudes predicted by the synthetic seismograms. Some of the first arrivals on EX2R are too poor to be picked with any confidence even though the shot sizes are as large as 270 kg. The moderately strong secondary arrivals with apparent velocity of 6.5 km/s on profile EX2 are also attenuated beyond 45 km. Interpretations of these anomalous features of EX2 and EX2R will be discussed later.

Airgun profile AG2N (figure 8a), shot along the same direction as EX2, is a good quality one; clear first arrivals can be seen to distances of 18 km. Profiles AG3S and AG1S (figures 3.8a and 3.9a), with a maximum range of approximately 14 km, are more typical of the other airgun profiles recorded. The constant source signature of the airgun and the close shot spacing allow the subtle variations in amplitude with distance to be observed. By modelling the airgun amplitude and travel time data, much better constraints on the upper crustal structures to depths of 2 km subsediment at each receiver site can be obtained than would be possible with the explosive sources alone.

3.2 Inversion of Travel Time Data

In the field of seismic travel time inversion, a number of significant developments have taken place in the last few years. Most of
the new advances are based on the reparameterization of travel time data, \( T(X) \), to the delay time function \( \tau(p) = T(p) - px(p) \), where \( p \) is the ray parameter. There are several advantages in using the function \( \tau \). For example, since \( \tau \) is a single-valued, monotonically decreasing function of \( p \), difficulties associated with triplications in the travel time versus distance function are avoided. Also first order errors in \( p \) result in only second order errors in \( \tau \) which are of the same order as errors in \( T(X) \); therefore, accurate estimate of \( \tau(p) \) can be obtained from the data \( T(X) \). Perhaps the most important aspect of the tau function is that the inverse problem can be formulated as a linear problem with a simple change of variable (Garmany, 1979).

In the present study, the interpretation of travel time data utilizing the tau-p function proceeds in two ways. First, extremal bounds on the velocity-depth function are obtained from the bounds of \( \tau(p) \) using the methods of Bessonova et al. (1974). Second, a particular \( v-z \) model which satisfies the tau-p data to within the errors in the observations is found by linear programming techniques (Garmany et al., 1979).

Before carrying out the tau inversion procedures, the tau-p data must first be calculated from the T-X travel time points. There are a number of methods for doing this reparameterization (eg. Kennett, 1976; Bates and Kanasewich, 1976) but after experimentation with the existing data, the technique described by Kennett and Orcutt (1976) was chosen for its superior numerical stability and ease of application. Figure 3.10 illustrates this method schematically. First, a smooth cubic-spline
Figure 3.10. Schematic diagram illustrating the calculation of \( \tau(p) \) from \( T(X) \). The maximum of the reduced travel time curve is the value of \( \tau \) at \( p_i \). The procedure is repeated for the range of values of \( p_i \) consistent with the data.
is fitted to the reduced travel time data. Then the tau value for a given value of $p$, say $p_i$, is obtained by finding the maximum of the reduced travel time curve $T(p_i X)$. The error associated with this estimate of tau is the same as the uncertainty in $T(X)$ and is set at $\pm 0.06s$.

The upper and lower bounds of $\tau(p)$ from the combined data set of airgun and explosive lines are then mapped into bounds of the velocity-depth function by the methods of Bessonova et al. (1974). An example of the extremal bounds solution is shown in figure 3.11d. The bounds delineate a change in the velocity gradient at approximately 3.5 km depth; however, no details of the velocity-depth structure are defined. Extremal bounds for the other profiles are similarly broad and bracket the final velocity-depth models in much the same manner as for EX2.

This lack of resolving power implies that very few results of geophysical interest can be obtained from the extremal bounds. Similar views have been expressed by others in the course of interpreting actual data by the tau method (eg. Spudich and Orcutt, 1980a). Recently a number of authors have shown that greater constraints on the extremal bounds can be achieved with the addition of $X(p)$ data (Orcutt, 1980 and Jurkevics et al. 1980). However, the extraction of $X(p)$ data from the travel time data $T(X)$ is a much more unstable process than the estimation of $\tau(p)$. Since the aim of the present study is to obtain detailed characteristics of the velocity-depth structure by modelling the amplitude data, a more desirable approach than obtaining refinements to the extremal bounds would be the calculation of a suitable starting model for synthetic seismogram calculations. The linearized inversion
formulation of tau-p data, described by Garmany et al. (1979), is ideal for this purpose.

By solving a set of linear tau-p equations using linear programming techniques, a particular velocity-depth model which satisfies the tau-p data to within the errors in tau can be easily derived (see Appendix 1). The linear programming solution for EX2 (LINP) is shown in figure 3.11d along with the final model (EX2) for comparison. As in the case of the extremal bounds calculation, an error of ±0.06 s in tau is assumed. The major advantage of the linear programming approach is that the computation is extremely fast and economical and a useful starting model for amplitude studies is readily obtained. Only minor changes to LINP are needed to arrive at the final model (figure 3.11d) which is obtained by amplitude comparisons using synthetic seismogram sections.

3.3 Synthetic Seismograms and Amplitude Modelling

In the last section, it has been shown that by using the first arrival travel time data alone, the extremal bounds thus obtained provide us with very little detailed information concerning the velocity structure of the crust. The linear programming technique does give models with considerably more structure, but the inherent uncertainty in travel time inversions, expressed clearly by the poor resolution in the extremal bounds, discourages us from putting too much faith in the LINP solutions.

Considerably more information about the crustal velocity structure can be extracted from seismic refraction data if the observed amplitudes
as well as travel times of the various seismic phases are considered. Recent advances in theoretical and computational techniques have allowed the modelling of amplitude data by synthetic seismograms to become a routine procedure in the interpretation of marine refraction data (eg. Malecek and Clowes, 1978; Spudich and Orcutt, 1980a). In the present study, synthetic seismogram modelling of the data is one of the most important steps in obtaining detailed velocity-depth information for the oceanic crust and upper mantle in the study site.

There are a number of methods for computing synthetic seismograms given the velocity and density structure of an earth model. Extensive reviews of the various methods can be found in Chapman (1978) and Richards (1979) while a comparison of the practical application of some of the methods is given by Spudich and Orcutt (1980b). Almost all of the techniques require lateral homogeneity in the velocity structure and they differ mainly in their degree of accuracy and the expense of computation (Aki and Richards, 1980).

For the present study, the WKBJ synthetic seismogram algorithm of Chapman (1978) is used in the modelling of the amplitude data. Chapman has shown that the WKBJ seismogram correctly predicts arrivals for turning rays, partial and total reflections and head waves of reflections. There are, however, a number of limitations in using the WKBJ algorithm. For example, WKBJ seismograms are not strictly correct for waves turning in the vicinity of a low velocity zone or a velocity discontinuity. Diffraction beyond triplication or precritical reflection from velocity gradient zones also are not correctly calculated (Spudich
and Orcutt, 1980b). However, numerical experiments carried out by the writer and others (Spudich and Orcutt, 1980b) have shown that WKBJ seismograms differ from the more accurate reflectivity method (RM) synthetic seismograms by negligible amounts for velocity-depth models typical of the oceanic crust. Significantly, the expense of computation for the reflectivity method is an order of magnitude higher than that for the WKBJ algorithm due to the fact that velocity gradients must be modelled as stacks of constant velocity layers in the RM approach whereas the WKBJ algorithm handles velocity gradients efficiently by analytic approximations. The relatively inexpensive WKBJ method allows a large number of trial models in the process of fitting the amplitude data. Except for some explainable lateral variations in structure to be discussed in detail in the following sections, the WKBJ approximation appears to be valid for the present data set.

The best fit synthetic record sections are plotted as figures 3b through 9b. In generating the synthetic seismograms, only the primary ray paths are considered since this approach generally produces acceptable fits to the data. An exception occurred for profile EX3R (figure 3.4) where a multiple reflection from the underside of the sediment-basement interface generated significant amplitudes at distances from 15 to 20 km. This multiple is modelled in the synthetic section and it is found to have negligible contribution at greater distances. Coherent arrivals of this phase are not observed on the other profiles. Reverberation within the sediment can also cause later arrivals which may explain the extended wave-train on the traces of profile EX2R (figure 3.6a). However, synthetic seismograms incorporating
the sediment layer multiples do not give significant amplitude for the assumed sediment velocity structure. An additional complication is the strong effects of varying sediment thickness on the arrival times of these phases with respect to the primary arrivals. For these reasons, only the primary ray paths are used.

The modelling procedure consists of calculating a synthetic seismogram section from a starting velocity-depth model (LINP solution) and comparing it with the data record section. Perturbations to the model are made by a trial-and-error approach and new synthetic sections are calculated until an acceptable fit to the amplitudes and travel times of the observed section is found. No attempt is made to match the observed waveforms exactly; the goodness of fit is determined subjectively by visual inspection of the relative amplitudes of the traces and using the travel time data as constraints.

With the development of automated schemes for synthetic seismogram fitting, such as the method of Chapman and Orcutt (1980), objective interpretation of amplitude data will have some advantages over the more tedious trial-and-error approach. As pointed out by Spudich and Orcutt (1980a), the drawback in the trial-and-error approach in amplitude modelling is the inability to define adequate bounds on the velocity-depth function. Though it is impossible to sample all models within the model space, one may display a range of acceptable models to give an idea of the extent of the uncertainty in the procedure, as has been done by Spudich and Orcutt. For the sake of simplicity, the writer has chosen to display only the final preferred models (figures 11a, b, and c) while
bearing in mind that there will be other acceptable models that can also satisfy the data to within the errors of the observations.

3.4 Results

With the acquisition of more data and the improvement in seismic interpretation techniques, the simple three layer model of the oceanic crust suggested by Raitt (1963) has undergone a number of modifications. Fine structures such as sublayering and velocity gradients within the layers are commonly found. In addition, there appears to be a systematic variation of the crustal structure with age, location, and tectonic regime of the site.

In this section, the P wave velocity-depth models for each of the three profiles and their reversals, interpreted by modelling the data record sections with WKBJ synthetic sections, are discussed. It will be shown that the oceanic crust at the study site as defined by the P wave data does contain some of the fine structures as well as lateral variations mentioned. Since the present results are derived by using techniques that require laterally homogeneous structure, the existence of lateral variations will have to be explained carefully.

The final velocity-depth models, shown in figures 11a, b, and c, indicate that the crustal structure in the Nootka fault zone region consists of zones of differing velocity gradients rather than distinctive layers. Still, the notion of a layered structure is useful for the purpose of comparison of seismic results; therefore, layer
Figure 3.11. (a), (b), and (c): Velocity-depth curves for all the profiles. Depths are measured from the seafloor. The sedimentary layer on all profiles is constrained to be 1 km thick and has a velocity of 1.8 km/s. The upper crustal models to a subsediment depth of 2 km are derived from the air gun data. (d) Extremal bounds for profile EX2 are denoted by squares. The linear programming solution LINP (dashed line) can be compared with the final model for EX2 (solid line).
divisions are assigned to the final velocity-depth profiles in accordance with the characteristics of the velocity gradient in each depth range. Figure 3.12 is such a schematic representation.

Upper Crust

A relatively consistent picture of the upper crust is shown by the velocity-depth profiles in figure 3.11. Immediately below the sediment layer, the velocity ranges from 3.7 to 4.7 km/s and increases with depth at a moderate average velocity gradient of 0.5 km/s/km (hereafter abbreviated to /s) to a depth of approximately 1.9 km. Then a zone of very high velocity gradient is found which marks the transition from layer 2A to layer 2B. The average gradient of 1.6/s in this zone is the highest that has been encountered. Such a zone is not found in the velocity-depth profile of EX2R; instead, a velocity discontinuity is found at a depth of 1.4 km. However, this feature of EX2R is not well constrained.

Below the transition zone lies layer 2B where the velocity gradient decreases to a lower value of 0.3/s except in EX2 and EX2R, for which the velocity gradient remains high at 1.0/s. At the base of layer 2B, velocities are found to be in the range of 6.0 to 6.4 km/s. The combined thickness of layer 2A and layer 2B ranges from 2.3 to 3.1 km but a lower value of 1.6 km is found for profile EX2R.
Figure 3.12. Schematic representation of the velocity-depth profiles of figure 3.11 in terms of the conventional layered model of the oceanic crust. The depth is measured from the seafloor. The numbers are the velocity at the top of each layer in km/s. The sediments, layer 1, are constrained to have a thickness of 1.0 km and a velocity of 1.8 km/s. Layer designations follow those of Christensen and Salisbury (1975). The question marks on the layer 2B-3A boundary of EX3R indicate that no boundary is evident in the velocity depth curve; the velocity of 6.1 km/s is that at the depth marked.
Lower Crust

Beneath layer 2B, a velocity discontinuity, varying in its degree of sharpness, is interpreted for all profiles except for EX3R where no discontinuity in either the velocity or the gradient is present. Where the discontinuity is observed, it is interpreted to be the boundary between layer 2B and 3A. The structure of the lower crust is sufficiently different among the various profiles to warrant an individual discussion of each.

EX1: As shown in figure 3.11a, layer 3A starts at a depth of 3.8 km and a velocity of 6.6 km/s, increasing with a small velocity gradient (0.1/s) to 7.6 km. At this depth, a sharp increase in the velocity gradient is found in a zone approximately 1.8 km thick. This zone marks the transition from layer 3A to the upper mantle and has been identified as layer 3B. Such a transition zone has also been found in a number of refraction studies of the oceanic crust (Malecek and Clowes, 1978; Spudich and Orcutt, 1980a). At the depth of 9.4 km, an upper mantle velocity of 8.0 km/s is found.

EX3: The velocity-depth profile of EX3 (figure 3.11c) indicates a more complex structure for layer 3 than that of EX1. At a depth of 6.0 km, a 0.6 km thick zone of high velocity gradient (0.9/s) is required to produce the large amplitude arrivals observed on the EX3 record section at the distance range of 30 km (figure 3.3a). The second occurrence of large amplitude arrivals at 45 km range on the record section is caused by the triplication from layer 3B. Layer 3B of profile EX3 is somewhat thicker than that of EX1 and the velocity gradient found
within is not as high. The velocity of 8.3 km/s derived for the depth of 11.2 km is interpreted as the upper mantle velocity.

EX3R: As mentioned earlier, the velocity-depth profile of EX3R does not show any discontinuity between layer 2B and layer 3A. Instead, the velocity remains nearly constant at 6.0 km/s to the depth of approximately 6.0 km and between 6.0 and 8.6 km the velocity gradient begins to increase rapidly (figure 3.11c). This zone of high velocity gradient gives rise to the large amplitude arrivals at the distances of 27 to 35 km on the record section of profile EX3R (figure 3.4a). These arrivals even dominate the triplication caused by layer 3B. This difference in amplitude behaviour between EX3 and EX3R is quite apparent when comparing the two record sections. Such a difference can be explained by the fact that the two profiles are not exact reversals; that is, seismic waves from shots at a range of 30 km from the respective receivers did not sample the same crustal material. The common depth region for the two profiles is at upper mantle depths, where the same velocity of 8.3 km/s is found for the reversed profiles EX3 and EX3R.

EX2: Unlike the other profiles, there is no apparent velocity higher than 7.5 km/s observed on profile EX2 below a depth of 6.4 km (figure 3.11b). A zone of high velocity gradient similar to layer 3B of the other profiles is found at a much shallower depth of 5.0 km. The shallow depth of this transition zone on EX2 reflects the fact that the large amplitude triplication occurs at the much closer distance of 25 km (figure 3.5a) than that observed on the other profiles. There are no
other large amplitude triplications observed to distances as far as 80 km. This implies that the velocity must remain constant below 6.4 km to some depth, a fact further confirmed by the extremal bounds of EX2 (figure 3.11d) which show that the maximum velocity below 6.4 km is well constrained to be less than 7.8 km/s. The 7.5 km/s velocity is interpreted to be that of the upper mantle along EX2 (compared with 8.0 and 8.3 km/s for EX1 and EX3). The maximum velocity observed on the reverse profile EX2R is also 7.5 km/s so that a dipping structure cannot be used to explain the low upper mantle velocity. A more likely explanation may be anisotropy in the upper mantle, similar to that observed in other refraction studies (Snydsman et al., 1975; Malecek and Clowes, 1978). This point will be discussed further. Aside from the low velocity of the upper mantle, the second anomalous aspect of profile EX2 is the shallow upper mantle depth of 6.4 km compared to 9.4 km and 11.2 km for EX1 and EX3 respectively. The thinner crust of EX2 has resulted mainly from the reduced thicknesses of layers 3A and 3B. The combined thickness of layers 3A and 3B for EX2 is 4.0 km compared to the average value of 6.7 km for EX1, EX3 and EX3R.

EX2R: For profile EX2R, an upper mantle velocity of 7.5 km/s at a depth of 8.1 km is found. This depth, while shallower than those of EX1 and EX3, is considerably greater than the 6.4 km upper mantle depth of EX2. The greater upper mantle depth also means a greater Pn cross-over distance - 35 km for EX2R compared to 25 km for EX2. This discrepancy between EX2 and EX2R raises a question concerning the validity of the assumption of lateral homogeneity required by the interpretation techniques. Certainly, part of the difference is due to the fact that
EX2 and EX2R do not form exact reversals of each other. Another possibility is the existence of a vertical fault along EX2, which means that the results obtained by assuming lateral homogeneity must be used with care.

3.5 Discussion

Hyndman et al. (1979) presented a model for the development of the Nootka fault zone derived from calculations based on the observed magnetic anomalies and other data. They showed that the orientation of the fault zone has rotated from an east-west to a northeast-southwest direction then returning closer to east-west, all within the past 8 Myr. Figure 3.13 shows the three seismic profiles superimposed on the Hyndman et al. (1979) model; two of the most recent orientations of the Nootka fault zone, for average ages of 0.5 and 1.5 Myr, are also shown. According to this model, the age of the Juan de Fuca plate in the study area ranges from 1 to 3 Myr and the adjacent crust on the Explorer plate across the fault zone is older by approximately 4 Myr. Much of the lithosphere older than 6 Myr on both plates has now been subducted beneath the America plate.

In the discussion that follows, it will be illustrated that this model can explain some aspects of the seismic results. Also, the lithology of ophiolite complexes, which are believed to be segments of oceanic crust emplaced on land, will be shown to be related to the seismic velocity structure of the crust in this region. A summary of the petrology and seismic velocity structure of the Bay of Islands
Figure 3.13. Tectonic model of Hyndman et al. (1979) showing the present configuration of the Explorer and Juan de Fuca plates using spreading parameters from observed magnetic anomalies. The thin solid lines with small numbers show seafloor ages; the two heavier lines with large numbers show fault locations at the time; both number sets in millions of years before the present. The approximate continental margin is indicated by the dashed line. The location of the explosive lines and OBS sites are shown by the heavy solid lines and solid circles.
ophiolite complex is given in table 3.1 (Salisbury and Christensen, 1978; and Christensen and Salisbury, 1979).

Layer 2

Figure 3.12 shows that the top of the igneous crust has a velocity ranging from 3.7 to 4.7 km/s which is in good agreement with other studies (eg. Cheung and Clowes, 1981; Spudich and Orcutt, 1980a). Deep sea drilling (eg. Hyndman et al., 1976), dredging, and studies on ophiolites (Salisbury and Christensen, 1978) have established that the subsediment material near spreading ridges is generally composed of fractured basalt flows and pillows. The velocity of the upper layer found by seismic refraction surveys is usually much lower than that observed from velocity measurements on laboratory samples (Hyndman and Drury, 1976). This discrepancy is most likely due to the fact that the actual crust contains fractures, voids and intercalated sediments of a scale much larger than the size of the laboratory samples (Hyndman et al., 1976). These factors tend to decrease the seismic velocity of the upper crust while the coherent samples used in laboratory measurements tend to give the maximum velocity. The increase in velocity with depth in layer 2A is probably a result of the closing of cracks and fractures within the basalt (Malecek and Clowes, 1978).

In the transition zone from layer 2A to layer 2B, the velocity increases from an average of 4.5 km/s to the 6.0 km/s range. This sudden increase in velocity probably marks the change from extrusive to intrusive levels of the crust. Spudich and Orcutt (1980a) have argued
Table 3.1. Petrology and measured seismic velocities (km/s) of the Bay of Islands ophiolite complex (from Salisbury and Christensen, 1978 and Christensen and Salisbury, 1979)

<table>
<thead>
<tr>
<th>DEPTH (km)</th>
<th>LAYER</th>
<th>LITHOLOGY</th>
<th>METAMORPHIC ASSEMBLAGES</th>
<th>MEASURED Vp</th>
<th>Vs</th>
<th>σ</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>2B</td>
<td>metabasalt</td>
<td>prehnite-pumice facies</td>
<td>6.25</td>
<td>3.40</td>
<td>0.29</td>
</tr>
<tr>
<td></td>
<td></td>
<td>brecc. dikes</td>
<td>green schist facies</td>
<td>6.20</td>
<td>3.30</td>
<td>0.30</td>
</tr>
<tr>
<td>2</td>
<td>3A</td>
<td>sheeted dikes (metadolerite)</td>
<td>epidote-amphibolite facies</td>
<td>6.75</td>
<td>3.75</td>
<td>0.28</td>
</tr>
<tr>
<td>3</td>
<td></td>
<td>late differentiates metagabbro</td>
<td>epidote-amphibolite facies</td>
<td>6.35</td>
<td>3.65</td>
<td>0.25</td>
</tr>
<tr>
<td>4</td>
<td></td>
<td>pyroxene gabbro</td>
<td>epidote-amphibolite facies</td>
<td>6.90</td>
<td>3.80</td>
<td>0.28</td>
</tr>
<tr>
<td>5</td>
<td></td>
<td></td>
<td>overprint</td>
<td>7.00</td>
<td>3.80</td>
<td>0.29</td>
</tr>
<tr>
<td>6</td>
<td>3B</td>
<td>olivine gabbro and troctolite</td>
<td></td>
<td>7.40</td>
<td>3.90</td>
<td>0.31</td>
</tr>
<tr>
<td>7</td>
<td></td>
<td>mantle</td>
<td>ultramafic</td>
<td>8.70* (\parallel)</td>
<td>4.90* (\parallel)</td>
<td>0.27* (\parallel)</td>
</tr>
<tr>
<td>8</td>
<td></td>
<td></td>
<td>serpentinization</td>
<td>8.20(\perp)</td>
<td></td>
<td>0.23(\perp)</td>
</tr>
<tr>
<td>9</td>
<td>(\perp)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

* average value \(\parallel\) parallel to spreading direction \(\perp\) perpendicular to spreading direction
that the reduction in pore space from a porosity of 24% to 2% can only account for the velocity increase within the top 0.6 km of the igneous crust. Any further increase in velocity with depth must result from other factors such as changing composition or increasing pressure. From the results of Salisbury and Christensen (1978), the increase in velocity from layer 2A to layer 2B observed in this study would correspond to the change from pillow and flow basalt to greenschist facies metabasalt and brecciated dikes found in the upper level of the Bay of Islands ophiolite complex in Newfoundland (see table 3.1). The reduced velocity gradient in layer 2B probably reflects the fact that the cracks and fractures at this depth are fewer in number so that their effect on the velocity gradient is diminished.

The boundary between layer 2B and layer 3A is highly variable. The velocity change across it ranges from a sharp jump in profile EX1 to a smooth transition in profile EX3R. Salisbury and Christensen (1978) suggested that the velocity discontinuity between layer 2 and layer 3 marks a metamorphic boundary between greenschist facies and epidote-amphibolite facies metabasalt and is related to the downward migration of water along joints and fractures. Lithologically, it coincides with the transition from brecciated dikes to metadolerite sheeted dikes. From the study of the Chilean ophiolite, Stern et al. (1976) proposed a similar metamorphic boundary between layer 2 and 3. If this is the case, the abnormally deep boundary between layer 2B and 3A (~3 km subsediment) may be a consequence of the complex plate motions that took place in the region within the past 8 Myr as described by Hyndman et al. (1979). The intense shear and fracturing associated with the constantly changing
plate boundary, exemplified by the present day Nootka fault zone, would allow seawater to migrate deeply into the crust. The variability of the nature of the boundary between layer 2 and layer 3 would be a result of the varying degree of water penetration and metamorphism at the different locations. One possible contradiction to this hypothesis is profile EX2R which has the shallowest layer 2 - layer 3 boundary, yet a great number of active faults are found along the profile (Hyndman et al, 1979). This suggests that the nature of the boundary between layer 2 and layer 3 may be much more complicated than a simple metamorphic boundary associated with water circulation.

Layer 3A

Previous studies have shown that layer 3 of the oceanic crust has a well-defined velocity range (6.4 to 7.0 km/s) but its thickness varies considerably (for example, see the compilation of Christensen and Salisbury, 1975). The present study is no exception. The velocity at the top of layer 3A has a narrow range of 6.4 to 6.6 km/s and the total thickness of layer 3 ranges from 3.1 to 7.4 km. The low 6.1 km/s layer 3A velocity of profile EX3R can be explained by an interface with medium velocity of 6.4 km/s which dips at an angle of 4° from the horizontal away from the Juan de Fuca ridge. Such a dip near an active spreading ridge is reasonable in view of the fact that most models of the oceanic crust indicate sloping structures near the ridge crest.

Besides the well-defined velocity range, the other distinctive aspect of layer 3A is its small velocity gradient (~0.1/s) in contrast
with the larger and more variable gradient of layer 2. The uniformity of
the velocity structure of layer 3A has led to the postulation of uniform
composition for this part of the crust (Malecek and Clowes, 1978). The
lithology of the Bay of Islands ophiolite shows that the top 1.0 km of
layer 3 is composed of metadolerite sheeted dikes with a compressional
wave velocity of 6.4 to 6.8 km/s measured at pressures of 0.9 to
1.2 kbar (Salisbury and Christensen, 1978). This certainly correlates
well with the results for the top of layer 3A. However, layer 3A of the
Bay of Islands ophiolite does show some changes in composition with
depth (table 3.1): a thick layer of intrusive metagabbro and gabbro
followed by a layer of pyroxene gabbro are found below the sheeted
dikes. Velocities within the metagabbro vary widely between 6.0 and
6.8 km/s where the low velocities are associated with the discontinuous
presence of late differentiates. Velocities in the pyroxene gabbro
increase slowly with depth from 6.9 to 7.1 km/s. Salisbury and
Christensen (1978) pointed out that these detailed changes in the
velocity structure are probably too small to be detected by routine
seismic refraction surveys, resulting in the interpretation of a uniform
layer 3A. This can explain the simple layer 3A structure interpreted for
profile EX1 where some gaps in shot spacing existed between 20 and 40 km
due to misfired shots. On the other hand, profile EX3 had a dense shot
spacing and revealed more details within layer 3A. Similar structures,
however, are not present in the reverse profile EX3R, rendering it
difficult to make any significant inferences on the fine structure of
layer 3A.
Layer 3B

The existence of a transition layer between layer 3A and the upper mantle is well documented in both the Pacific (Sutton et al., 1971; Malecek and Clowes, 1978; Spudich and Orcutt, 1980a) and the Atlantic (Steinmetz et al., 1977; Fowler and Keen, 1979; Detrick and Purdy, 1980). The velocity in this transition layer (3B) generally varies from 7.0 to 7.7 km/s, with the thickness ranging from 2.0 to 5.0 km. The velocity-depth profiles in figure 3.11 indicate that layer 3B of the crust in this region of the Pacific is best described as a zone 1.4 to 2.3 km thick, in which the velocity increases gradually from that of the lower crust to that of the upper mantle, with no sharp discontinuity in velocity at the crust-mantle interface. Christensen and Salisbury (1975) suggested that the transition from layer 3A to 3B corresponds to the transition from metagabbros to fresh gabbros. In the model of Stern et al. (1976), this boundary marks the maximum depth of hydrothermal circulation and layer 3B is also composed of fresh gabbros. The observed velocity in the lower half of layer 3B, however, is higher than that of fresh gabbro and the presence of moderate to strong velocity gradients implies that the composition within this layer must be changing gradually. Layer 3B, therefore, can not be composed of fresh gabbro throughout. The zone of interlayered olivine gabbro, troctolite and peridotite found in the Bay of Islands ophiolites (Salisbury and Christensen, 1978) would be more likely to give the required velocity and gradient inferred from the data (see table 3.1).
Upper Mantle

Normal upper mantle velocities of 8.0 and 8.3 km/s are observed for profile EX1 and the reversed profile EX3,3R. An anomalously low velocity of 7.5 km/s has been interpreted as the upper mantle velocity for the reversed profile EX2,2R. Although these values of upper mantle velocity are within the range found during other studies on the Juan de Fuca plate, some explanation of the difference is necessary.

Davis et al. (1976) reported the results of an OBS refraction survey on the Juan de Fuca ridge. Their refraction line 'east' partially reverses the profile EX1. For this line, they interpreted an upper mantle velocity of 8.3 km/s which is similar to the 8.0 km/s found for EX1. For lines 'north' and 'south', which are parallel to our profile EX2, they observed no upper mantle refracted arrivals. From a refraction survey carried out on Explorer plate near Explorer ridge, in an area northwest of the Nootka fault zone, Malecek and Clowes (1978) found a velocity of 7.9 km/s for the upper mantle along a reversed profile that ran perpendicular to the spreading ridge. On the profile that ran parallel to the ridge, the upper mantle velocity was found to be 7.3 km/s. Anisotropy of the upper mantle was invoked to explain the difference in the velocities.

Figure 3.13 shows that profile EX2 is parallel to the Juan de Fuca ridge and is only 23° from the magnetic lineation on the Explorer plate. It is along this profile that we found the upper mantle velocity to be low (7.5 km/s). For profiles EX1 and EX3, the strikes of the magnetic anomaly pattern vary from 55° to 80°. Here, we found the upper mantle
velocities to be high. This apparent anisotropy agrees well with the observations of Malecek and Clowes (1978). Anisotropy of the upper mantle has been well documented by other studies and is summarized by Christensen and Salisbury (1975). Generally, it is found that the upper mantle velocity is high in a direction perpendicular to the ridge and low parallel to it (Raitt et al., 1969; Keen and Barrett, 1971; Snydsman et al., 1975). Hess (1964) suggested that mantle anisotropy is produced at the ridge crest by the alignment of olivine crystals in the direction of flow. Subsequently, a number of authors proposed similar mechanisms which required a preferred alignment of the a-axis of olivine crystals to create the anisotropy observed. It is suggested that the 10% variation (defined by \((1 - V_{\text{min}}/V_{\text{max}}) \times 100\%\)) in upper mantle velocity observed in this study is due to the anisotropy of the upper mantle. Since the crustal ages are so young, this implies that the process which causes formation of the anisotropy effect takes place while new crust is being created at or near the ridge and this effect is preserved thereafter.

Crustal Thickness

Except for profile EX2, the upper mantle depths shown in figure 3.12 are unusually deep for crust less than 6 Myr old. The result is in line with other crustal studies undertaken on the Juan de Fuca and Explorer plates. For example, Davis et al. (1976) interpreted a crustal thickness in excess of 11.2 km in their refraction study on the Juan de Fuca ridge. Malecek and Clowes (1978) reported abnormally thick crust (8-10 km) on the Explorer plate within 50 km of the spreading ridge.
crest. 'Bunching up' of the young crust on 'collision' with the America plate was suggested by Malecek and Clowes to explain the anomalous crust in the region. On inspection of figure 3.12, it can be concluded that whatever the mechanism is for producing such thick crust, it must do so by the thickening of layer 3, the layer that is associated with metagabbros and cumulate gabbros.

It was remarked earlier that the differences in the upper mantle depths interpreted for the profiles must be reconciled so that the interpretation using laterally homogeneous models would be meaningful. This is especially true for the reversed profiles EX2,2R and EX3,3R. The following arguments are offered to explain the differences in crustal thickness observed for these two profiles. It is asserted that the velocity-depth structure of EX2 (figures 3.11b, 3.12) is more representative of the 1 Myr old Juan de Fuca plate and its reversal, EX2R, is more indicative of the crustal structure of the 5 Myr old Explorer plate. The differences of the upper crustal structures and the thicknesses are real. They can be attributed to differences in crustal formation and/or evolution with time, as described later, or to the Nootka fault where a 1.6 km vertical offset in the upper mantle depth may be present. Such an offset is inferred from the difference between the interpreted depths to Moho for profiles EX2 and EX2R.

Consider the tectonic model of Hyndman et al. (1979) in figure 3.13. For profile EX2, shots from distances to 30 km would have travelled through only the 1 Myr old crust of the Juan de Fuca plate while for EX2R shots over the same distance range would have travelled
through only the 5 Myr old Explorer plate. Interpretation of the data up to the fault would produce the velocity-depth models of figure 3.11b for the respective plates. What about crossing the fault zone? Ray tracing through a faulted model consisting of the two laterally homogeneous models of figure 3.11b, joined by a vertical fault, produces negligible differences in the travel times in both directions. However, the amplitudes of arrivals that have crossed the fault are expected to be somewhat attenuated. This is consistent with the amplitude characteristics of EX2 and EX2R where the Pn amplitudes at distances greater than 40 km are significantly lower than those predicted by the synthetic seismograms.

Additional support for the interpretation of different crustal thicknesses on either side of the Nootka fault is provided by EX1 which also crosses the fault zone. The layer 2 structure of EX1 interpreted from travel paths through the 2 Myr old part of Juan de Fuca plate is similar to that of EX3, as we would expect. For distant arrivals, the ray paths travel principally through the Explorer plate and Nootka fault zone. Thus the lower crustal structure should be similar to that interpreted for EX2R. As shown in figure 3.12, this is the case. The final result is a crustal thickness intermediate between that of EX2R and EX3. It can be noted that there is some degree of attenuation of energy beyond 56 km, a distance for which ray paths are bottoming in the fault zone.

Unlike profile EX2, the reversed profile EX3,3R traverses only the Juan de Fuca plate; however, the crustal age along the profile varies
from 1 to 3 Myr. The obvious differences between the velocity-depth structures of EX2 and EX3 imply that there must be variation of the crust with age. Since the upper mantle depths of both EX3 and EX3R are the same at 11.2 km, it is reasonable to assume that this depth is representative of the crust at the midpoint of the reversed profiles where the crustal age of the Juan de Fuca plate is 2 Myr. A comparison with profile EX2, where the 1 Myr Juan de Fuca plate has a crustal thickness of 6.4 km, would imply that the crust thickens by 4.8 km within a time interval of only 1 Myr. Such a rapid rate of 'crustal maturing' would mean that, except for very young regions such as that of EX2, differences in crustal thicknesses would be small between crusts of different ages on adjacent sides of transform faults. This is confirmed by other studies on fracture zones of the Mid-Atlantic ridge (eg. Detrick and Purdy, 1980) where no change in structure is detected on either side of the fault. Alternatively, the differences in crustal thickness could represent real variations with time of the processes of crustal formation at the ridge. Then the inferences are that the formation process is time varying on a time scale of less than 1 Myr and the variations are established at the ridge and carried away with the spreading process. Of course, the effects of both time variations in the formation process and subsequent changes as a result of 'crustal maturing' with age could be involved.

Variations of 4.8 km in crustal thickness over distances of tens of kilometers could call into question the use of the WKBJ synthetic seismogram algorithm which assumes a laterally homogeneous earth model. A recently developed approximate method for calculation of synthetic
seismograms, combined with ray tracing, in laterally varying structures (McMechan and Mooney, 1980) has enabled a check of the consistency of the interpretation against the observed seismic sections. The laterally varying model (figure 3.14) is constructed from the homogeneous models derived from profiles EX3 (range 0 to 15 km), EX3R (range about 30 km) and EX2 (range 45 to 60 km), with smooth variations to provide model continuity.

This model gives good agreement in both the amplitudes and travel times of the forward and reverse profiles EX3,3R. Although head wave contributions are not calculated by this approximate method, resulting in the fact that traces beyond 45 km have zero amplitude, the laterally varying model of figure 3.14 and 3.15 does offer some qualitative explanation to the observed amplitudes at far distances on both profiles. The pronounced change in upper mantle depth at the distance of 45 km on profile EX3 implies that propagation of Pn phases would be hindered. The sharp drop-off in amplitude beyond 50 km on the record section of EX3 (figure 3.3a), which the WKBJ synthetic seismograms were unable to model, could be the manifestation of this lateral variation in structure. For profile EX3R, however, the geometry of the ray paths indicates that the Pn phases would not be affected in a similar manner. This is consistent with the relatively large Pn amplitudes on profile EX3R at distances ranging from 50 to 60 km (figure 3.4).

No attempt is made to match precisely the synthetic sections from the laterally varying models with the observed data. The main purpose of this exercise is to show that the observed characteristics of the record
Figure 3.14. Synthetic seismogram section and ray paths for profile EX3 calculated by the methods of McMechan and Mooney (1980). The model is constructed from the laterally homogeneous models of EX3 (range 0 to 15 km), EX3R (about 30 km) and EX2 (range 45 to 60 km), with smooth interpolation of boundaries and velocities. Generally, no velocity discontinuity existed across the boundaries. The crosses are observed travel times. VR, the reducing velocity, is equal to 6 km/s.
Figure 3.15. Same as figure 3.14 for the reverse profile EX3R. Note the differences in both the ray paths and the amplitude characteristics of the synthetic seismogram section as compared with profile EX3.
sections can be explained by the strong lateral variations in crustal structure which have been interpreted along the profile. A comparison of figures 3.14 and 3.15 with figures 3.3a and 3.4a exemplifies this consistency.

### 3.6 Concluding Remarks

The following summarizes the conclusions based on the results of the P wave interpretation presented in this chapter:

1. As crustal age increases from 1 to 2 Myr on the Juan de Fuca plate, the crustal thickness increases from 6.4 to 11.2 km. This difference is due primarily to differences in the thickness of layer 3. A rapid process of 'crustal maturing' within a 1 Myr time interval, variations with time in the process of crustal formation at the ridge, or both could account for this significant change.

2. On average, sub-sediment crust in this region is abnormally thick compared to crustal sequences from other studies near spreading centres. This may be a result of the complex interaction of the small and young Juan de Fuca and Explorer plates with the larger and older America and Pacific plates.

3. The velocity-depth curves interpreted for the different profiles are consistent in general terms with the model of oceanic crust represented by ophiolite complexes. Variations in detail exist, as would be expected for materials formed in different regions.

4. Velocity anisotropy within the upper mantle is found. The velocity varies from 7.5 km/s in a direction parallel to the ridge to 8.3 km/s in a direction approximately perpendicular to it, giving a
10% anisotropic effect. This concurs with previous studies in the region.

(5) Seismic energy is attenuated by the Nootka fault zone. The three profiles which cross the fault zone show a noticeable decrease in the amplitude of seismic phases at distances corresponding to ray paths traversing it.

In the following chapter, it will be shown that some of the findings based on the P wave interpretation will be substantiated by the independent interpretation of the shear wave data.
4. INTERPRETATION OF SHEAR WAVE DATA

4.1 Introduction

The importance of shear wave information in determining the physical properties and mineralogy of the oceanic crust has been propounded by a number of authors (Christensen and Salisbury, 1975; Hyndman, 1979; and Spudich and Orcutt, 1980a). Still, very little is known about the S wave velocity structure of the oceanic crust and upper mantle, despite the fact that ocean bottom seismometers have been in widespread use for refraction studies in recent years. The first part of this chapter outlines some of the problems in observing shear wave arrivals in marine refraction surveys and discusses methods for improving the signal quality of these arrivals.

Out of the five refraction profiles in this study, two have recorded useful shear wave arrivals. In the second part of this chapter, the interpretation of these two S wave profiles is discussed. It is shown that the results of the shear wave interpretations complement those of the P waves presented in the previous chapter. On the other three profiles, shear arrivals can be observed; however, due to either poor arrival picks or erratic amplitude behaviour of the S phases, no interpretation of the refracted S waves is attempted. The next section discusses some of the factors that affect the amplitude of the converted S waves.
4.2 Shear Wave Conversion

In marine refraction experiments, only compressional energy is generated in the water. For shear waves to be observed, the P waves from the source must be converted to S mode at some boundary in the media. The most likely location in the oceanic crust for this conversion to take place is at the water-sediment or the sediment-basement interface where the contrast in seismic properties across the boundary is large. However, the efficiency of conversion and transmission of shear waves in the crust is highly variable, leading to the unpredictable nature of the S wave arrivals. Spudich and Orcutt (1980a) and White and Stephen (1980) have discussed in detail some of the factors affecting the conversion of P to S waves. A summary of the more important aspects is given in the following.

(a) If Poisson's ratio of the sediment is high, such as in unconsolidated marine sediment, there will be very little conversion of P to S waves at the water-sediment interface. However, the conversion will be strong at the sediment-basement interface (Spudich and Helmberger, 1979).

(b) Efficiency of conversion is reduced if the basement Poisson's ratio is high. Weathering and cracks can increase the Poisson's ratio of the igneous basement; therefore lateral variation in weathering or non-uniform distribution of cracks can cause drastic variations in the amplitude of the converted S waves (White and Stephen, 1980).
(c) Phase coherency of the converted S waves is affected by rough topography on the basement interface, especially if the scale of the topographical variation is in the order of a seismic wavelength (~300 m). Even in regions where seismic velocity contrast is favourable, i.e. Poisson's ratio being high in the sediment and low in the basement, rough topography can cause poor S arrivals (Spudich and Orcutt, 1980a).

(d) Conversion efficiency is strongly dependent on the incident phase velocity, which is a function of angle of incidence, as well as the basement P velocity. For phase velocity equal to the basement P velocity, no conversion will take place. Conversion remains low for phase velocity higher than the basement P velocity (White and Stephen, 1980).

(e) If the P velocity of the sediment exceeds the S velocity of the basement, no conversion will take place (Spudich and Helmberger, 1979).

Since converted S waves are observed in all of the profiles in this study, this last remark probably does not apply to this area. It remains to be determined where the P and S conversion takes place. Two sets of evidence indicate that the main refracted S arrivals are converted from the down-going P waves at the sediment-basement interface. First, if the conversion had taken place at the water-sediment interface, the intercept time of the refracted S arrivals would be 1.5 s later than
that observed due to the low S velocity of the sediment ($\sim 0.5 \text{ km/s}$). Second, it is shown in section 4.4 that Poisson's ratio of the sediment layer in this region is high ($> 0.45$) which implies that strong conversion of P to S waves is to be expected from the sediment-basement interface with little or no conversion occurring at the seafloor.

Counteracting the favourably high Poisson's ratio of the sedimentary layer is the rough topography of the basement in the study area. Large scale ($\sim 1 \text{ km}$) variations in the basement depth have been shown in Chapter 2 to have caused large travel time anomalies in the P waves. Detailed examinations of the reflection profiles of Davis and Lister (1977) and Hyndman et al. (1979) reveal many small scale ($\sim 300 \text{ m}$) basement topographical disturbances, especially in the vicinity of the Nootka fault zone. This could explain the poor S wave arrivals in some of the profiles, although except for large features such as the Nootka fault zone, detailed correlation of topographical features with the converted S wave amplitude variation is difficult to make.

4.3 Optimization of Shear Wave Observations

4.3.1 Shear Wave Enhancement

Where three component seismogram data are available, seismic arrivals can be enhanced by the use of polarization filters. Cheung (1978) described in detail a particular application of such filters to OBS refraction data to improve the S wave arrivals. However, due to the
large uncertainties in the estimation of the emerging angle and the azimuth of the seismic phases, the improvement in data quality was limited. Only one out of four sections that he processed benefited slightly from the application of the two polarization filters. No further attempt is made in this study to utilize such filters.

However, some improvement in signal-to-noise ratio for the S wave data can be achieved by simply constructing SV (vertically polarized shear wave) seismograms from the two horizontal components. Figure 4.1 illustrates the projection of rectilinear SV motion on the horizontal plane where X and Y are the two orthogonal horizontal components. Knowing the angle \( \alpha \), SV(t) may be obtained from X(t) and Y(t) as

\[
SV(t) = \frac{X(t)}{\cos(\alpha)} + \frac{Y(t)}{\sin(\alpha)} / 2 \tag{4.1}
\]

In this process, the coherent signal is expected to be reinforced while any random noise will be cancelled out. This operation can also be regarded as a rotation of the horizontal seismograms into the radial direction of motion. It has been assumed that the vertical component of the SV motion is negligible due to the near vertical incident angles which are generally less than 10° from vertical.

In order to find the angle \( \alpha \), the orientation of the horizontal seismometers in the free fall OBS's must first be determined. This may be achieved by using the relative amplitude of the two horizontal seismograms for a refracted P arrival. This approach, however, has two disadvantages: the refracted ray path is strongly affected by near-
Figure 4.1. The horizontal plane projection of SV motion. $\alpha$ is the angle that the incident plane makes with respect to the $X$ component.
surface inhomogeneities and the near-vertical P arrivals have very small horizontal components of motion. A better method is to use the relative amplitude of the direct water wave arrivals which are not affected by any lateral variations in the sub-bottom structures and usually have adequate amplitudes for reliable estimates of the seismometer orientations.

Figure 4.2 shows the particle motion on the horizontal plane for a number of direct water wave arrivals on OBS1. The angle is found by fitting a regression line through the locus of the particle motion for a one half second duration. When shots from various azimuths are used to check for the consistency of $\alpha$, a variance in the order of $5^\circ$ is found. This is taken to be the uncertainty in the estimate.

A set of seismograms for shot 23 of profile EX3 is shown in figure 4.3. The dashed lines denote a number of arrivals whose ray paths are shown schematically in figure 4.7. Trace (a) of figure 4.3 is the vertical component while traces (b) and (c) are the horizontal components, Y and X respectively. Trace (d) is produced from (b) and (c) via equation (4.1) and it represents the radial component of motion. Two distinctive S arrivals can be seen on the horizontal seismograms. As expected, these arrivals register very small amplitudes on the vertical component seismogram.

The main S phase, PSS, originates from conversion of P to S waves at the sediment-crust interface below the shot point while the PPS phase travels as a P wave through most of its travel path and undergoes a conversion to an S wave at the sediment-crust interface immediately
Figure 4.2. Horizontal particle motion plots of four direct water wave arrivals. On each plot, the '+' symbols define the regression line fitted to the locus of the motion. The orientation of the horizontal seismometers is determined from the angle that this line makes with respect to the X axis.
Figure 4.3. Trace (a) is the vertical component while traces (b) and (c) are the horizontal components of a 30 kg shot recorded at a distance of 25 km. Trace (d) is the radial component of motion constructed from traces (b) and (c). Trace (e) is the bandpass filtered version of (d). Bandpass frequency limits are 2 and 10 Hz.
below the receiving OBS. Both S arrivals are slightly enhanced in trace (d), although the improvement is not very obvious due to the already high signal-to-noise ratio of the original data.

In contrast, figure 4.4 shows a set of seismograms which have a higher noise level than the previous example. Here, the main PSS phase is clearly seen in the SV seismogram of (d); whereas, the same arrival on the two horizontal traces is interfered with by the wave coda of earlier arrivals.

These two examples illustrate that the rotational process described by equation (4.1) can provide some enhancement of the shear arrivals. It should be noted that in practice, if the angle is near either 0 or 90°, as in the case of profile EX2, equation (4.1) is not used. Instead, the horizontal component with the larger amplitudes is considered as the SV trace and no rotation is necessary.

Further improvement in the appearance of the seismogram can be obtained by bandpass filtering. Trace (e) on both figures 4.3 and 4.4 are the SV seismograms bandpass filtered from 2 to 10 Hz. Figure 4.5 shows the relative Fourier power spectra of the two S phases and it is evident that they have well-defined peaks in the frequency band so that filtering would help to reduce the noise. The waveforms in the filtered traces are smoothed and most of the high frequency noise has been removed. This will facilitate the comparison with synthetic seismograms for amplitude interpretation. However, as in the case of the P wave data, bandpass filtering usually does not help in the picking of the shear wave arrival times.
Figure 4.4. Same description as figure 4.3 for a 5 kg shot recorded at a distance of 45 km. The main PSS phase shows clearly in (d) and (e), whereas the arrival about 0.4 s earlier interferes with PSS in (b) and (c).
Figure 4.5. Relative Fourier power spectra of the PSS and PPS phases (see trace (c) of figure 4.3). Spectra are calculated from signal durations of 1 second.
4.3.2 Three Dimensional Particle Motion Section

One of the major difficulties in observing shear waves in marine refraction experiments is the fact that the refracted shear waves often arrive in coincidence with some of the P phases, most noticeable of which are the water multiples of the main refracted P wave. This makes it difficult to identify the onset of the S arrivals. In order to discriminate between the various arrivals, it is necessary to compare arrivals on the horizontal component section with those on the vertical component. This can be a tedious process.

A possible solution proposed here is to construct a three dimensional particle motion section on which information from the three components are displayed simultaneously. An example of such a section is given in figure 4.6 for profile EX3. In isotropic media, particle motion is restricted to the incident plane; therefore, only two components, the vertical and the radial, are needed to define the particle motion. The construction of the radial component from the two orthogonal horizontal seismograms has been described in section 4.3.1. In figure 4.6, the time axis is horizontal and increasing to the right. Vertical motions are plotted up and down with respect to this axis while horizontal motions are plotted at a perspective of 45° with respect to this axis.

Some of the major arrivals are labelled on the section and the ray paths corresponding to these arrivals are shown in figure 4.7. Identification of the various phases is facilitated in two ways by the particle motion section. First, distinction between compressional and shear waves can be made by considering the sense of motion of an
Figure 4.6. Three dimensional particle motion section for profile EX3. The reduced travel time axis is horizontal and increasing to the right. Vertical motions are plotted up and down while horizontal motions are plotted at a perspective of 45° with respect to the time axis. The distance axis is also aligned at 45° with respect to the time axis. Topographical corrections to the travel times are calculated using the phase velocities of the PSS arrivals. See figure 4.7 for definition of the ray paths of the various phases.
Figure 4.7. Schematic diagram showing the ray path of the various seismic phases that can be seen in the section of figure 4.5.
arrival, i.e. motion would be predominantly vertical for P waves versus the predominantly horizontal motion for S waves. Second, due to the perspective plot, the coherency between traces is enhanced so that the differences in the phase velocities of the arrivals are more obvious. This serves to further identify the different phases.

One good example of these two points is illustrated by the converted shear phase PPS. This phase lags behind the first P arrival by an almost constant amount of time which is equal to the difference in the one way travel time through the sediment for P and S modes. The PPS arrivals have the same phase velocity as the P, yet the particle motion is clearly horizontal, identifying them as shear waves. The PPS phases are important in defining the Poisson's ratio of the sedimentary layer. This will be discussed in the next section.

4.4 Poisson's Ratio of the Sediments

It will be shown in this section that an accurate estimate of Poisson's ratio in the sediment layer can be obtained from the travel times of the PPS phases, without explicitly calculating the P and S velocities. This new procedure assumes constant velocities within the sediment and the Poisson's ratio thus determined is an average value.

Consider that the sedimentary layer beneath the OBS has a thickness of h and that the P and S velocities, \(v_p\) and \(v_s\), are constant within the layer. Let \(t_p\) and \(t_s\) be the one way vertical travel times through the sediment for the P and S waves respectively. We have
\[ t_p = \frac{h}{v_p} \]
\[ t_s = \frac{h}{v_s} \]

If we define \( \Delta t = t_s - t_p \)

then \( \frac{v_p}{v_s} = \frac{t_s}{t_p} = 1 + \Delta t/t_p \)

(4.2)

Now Poisson's ratio in terms of \( \frac{v_p}{v_s} \) is

\[ \sigma = \frac{(\frac{v_p}{v_s})^2 - 2}{2(\frac{v_p}{v_s})^2 - 2} \]

(4.3)

Substituting (4.2) into (4.3) we obtain

\[ \sigma = \left( 1 - \frac{\Delta t}{t_p} \right) / 2 \]

(4.4)

where \( \frac{\Delta t}{t_p} = \left\{ 2\left( \frac{t}{t_p}\right) + \left( \frac{t}{t_p}\right)^2 \right\}^{-1} \)

(4.5)

So Poisson's ratio of the sediment can be calculated by knowing only the values of \( \Delta t \) and \( t_p \). If we assume that the P and PPS phases have vertical travel paths, then \( \Delta t \) can be easily measured from the travel time lag between the two phases. The error introduced by this assumption is in the order of 1\% which is negligible compared to the uncertainties of other factors. The variable \( t_p \) in equation (4.5) can be obtained from the reflection profiles over the OBS site, and in this study, the values of \( t_p \) for the three sites are interpreted from the reflection profiles of Davis and Lister (1977) and Hyndman et al. (1979).

Table 4.1 tabulates Poisson's ratio for the sedimentary layer
beneath the three OBS sites, calculated with the value of $\Delta t$ and $t_p$ given in the same table. $N$ denotes the number of observations averaged to give $\Delta t$.

**TABLE 4.1 Poisson's ratio of the sediment layer.**

<table>
<thead>
<tr>
<th>OBS</th>
<th>N</th>
<th>$\Delta t$ (s)</th>
<th>$t_p$ (s)</th>
<th>Poisson's Ratio</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>15</td>
<td>$1.50 \pm 0.01$</td>
<td>$0.67 \pm 0.05$</td>
<td>$0.45 \pm 0.01$</td>
</tr>
<tr>
<td>2</td>
<td>7</td>
<td>$1.22 \pm 0.01$</td>
<td>$0.44 \pm 0.05$</td>
<td>$0.46 \pm 0.01$</td>
</tr>
<tr>
<td>3</td>
<td>15</td>
<td>$1.42 \pm 0.01$</td>
<td>$0.56 \pm 0.05$</td>
<td>$0.46 \pm 0.01$</td>
</tr>
</tbody>
</table>

The values of $\sigma$ given in Table 4.1 for the three sites are similar and are consistently high. This indicates that the sediment in this region is uniform and poorly consolidated, a fact also verified by other studies in the area. For example, Cheung (1978) found values of $\sigma$ in the neighbourhood of 0.49 for the sediment column in the Explorer Ridge region. Based on the finding of low $P$ velocities within the sediment layer from an OBS refraction survey, Davis et al. (1976) concluded that the sediment in the Juan de Fuca Ridge region is only slightly compacted. Rapid sedimentation rates (up to 1 cm/yr) were given as the cause of the poor consolidation. So for this near coastal region of the Pacific Ocean, high values of $\sigma$ are to be expected.
4.5 Shear Wave Record Sections and Amplitude Modelling

Despite the considerable efforts made in enhancing the shear wave arrivals, only two out of the five profiles give sufficient data for the interpretation of the PSS phases. The SV record sections of profiles EX3 and EX2 are plotted with a reducing velocity of 4 km/s in figures 4.8a and 4.9a. The timings of the traces have been corrected for topographical variations using the phase velocity of the PSS phases derived from the slope of the uncorrected travel time curves. Amplitude normalization for the different charge sizes and the application of an $r^2$ spreading factor follow the same procedures as those discussed for the P wave record sections. The data traces are bandpass filtered between 2 and 10 Hz. The choice of frequency limits lower than those chosen for the P wave data stems from the fact that the S arrivals generally have lower frequency content than the P waves.

By glancing at the two record sections of figures 4.8a and 4.9a, one can easily detect the obvious difference in their amplitude behaviour; namely, the large amplitude arrivals associated with the near critical reflection from the gradient zone immediately above the crust-upper mantle boundary are found at a distance range of 45 km for profile EX3 versus 23 km for profile EX2. This difference is a consequence of the different crustal thicknesses, already detected by the P wave data, for the two profiles. A similar contrast in the amplitude characteristics of the two profiles can be seen also in the P wave record sections (figures 3.3a and 3.5a).

Other points to be noted about the S wave record sections are the
Figure 4.8. (a) SV record section and (b) best fit synthetic seismogram section for profile EX3. Synthetic seismograms are calculated using the S velocity-depth model of figure 4.10a.
Figure 4.9. (a) SV record section and (b) best fit synthetic seismogram section for profile EX2. Synthetic seismograms are calculated using the S velocity-depth model of figure 4.10b.
emergent character of the first breaks, and the low amplitudes of the arrivals at near distances and at distances beyond the cross-over point of the Sn phases. The emergent first breaks introduce greater uncertainties into the picking of the S arrivals. Travel time errors are in the order of ±0.08 for the S waves versus ±0.06 s for the P waves.

The lack of energy for the near distance arrivals means that the airgun profiles, so vitally important in providing the constraints on the upper crustal structures, are unfortunately useless for shear wave interpretation. Additional complication for the observation of near distance S arrivals is the interference from the large amplitude water waves which arrive in approximately the same time window.

Having illustrated some of the limitations of the S wave data, interpretation of the two profiles is presented in the following. Inversion of the travel time data, using the \( \tau(p) \) function, proceeds in the same way as for the P waves. Due to the greater uncertainty in the travel times, the extremal bounds calculated for the S velocity-depth functions are even more lacking in resolving power than the P velocity bounds. It is found that constraints on the S velocity structure come mainly from the amplitude data; therefore, the discussion that follows will concentrate on the modelling of the S wave amplitude, using synthetic seismograms calculated by the WKBJ algorithm.

The WKBJ algorithm allows inter-mode conversion at any boundary of the model so that the PSS phases can be modelled directly both in travel time and amplitude. This is particularly important since the conversion of P to S at the sediment-basement interface is strongly dependent on
the phase velocity and the velocity contrast present.

Certain assumptions concerning the S velocity structure at shallow depths have to be made since no refracted S arrivals are observed from these depth regions. A good estimate of the S velocity in the sediment comes from the PPS phases (see discussion in the previous section). With a Poisson's ratio of 0.45, the P velocity of 1.8 km/s in the sediment would give 0.5 km/s for the S waves. For this reason, the sediment layer is constrained to have a constant velocity of 0.5 km/s and a uniform thickness of 1.0 km.

The existence of converted shear waves places the lower limit of the basement S velocity at 1.8 km/s, equal to the sediment P velocity. Basement S velocity of 2.2 km/s is chosen to give reasonable amplitudes for the observed PSS phases, which are partially controlled by the velocity contrast at the sediment-basement interface. This value of basement S velocity is similar to those found for the Gaudalupe site by Spudich and Orcutt (1980a).

With the exception of the constraints mentioned above for the shallow structures, the S wave data are initially modelled independently of the P wave data. The trial-and-error approach employed in modelling the P wave amplitude data is used also for the S waves. It is found from the independent modelling that the interfaces marking zones of different velocity gradients coincide to within 200 m with those found for the P velocity-depth functions. This difference is insignificant considering that the typical wavelength of crustal shear waves is approximately 0.8 km, so it is deemed justifiable to constrain the interfaces to be at
the same depths for the ease of interpreting the results.

The best fit synthetic seismogram sections are plotted below the data sections in figures 4.8b and 4.9b. As with the P wave presentation, the travel time curves from the synthetic sections are transferred onto the data section. Generally, the synthetic record sections give excellent fit to the observations. An exception is found in profile EX3 where the traces at distances beyond 48 km show a sudden decrease in amplitude which is not modelled adequately by the WKBJ seismograms. This phenomenon is similar to that observed in the P wave record section of profile EX3 and it lends support to the inferred lateral variation in crustal thickness at approximately 45 km distance range, as discussed in the previous chapter.

4.6 Results and Discussion

In this section, the shear wave velocity-depth models for profiles EX2 and EX3, interpreted by modelling the travel time and amplitude data with the WKBJ synthetic sections, are discussed. The relationship between the seismic velocity of oceanic crust and the lithology of ophiolite complexes, expounded upon in detail in section 3.5, is re-examined here in light of the added information from the shear wave data.

Figure 4.10 shows the final $v_p$ and $v_s$ models for profiles EX2 and EX3. The zones of differing velocity gradients, more obvious in the $v_p$ curves, are also evident in the $v_s$ curves. A schematic representation of the velocity structures, similar to figure 3.12 of the last chapter but
Figure 4.10: $v_p$ (solid line) and $v_s$ (dashed line) models for profiles EX2 and EX3.
Figure 4.11. Schematic representation of the velocity-depth models of figure 4.10. The depth is measured from the seafloor. The numbers are the velocity at the top of each layer in km/s. Layer 1, the sediment layer, is constrained to have the given velocities and a thickness of 1 km.
with $v_S$ information added, is shown in Figure 4.11 to facilitate the discussion of the results. In Figure 4.12, the apparent Poisson's ratio versus depth for the two profiles are plotted. The values of Poisson's ratio are only apparent because the relationship between $\sigma$ and $v_p/v_S$ (Equation 4.3) used in the calculation is appropriate only for isotropic media. In the last chapter, an anisotropic effect of 10% is found for the upper mantle $P$ velocity; therefore, in the neighbourhood of upper mantle depths, values of Poisson's ratio in Figure 4.12 are for the purpose of comparison only, and they do not represent the actual elastic parameter at these depths. Further discussion of velocity anisotropy will be given later.

Spudich and Orcutt (1980a) have suggested a method of displaying seismic velocity models in which the $v_p(z)$ and $v_S(z)$ are plotted as a continuous path through the $v_p$-$v_S$ plane, parameterized by depth $z$. The main advantage of this presentation is that results from laboratory samples can be conveniently plotted on the same diagram. In Figure 4.13 the $v_p$-$v_S$ paths of EX3 and EX2, with subsediment depths labelled along the paths at 1 km intervals, are shown. Paths of constant Poisson's ratio which form straight lines in the $v_p$-$v_S$ plane also are plotted. An enlarged section of the two paths is shown in Figure 4.14 along with some laboratory measurements of ophiolite samples. The laboratory data are compiled by Spudich and Orcutt (1980a) who have adjusted the original measurements of Christensen (1978) and Salisbury (1974) on ophiolite samples to pressure conditions appropriate for comparison with their seismic models interpreted for the Gaudalupe site. These same laboratory data can be used for comparison with the results of the
Figure 4.12. Apparent Poisson's ratio versus depth for profiles EX2 and EX3 calculated from the velocity-depth models of figure 4.10. The question mark emphasizes the uncertainty of the estimate of Poisson's ratio in layer 2A. The symbol 'M' denotes the crust-mantle interface.
Figure 4.13. $v_p$-$v_S$ paths for EX2 and EX3. Subsediment depths in 1 km intervals are marked along the paths with numbers above the paths for EX3 and numbers below the paths for EX2. Paths of constant Poisson's ratio, which form straight lines in the $v_p$-$v_S$ plane, are plotted at intervals of 0.05.
Figure 4.14. $v_p$-$v_S$ paths for EX2 and EX3 compared with ophiolite sample velocities. The rock types are: E - epidote-amphibolite facies metadolerite and metagabbro, S - partially serpentized peridotite, G - greenschist facies metadolerite, P - pyroxene gabbro, O - olivine gabbro, T - trondhjemite, and M - metabasalt.
Figure 4.15. A comparison of EX2 and EX3 with the seismic models, FF2 and FF4, of Spudich and Orcutt (1980a).
present study since the small changes in velocity ($\sim 0.01$ km/s for $V_p$) due to the slight variations in depths between the seismic models are negligible. Similarities between the seismic models interpreted by Spudich and Orcutt for the Gaudalupe site (FF2 and FF4) and profiles EX2, EX3 can be seen in figure 4.15 where the $V_p$-$V_s$ paths coincide in the $V_p$ range of 6.0 to 7.0 km/s. In the following, a discussion of the seismic results of the present study and their relationship with the laboratory results is given in order of increasing depths.

Layer 2A

There are no direct constraints on the $S$ velocities in layer 2A (1.0 to 2.0 km sub-bottom depth). The choice of 2.2 km/s for $V_s$ (see figure 4.11) at the top of this layer is based on the fact that it gives reasonable conversion of P to S energy at this interface. By a similar argument, Spudich and Orcutt (1980a) also interpreted a basement $V_s$ of 2.2 km/s for the Gaudalupe site. However, changing the value of basement $V_s$ by $\pm 0.3$ km/s produces only minor differences in the amplitudes of the PSS phases.

Since no strong subcritical reflections are observed from the top of this layer, the existence of an $S$ velocity gradient within the layer is required. However, the exact nature of this gradient is not defined by the data so that the $V_s$ models for this depth region are constructed in a rather ad hoc manner to give reasonable intercept time for the deeper arrivals. Due to the uncertainty of $V_s$, the value of Poisson's ratio in this layer is not reliable; a question mark is placed on figure
4.12 to emphasize this uncertainty.

The inference that this layer is composed of fractured basalt flows and pillows is consistent with the $v_S$ results. Typical laboratory measurements of basalt samples dredged from the seafloor give values of $v_S$ in the neighbourhood of 2.5 km/s (Hyndman and Drury, 1976). The presence of large scale cracks and voids in the in situ oceanic crust can be expected to reduce $v_S$ to the order of 2.2 km/s.

Layer 2B

In the transition zone from layer 2A to 2B, both $v_P$ and $v_S$ increase rapidly with depth. This increase in velocities is thought to correspond to the change from pillow and flow basalt to greenschist facies metabasalt and brecciated dikes found in the upper level of the Bay of Islands ophiolite complex (Christensen and Salisbury, 1978; table 3.1).

Within layer 2B, the most striking aspect of the seismic results is the very low values of Poisson's ratio which are determined there (see figure 4.12 for example). Upon comparing the $v_P$-$v_S$ paths with the results from ophiolite samples (figure 4.14), we see that the correspondence between the $v_P$-$v_S$ paths and the metabasalt samples is only marginal, since the values of $\sigma$ for the metabasalts are in the neighbourhood of 0.28 while EX2 and EX3 have values of $\sigma$ near 0.25. Only towards the lower depths of layer 2B is a rock type, the trondhjemites from the Bay of Islands ophiolite complex, found that has similar $v_P$-$v_S$ values. This is rather surprising since trondhjemites are seldom found in ocean bottom dredge samples, and in the ophiolite suites,
trondhjemites are usually found beneath the sheeted dikes, at depths well below that of layer 2B.

Spudich and Orcutt (1980a) also found low values of Poisson's ratio ($\sigma < 0.27$) at depths of 0.8 to 1.5 km subsediment which they were not able to explain. They suggested that the observed low values of $\sigma$ could be due to the presence of pore fluid; however, they have done theoretical calculations which indicated that pore fluid effects are not sufficient to lower Poisson's ratio from approximately 0.28 measured in laboratory samples to a value of 0.24 observed in layer 2B. The results remain unexplained.

Layer 3A

In the transition from layer 2B to 3A, the $v_p$-$v_S$ paths in figure 4.14 move into accordance with the laboratory measurements. Although there is some overlapping in the laboratory results for the greenschist facies and epidote-amphibolite facies metadolerites, a transition in the $v_p$-$v_S$ plane between the two groups can be seen in the vicinity of $v_p=6.6$ km/s and $v_S=3.6$ km/s. The trajectory of the $v_p$-$v_S$ paths for both EX2 and EX3 in this range of velocities supports the hypothesis that the interface between layer 2B and 3A marks a metamorphic boundary between the greenschist facies and epidote-amphibolite facies metadolerites (Stern et al., 1976).

Velocity gradients within layer 3A are much more moderate than those at shallower depths. Poisson's ratio, however, increases with depth to a value of approximately 0.29 as the $v_p$-$v_S$ paths move from the
epidote-amphibolite facies metadolerite and metagabbro to pyroxene gabbro, then to olivine gabbro (figure 4.14). The good agreement between the velocities of the lab samples and the observed seismic results allows us to infer the composition of layer 3A from the lithological sequence of the Bay of Islands ophiolite complexes, where the corresponding layer 3A consists of metadolerite sheeted dikes underlain by a layer of intrusive metagabbro and gabbro, followed by a layer of pyroxene gabbro (table 3.1). For this particular study site on the Juan de Fuca plate, the correspondence between the in situ oceanic layer 3A and the lithology of the Bay of Islands ophiolites is excellent.

Layer 3B and Upper Mantle

In layer 3B, velocities increase from those of the base of layer 3A to those of the upper mantle, at velocity gradients that are higher than those found in layer 3A. There are no sharp discontinuities in either $v$ or $v$ at the crust-mantle interface. The increase in velocities with depth in this layer is accompanied by a decrease in the apparent Poisson's ratio (figure 4.13). These are the only common features of the lower crust between profiles EX2 and EX3.

In the results displayed in figures 4.10 to 4.14, a number of disparities between the two profiles, at depths near the crust-mantle boundary, can be detected. The most obvious one is the 4.6 km difference in the interpreted upper mantle depths. This aspect has been discussed in detail in the last chapter and it suffices here to say that the independent interpretation of S wave data confirms the large variation
in crustal thicknesses between EX2 and EX3.

The other major dissimilarity between the profiles is the difference in the values of apparent Poisson’s ratio at upper mantle depths (see figure 4.12). The term ‘apparent’ has to be stressed here since, as mentioned earlier, equation 4.3 used in the calculation of σ is appropriate only for isotropic media. Effects of anisotropy for upper mantle P waves is significantly large (10%) but the interpreted upper mantle shear wave velocities, 4.5 and 4.6 km/s respectively for EX2 and EX3, give an anisotropic effect of only 2%. This combination of high degree of P wave anisotropy and low degree of S wave anisotropy results in the unusually low apparent σ of 0.22 for profile EX2, versus the more common value of 0.28 for profile EX3.

The results given so far are based on the best fit synthetic seismogram modelling of the data. It is reasonable at this point to ask the question of how significant is the observed difference between the two apparent Poisson’s ratios. If we consider the extremal bounds calculated from the travel time data, the resulting bounds on σ are certainly broad enough to encompass both values of apparent σ. All this means is that travel time data alone can not resolve the difference in σ. Modelling the amplitude data, on the other hand, involves a trial-and-error procedure so that a rigorous answer can not be given to the above question. However, a numerical experiment discussed below will shed some light on the possible uncertainties of the S wave velocity models which are expected to dominate the errors in the calculation of σ.
Since the Sn head waves have extremely small amplitudes, their arrival times do not provide sufficient constraints to the upper mantle $v_S$. The principal features of the data set which allow the upper mantle S wave velocity to be determined with adequate resolution are the relative amplitudes and positions of the near-critical-angle reflections from the crust-upper mantle interface. Figure 4.16 shows three synthetic seismogram sections calculated by the WKBJ method and the three velocity-depth curves used for the calculation, for profile EX2. The best-fit section is produced with an upper mantle S-wave velocity of 4.53 km/s (model b). The synthetic seismograms give less satisfactory fits (Figure 4.16a, c) when the upper mantle velocity deviates by 0.2 km/s from the preferred value, a fact demonstrated more clearly by the plots of amplitude versus distance in Figure 4.16e. These observations indicate that the probable error in the estimate of the upper mantle S wave velocity is in the order of $\pm 0.1$ km/s. Uncertainties in the upper mantle $v_p$ are somewhat less than that of $v_S$, but even assuming errors of $\pm 0.1$ km/s for both $v_p$ and $v_S$, the difference in $\sigma$ of 0.06 is detectable and can not be attributed wholly to uncertainties in the interpretational techniques.

Apparent Poisson's ratio of the two profiles not only differs for the upper mantle, but from figure 4.13, we see that the $v_P-v_S$ paths actually start to diverge within layer 3B, indicating that the effects of anisotropy are present, but to a smaller degree, in the lower crust also. This may have important implications for the possible cause of anisotropy in the oceanic upper mantle. Further discussion of this
Figure 4.16. (a), (b), and (c) Synthetic seismogram sections modelling the near-critical-reflections from the crust-mantle interface, calculated from the respective velocity models of (f). Model b is the preferred model. (d) Relevant portion of the data record section of profile EX2. (e) Plot of relative amplitude versus distance for the three model synthetic sections, compared with the data. The numbers denote the upper mantle S velocity of the given velocity model.
aspect is given in the next chapter on anisotropy.

In summarizing the seismic results of layer 3B and the upper mantle, it is noted here that any postulated petrological model of the oceanic crust for this study site would have to be consistent with the following observations: (1) velocities in layer 3B increase with depth at a relatively pronounced gradient; (2) no sharp velocity discontinuities are present at the crust-upper mantle interface; (3) Poisson's ratio decreases with depth in layer 3B; and (4) anisotropy, mainly affecting the P wave velocities, is present in the upper mantle and possibly in the lower half of layer 3B.

Again, the ophiolite analogy provides a plausible model to account for the above observations. Corresponding to layer 3B, the zone of interlayered olivine gabbro, troctolite, and peridotite found in the Bay of Islands ophiolite complex (table 3.1) has the required medium velocities ($v_p \sim 7.4$ km/s, $v_s \sim 3.9$ km/s) compared to those observed in layer 3B. The velocity gradient and decreasing $o$ in this layer can be explained by an increasing olivine content with depth. A gradual transition from gabbroic rocks to ultramafic rocks, similar to that observed in the Bay of Islands ophiolite would correlate well with the crust-upper mantle interface observed here. Laboratory derived velocities of the ophiolite ultramafic samples at 25°C are 8.4 and 4.9 km/s respectively for $v_p$ and $v_s$ which are slightly too high, but higher temperatures (~200°C) at depths corresponding to the crust-upper mantle interface are likely to reduce the laboratory determined velocities to those observed for the upper mantle. Preferential
alignment of the olivine crystals, on the other hand, may cause the observed anisotropy in layer 3B and the upper mantle. This point is discussed further in the next chapter.

4.7 Concluding Remarks

In terms of structural definition, the shear wave data is found to have less resolution than the P wave data. This is especially true for shallow depths where little is known about the shear wave velocities from the present refraction techniques. Nonetheless, the interpretation of the shear wave data is able to confirm the variation in crustal thicknesses, interpreted previously from the P wave data, between profiles EX2 and EX3. This adds support to the inferred tectonic processes discussed earlier in Chapter 3.

More important is the fact that the added $v_s$ information has greatly reduced the ambiguities in the interpretation of the petrology of the oceanic crust. For example, the velocity structures of profiles EX2 and EX3 appear to be quite different at first glance due to the varying thicknesses of various layers. Their $v_p$-$v_s$ paths, however, show remarkable similarities over much of the crust, implying that the layers have similar composition even though their thicknesses vary. The obvious disparities in the lower crust and upper mantle are explained by invoking velocity anisotropy.

It is shown that the ophiolite model of the oceanic crust and upper mantle is consistent with the results presented here. Variations exist, as exemplified by the low Poisson's ratio in layer 2B which still lacks
satisfactory explanation. Spudich and Orcutt (1980a) suggested that the low $\omega$ in the upper crust is a phenomenon related to young oceanic crust. The finding of low Poisson's ratio in layer 2B in this study supports their conjecture.
5. AREAL TRAVEL TIME DATA AND VELOCITY ANISOTROPY

5.1 Introduction

Compressional wave anisotropy in the oceanic upper mantle has been well documented by numerous marine refraction studies (Raitt et al., 1969; Keen and Barrett, 1971; Snydsman et al., 1975; and Malecek and Clowes, 1978). In all cases, the direction of maximum P-wave velocity is approximately parallel to the direction of sea-floor spreading as inferred from magnetic anomaly patterns. In the present study, a similar anisotropic effect is indicated by the interpretation of the profile data. The principal explanation for this anisotropy is preferential alignment of the a-axis of olivine crystals in the direction of flow during formation at the ridge crest (Hess, 1964). That olivine is seismically anisotropic with the direction of maximum compressional velocity parallel to the a-axis is well established by laboratory experiments (Verma, 1960; and Birch, 1960, 1961).

In this chapter, further evidence for P wave velocity anisotropy in the upper mantle is given. The low degree of shear wave anisotropy, discussed in Chapter 4, is related to two recent laboratory studies on seismic anisotropy of ophiolite samples. Implications for seafloor spreading processes are also discussed.
5.2 Areal Data

Although the present experiment was not designed for the purpose of a study of velocity anisotropy, the shot-receiver configuration (figure 2.1) does provide some useful travel time data for examining the velocity variation as a function of the azimuthal angle. This is illustrated in figure 5.1 where the possible distribution of shot-receiver distances and azimuths is shown. It is clear from this figure that for distances over 30 km, which correspond to upper mantle arrivals, the data cover a relatively wide range of azimuthal angles. This indicates that it may be feasible to obtain information concerning the azimuthal variation of velocity in the upper mantle from the 'off-profile' travel time data.

There are, however, several problems related to the interpretation of such travel time data; the most important of which is the effects of lateral variations in structure which can also produce azimuthally varying apparent velocities. To overcome the effects of large scale variation in crustal thicknesses, the data are first partitioned judiciously into groups according to the different crustal regimes defined broadly by the profile interpretations (see Chapter 3). Within each group, the variation in the shot-receiver azimuth is limited to less than 25°. After correcting the travel times for topography, an apparent velocity is calculated for each group and this velocity is assigned to the average azimuth of the group. Of course, only upper mantle arrivals are selected for the calculation.

Effects of unknown local variations near the surface at the shot or
Figure 5.1. Distance versus azimuth for all possible arrivals given the shot-receiver configuration of figure 2.1. Azimuths are measured in degrees from North.
receiver sites cannot be readily accounted for. It is assumed that contaminations by such effects are random and can be minimized by calculating the apparent velocity using a sufficiently large number of travel time points. This requirement and the limit placed on the range of azimuths within the groups are mutually incompatible so that compromises are often necessary.

Given the above limitations, only a small number of apparent velocities of the upper mantle are obtained from the P wave data and no useful data is available from the S wave data due to poor arrival picks. The azimuthally varying Pn velocities are plotted in figure 5.2 where Pn velocities from the profile interpretations are also included. In this presentation, the azimuth is measured with respect to the direction of spreading, as defined by the magnetic pattern based on the Hyndman et al. (1979) model (figure 3.13), at the midpoint of a given shot-receiver path. The scatter in the data is considerable; however, the existence of a velocity maximum near 0° azimuth and a minimum near 90° azimuth is definitely indicated. The interpretation of this data set is discussed in the following section.

5.3 Interpretation of Areal Data

Backus (1965) has shown that for weakly anisotropic media (< 10% anisotropic effect), the square of the P wave velocity as a function of azimuth is given by

\[ v^2(\theta) = A + B \cos(2\theta) + C \cos(4\theta) + D \sin(2\theta) + E \sin(4\theta) \] (5.1)
Figure 5.2. Apparent upper mantle P velocities as a function of azimuth, determined from areally distributed data, are plotted as solid circles. Velocities determined from the profiles, with smaller estimated errors, are plotted as solid squares. Errors in velocities and azimuths are subjective estimates based on the reliability of individual data sets. The azimuths are measured in degrees from the seafloor spreading directions inferred from the tectonic model of Hyndman et al. (1979). The solid line is a least-squares-fit of the data to equation (5.2). ORTHOR represents the velocity variation of a mixture of 29% orthorhombic olivine and an isotropic material with a velocity of 7.2 km/s. TRANSV represents the velocity variation of a mixture of 37% transversely isotropic olivine and an isotropic material with a velocity of 6.8 km/s.
where the coefficients $A$ to $E$ are linear combinations of six elastic constants (see Appendix 2). If the anisotropic media contains at least one vertical symmetry plane, which is a reasonable assumption for the upper mantle where anisotropy is caused by the alignment of the olivine crystals, equation (5.1) reduces to

$$v^2(\theta) = A + B \cos(2\theta) + C \cos(4\theta)$$  \hspace{1cm} (5.2)

where $-\theta$ is measured from the direction possessing sagittal symmetry (Crampin, 1977). For the case where the $a$-axis of the olivine crystal is aligned in the direction of seafloor spreading and the $b$- and $c$-axes are randomly oriented, the direction of sagittal symmetry is the spreading direction.

If the functional form of equation (5.2) is assumed for a given set of velocity versus azimuth data, the three coefficients can be obtained by a least-squares fitting procedure. This is done for the $Pn$ velocity data shown in figure 5.2. The solid curve (LSQ) represents the least-squares-fit solution and is given by

$$v^2(\theta) = 62.28 + 5.53 \cos(2\theta) - 2.77 \cos(4\theta)$$  \hspace{1cm} (5.3)

The coefficients $A$, $B$, and $C$ thus obtained do not allow immediate identification of the composition of the media since the resulting elastic constants are only the apparent values of the parameters. Crampin and Bamford (1977) suggested a possible interpretational technique in which the upper mantle is assumed to be a mixture of
isotropic material and a likely anisotropic material such as olivine. The ratio of these two components and the parameters of the isotropic material can then be determined from the coefficients of equation (5.2) using the elastic constants of the assumed anisotropic material (see Appendix 2). Such information can be valuable for the interpretation of the petrology of the upper mantle.

Using the results of the least-squares-fit equation (5.3), the velocity variations of two mixtures are calculated and are shown in figure 5.2. If the anisotropic material is assumed to be orthorhombic olivine, it would constitute 29% of the mixture and the isotropic portion would have a velocity of 7.2 km/s. If the anisotropic material is assumed to be transversely isotropic olivine, a more likely candidate for the constituent of the upper mantle than orthorhombic olivine, it would make up 37% of the mixture, giving a velocity of 6.8 km/s for the isotropic portion. The differences in the velocity variations of the two mixtures are slight (figure 5.2) which implies that with this type of data, the two different olivine fabrics are probably indistinguishable.

However, it is evident from figure 5.2 that the velocity variations of both mixtures differ significantly from that exhibited by the data. There are a number of possible causes for the misfit and they are discussed in the next section. For the moment, one particular aspect of the misfit is addressed.

Given a two-component mixture of the type described above, equation (5.2) predicts that the maximum of the velocity should occur at $0^\circ$ azimuth while the minimum should occur at $90^\circ$ since the coefficient of
the \( \cos(4\theta) \) term is positive for the two olivines. The mixture model curves derived from the coefficients of the least-squares-fit equation should, and do, coincide with the LSQ curve at these two values of azimuth. However, the maximum of the LSQ curve is not located at 0° but rather at approximately 30°. This suggests that a more reasonable approach in modelling the data is to abandon the least-squares-fit curve and instead, take the well-determined maximum and minimum velocities (8.3 and 7.5 km/s respectively) from the profile interpretations to be the end points of the model curves.

The results of the second modelling approach are shown in figure 5.3. The velocity variations of the two model mixtures are now more in line with the trend of the data. The orthorhombic olivine mixture consists of 33% anisotropic material and the isotropic material has a velocity of 7.4 km/s. For the transversely isotropic olivine mixture, the percentage of anisotropic material is 42% while the velocity of the isotropic material is 7.0 km/s.

5.4 Discussion

The results obtained from the interpretation of the P wave velocity anisotropy data, using the relatively simple techniques described in the last section, can only be regarded as a rough indication of the possible composition likely to be found in the upper mantle. It is important to stress the fact that seismic waves actually sample the media over several wavelengths so that the value of the velocity is only an average over a distance of several kilometers; whereas, the proposed mixtures
Figure 5.3. Description is the same as figure 5.2 except ORTHOR now represents the velocity variation of a mixture of 33% orthorhombic olivine and an isotropic material with a velocity of 7.4 km/s, and
TRANSV represents the velocity variation of a mixture of 42%
transversely isotropic olivine and an isotropic material with a velocity of 7.0 km/s.
are modelled after rock crystals much smaller in dimension and are assumed to have idealized configurations. The real earth is probably much more complex and equation (5.2) is only an approximation.

Complication also arises from the inherent uncertainties in the data. For example, the determination of the azimuthal angle is far from precise due to the large uncertainties in the inferred direction of seafloor spreading and the fact that the velocity calculations are done over a finite range of azimuths. Even if the orientations of the ray paths with respect to the spreading direction are known exactly, the alignment of the olivine crystals in the upper mantle is not necessarily perfect. Given these considerations and the sparsity of data, more sophisticated modelling than what has been done is probably not warranted.

Before discussing in more detail the anisotropy results of this and the previous two chapters, a brief summary of the observations is given in the following. For the upper mantle, P wave velocity anisotropy is pronounced with velocities of approximately 8.3 and 7.5 km/s parallel and perpendicular, respectively, to the direction of seafloor spreading. The S wave velocities, 4.6 and 4.5 km/s, determined from two shear wave profiles are nearly isotropic. This results in a prominent apparent anisotropy of Poisson's ratio for the upper mantle, with values ranging from 0.22 to 0.28, the high value being in the direction of plate motion. The upper mantle P wave anisotropy can be crudely modelled by a mixture of 42% transversely isotropic olivine and an isotropic material with a velocity of 7.0 km/s. There are hints that the lower half of
layer 3B also possesses velocity anisotropy of the type exhibited by the upper mantle, but the magnitude is smaller.

If the hypothesis that ophiolites represent obducted fragments of oceanic lithosphere is to be accepted, then evidence from laboratory studies of ophiolite samples must be consistent with the observations of anisotropy in the oceanic crust and upper mantle reported here and elsewhere. Christensen and Salisbury (1979) studied the petrofabrics of fifteen ultramafic samples consisting entirely of tectonites from three widely spaced traverses in the Bay of Islands, Newfoundland ophiolite complex. Their comprehensive analysis yielded a compressional wave anisotropy of 6% (percentage defined by \((1 - V_{\text{min}}/V_{\text{max}}) \times 100\)) with the maximum velocity of 8.7 km/s in the plane of the Mohorovicic discontinuity and parallel to the inferred spreading direction, and the minimum velocity of 8.2 km/s perpendicular to this direction (table 3.1). For shear waves, for which the propagation velocity depends on vibration direction resulting in two different S-wave velocities, the results yielded nearly isotropic S wave velocities for both polarizations of the shear wave (average \(v_s = 4.9 \text{ km/s}\)). These velocity values were calculated for a temperature of 25°C and for appropriate upper mantle pressures. As pointed out by Christensen and Salisbury (1979), increases in temperature of 200°C will lower the velocities into good agreement with observed oceanic upper mantle velocity with little effect on anisotropy. They found Poisson's ratio ranging from 0.23 to 0.27 with the high values generally in the direction of spreading. The agreement between the results of the present study and those of Christensen and Salisbury (1979) is excellent.
In another study, an harzburgite sample from the Antalya ophiolite complex in Turkey was analysed in detail (Peselnick and Nicolas, 1978). Calculations of P wave velocities for the uppermost mantle at $250^\circ$C and 5 kbar, based on the elastic constants obtained from the sample, gave values of 8.46 km/s in a direction normal to the inferred ridge crest and 8.16 km/s parallel to it. The 3.5% anisotropy of 0.3 km/s agrees with some in-situ experiments (Raitt et al., 1969), but is less than the anisotropy usually determined in marine refraction studies. For the two modes of shear wave propagation, anisotropic effects of 2.3% and 1.6% were found.

Although there are differences between these two ophiolite studies, the general characteristics of anisotropic effects, such as the directions of maximum and minimum velocity with respect to spreading direction and the higher anisotropic effect for P waves compared with S waves, are shared by the two sets of results. These characteristics are also consistent with the results determined in the present study. This furnishes yet another positive argument for the ophiolite hypothesis.

A puzzling aspect of the present results is the anisotropy observed in the lower half of layer 3B where the primary composition is thought to be cumulate olivine gabbro (Christensen and Salisbury, 1975) which generally is seismically isotropic. Christensen (1972) suggested that dipping of amphibolite foliation planes in the lower crust may cause some of the observed velocity variations interpreted by Keen and Barrett (1971) for that part of the crust. However, it is not clear what
mechanism would cause dipping of the foliation plane of the amphibolite.

Another possible explanation for the anisotropy observed in layer 3B is given, again, by the ophiolite samples. Christensen and Salisbury (1979) found that for the gabbros immediately overlying the ultramafics of the Bay of Islands ophiolites, the olivine fabric is identical with that in the ultramafic tectonites, indicating that tectonization has extended beyond the ultramafics into the overlying gabbros. This reorientation of olivine originally formed by cumulus processes may be caused by translation gliding presumably induced by the same stress field that is responsible for the anisotropy in the upper mantle. The fact that the orientation of olivine is the same in the lower crust as it is in the upper mantle would suggest the possible existence of anisotropy in the lower crust of the same type as observed for the upper mantle but the lower olivine content of the lower crust would imply a lower degree of anisotropic effect. To further substantiate this assertion would require a more extensive data set than that available in the present study.

In conclusion, ophiolites provide a number of explanations for the observed anisotropy in the oceanic crust and upper mantle, as discussed above.
6. SUMMARY

The first objective of this study is to investigate the crustal structure in the region of the Nootka fault zone through the use of seismic refraction methods. This has been accomplished and the interpretations of the data show remarkably detailed changes in crustal properties with depth and significant lateral variations in structure.

The results in this study indicate that layers 2A and 2B in the upper crust are characterized by rapid increases in seismic velocities. For example, P wave velocity increases from a range of 3.7 to 4.5 km/s at the top of layer 2A, to values of 6.0 to 6.4 km/s at the base of layer 2B. Similarly, shear wave velocity increases from 2.2 to 3.6 km/s over the same depth interval of approximately 3.5 km; although the exact nature of the shear wave velocity gradient is uncertain.

In contrast, velocities in layer 3A are relatively uniform, with $v_p$ lying in the range of 6.4 to 6.8 km/s and $v_s$ remaining fairly constant with depth at values near 3.6 km/s. No velocity discontinuity is observed at the crust-upper mantle interface. Instead, the transition from crust to upper mantle is marked by the presence of a velocity gradient zone, layer 3B, in which the velocities increase gradually from those at the base of layer 3A to those of the upper mantle. P wave velocity in the upper mantle varies significantly with azimuth from 8.3 km/s in a direction parallel to the inferred direction of seafloor spreading to 7.5 km/s in a direction parallel to the spreading ridge. However, corresponding variations in the observed upper mantle shear wave velocities, 4.6 and 4.5 km/s, are small and lie within the
uncertainties of $\pm 0.1 \text{ km/s}$ in the $\text{Sn}$ velocity determinations.

The interpretation of shear wave data presented in this study provides the first in situ evidence for a low degree of shear wave velocity anisotropy in the upper mantle, a phenomenon which has been suggested by laboratory studies of ophiolite ultramafic samples. On the other hand, upper mantle $P$ wave velocity exhibits a 10\% anisotropic effect, resulting in a prominent anisotropy in the values of Poisson's ratio for the upper mantle. These aspects of velocity anisotropy can be verified by experiments suitably designed for the study of anisotropy in the crust and upper mantle. Details of the crustal structure outlined in this study should provide relevant guidelines for designing such an experiment for this region.

The most notable feature of the crustal structure in this region is the sudden increase in crustal thickness from 6.4 to 11.2 km over a lateral distance of approximately 30 km. A rapid process of 'crustal maturing' within a 1 Myr time interval, variations with time in the process of crustal formation at the ridge, or both could account for this significant variation in structure. Such an abrupt change in crustal thickness should show clearly on a gravity anomaly map, but unfortunately, the existing gravity data for the west coast of Canada do not cover this region. Thus an independent check on this aspect of the interpreted results could be provided by a gravity survey over the area.

The second objective of this study is to relate the seismic results to the petrology of the oceanic crust and upper mantle. In this respect, a convenient unifying theme is supplied by the ophiolite hypothesis for
the oceanic lithosphere. Correlations between the seismic results and the properties of ophiolite complexes, in particular the Bay of Islands ophiolites, consistently surface throughout the study. For example, the observed velocities and gentle velocity gradients in layer 3A agree very well with laboratory measurements of seismic velocities for the sequence of metadolerite sheeted dikes, metagabbros, and pyroxene gabbros from the Bay of Islands ophiolites. Below the pyroxene gabbros, the lithology of this ophiolite complex shows increasing content of pyroxene and olivine as troctolites and olivine gabbros become more abundant with depth. This has its correspondence in the more pronounced velocity gradient in layer 3B compared with layer 3A as interpreted in this study. Yet another correlation between the properties of the ophiolites and the seismic characteristics of the oceanic upper mantle is provided by the observation of velocity anisotropy. Petrofabric studies of ultramafic samples from ophiolites indicate that the P wave velocity in the upper mantle would exhibit azimuthal variations, with maximum and minimum velocities in directions parallel and perpendicular to the directions of plate motion, respectively. The petrofabric studies also show that the corresponding S wave velocity is only mildly anisotropic. These results are consistent with those found in this study.

The correspondence between data measured from ophiolites and the observations in this study provide support for the supposition that ophiolites are segments of oceanic crust and upper mantle emplaced on land. By comparing the seismic results with the lithology of ophiolites, the enormous non-uniqueness in the petrological interpretation of seismic velocities is thus reduced, so that inferences on the petrology
of the oceanic crust and upper mantle are possible. More detailed
determination of the lithological sequences of the in-situ oceanic crust
will require high quality P and S wave data recorded on three-component
instruments with good amplitude control. Shot spacings of 1 km or less
will be needed to provide the dense data coverage necessary for the
observation of subtle amplitude variations with distance. For studies of
velocity anisotropy, adequate azimuthal distribution of ray paths will
be essential.


Wilson, J. T., 1965. Transform faults, ocean ridges and magnetic anomalies southwest of Vancouver Island, Science, 150, 482-485.
APPENDIX 1: Linearized Inversion of Tau-P Data

Assuming that the velocity is a monotonically increasing function of depth, the integral expression for $\tau$ is

$$\tau(p) = 2 \int_{z}^{z(p)} (u^2 - p^2)^{1/2} \, dz \quad (A.1.1)$$

where $u(z) = 1/v(z)$

It was shown by Garmany (1979) that by changing the dependent variable to $Z=Z(u)$ and defining the model to be $m(u)=dZ/du$, a linear relationship exists between $\tau(p)$ and $m(u)$. Equation (A.1) becomes

$$\tau(p) = \int_{0}^{u_{max}} m(u) \cdot h(u,p) \, du \quad (A.1.2)$$

where $m(u) = dZ/du$

$$u_{max} = 1/v(z=0)$$

and

$$h(u,p) = \begin{cases} 
-2(u^2 - p^2)^{1/2} & u > p \\
0 & u \leq p 
\end{cases}$$

If we partition $(0, u_{max})$ into $M$ intervals and assume that $m_i=dZ/du$ is constant within each interval $(u_{i-1}, u_i)$, we can rewrite (A.1.2) as

$$\tau(p) = \sum a_i m_i \quad (A.1.3)$$
where \[ a_i = \int_{u_{i-1}}^{u_i} h(u, p) \, du \]

with \[ u_0 = 0 \]

Given \( N \) discrete data points \((\tau_j, p_j), j=1,N \), with error \( \Delta \tau_j \), then (A.1.3) becomes the \( 2N \) inequalities:

\[
\tau_j - \Delta \tau_j \leq \sum a_i m_i \quad j=1,N
\]

\[
\tau_j + \Delta \tau_j \geq \sum a_i m_i
\]

from which the \( m_i \)'s can be easily found using linear programming techniques. The model \( m(u) \) is then integrated to give \( Z(u) \) which in turn is transformed to \( V(z) \) (LINP curve in figure 3.11d). The approach taken in this study is to find the \( m_i \)'s by minimizing the depth to the highest velocity subjected to the constraints of (A.1.4). This is slightly different from the approach of Garmany et al. (1979) who obtained the bounds on the \( V(z) \) function by alternately maximizing and minimizing the depth to each velocity interval instead of finding a particular set of \( m_i \)'s which satisfies the constraints (A.1.4).
APPENDIX 2: Inversion of P Wave Anisotropy

Backus (1965) has shown that for weakly anisotropic media (< 10% anisotropic effect), the square of the P wave velocity as a function of azimuth is given by

\[ \varphi v^2(\theta) = \left\{ 3C_{1111} + 3C_{2222} + 2(C_{1122} + 2C_{1212}) \right\}/8 + (C_{1111} + C_{2222})/2 \cos(2\theta) + (C_{2111} + C_{1222})/2 \sin(2\theta) + \{C_{1111} + C_{2222} - 2(C_{1122} + 2C_{1212})\}/8 \cos(4\theta) + (C_{2111} - C_{1222})/2 \sin(4\theta) \]  

(A.2.1)

where \( \theta \) is measured from the \( x_1 \) axis. If the media contains at least one vertical plane of symmetry, Crampin (1977) has shown that equation (A.2.1) can be reduced to:

\[ \varphi v^2(\theta) = \left\{ 3C_{1111} + 3C_{2222} + 2(C_{1122} + 2C_{1212}) \right\}/8 + (C_{1111} + C_{2222})/2 \cos(2\theta) + \{C_{1111} + C_{2222} - 2(C_{1122} + 2C_{1212})\}/8 \cos(4\theta) \]  

(A.2.2)

where \( \theta \) is measured from the direction possessing sagittal symmetry. Rewriting equation (A.2.2) as

\[ v^2(\theta) = A + B \cos(2\theta) + C \cos(4\theta) \]  

(A.2.3)

then from equations (A.2.2) and (A.2.3) we have
\[ v_1^2(\Theta=0) = A + B + C = C_{1111}/\varphi \]  \hspace{1cm} (A.2.4)

\[ v_2^2(\Theta=\pi/2) = A - B + C = C_{2222}/\varphi \]  \hspace{1cm} (A.2.5)

Given a set of velocity versus azimuth data, a least-squares fitting of the data to the functional form of equation (A.2.3) would yield the coefficients \( A, B, \) and \( C \). Following Crampin and Bamford (1977), if we assume that the observed velocity variation is due to a mixture of anisotropic material with the following velocity behaviour

\[ v_A^2(\Theta) = A' + B' \cos(2\Theta) + C' \cos(4\Theta) \]  \hspace{1cm} (A.2.6)

and an isotropic material with unknown constant velocity \( v_I \) in the ratio of \( x/(1-x) \) (anisotropic/isotropic), the velocity of the mixture is then given by

\[ v^2(\Theta) = x v_A^2(\Theta) + (1-x) v_I^2 \]  \hspace{1cm} (A.2.7)

Equating this to the least-squares-fit equation (A.2.3) and solving for \( x \) and \( v_I^2 \) we have

\[ x = \frac{2B}{2B'} \]  \hspace{1cm} (A.2.8)

and

\[ v_I^2 = A + B + C - x (A' + B' + C') / (1-x) \]  \hspace{1cm} (A.2.9)

where \( A', B', \) and \( C' \) are the appropriate coefficients for the anisotropic material and are assumed to be known from laboratory
measurements and A, B, and C are the coefficients from the least-squares-fit equation. In terms of $v_1$ and $v_2$ we have

$$ x = \frac{(v_1^2 - v_2^2)}{2B'} \quad \text{(A.2.10)} $$

and

$$ v_T^2 = \frac{v_1^2 - x (A' + B' + C')}{(1-x)} \quad \text{(A.2.11)} $$

For convenience, the density has been assumed to be the same in all cases for the above derivation, although this is not necessary. Using the results of Verma (1960), Crampin and Bamford (1977) calculated $A'=75.256$, $B'=18.953$, and $C'=3.262$ for orthorhombic olivine and $A'=79.012$, $B'=15.117$, and $C'=3.342$ for transversely isotropic olivine (with velocities measured in units of km/s). These are the values used in Chapter 5 for the calculation of $x$ and $v_T$. 