Ophiolites, synthetic seismograms, and oceanic crustal structure

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I would like to thank my wife Bonnie for her support, patience, and friendship which has grown through the years. I dedicate this dissertation to her and our children Joni and Casey.
ABSTRACT

Synthetic seismograms for five ophiolites have been generated using the full reflectivity method. Comparison of the synthetic seismograms from the Samail ophiolite, Oman, (5-8 m.y. age) with OBS refraction data acquired during the ROSE Experiment suggests that the Samail ophiolite is an excellent model for 4.5 m.y. old oceanic crust. Similarly, seismic refraction data from 60 m.y. old crust in the northeast Pacific is in good agreement with synthetic seismograms of the Bay of Islands ophiolite suite (40-45 m.y. age). Using these results it is suggested that that an aging process exists whereby young crust with slow varying seismic velocity gradients is altered to crustal sections with higher seismic velocity gradients which, at seismic frequencies, can be fit by discrete layer velocity-depth models. In particular, older crustal sections have a seismically distinct "layer 3" composed of 3 sublayers. Layer 3a has a P-velocity of 6.7 km/sec and corresponds to rocks of metagabbro composition. The Bay of Islands ophiolite petrology suggests that the 7.1 km/sec horizon (layer 3b) is composed of cumulate pyroxene gabbros. The 7.4 km/sec layer 3c horizon represents the more ultramafic section of the cumulates at the base of the ophiolite. Serpentinization of the olivine component of the lower crust, and the appearance of a P-low-velocity zone at the extreme base of the crust is suggested by the seismic observations. Constraints have been placed on the crust-mantle transition for older crust through detailed comparison of
ophiolite models and seismic observations. The seismic structure of
the crust and the resolution of these sublayers is particularly
dependent on the shot spacing, frequency bandwidth, and experimental
design. Modeling of the Troodos complex, the Vourinos ophiolite, and
the Papua ophiolite show that the velocity-depth functions for some
ophiolite models are not capable of producing synthetic seismograms
that are "oceanic" in seismic character. This implies that all
ophiolites cannot be considered as "normal" oceanic crustal sections.
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This dissertation is a composite of three papers co-authored with J. Gettrust. In these instances it is customary to explain the contributions of each author. The initial ideas for the dissertation developed from my previous work on the Point Sal ophiolite in California and the inversion techniques learned during the analysis of seismic refraction data at the Hawaii Institute of Geophysics (HIG). I formulated and carried out the basic structure of the dissertation; J. Gettrust was instrumental in every phase of the project, particularly because of his knowledge in applied seismology and his accessibility to computer software, but more important was his guidance, support, and direction. In a project like this it is very hard to evaluate the contribution of each individual because the complexity and dimension of the problem is beyond the limitation of a single person. Without the cooperation of both people this project could not have been completed.
CHAPTER I

INTRODUCTION

The term ophiolite refers to a distinct assemblage of ultramafic and mafic rocks including peridotite, gabbro, diabase, and spilite, which recently, has taken on implications of a genetic relation to oceanic crust. Many lines of evidence suggest that ophiolites should be considered representative allochthonous sections of oceanic crust tectonically emplaced along modern and ancient plate boundaries. But there is little agreement on whether these petrologically distinct suites are generated in back-arc basin environments or at "typical" oceanic spreading centers. This brings us back to question the original assumption. Do ophiolites represent oceanic crust?

Most comparisons of ophiolites and oceanic crust are based on indirect comparisons of laboratory compressional ($V_p$) and shear ($V_s$) velocity measurements of ophiolite samples with marine seismic refraction results. Although most ophiolites are generally thinner than oceanic crustal sections, the indirect evidence shows that the simple layered velocity models for the oceanic crust are generally compatible with the velocity structure imposed on ophiolite sections. The fallacy in this argument lies in the fact that a similar velocity-depth function does not necessarily imply that the ophiolite section would produce a seismic refraction record comparable in travel time or energy distribution to data obtained from seismic refraction experiments from oceanic crust.
With the advent of high-speed computer technology synthetic seismograms can now be generated for ophiolite velocity-depth models. This provides a direct method for comparing ophiolites to oceanic crust. The purpose of this dissertation is to directly compare oceanic crustal refraction data with synthetic seismograms for detailed ophiolite models and test the hypothesis that some ophiolites may represent oceanic crustal sections.

Synthetic seismograms for five ophiolites have been generated using the reflectivity method of Fuchs and Muller (1971) and Kennett (1975). These ophiolites include: (1) the Samail ophiolite in Oman; (2) the Bay of Islands complex (BOI), in Newfoundland; (3) the Vourinos complex of northern Greece; (4) the Troodos ophiolite in Cyprus; and (5) the Papua complex of New Guinea.

The details and/or comparisons of the synthetic seismograms with marine refraction data are described in three chapters.

Chapter II compares ocean bottom seismometer (OBS) refraction data for crust of 60 m.y. age with six synthetic seismograms generated from the velocity depth model of the Bay of Islands ophiolite. This ophiolite represents a petrologic suite with a formation to emplacement age of approximately 45 m.y. The comparison of the OBS data and the synthetic models supports the hypothesis that this ophiolite represents a mature oceanic crustal section. The modeling also provides an explanation for the variability found in previous refraction interpretations for mature oceanic crust. In addition, a new crustal
interpretation for the lower crust is developed and constraints are placed on the degree serpentinization and alteration of the lower most crust and upper mantle.

In Chapter III the synthetic seismograms for the young Samail ophiolite section (formation to emplacement age of 5 to 8 m.y.) are directly compared to the OBS refraction data for crust of 0.5 to 4.5 m.y. age obtained in the Rivera Ocean Seismic Experiment (ROSE). The correlation of the synthetic seismograms and the refraction data, again supports the conclusion that some ophiolites may be representative sections of oceanic crust. Furthermore, it is possible, with the results from both Chapter II and III to address the problems concerning the aging processes that alter the ocean P-velocity structure. A new model is proposed to explain the variation in oceanic crustal structure with age.

The synthetic seismograms presented in Chapter IV for the remaining three ophiolites: Vourinos, Troodos and the Papuan ophiolite, are based on estimated velocity-depth profiles (Christensen, 1978). The synthetics do not warrant a direct comparison with marine refraction data because of the speculative nature of the data. But they do offer insight into the problems in inversion of marine refraction data, as well as a basis for further modeling of ophiolites. It can be concluded from the modeling of these three sections that all ophiolites cannot generate synthetic seismograms that are "oceanic" in seismic characteristics. This suggests that either the assumptions for the modeling are incorrect, or that not all ophiolites are representative
sections of obducted "oceanic" crust formed at spreading ridges.

Chapter V further discusses the assumptions upon which this dissertation is based and implications for the origin of ophiolites as suggested by modeling all 5 distinctly different ophiolites.
CHAPTER II
COMPARISON OF OBS DATA AND SYNTHETIC SEISMOGRAMS
FOR THE BAY OF ISLANDS OPHIOLITE

Introduction

The acceptance of ophiolites as obducted sections of oceanic crust is demonstrated by the application of the ophiolite model to the study of oceanic crustal dynamics, physical properties, thermal and petrologic structure, mineralization, and marine magnetic anomaly sources [Cann, 1968, 1974; Coleman, 1971, 1977; Aumento, 1972; Vine and Moores, 1972; Banerjee, 1973; Wenner and Taylor, 1973; Moores and Jackson, 1974; Bonatti and Zerbi, 1976; DeWit and Stern, 1976; Salisbury and Christensen, 1978; Manghnani et al., 1981; and others]. Ophiolites should give us a framework within which to integrate these data from the various disciplines. However when oceanic crustal models based on marine seismic refraction analysis are compared with laboratory velocity-depth models for ophiolites, the evidence suggests that many ophiolites are anomalous both in thickness and physical properties [Moores and Jackson, 1974; Christensen and Salisbury, 1975]. Ophiolites are generally thought to be too thin; further, many lack a 3 km thick basal section of rocks with velocities of 7.2-7.6 km/sec often reported in airgun-sonobouy crustal refraction (ASPER) interpretations [Maynard, 1970; Sutton et al., 1971; Hussong, 1972]. The objective of this study is to directly compare synthetic seismograms based on the physical properties of ophiolites with good-quality marine seismic refraction
observations. It is clear that an ophiolite complex must be carefully chosen if it is to be used to test ocean crust models. Factors that must be considered include spreading rates, the age of the section, and the extent of deformation that occurred during the emplacement of the ophiolite. These data are available for very few ophiolites; the Bay of Islands ophiolite complex appears to fit parameters for Pacific Ocean basin crust formed at a moderately-fast (3 to 5 cm/yr) spreading center. It is important to note that we must assume that modification of the oceanic crust ceases when the crustal section is obducted [Dewit and Stern, 1976; Salisbury and Christensen, 1978]. Further evidence that the crustal velocity structure can be "frozen" following obduction is given by Kempner and Gettrust [1981] who show that the Oman ophiolite which has a young (5 to 8 m.y.) formation to emplacement age is seismically similar to young (4.5 m.y.) oceanic crust but is not compatible with mature oceanic crust. Specifically, synthetic seismograms based on the Bay of Islands ophiolite are used to reinterpret OBS refraction profiles from the northeast Pacific that sample oceanic crust of approximately similar crustal age. We emphasize the use of the synthetic models for ophiolites as a guideline to interpret seismic refraction data and thereby to account for the variations reported for "layer 3". Because of the varied interpretations of the lower crustal structure based on seismic refraction data and the importance of these interpretations to the acceptance of ophiolites as representative sections of oceanic crust, we will focus our attention directly on the lower crust.
Synthetic seismograms, based on six variations of the velocity-depth models for the Bay of Islands (BOI) ophiolite sequence [Salisbury and Christensen, 1978], have been generated for this study. Using these models we can determine the effects of the variations in the seismic velocity structure on the travel-time curves and amplitude distributions of the synthetic seismograms; these synthetic seismograms are directly compared with the seismic refraction observations to constrain the interpretation of the data. The models allow us to test for the presence of a low velocity zone associated with the late stage differentiates in the upper crust, and serpentinization of the basal cumulates. The hypothesis of a thick, high-velocity basal layer is also examined and constraints are placed on the structure of the lower crust. The results of this comparison of synthetic seismograms for the BOI ophiolite with the OBS refraction data suggests that the lower crust is composed of 3 "sublayers". The resolution of these zones is dependent upon the signal to noise ratio of the data, the shot spacing, and the inversion technique employed. The following analysis accounts for most of the seismic refraction interpretations of old oceanic basin crust without any need for great lateral inhomogeneity in apparent composition and thickness. Our results suggest that the most prominent refracting horizon associated with "layer 3" has a P-velocity of 6.7 km/sec, similar to that reported by Raitt [1963] and Shor et al. [1970]. The apparent thickness of this layer will vary depending on the quality and density of the data. Interpretation of observed amplitudes shows that there is no support for a low velocity zone associated with late stage
differentiates within the 6.7 km/sec horizon. Second arrivals define a second refraction horizon, with a P-velocity of 7.1 km/sec, that lies below the 6.7 km/sec refractor. This refractor may be responsible for most of the variations found in seismic interpretations of layer 3. Finally a third sublayer, a basal high P-velocity layer may be present. However, support for this layer which is identified using second arrivals is tenuous. Serpentinization of the olivine component of this basal layer is supported by this analysis. The P-velocity inversion associated with serpentinization appears to be limited to a thickness of approximately 1 km.

Previous Work

Although synthetic seismograms have been used extensively in the inversion of marine seismic refraction data [Helmberger and Morris, 1969, Lewis and Snoysman, 1977, 1979; Meeder et al., 1977; Helmberger and Engen, 1978; Whitmarsh, 1978; Fowler and Keen, 1979], the application of this technique to ophiolite models is relatively new. Spudich [1979] and Spudich and Orcutt [1980] used the BOI ophiolite to point out the general similarities between the synthetics for the BOI ophiolite and the Fanfare seismic refraction line obtained near the Guadalupe Island area in the Pacific. Both papers focus attention on shear wave information to determine the composition of the crustal section. Nichols et al. [1980] generated synthetic seismograms using the velocity-depth function of the Point Sal ophiolite, a section of the
California Coast Range ophiolite complex. Because this ophiolite sequence is too thin to fit acceptable oceanic crustal models, they added a 2 km thick serpentinized basal layer to the Point Sal data and inferred a process of continuous serpentinization of mantle to account for this amount of altered material. They did not test this model by comparing the synthetics with seismic refraction observations.

In the past, three basic models for mature ocean crust have been proposed. Each is based on a different refraction technique. These interpretations have been summarized by Christensen and Salisbury [1975] and are shown in Figure 1. A review of these models will be helpful for later discussion. The 3 layer model of Raitt [1963] consists of a 1.5 km thick layer 2 with a P-velocity of 5.13 km/sec overlying layer 3 which has a thickness of 4.97+1.25 km with a P-velocity of 6.73+.19 km/sec. This interpretation, referred to as the "standard crustal section", is usually based on seismic first-arrival data taken with shot spacing of 2.0-2.5 km or greater. We will refer to this model as the Type 1 section. By increasing shot density, using second arrival amplitude information, and extending marine seismic refraction profiles to greater distances, it becomes possible to refine the interpretation of layer 3 into a thin 6.4 km/sec refractor overlying a 5.0 km thick 7.1 km/sec refracting horizon. These crustal models are referred to as the Type 2 sections. Airgun source and sonobouy receivers profiles (ASPER) [Maynard, 1970; Sutton et al., 1971; and Hussong, 1972], have been used to interpret a unique layer 3 composed of a 6.8 km/sec refracting horizon overlying a basal layer with a P-velocity varying from 7.2 to
Figure 1  Comparison of velocity-depth structures for mature oceanic crust. Type 1 represents crustal refraction data from Raitt [1963] and summarized by Christensen and Salisbury [1975] for mature crust. Type 2 and Type 3 are interpreted from sonobuoy data. \(v_p\) and \(v_s\) represent compressional and shear velocities (km/sec), respectively; \(t\) is the thickness of each layer in kms; and ( ) is Poisson's ratio.
OCEANIC CRUSTAL STRUCTURE

**TYPE 1**

\[
\begin{align*}
V_p &= 2.20 \pm 0.31 \\
t &= 0.68 \pm 0.90 \\
V_p &= 5.04 \pm 0.69 \\
t &= 1.39 \pm 0.50 \\
V_p &= 6.73 \pm 0.19 \\
t &= 4.97 \pm 1.25 \\
V_p &= 8.15 \pm 0.31
\end{align*}
\]

**TYPE 2**

\[
\begin{align*}
V_p &= 2.0 \\
t &= 0.8 \\
V_p &= 3.7 \\
V_p &= 4.4 \\
V_p &= 5.8 \\
t &= 1.6 \\
V_p &= 6.4 \\
t &= 1.2 \\
V_p &= 7.1 \\
t &= 4.8
\end{align*}
\]

**TYPE 3**

\[
\begin{align*}
V_p &= 2.0 \\
t &= 0.8 \\
V_p &= 3.7 \\
V_p &= 4.4 \\
V_p &= 5.8 \\
t &= 1.6 \\
V_p &= 6.8 \\
V_s &= 3.75 \\
\sigma &= 0.29 \\
t &= 3.0 \\
V_p &= 7.5 \\
t &= 2.6 \\
V_p &= 8.3 \\
V_p &= 4.7 \\
\sigma &= 0.26
\end{align*}
\]

(AFTER CHRISTENSEN AND SALISBURY, 1975; RAITT, 1963)
7.6 km/sec. Both of these layers are approximately 3.0 km thick. This results in an increase in the average oceanic crustal thickness. The ASPER technique, which uses phase coherence without amplitude information often yields mantle P-velocities greater than 8.3 km/sec. We refer to this model as Type 3. See Christensen and Salisbury [1975] for an in depth comparison of the crustal types.

Our comparison of the ocean basin refraction data with the synthetics for the BOI ophiolite show that the apparent differences between these crustal models can be explained by differences in the data density and interpretative techniques.
Seismic Observations (The NEPAC Refraction Profiles)

The NEPAC OBS Refraction Experiment was carried out in the northeast Pacific in July 1978 by the Hawaii Institute of Geophysics aboard the R/V KANA KEOKI. As part of this experiment, sixteen OBSs were deployed at 25 km intervals between the Murray and Molokai Fracture Zones as a linear N-S array (Figure 2). Shots ranging in size from 15 to 120 lbs were fired along the array at approximately 1 km intervals. The linear array and shotline parallel anomaly 26 (60-62 m.y.). The apparent spreading rate at the time of formation for this crustal section, based on magnetic anomalies has been given as 4.9 cm/yr to 5.6 cm/yr [Malahoff and Handschumacher, 1971; Harrison and Sclater, 1972; Goslin et al., 1972]. The average water depth in the region is 5.0 km, with little topographic variation. A thin veneer of sediment, approximately 100 meters thick, has been inferred from the 3.5 kHz reflection records. Preliminary analysis of the OBS data indicated a "normal" 3-layer crust showing little inhomogeneity over the 400 km OBS array. Detailed interpretation of these data support a Type 2 structure with a 6.2-6.4 km/sec layer overlying a 7.1 km/sec section (Figure 3). The actual data for OBS "Locris" is presented in Figure 4. The presence of the 6.4 km/sec horizon of layer 3 is poorly supported by first-arrivals alone. However, the high-amplitude second-arrivals show significant coherence that can only be resolved because of the high shot density and good signal to noise ratio of the data from the hydrophone. This second arrival information allows us to discriminate the 6.4 km/sec refractor. The 7.1 km/sec refraction horizon is based on the
Figure 2  
(a) Location map for the NEPAC refraction experiment showing position and location of shotlines (solid lines) and shots greater than or equal to 600 lbs (dark circles). The magnetic anomalies are represented by the numbered dark bands.

(b) Enlargement of inset in (a) showing the details of the shotline (solid line) and the location of the OBSs. Solid circles represent OBSs that were retrieved; open circles are OBSs that were lost during the experiment.
Figure 3  P-Velocity depth models for 3 OBSs.
Figure 4  Comparison of the observed refraction data with the spherical earth raytrace interpretation and the synthetic seismogram. This illustration shows (a) a hydrophone record section for OBS "Locris" (south); (b) the same as (a) with the raytrace interpretation superimposed on the data; (c) synthetic seismogram computed using the velocity depth model from "Locris" (south). All sections are reduced to 8.0 km/sec. Amplitudes are scaled to maximum amplitude in each trace.
first-arrivals from that refractor observed between 15 and 30 km range, and by the large amplitude reflections observed between 40 and 65 km range. Comparison of these data with synthetic seismograms for this simple velocity-depth model (Figure 4a, b, and c) suggests a satisfactory inversion of the data which, in most cases, would be acceptable. But closer inspection of the record section reveals two details that warrant further modeling. These are: (1) at 29 to 31 kms, high-amplitude first-arrivals appear which cannot be justified as part of the refraction branch from the 6.7 km/sec refractor, and (2) the reflection branch from the 7.1 km/sec layer is abruptly offset at the 65 km range. While it is possible that these features are caused by lateral heterogeneity, it is reasonable to consider alternative mechanisms that do not require the assumption of a lateral change in structure. We will show that the "fine structure" of the seismic arrivals can be satisfied by a laterally homogeneous velocity-depth function that is consistent with the BOI ophiolite.

Modeling The Bay of Islands Ophiolite

The work of Salisbury and Christensen [1978] on the physical properties of the Bay of Islands ophiolite is an essential prerequisite to this study. Our work is a logical outgrowth of their efforts. Their complete velocity-density-depth model (Figure 5) of a nearly continuous ophiolite section, with the distinction of being similar in thickness (within one standard deviation) to standard oceanic crustal models,
Figure 5  Envelopes of compressional ($V_p$) and shear ($V_s$) velocity, and density ($\rho$) versus depth for the Bay of Islands ophiolite complex as described by Salisbury and Christensen [1978]. Heavy line represents the preferred fit; dashed line shows areas of discontinuous velocity inversions.
minimizes the need for assumptions concerning the continuity of the stratigraphic section. The authors do suggest that a few hundred meters of basalt may be missing from the uppermost section of the ophiolite. Jacobsen and Wasserburg [1980] have used Nd and Sm aging techniques to suggest an age of 45 m.y. from formation to emplacement for the BOI ophiolite. Under the assumption that the oceanic crustal structure does not change significantly after 30 m.y. [Goslin et al., 1972; Woollard, 1975; Trehu et al., 1976] comparison of BOI with the NEPAC refraction line (60–62 m.y.b.p.) is quite reasonable. Although no direct control for the spreading rate at the time of formation of the BOI ophiolite is available, we assume that the BOI ophiolite was formed at a ridge having a minimum spreading half-rate of 3.0 cm/yr based on the thickness of the dike and plutonic section [Kusznir, 1980]. Both Cann [1974] and Kusznir [1980] have pointed to this direct dependence on spreading rate of layer thickness for the intrusive as well as the plutonic structure of the crust.

A test of the assumptions concerning the age of the ophiolite as well as the spreading rate is shown in Figure 6. Here, the velocity-depth function for the BOI ophiolite has been plotted together with the limits of oceanic crust of 12 m.y. and 47 m.y. age taken from the Tau-P inversion of refraction data from the Pacific [Kennett et al., 1977]. Note that the ophiolite model fits only the bounds for the older crustal section. A comparison of the BOI velocity depth structure with the results from the preliminary initial inversion of the NEPAC data (Figure 5) also shows that there is a very strong similarity between the
Figure 6  The P-velocity-depth function from the OBS "Locris" (south) and the model for the Bay of Islands complex from Salisbury and Christensen [1978] are superimposed in (a). In (b) the Bay of Island P-Velocity model is shown superimposed on the P-velocity bounds for oceanic crust greater than 45 m.y. old. Similarly, (c) shows the ophiolite P-velocity-depth function superimposed on the P-velocity bounds for oceanic crust greater than 12 m.y. The observational bounds for oceanic crust are after Kennett et al. [1977].
Synthetic seismograms for the BOI ophiolite were generated using the reflectivity method of Fuchs and Muller [1971] and Kennett [1975]. Because of the difficulty in correcting for the large variations in shot size and automatic gain changing (AGC) within the OBS, the record sections are shown with each trace normalized to the maximum amplitude in that trace; the synthetic seismograms are plotted in a similar fashion.

Six synthetic seismograms have been generated using the velocity-depth function shown in Figure 6 as a starting point. Of these six models (Figures 7b, c, d; and 8b, c, d), five lie within the velocity bounds shown in Figure 6. Model 1 shows the $V_p$ velocity structure preferred by Salisbury and Christensen [1978]. Model 2 has been generated without the P-low-velocity zone in the upper part of layer 3 that represents the late-stage differentiates. Model 3 is similar to Model 2 with respect to crustal structure, but the mantle P-velocity has been reduced to 8.0 km/sec, as have all the mantle P-velocities for the remainder of the models. This is to insure a better fit to the NEPAC OBS data which has a mantle P-velocity of 8.0 km/sec.

In order to test for the serpentinization of the lower crust, Model 4 has a 1 km thick P low-velocity zone at the base of the crust which can be associated with the cumulate ultramafics. Model 5, on the other hand, places a continuously increasing P-velocity gradient at the base of the crustal section. The only model which does not fit within the
Figure 7  Record section for observations from OBS "Locris" (south) and synthetic seismograms generated using Models 1, 2 and 3 velocity depth functions are shown. None of the synthetic seismograms produce a completely satisfactory comparison with (a). Comparison of Models 1 and 2 (b and c) shows the effects of an upper crustal low velocity zone on arrival amplitudes in the synthetics. Similarly a comparison of Models 2 and 3 (c and d) shows the effects of mantle anisotropy on arrival amplitudes. Amplitudes are scaled to maximum amplitude in each trace. All sections are reduced over 8.0 km/sec.
Figure 8  Record section for observations from OBS "Locris" (south) and synthetic seismograms generated using Models 4, 5 and 6 velocity depth functions are shown. Model 4 in (b) shows many similar features in the amplitude distribution of arrivals. Model 6 has to be rejected because the reflection branch which dominated the second arrivals from 30 to 90 kms has a velocity of 7.4 km/sec which is not characteristic of the observations. All plots are scaled as in Figure 7.
BOI P-velocity bounds is Model 6. This model, with a thickened high P-velocity basal layer was generated to test the Type 3 (ASPER) crustal model. Models 1 and 2 are used to evaluate the upper crustal low velocity zone and Models 3 thru 6 are used to examine the lower crustal structure.

Comparison of the NEPAC Refraction Data with the BOI Synthetics

The first-arrivals for the Bay of Islands ophiolite (Models 3, 4, and 5) are near perfect fits to those for the "Locris" refraction line. This is shown in Figure 9 (a, b, and c) where the travel-time curve of ophiolite Model 3 is seen to be nearly identical to the first arrival picks of the "Locris" data. Using first-arrivals alone, the BOI synthetic models could not be distinguished from the "Locris" data. Amplitude information is the key to unravelling the details of the crustal structure.

In their description of the BOI section Salisbury and Christensen [1978] noted the possibility that the velocity structure for the uppermost basaltic layer may be too fast because their sampling technique cannot correct for tectonic fracturing. Field relations indicate that the stratigraphic sequence may also be missing roughly 400 meters of the uppermost basalt; this would further modify the upper P-velocity function. Because of these problems we will not be concerned with the short range seismic arrivals in the synthetic seismograms, but will begin our analysis of the arrivals associated with the base of
Figure 9  A typical travel time curve for the Bay of Islands ophiolite. The 6.7, 7.1 and 7.4 km/sec refractors and the multiple reflected-refractions (i.e., 2PmP) are shown in (b). Reinterpreted OBS refraction line in (c) shows a strong similarity to (b). First-arrivals travel-time paths are indistinguishable for the BOI data synthetics and the OBS refraction observations. Second arrivals show slight variation between the observations and the synthetic seismograms.
layer 2. We will show that only one of these models is consistent with
the NEPAC OBS data, and slight differences between synthetic seismograms
for our preferred crustal model and the observed data can be eliminated
by modification of the crust-mantle transition zone.

In Figures 7 and 8, the amplitude of the second-arrivals in the
10-15 km range define the 6.2 layer in the ophiolite models. These
arrivals are quite similar in character to those seen in the refraction
data. This implies that the 6.2-6.4 km/sec refractor found in the NEPAC
data is not necessarily part of the lower crust as we had assumed in our
preliminary interpretation which yielded a Type 2 crustal model.
Therefore these refractors probably correspond to P-velocities and rock
types of layer 2 in the upper crust. The amplitude information suggests
a strong gradient or discontinuity at this boundary. Without the
constraints of amplitude information and high shot density at these
particular ranges, first-arrival picks would not resolve the details of
this layer and the inversion would have regressed to a typical Type 1
(Raitt) model. Between the 24 to 28 km range in the NEPAC record
section, a partial shadow zone (a decrease in first-arrival amplitudes)
suggests that a P-low-velocity zone may exist in the upper crust. The
appearance of P-low-velocity zones could be associated with late-stage
differentiates common to many ophiolites [Coleman, 1977], but these
plagiogranites are not particularly important components in dredged
oceanic rocks [Bonatti, 1976; Christensen, 1976]. Various processes
have been proposed to account for the formation of plagiogranites in
localized sections of the oceanic crust [Christensen and Salisbury,
1975; Coleman, 1975; Sinton and Byerly, 1980]. Synthetics for Models 1 and 2 compare the effects of the P-low-velocity zone in the upper crust on the seismic arrivals. Without the LVZ there is no low amplitude shadow zone but rather a series of arrivals with little variation in amplitude (Model 2). The LVZ in Model 1 produces a decrease in first-arrival amplitudes in the 20 to 22 km range but does not fit first-arrivals seen in the NEPAC data. The first-arrival shadow zone at 24 to 28 km range that is found in the NEPAC observations can also be caused by deep crustal structure rather than an upper crustal P-low-velocity zone. This explanation is supported by the synthetic seismogram (Model 4) which show the same relative amplitude changes as the NEPAC data. The decrease in amplitude at 24 to 28 km is caused by interference effects from seismic arrivals that have penetrated into the lower crust. This interference arises because a P-low-velocity zone at the base of the crust forces the mantle reflection branch closer to the origin, where these larger amplitude arrivals dominate the signals that have traveled only within the upper crust.

It is obvious that the relatively simple BOI velocity-depth function generates a complex series of seismic arrivals from the lower ophiolite structure. The distinguishing characteristics of the lower crustal structure are evident in seismic arrivals seen at long ranges, particularly as second arrivals. For example, the travel-time curve and synthetic seismogram of Model 5 (Figure 10) show three identifiable seismic refractors associated with the lower crust. A 6.7 km/sec refracting horizon from the upper gabbro layer and a 7.1 km/sec
Figure 10 Uninterpreted and interpreted synthetic seismograms generated using Model 5. The uninterpreted section in (a) shows high amplitude second arrivals at 25-43 km range often interpreted as a single mantle reflection branch. In (b) the complexity of the synthetic seismogram is revealed by superimposing the travel time curve for the model on the synthetic seismogram. Multiple reflected-refractions begin to affect second arrivals at a range of 10 km and dominate the second arrivals after 50 km range. The synthetic seismograms are reduced to 8.0 km/sec. Amplitudes are scaled to the first shot with a range scaling factor of $1/r^2$. 
refractor associated with the upper cumulates produce distinctive wide-angle reflection branches extending to 60-70 km range. A 7.4 km/sec refraction horizon, from the basal cumulate ultramafics, generates a second-arrival branch in the 35-45 km range. The postcritical reflection branches from the 6.7, 7.1, and 7.4 km/sec refractors can also be found in the NEPAC data (Figure 9). In our preliminary interpretation, we had interpreted these arrivals as a single reflection branch from a 7.1 km/sec refractor. This interpretation is shown in Figure 11. The observations are, of course, more difficult to interpret than the synthetics. For example, the first-arrival branch between 20 and 30 km range definitely supports the 6.7 km/sec refractor, but postcritical reflections from the 6.7 km/sec refracting horizon are masked by the reflection branch from the 7.1 refractor between 40 and 65 km range. Combining the 7.1 km/sec reflection branch data (40 to 60 km range) with the 6.7 km/sec refracted arrivals at 20 to 30 km can produce a Type 1 crustal section with a P-velocity near 6.9 km/sec. The actual value of this (incorrect) average P-velocity will depend, of course, on the weight given to the 6.7 km/sec and the 7.1 km/sec arrivals. The degree to which one is biased depends on the experimental format including the length of the refraction experiment, with the longer refraction lines biasing the inversion towards the higher velocity. The Type 1 section appears to be an artifact of picking the wrong reflection branch from a complex group of second arrivals. Helmberger [1977] discussed this problem in his analysis of seismic refraction data where two adjacent refractors, with
Figure 11 Summary showing the uninterpreted record section from OBS "Locris" (south) (a) together with initial interpretation of observations based on linear inversion methods (b), and the new interpretation based on the results from modeling the BOI ophiolite (c). Rather than interpreting a single mantle reflection branch as shown in (b), the ophiolite modeling satisfactorily explains many of the complex arrivals in the refraction data. In (c) the multi-layered structure of the lower crust is described through the many refraction branches. The multiple reflected-refractions begin to define the previously uninterpretable second arrivals.
P-velocities of 6.6 and 7.0 km/sec, were averaged to produce a single refracting horizon with a P-velocity of 6.8 km/sec.

The high-amplitude reflection branch from the 7.1 km/sec refractor that dominates the synthetic seismograms at 28-30 km range and 29-31 range in the "Locris" data are often misinterpreted as being part of the 6.7 km/sec refracted arrival. The synthetic seismograms and travel-time curve for Model 5) reveal how this reflection branch becomes nearly coincident with the 6.7 km/sec refraction at ranges of 28 to 31 km. The distinct, high-amplitude nature of these arrivals can be used to define the 6.7-7.1 km/sec boundary. Petrologically, this represents the layer 3A-3B interface marking the boundary between the massive hornblende gabbros and the cumulate mafic gabbros.

Postcritical reflections (at ranges greater than 70 km), also can significantly constrain interpretation of seismic data. Second-arrivals picked as postcritical reflections from the 6.7 or a 7.1 km/sec reflectors at ranges greater than 65-70 km may, in fact, be multiple reflected refraction branches [Meissner, 1969]. This phenomena is diagrammatically illustrated in Figure 12, and can easily be seen in the synthetic seismogram presented in Figure 10 and the observations (Figure 11). Thus, extension of a primary reflection branch to excessive ranges can lead to interpretations that include unnecessary low velocity zones in the lower crust. If used properly, these reflected-refractions can be used to distinguish between possible crustal models. For example, the synthetics for Models 3, 4, and 5 exhibit distinctly different multiple reflected-refractions, in the distance range of 45-90 km, which
Figure 12 A diagrammatic illustration of the raypaths and travel-time curves for reflected refractions generated by a layered velocity depth model. These second arrivals are sensitive to the velocity structure and can be used to constrain acceptable velocity depth models. (R) signifies refracted arrivals; (r) denotes reflected arrivals.
when compared with similar phases in the NEPAC data (Figures 7 and 8) allows us to reject Model 3 as an acceptable model for the observations.

The Basal Layer

Since Models 1, 2, and 3 produce synthetic seismograms that are not consistent with the NEPAC data, we can turn our attention to testing Models 4 and 5. There are two regions in the NEPAC data where second-arrival energy effects the first-arrivals and creates shadow zones in the first arrival-coda. These can be seen at 27-29 km range and 34-36 km range in the observations. Model 5, which is equivalent to Model 4 except that it lacks a P-velocity inversion at the base of the crust, does not properly emulate the "shadow zones" at 34-36 km. Model 4, on the other hand, produces similar arrival patterns as the observations in the NEPAC data, but the shadow zones are shifted towards the origin by approximately 2 km. These results can be used to show that Model 5 is not correct, but we must explain the discrepancies between Model 4 and the data. The clue to the solution to this problem can be found in the fact that large-amplitude first-arrivals from the mantle (47 km range for Model 4) are approximately 5 km closer to the origin than similar arrivals in the data. This premature focusing of the seismic energy is due to the fact that Model 4 has a first-order M-discontinuity. Model 4a (Figure 13) shows synthetic seismograms for the Model 4 crustal section, but the crust-mantle transition is modified to include a half-kilometer velocity gradient rather than a first-order
Figure 13 Record section for observations from OBS "Locris" (south) and synthetic seismograms generated using Model 4a. The crust-mantle transition has been modelled with a velocity gradient rather than a first-order discontinuity as seen in Model 4.
discontinuity. This change in the crust mantle boundary not only correctly positions the larger-amplitude mantle arrivals at 50 km range, but shifts the shorter range, "shadow zones", discussed previously, to the correct range of approximately 35 km.

By carefully considering second-arrivals, and the character of the first-arrival coda, we have been able to fit seismic observations extremely well with a velocity-depth function that is consistent with the BOI ophiolite. Further, we have been able to use these data and synthetic seismograms to constrain the fine structure (and hence, the petrologic model) of mature oceanic crust.

Evaluation of the Type 3 Crustal Section

The close shot spacing used in airgun sonobuoy (ASPER) profiles allowed Sutton et al. [1971] to resolve 2 sublayers, a 6.8 km/sec refractor overlying a 7.5 km/sec refractor, within "layer 3". Each of these refracting horizons is approximately 3 km thick. The 6.8 km/sec refractor is essentially equivalent to the 6.7 km/sec refracting horizon reported by other investigators. The inferred 7.5 km/sec refracting horizon is of greater importance since it represents a significant departure from previous models. Christensen and Salisbury [1975] point out that while rocks with compressional velocities of 7.5 km/sec are found within ophiolites, they represent less than 1 km of the basal crustal structure within the ophiolite section. This is not consistent with the interpretation of the ASPER data. Interpretations of ASPER
data also tend to have mantle P-velocities of 8.3 km/sec and greater, and a crust 1 to 2 km thicker than standard oceanic sections. Morris, Raitt, and Shor [1969] have shown a distinct azimuthal dependence of the mantle velocities which is not apparent in the ASPER interpretations. Helmberger [1977] questioned the existence of this 3 km thick basal section on the grounds that there should be a distinct reflection branch associated with this high velocity layer that is not observed and concluded that the basal layer could not be as thick as suggested by Sutton et al. [1971]. Spudich [1980], using synthetic seismograms of ASPER records, suggested that the 3 km thick 7.5 km/sec layer may be the result of picking the mantle reflection branch (PmP) as a 7.5 km/sec refraction (Figure 14). Curvature of the apparent 7.5 refractor beyond the critical refraction point at 20 km range also implies that this phase is a reflection branch rather than a refractor.

Since the NEPAC OBS line is very close to one of the regions where ASPER data were interpreted to support a thick high P-velocity basal layer, we can use the NEPAC data with its close shot spacing amplitude information to help investigate this problem.

Synthetic seismograms for the BOI ophiolite with a 3 km thick 7.5 km/sec layer at the base of the crust (Model 6) are shown in Figure 8d. Comparison of these synthetics with the observations from the NEPAC data shows that there is considerable disagreement in the amplitude of arrivals in the 15-30 km range between these synthetics and the observations. A comparison of the appropriate travel-time curves and seismic raypaths for Model 4 (Figure 15), which fits the observations,
Figure 14 Uninterpreted ASPER record section together with interpretations of the coherent refractor arrivals. In (b) the ASPER record section is shown for sonobuoy #93 from the Pacific (made available by D. Hussong). In (a) the single arrowed line marks the upper crustal refraction arrival with a typical P-velocity of 6.74 km/sec. The double arrowed line shows the highly coherent arrivals in the 22 to 30 km range which produce a 7.38 km/sec intracrustal refractor. The triple arrowed line is another interpretation producing an 8.25 km/sec mantle arrival by picking the near critical extension in the range of 18 to 22 km of the previous arrivals. D1 indicates the direct water wave. R1 and R2 represent the first and second reflections, respectively.
Figure 15  Travel time curve and raypaths for Model 4. In (a) the P-velocity function for BOI Model 4 is shown. The crosses (x) on the travel time curve in (b) display the time of arrival associated with each raypath shown below in (c). The dashed line represents the mantle refractor.
Figure 16 The same as Figure 15 except that the travel time curves and raypaths are for Model 6.
BAY OF ISLANDS
OPHIOLITE
DISTINCT 7.5 KM/SEC
BASAL LAYER,
AFTER SUTTON ET AL.,
8.0 KM/SEC MANTLE

RANGE (KM)

0.000

-6.000

-10.000

2.000 4.000 6.000 8.000 10.000

VELOCITY (KM/SEC)

MODEL 6

RV = 8.00

(a)

(b)

(c)
and Model 6 (Figure 16) illustrates the differences in seismic energy
distribution from these models. The effects of the reflection branch
from the 7.5 km/sec refracting horizon in Model 6 dominates arrivals
from 19 km to 80 km, thereby creating a distinct 7.5 km/sec postcritical
reflection branch. Model 4, on the other hand, produces maximum
amplitude second-arrivals (beginning at 20 range) which are reflections
from the mantle (Compare Figure 8b with 8d). This result coupled with
the fact that the "Locris" OBS refraction data do not have high
amplitude second-arrivals with a 7.5 km/sec P-velocity extending out to
80 km range (but rather a 7.1 km/sec layer as seen in Model 4) suggests
that the 7.5 km/sec arrival branch is better interpreted as subcritical
reflections for the mantle in the ASPER data. This suggests that the
layer 3A-3B boundary observed in seismic data from older oceanic basins
is marked by a 6.7-7.1 km/sec velocity transition. This result is
consistent with ophiolite data.
Conclusions

The synthetic modeling of the BOI ophiolite has proven to be useful in interpreting and re-evaluating seismic refraction data that samples crustal structure in ocean basins. These results point out that experimental design and instrumentation can significantly affect resulting crustal models. High shot densities, wide-bandwidth instrumentation, and the dynamic range to properly record amplitude information are essential to obtaining realistic models.

The synthetic models support the concept that some ophiolites are ocean crustal sections. The application of the BOI ophiolite models to the inversion of NEPAC OBS data has allowed us to better resolve details of the ocean crustal structure. At the same time this technique has allowed us to rationalize often conflicting models of oceanic crust. The simple 3-layer models have given way to more complex models as better data has become available; as the seismic models become better constrained they tend to converge towards the ophiolite models. This convergence to a directly observable geologic section, in turn, allows us to associate seismic refraction data with distinct petrologic units.

Under the assumptions outlined previously, the Layer 2-Layer 3 boundary corresponds to the metamorphic facies change from greenschist to amphibolite. This transition is associated with the rapid change in P-velocity from 6.2 to 6.7 km/sec which occurs over a few tens to hundreds of meters in the sheeted dike complex within the BOI ophiolite. It has been suggested that a distinct layer 2-layer 3 boundary may not exist [Spudich, 1980]. It is possible that this boundary, which appears
to be associated with the hydrothermal metamorphism of the crust, is age dependent. Younger crust has a smooth transition between layer 2 and layer 3 while older crust such as that examined in this paper exhibits the more distinct layering.

This study shows how various interpretations of seismic data may affect the apparent structure of ocean crustal layer 3. Historically, layer 3A has had some of the widest variability in interpretations, despite the fact that the actual refraction first-arrivals from that seismic horizon mark a nearly homogeneous layer with a P-velocity of 6.7 km/sec. The P-velocity of Layer 3A (6.7 km/sec) is appropriate for metadolorites and metagabbros [Salisbury and Christensen, 1978; Spudich, 1979].

The synthetic models have resolved some of the apparent problems concerning the nature of the lower crust. Rather than having a 3 km thick high velocity basal section with a 7.5 km/sec velocity, or a thick homogeneous 6.7 km/sec layer, our results indicate the basal section may be composed of two distinct horizons, a 2.0 km thick 7.1 km/sec layer and a 1.0 km thick basal layer with a high velocity lid and serpentinized cumulates below. Correspondence of the seismic velocity structure to the BOI ophiolite petrologic structure shows that the 7.1 km/sec horizon corresponds to the cumulate pyroxene gabbros, and the 7.4 km/sec horizon represents the more ultramafic component of the cumulates at the base of the ophiolites. Serpentinization of the olivine component and the resultant low velocity zone at the extreme base of the crustal structure is consistent with our observations.
We suggest that the Bay of Islands ophiolite is seismically indistinguishable from standard refraction data from older ocean crust. Not only does it compare well with refraction results but it has been useful in clearing up many of the ambiguities found in refraction interpretations, particularly where interpretations are a product of the refraction methodology rather than the seismic structure of the oceanic crust.
CHAPTER III

COMPARISON OF SYNTHETIC SEISMOGRAMS OF THE SAMAIL

OPHIOLITE, OMAN AND THE ROSE REFRACTION DATA FROM THE EAST PACIFIC RISE

Introduction

In the last decade, research on ophiolite complexes has broadened our understanding of ocean crustal structure and tectonic processes. Through the study of the physical properties of these ophiolite suites analogies have been drawn with marine seismic refraction data to suggest the composition and structure of the ocean crust [Coleman, 1971, 1977; Moores and Jackson, 1974; Peterson et al., 1974; Salisbury and Christensen, 1978; Spudich and Orcutt, 1980a]. Recently, with the availability of complete physical property data for select ophiolites and the advent of relatively inexpensive high-speed computer technology, synthetic seismograms for ophiolite sections have been directly compared with marine seismic refraction data [Spudich et al., 1978; Nichols et al., 1980; and Kempner et al., 1981]. Results from this research are encouraging.

In this study we apply the reflectivity method of Fuchs and Muller [1971] and Kennett [1975] to generate a series of synthetic seismograms for the Samail ophiolite complex in Oman. We emphasize the use of long-range observations of reflected-refractions [Kempner et al., 1980] to determine the fine structure of oceanic crust by comparing these features with those found in the synthetic seismograms. The crustal
models used are based on velocity-depth functions from both Manghnani et al. [1981] and Christensen and Smewing [1979, 1981; and pers. comm.] for the Ibra and northern section of the ophiolite complex, respectively. The seismograms are compared with ocean bottom seismometer (OBS) refraction data from Phase I of the Rivera Ocean Seismic Experiment (ROSE) that sample oceanic crust of similar (4.5 m.y.) and younger (0.5 m.y.) age [see Ewing, 1979].

Our results suggest that within the range of measured limits of compressional and shear velocities, and densities for the Samail ophiolite, synthetic seismograms can be generated which offer good approximations to seismic refraction observations from young oceanic crust. Some variation between the synthetics seismograms and the real data occur in the second arrivals. Our results suggest this problem arises because reflected-refractions are particularly sensitive to the P-velocity gradient in the uppermost crust. Unfortunately, this portion of the oceanic crust and the ophiolite suites is poorly constrained. However, the agreement between synthetics for the Samail ophiolite, and seismic refraction observation from young crust is striking; the differences noted above are well within the expected variability of oceanic crustal structure.

We have investigated the Bay of Islands ophiolite complex (BOI) of Newfoundland in a similar manner [Kempner and Gettrust, 1981] and show that synthetic seismograms for the BOI ophiolite fit marine crustal-refraction data from older (60 m.y.) ocean basin crust. That investigation is based on a comparison of seismic refraction
observations from the northeast Pacific with synthetic seismograms the BOI ophiolite complex. These similarities between seismic observations sampling young and mature oceanic crust with synthetic seismograms for ophiolite suites of similar ages suggests that it is reasonable to use these ophiolites to study the effect of aging on oceanic crustal structure. We have used this technique to develop an aging model for oceanic crust that explains the transformation of the velocity-depth function from smooth gradients to the more rapid seismic velocity transitions required to fit seismic observations from mature oceanic crust.

Previous Work

Most recent analyses of young ocean crustal refraction data from the Pacific suggest that the velocity-depth structure is best fit by continuous gradients. Furukawa et al. [1981], using seismic data sampling crust of 0.5, 2.5 and 4.5 m.y.b.p. (obtained during the ROSE experiment) show that velocity gradients of 1.0-2.0 sec⁻¹ for the upper 2 km and 0.1-0.2 sec⁻¹ for the deeper crust fit the observations quite well. In general, no variation in crustal P-velocities with age was found; however, the development of a distinct reflection branch from the crust-mantle transition occurs by 4.5 m.y. age. Orcutt et al. [1976], working in an area near Siquieros Fracture Zone, obtained similar results, although they do suggest that slight stratification (high P-velocity gradients) occurs within the crust by 5 m.y. age. Neither Orcutt et al. [1976] nor Furukawa et al. [1981] found evidence for a
basal P low-velocity zone. Lewis and Snydsman [1977; 1979] presented interpretations of seismic data from oceanic crust as young as 2.5 m.y.b.p. from the Cocos Plate that include P low-velocity layers within crust. They used these data to suggest a process of continual serpentinization of the lower crust and upper mantle to account for thickening of the crust with age [Goslin et al., 1972; Christensen and Salisbury, 1975].

In an evaluation of the long-term evolutionary processes, Houtz and Ewing [1976] used sonobuoy data to suggest an increase in the P-velocity of the upper crust as a function of time. This increase in P-velocity is thought to result from the filling of voids and cracks within the uppermost basalts. The lower crust shows an inverse relation of mean P-velocity with age [Christensen and Salisbury, 1975], although both high P-velocity basal layers [Sutton et al., 1971] and P-low velocity zones [Meeder et al., 1977] have been proposed for older crust.

Historically, synthetic modeling of seismic refraction data has proven to be useful in resolving ambiguities in interpretations. For example, Helmberger and Engen [1978] directly approached the problem of the lower crustal P-velocities by using synthetic seismograms and concluded that dipping layers can generate the same signal characteristics as low P-velocity zones. Their modeling of 20 m.y. old crust also suggested that the oceanic crust thickens with age; to account for this they appeal to the serpentinization processes outlined by Woollard [1975]. Lewis and Snydsman [1977; 1979] also used synthetic seismograms to reach similar conclusions for the Cocos Plate data.
The application of ophiolite models to interpret ocean crustal structure is not new. Moores and Jackson [1974] pointed out similarities between a complete series of ophiolite models and ocean crust. Peterson et al. [1974], Salisbury and Christensen [1978], and DeWit and Stern [1976] have made comparisons between velocity-depth function for selected ophiolites and published velocity-depth functions from marine seismic investigations. More recently, synthetic seismograms for ophiolites based directly on the physical properties data of ophiolites have been used by Spudich et al. [1978] to model oceanic crust. Their results showed that there is excellent agreement between the BOI ophiolite synthetics and Fanfare data from 12 m.y. old crust in the Pacific. Nichols et al. [1980] used synthetics based on the P-velocity depth function for the Point Sal ophiolite in California to evaluate the proposition that serpentinization of the lower crust and upper mantle is the mechanism for crustal thickening with age.

The Ophiolite Models

The Samail ophiolite complex in Oman is considered to be a section of the Tethyan ocean crust formed at a Cretaceous spreading center. Although interpretations of the actual obduction process may vary [Gealey, 1977; Welland and Mitchell, 1977; and Coleman, 1977], it is generally accepted that the petrologic suite was formed at an ocean basin spreading center rather than within a back-arc basin. The southern Ibra section of the ophiolite as described by Hopson et al. [1981] has a 9 to 12 km thick tectonite peridotite layer at the base
of the section. This unit, considered to be depleted mantle material composed of harzburgite and dunite, is overlain by a 3 to 5 km thick region composed of cumulate gabbro showing evidence of periodic replenishment during crystallization. The overlying non-cumulate gabbros (with a thickness of 0.2 to 1.0 km) extend upward to the diabase dike complex which varies from 1.2 to 1.6 km in thickness. Occasional late-stage differentiates are found to be associated with the gabbro-dike boundary. The dikes progress upward into a thin (approximately .5 km) section of pillow lavas. The actual thickness of the pillow lava layer, based on stratigraphic relations seen in the northern sections of the ophiolite, may originally have been greater than 1 km.

Generally, crustal and petrologic interpretations indicate variations of crustal structure with age [Goslin et al., 1972; Woollard, 1975; Christensen and Salisbury, 1975; Lewis, 1978] and spreading rates [Cann, 1968, 1974; Kusznir, 1980]. Therefore, before useful comparisons of ophiolites with oceanic crust can be made, it is necessary to have some control on the formation and emplacement age of the ophiolite. Uranium lead isotope ages from zircons within the plagiogranites at Oman define a formation age of 95+/-1 m.y. in the southeast section and 98+/-1 m.y. in the northern province of the ophiolite [Tilton et al., 1979]. These ages agree with the biostratigraphic age of Tippit and Pessagno [1979]. Lamphere [1979] gives an initial emplacement age of approximately 90 m.y. based on K-Ar dating of amphibolites at the base of the allochthonous section; this suggests that the ophiolite
represents a section of ocean crust of no greater than 5 to 8 m.y. age. Tilton et al. [1979] present evidence for a spreading half rate of 2-5 cm/yr. Thus the age and spreading rate of the Samail ophiolite appears to be acceptable for a comparison with the Rose refraction line sampling 4.5 m.y. old crust which was formed at the EPR with a spreading half-rate of 5 cm/yr [Hey et al., 1977].

The physical properties \( (V_s, V_p, S) \) used in the seismic modeling of the ophiolite are based on laboratory measurements of samples from the ophiolite that were corrected for pressure conditions appropriate for an oceanic crustal environment. Manghnani et al. [1981] have created a velocity-depth function for the Muscat-Ibra transect based on a series of closely spaced traverses through the ophiolite complex (Fig. 17a). Christensen and Smewing [1979] worked on the northern section of the Samail complex and generated physical properties data for another complete stratigraphic section (Fig. 17b). These velocity-depth models (Figure 18a) indicate the extent of lateral variation in crustal structure that has been found in the Samail ophiolite complex.

The Manghnani et al. [1981] velocity-depth function was used as the starting model for generation of synthetic seismograms; this model was perturbed until the synthetic seismograms were in good agreement with the observations. The modifications to the original Manghnani velocity-depth function that were required to obtain a satisfactory fit were minimal. For example, the basaltic depth section of the velocity depth function was proportionately thickened to 1.3 km, and the gabbroic section had to be foreshortened approximately .7 km to make the mantle
Figure 17 Density, compressional and shear velocity-depth functions for the Samail ophiolite. In (a) the values determined by Manghnani et al. [1981] for the southern part of that complex. The values determined by Christensen and Smewing [1981] from the northern portion of the same ophiolite suite are presented in (b).
(a) 

(b)
crossover of the synthetic fit the ROSE seismic refraction observations. This velocity-depth function is referred to as Model 1. All changes in the thickness of the stratigraphy and velocity-depth profile remain within the limits for the ophiolite as described by Hopson et al. [1981].

To test our assumptions on the age of the ophiolite as well as our modifications of the stratigraphic column, the Manghnani et al. [1981] velocity-depth function was initially compared with Tau-P inversion limits of crustal refraction data taken in the southeast Pacific over crust of <5 m.y. and 5-6 m.y. [Kennett et al., 1977]. Data shown in Figures 18b and 18c support the age dating results for Oman by placing the velocity depth function exactly within the boundaries for similar age ocean crust. This result suggests that the Oman complex may be a working model for young ocean crust. However, agreement between travel-times for marine seismic observations and the ophiolite may not be sufficient evidence for this inference. Synthetic seismograms for the Samail ophiolite would, if the comparison is positive, provide better support for this proposition.

Synthesis Techniques

Synthetic seismograms for the Oman profiles were generated using the reflectivity methods as described by Fuchs and Muller [1971] and Kennett [1975]. The seismograms are parameterized to simulate the marine OBS case with an explosive source at the surface and a receiver positioned at the water-sediment interface. Full compressional ($v_p$) and
Figure 18 The P velocity-depth functions from Manghnani et al. [1981] and Christensen and Smewing [1979] are superimposed in (a). These models represent possible lateral variation in the Samail ophiolite complex. In (b) the Manghnani et al. [1981] P-velocity model is superimposed on P-velocity bounds for 5-6 m.y. old oceanic crust. In (c) the same ophiolite velocity-depth function is superimposed on the P-velocity bounds for oceanic crust < 5 m.y. old. The observational bounds for oceanic crust are after Kennett et al. [1977].
LATERAL VARIATION WITHIN SAMAIL OPHIOLITE
AFTER MANOOGIAN ET AL. (1981) (---)
FROM CHRISTENSEN AND SHEWING (1979) (---)

(a) VELOCITY (KM/SEC)

(b) VELOCITY (KM/SEC)

(c) VELOCITY (KM/SEC)

LATERAL VARIATION WITHIN SAMAIL OPHIOLITE
AFTER MANOOGIAN ET AL. (1981) (---)
FROM CHRISTENSEN AND SHEWING (1979) (---)

(a) VELOCITY (KM/SEC)

(b) VELOCITY (KM/SEC)

(c) VELOCITY (KM/SEC)
shear \( (V_s) \) wave response including multiples and converted phases is produced based on the \( V_p, V_s \), and density profiles from the ophiolite sections. Since this technique does not use continuous gradients, they must be approximated by a sequence of thin layers with small increments in seismic velocity between layers. Our criterion for the number of layers in the velocity structure was .1 km/sec/layer; this limit ensured that the reflectivity calculations properly modeled the seismic-velocity gradients encountered. A minimum phase velocity of 3 km/sec was used to eliminate water wave and sediment arrivals which would cause wrap-around problems and interfere with crustal arrivals at longer ranges. This limit does, however, decrease the amplitude of upper crustal shear arrivals with similar or lesser phase velocities. The maximum phase velocity was set at 15 km/sec. To match the dominant frequency of the Rose refraction data, a 10 Hz ricker wavelet source function was used in conjunction with a 3 to 15 Hz cosine-squared frequency window. The ophiolite sections were modelled for a water depth of 3100 m with a 100 m thick homogeneous veneer of sediments having a P-velocity of 1.7 km/sec. These parameters correspond to the Rose refraction line situation. Attenuation coefficients for the sediments are .0002 and .0004 for compressional and shear waves, respectively. Igneous rock values for \( 1/Q \) are .001 and .005. Typically, 650 wave numbers and 250 frequencies were calculated for each seismogram. All synthetics were computed on HIC's Harris S-135 computer which has a 48-bit floating point word length. The computation time for a typical reflectivity-matrix was approximately 10 hours.
The ROSE Refraction Data

As noted previously, a large seismic data base from young oceanic crust near the East Pacific Rise (between the Clipperton and Orozco fracture zones) was gathered during the ROSE experiment. These data are appropriate for comparison with the Samail ophiolite as they are young (0.5 to 4.5 m.y.b.p.) and are associated with a fast spreading center. We had originally used geophone data from RIG OBSs to test the Samail ophiolite. It was clear that the velocity-depth model developed by Manghnani et al. [1981] satisfied the travel-time observations [Furukawa et al., 1981]; however, we wished to exploit the use of second-arrivals to further constrain the model. To do this, we used Woods Hole Oceanographic Institute (WHOI) hydrophone data from the same refraction lines. These data have well defined second-arrival branches that could be compared with reflected-refractions that are clearly evident in the synthetic seismograms. Figure 19 shows typical reflected refraction paths and travel-time curves for a seismic-velocity model with constant gradient. The velocity-depth functions required to fit the ophiolite and seismic data produce more complex reflected refraction arrivals that can be unraveled using ray-tracing techniques [Kempner et al., 1980].

Data from ROSE refraction profiles 2S (4.5 m.y. age) and 4S (0.5 m.y. age) are shown in Figure 20. The shot spacing on both profiles is less than 1 km. The record sections have been reduced over 6.5 km/sec and all data have been filtered from 2-to-20 Hz using a zero-phase shift, 8 pole, Butterworth filter. All traces are scaled to the largest amplitude in that trace. Data from both refraction lines show nearly
Figure 19 A diagramatic illustration of the raypaths and travel-time curves for reflected refractions generated by a continuous velocity-depth model. These second-arrivals are sensitive to the velocity gradients and discontinuities and can be used to constrain acceptable velocity-depth models.
Figure 20 Uninterpreted record sections from ROSE seismic profiles
recorded by the Woods Hole Oceanographic Institute (WHOI)
ocean bottom hydrophones. Figure (a) shows data from line 4s
which sampled 0.5 m. y. age crust. In (b) data from 4.5
m. y. age crust (line 2s) is shown. The distinct
second-arrival branches in these data make it possible to use
reflected refraction phases to constrain possible
velocity-depth functions for these observations.
continuous first-arrival branches with no strong seismic energy concentrations that would be diagnostic of seismic velocity discontinuities or strong seismic velocity gradients in the crust. A first mantle-reflection branch (PmP) is seen in the data from line 28. These data fit the continuous seismic-velocity gradient crustal model proposed by Furukawa et al. [1981].

Seismic data from continuous velocity-gradient structures are not necessarily lacking in diagnostic second-arrival information. Synthetic seismograms (Figures 21 and 22) based on a velocity-depth function for the Samail ophiolite [Manghnani et al., 1981] offer convincing evidence of this fact. We will show that similar second-arrival branches are observed in the ROSE data, and that these reflected-refractions allow us to differentiate between velocity-depth models that produce equivalent travel-time curves.
Figure 21  The P velocity-depth function for Model 1 of the Samail ophiolite together with uninterpreted and interpreted record sections of synthetic seismograms generated using the Model 1 velocity-depth function. The $P_g$ branch represents crustal-path first-arrivals while $2P_g$ denotes the first reflected refraction multiple of that branch. The $PmP$ branch traces reflections from the M-discontinuity. Shear wave and multiply reflected shear wave branches are also shown.
SAMAIL OPHIOLITE, OMAN
MODEL I
(AFTER MANGHNANI, ET AL.)

![Graphs and Diagrams](image)
Figure 22 The P-velocity-depth function for Model 2 of the Samail ophiolite together with uninterpreted and interpreted record sections of the synthetic seismograms generated using the Model 1 velocity-depth function. The P-velocity inversion in the crust generates two reflected refraction branches rather than the single branch shown in Figure 21. Complexities in the first-arrival branch near 30 km range and the strong 2P2P branch distinguish these synthetic seismograms from those presented in Figure 21.
SAMAIL OPHIOLITE, OMAN
MODEL 2
LVZ IN LAYER 3

(a)

(b)

(c)
Comparison of Observations with Synthetic Seismograms

We compare synthetic seismograms for three acceptable Samail ophiolite velocity-depth models with the ROSE observations from 0.5 and 4.5 m.y. age crust. Models 1 (Figure 21) and 2 (Figure 21) are synthetic seismograms based on the Manghnani velocity-depth function. Model 2 is appropriate when late stage differentiates in the crust produce a P low-velocity zone; otherwise, it is equivalent to Model 1. Model 3 (Figure 23) is based on the velocity-depth function for the Samail ophiolite proposed by Christensen and Smewing [1979, 1981]. The synthetic seismogram record sections are presented in the same format as the observations.

Each of the synthetic seismograms has been interpreted using the ray-tracing technique discussed previously. This has allowed us to determine to what depths the second-arrival branches had penetrated, and to determine which crustal multiples should be observable. Perhaps the most diagnostic differences between the synthetics for Models 1 and 2 occur beyond 30 km range (Figures 21 and 22). Relatively simple intracrustal multiples that are produced by Model 1 are replaced by two sets of intracrustal multiples in Model 2; the splitting of the intracrustal multiples is caused by the P-velocity inversion in Model 2. To emphasize this fact, we have denoted the intracrustal multiples for Model 1 (Figure 21) as Pg and as P1 and P2 for Model 2 (Figure 22). While there are differences in the first-arrival branches of the synthetic seismograms from Model 1 and 2, they represent differences that may be difficult to identify in less closely spaced observations.
Figure 23 The P velocity-depth function for Model 3 of the Samail ophiolite together with uninterpreted and interpreted record sections of the synthetic seismograms generated using the Model 3 velocity-depth function. As in the record sections presented in Figure 22, the P-velocity inversion in the crust generates multiple reflected refraction branches, and complicates the first-arrival branch at 25 km range.
SAMAIL OPHIOLITE,
OMAN
MODEL 3
(AFTER CHRISTENSEN)
with excellent signal to noise ratios. The second-arrival branches, especially the 2P2P branch, (Figure 22) would be easier to identify in observations; we weight our comparison of these synthetics with ROSE observations on the more easily resolvable second-arrival branches as well as first-arrivals.

A comparison of synthetics for Model 1 and observations from line 4S (0.5 m.y. age crust) shows that data from extremely young oceanic crust differs from the synthetics for the Samail ophiolite models (Figure 24). However, it is not difficult to see that there are possible generic relationships between these synthetics and the observations. To point out these relationships, we have superimposed the travel-time plot for Models 1 and 2 on these data (Figures 24 and 25). We have not included the reflection branch from the M-discontinuity as there is no evidence in these data or in the data shown by Furukawa et al. [1981] that a crust-mantle discontinuity exists at this age.

In the distance range 0-to-12 km, the first-arrival branches for Models 1 and 2 clearly are delayed with respect to those in the observations. These differences may result, in part, from unresolved topographic effects; however, the primary cause is more likely due to differences in the upper crustal seismic-velocity structure. Beyond 12 km range, the first-arrival branches of the synthetics and observations converge. This indicates that there is little difference in the lower crustal seismic velocity-depth structure between the ophiolite model and extremely young oceanic crust. Our investigation of differences in the
Figure 24 Record sections for observations from ROSE line 4s (0.5 m. y. age) and synthetic seismograms generated using the Model 1 velocity-depth function. There are similarities in the second-arrivals in the distance range 15–30 km (the 2Pg and 3Pg branches). However, there are obvious discrepancies between the observations and synthetic seismograms. The PmP branch is not shown on the observations because there is no evidence of an M-discontinuity in the data.
Figure 25 The same as Figure 24 except that the synthetic seismograms and interpretation are from Model 2. These synthetic seismograms are similar to the observations but, as with the comparison shown in Figure 24, there are significant differences between the synthetics and the observations.
second-arrival branches between the ophiolite Models 1 and 2 and the observations shows that they can be accounted for by differences in the upper crustal P-velocity structure. A change in the amplitude of the first-arrivals in the observations between 17 and 20 km range and the apparent 2P2P branch in the observations suggests that Model 2 should be preferred. Since the ophiolite models represent older material, it is not surprising that there are differences between the synthetics for these models and the observations from 0.5 m.y. age crust. It is interesting, however, that the observations differ from the synthetics for the Samail ophiolite primarily in terms of the upper crustal seismic-velocity structure.

Figures 26 and 27 present synthetic seismograms for Models 1 and 2 together with observations from line 2S (4.5 m.y. age crust). Both models fit the first-arrival branch in the observations quite well. This agreement carries through to the more diagnostic second-arrival branches, where it is especially striking for Model 2. While the change in amplitudes in the first-arrival energy at approximately 30 km range is seen in both Model 2 and the observations, there is also strong support for the second-arrival branches including the 2P2P phase in the observations. This phase is not observed in the synthetic seismograms for Model 1.

Our comparison of observations with synthetic seismograms for probable oceanic velocity-depth structures has shown that multiply refracted arrivals place sensitive constraints on acceptable crustal models. Interpretation of these arrivals requires instrumentation that
Figure 26 Record sections for observations from ROSE line 2s (4.5 m.y. age) and synthetic seismograms generated using the Model 1 velocity-depth function. The similarities between these data and the synthetic seismograms are particularly good to a distance of approximately 30 km. The strong second-arrival branch that is found in the observations at distances greater than approximately 30 km are not evident in the synthetic record section.
Figure 27 Record sections for observations from ROSE Line 2s (4.5 m.y. age) and synthetic seismograms generated using the Model 2 velocity-depth function. These synthetics are in excellent agreement with the observations. The 2P2P branch, for example, satisfies the need for a strong second-arrival branch (roughly parallel to the first-arrival branch) at ranges beyond 30 km. The complexities in the first-arrival branch at approximately 28 km range are also fit by the synthetic seismograms for Model 2.
accurately reproduces the second arrivals. In addition, this technique requires that refraction profiles be extended beyond the range where first-arrivals sample the mantle. The signal-to-noise ratio from the "small shot" refraction profiles run during ROSE makes it impossible to interpret data recorded beyond 50 km range. We have generated synthetic seismograms for the ophiolite models to 90 km range (Figure 28), which indicate that longer seismic refraction profiles (with dense shot spacings) would provide much greater control on the interpretation of crustal structure.

The synthetic seismograms generated using the Christensen and Smewing [1981] velocity-depth function for the Samail ophiolite (Model 3, Figure 23) clearly are not as consistent with the ROSE observations as synthetics for Models 1 and 2. Since Model 3 represents a viable velocity-depth function for the Samail ophiolite, we believe that the three models presented in this paper represent expected lateral variation in the seismic velocity structure of young oceanic crust. The ROSE observations fit both Model 1 and Model 2 velocity-depth functions; other observations of young oceanic crust may be better fit by the velocity-depth function for Model 3.
Figure 28 Synthetic seismograms for Models 1, 2, and 3 computed to 90 km range. In this figure, the amplitudes of the traces are not normalized to the maximum amplitude in each trace but are scaled to the inverse square of the range. While there are obvious differences between the synthetic seismograms for Model 3 and those for Models 1 and 2 at all ranges, it is clear that extending refraction profiles to 90 km range would make it possible to differentiate between Models 1 and 2. This is particularly interesting since these velocity-depth functions differ only in that Model 2 has a small P-velocity inversion centered at slightly more than 6 km depth (see Figures 21 and 22).
Discussion

The velocity-depth structure of the Samail ophiolite consists of a high seismic-velocity gradient for the upper two kilometers overlying a less positive gradient in the lower crust. This model is consistent with seismic observations of young oceanic crust [Orcutt et al., 1976; Kennett et al., 1977; Lewis and Snydsman, 1977; Furukawa et al., 1981]. This is demonstrated in Figure 29b which shows the velocity-depth function for Model 1 superimposed on the velocity-depth bounds which Furukawa et al. [1981] computed for ROSE line 2 observations. On the other hand Kempner and Gettrust [1980; 1981] have shown that seismic refraction data from old oceanic crust (60 m.y.) and synthetic seismograms for the Bay of Islands ophiolite complex (BOI) in Newfoundland (45 m.y. formation to emplacement age) agree in terms of travel-time, and more importantly, contain similar reflected-refraction arrivals. The BOI model is notably stratified with distinct layering of the upper and lower sections. The excellent agreement between this ophiolite and observations suggests that it is appropriate to consider some ophiolites as representative of oceanic crustal sections. If this is so, it is then reasonable to use certain ophiolites of differing age to study the evolution of the oceanic crust. Figure 30 shows the similarities in the petrologic structure between a young and an old ophiolite suite. Figure 29a presents a comparison of the velocity-depth function for young and mature ophiolites normalized to a constant mantle depth. The upper crustal structure shows an increase in P-velocity with age as suggested by Houtz and Ewing [1976]. However, Salisbury and
Figure 29 The change in the P velocity-depth function between young and mature oceanic crust is shown in (a) which compares the Samail ophiolite from Oman (Model 1) with the BOI ophiolite. These models suggest that the P-velocity increases with age in the upper crust and may decrease with age in the lower crust. Also shown in (b) are the P velocity-depth bounds determined by Furakawa et al. [1981] for ROSE line 2s superimposed on the Model 1 velocity-depth function. The velocity-depth function lies outside the bounds for the ROSE data in the upper crust, the depth where both functions are least well known.
AGING OF THE OCEANIC CRUST
SAMAIL OPHIOLITE 5-8 MY. (---)
BOI OPHIOLITE 40-45 MY. (-----)

(a)

SAMAIL OPHIOLITE (---)
EAST PACIFIC RISE 4.5 MY. (-----)

(b)
Figure 30  The lithologies of the Bay of Islands (BOI) ophiolite (45 m. y. age) and the Samail ophiolite (approximately 5 to 8 m. y. age).
Bay of Islands Ophiolite Newfoundland

LITHOLOGY

basalt/metabasalt

metadolerite sheeted dike complex

(Late Stage Differentiates)
massive gabbro / metagabbro

cumulates
pyroxene gabbro

olivine gabbro

dunite

(SEISMIC MOHO PETROLOGIC MOHO)

(tectonite-Cumulate Transition)

tectonite peridotite

ultramafics

after Salisbury and Christensen, 1978

0.5-1.5 km

1.2-1.6 km

<1.0 km

2.6-5.5 km

after Hopson et al., 1980
Christensen [1978] suggest that the BOI P-velocity values may be slightly overestimated due to laboratory measuring techniques and the fact that a few hundred meters of material may be missing from the top of the section. The lower crustal sections of the ophiolites show a definite decrease in seismic velocity with age.

Petrologic evidence for both ophiolite sections show pervasive hydrothermal alteration and metamorphism which generally controls the seismic velocity of the ophiolites down through the massive gabbros. Retrograde metamorphism from zeolite through greenschist in layer 2 and finally amphibolite within the gabbros is common, but is definitely restricted from most cumulate gabbros. Only material adjacent to local fracturing in the cumulates section shows any hydrothermal alteration [Donato and Coleman, 1976]. Similarly, in the BOI section, amphibolites associated with deep hydrothermal circulation are found through the massive gabbros [Salisbury and Christensen, 1978]. Since the alteration effects are found in both young and old ophiolites we believe that the metamorphism (by hydrothermal circulation) is overprinted on the crustal section immediately upon formation. DeWit and Stern [1976] use $^{18}O$ data from the Chilean ophiolite to support a similar conclusion.

Further alteration of the upper crust is controlled by long term processes of continual hydrothermal convection, fracturing, and secondary low temperature alteration, mineralization and cementation of the upper 1 to 2 km of crust. The relative importance of these processes depends on the porosity and permeability of layer 2 [Spudich and Orcutt, 1980b], which of course, changes with time.
Modification of the lower crustal structure is not required by the Samail ophiolite, and is not reflected in the seismic observations to 4.5 m.y. age. Spudich [1978] showed that 12 m.y. old oceanic crust has a similar velocity profile similar to that of the BOI ophiolite. Since the BOI ophiolite sequence supports alteration of the lower oceanic crust, alteration of the lower oceanic crust must begin sometime between 4.5 and 12 m.y. age.

Anderson and Hobart [1976] suggest that the convective transport of heat within the ocean crust ends between 5 and 10 m.y. age; the time of cessation depends on the sedimentation rate in the area. Geochemical data based on $^{18}O$ values from basalts show thermal enrichment up to 10 m.y. age followed by a slower rise in $^{18}O$ [Muehlenbachs, 1980]. This supports a change from hydrothermal convection to a conductive heat transfer. With this transition from convective to conductive heat transfer we propose that a second process of alteration begins. Water can now slowly permeate the lower crust, serpentinizing the olivine in the basal crustal "layer", thus decreasing the seismic velocity in that region. The extent of alteration to the lower crust and uppermost mantle is uncertain. Woollard [1975]; Clague and Straley [1977]; Luyendyk and Nichols [1977]; Lewis [1978]; and Nichols et al. [1980] propose various degrees of serpentinization of upper mantle material based on petrologic, seismic refraction, and synthetic modeling data. However, no more than 1 km of the BOI (60 m.y. age) model had to be altered (i.e., P-velocity inversion) to fit that model to the NEPAC refraction data [Kempner et al., 1980; Kempner and Gettrust, 1981]. The
process of deep crust upper mantle alteration appears to continue for approximately 30 m.y. at which time the apparent thickening of the oceanic crust ceases [Goslin et al., 1972; Woollard, 1975; Trehu et al., 1976].

Our results, together with the geochemical and petrologic data, support a two-stage process during crustal aging. Late-stage serpentinization of the crust (and possibly, uppermost mantle) may occur; however, the alteration need not be extensive. The fact that the crust-mantle velocity transition becomes distinct by 4.5 m.y. age, before serpentinization of the lower crust-upper mantle takes place, suggests that the crust-mantle boundary may initially represent a compositional boundary rather than a phase boundary.

Conclusions and Summary

The striking similarities between synthetics for Samail ophiolite velocity-depth functions and observations from young oceanic crust suggest that ophiolites represent good working models for study of the oceanic crust. The technique used in this paper to identify and match second-arrival branches in the synthetics with second-arrivals found in the observations allows us to constrain the set of acceptable velocity-depth functions to a much greater extent than would be possible using inversion techniques based on first-arrival information. We have also demonstrated that longer seismic refraction lines (to 90 km range) allow us to use reflected-refractions to refine velocity-depth functions for the entire crustal section. These results suggest that it is
possible to mitigate the problems inherent in marine seismology where the shot-receiver geometry usually precludes sampling the upper crustal structure using first-arrivals.

Since ophiolites are "seismically acceptable" models for both young and mature oceanic crust, it is reasonable to use the physical properties of ophiolites to model ocean crustal aging processes. Using data presented in this paper and for the BOI ophiolite [Kempner and Gettrust, 1981], we propose a two-stage aging process. Initially, hydrothermal circulation associated with the dynamic oceanic-ridge environment controls alteration of the crust and, therefore, the seismic-velocity structure of the upper crust. A change from convective to conductive heat-flow mechanisms between 5 and 10 m.y. age initiates slow alteration of the lower crust, producing serpentinization of the basal crustal material and possibly the upper mantle. The BOI ophiolite and seismic refraction data from mature oceanic crust do not require extensive serpentinization of the lower crust or upper mantle. The two-stage aging model provides a mechanism to limit the extent of alteration that is consistent with the physical and chemical properties of both young and mature ophiolites.
CHAPTER IV

FURTHER MODELING OF OPHIOLITES

Introduction

Laboratory measurements of the compressional ($v_p$) and shear ($v_s$) velocities of rocks from stratigraphically complete ophiolite sections have been shown to produce velocity-depth functions comparable to interpretations of marine seismic refraction data for oceanic crust [Coleman, 1971, 1977; Moores and Jackson, 1974; Peterson et al., 1974; DeWit and Stern, 1976; Salisbury and Christensen, 1978; Christensen and Smewing, 1981; Manghnani, et al., 1981]. Recently, synthetic seismograms generated from velocity-depth profiles for ophiolites have been directly compared to ocean seismic refraction data [Spudich et al., 1978; Spudich, 1979]. Similarly, Kempner and Gettrust [1981a, b] have compared synthetic seismograms for both the Samail ophiolite in Oman and the Bay of Islands complex in Newfoundland with marine refraction data from oceanic crust of comparable age. They suggested a two stage aging process for the ocean crust based on the correlation of the synthetic seismograms and the marine refraction data. Nichols et al. [1980] developed a synthetic seismogram for the Point Sal ophiolite in California and used the synthetic to propose a model of mantle serpentinization as the crust ages to account for the difference between the Point Sal ophiolite thickness and marine seismic observations for oceanic crust. These examples point out the use of ophiolite data to
constrain the inversion of seismic refraction data and represent new techniques by which we can better understand oceanic crustal structure. Alternatively, marine seismic refraction data could be used to explain the origin of the various ophiolite structures if synthetic seismograms from the more stratigraphically complete ophiolite sections were available for comparison with seismic observations.

Because compressional and shear velocity information is quite limited for ophiolites, Christensen [1978] published estimated velocity-depth profiles of five of the more stratigraphically complete ophiolite sections. These estimates of the seismic velocity structure are based on laboratory $V_p$ and $V_s$ suites in California and Oregon, as well as on published data. Two of the five ophiolites, the Bay of Islands complex, Newfoundland, and the Samail section of Oman, have been described in more detail [Salisbury and Christensen, 1975; Christensen and Smewing, 1981; Manghnani et al., 1981] and have been modelled previously [Kempner and Gettrust, 1981a, b]. As a corollary to Christensen's [1978] work, synthetic seismograms of the three remaining ophiolites; the Vourinos complex of northern Greece, the Troodos complex of Cyprus, and the Papuan ophiolite of New Guinea, are modelled. The $V_p$ and $V_s$ velocity-depth functions for which the synthetics have been computed are presented in Figure 31. Our results show that the synthetic seismograms generated from these ophiolite models do not have the same seismic signature as marine refraction data for "normal" oceanic crust. This suggests that greater care must be taken not to impose similar generic interpretations for all ophiolite suites.
Figure 31 Stratigraphic columns and estimated compressional ($V_p$) and shear ($V_s$) wave velocities for 3 major ophiolites [from Christensen, 1978].
Methods

The synthetic seismograms were generated using the reflectivity method described by Fuchs and Muller [1971] and Kennett [1975]. All synthetic seismograms were generated for the marine case with the seismic source at the surface and the receiver at the ocean-sediment interface. The number of layers in each model is a function of the velocity gradient. The criterion for a "layer" was set at 0.1 km/sec/layer. This limit on the maximum allowable rate of change ensures proper reflectivity calculations. A cosine-squared frequency window from 3 to 12 Hz was used with an 8 Hz ricker wavelet source signal. Minimum and maximum phase velocities were set at 3 and 15 km/sec, respectively, for all models. This ensures that no sediment or water-wave arrivals will interfere with the crustal refractions. All synthetic seismograms are reduced over 6.5 km/sec. The record sections extend to 120 km range with a 2.5 km shot spacing. Density values are empirically formulated based on the equation \( P = 0.252 + 0.3788 * V_p \). Attenuation coefficients for both shear and compressional velocities were 0.0005 and 0.0001 for the sediment and 0.001 and 0.002 for the igneous rocks, respectively.
Synthetic Seismograms

The Vourinos Ophiolite

The Vourinos ophiolite complex of Greece is typical of many ophiolite suites in that the petrologic section is extremely thin compared to ocean basin refraction interpretations. This is demonstrated in the synthetic seismograms (Figure 32) and the travel-time curve of the ophiolite (Figure 33) by a mantle crossover range of less than 12 km. The lack of distinct second-arrivals in the seismogram, other than multiple reflected refractions [see Meissner, 1969; Lewis, 1978; and Kempner and Gettrust, 1981a, b; for an explanation of raypaths], is suggestive of the gradient structure found in the velocity-depth profile. The few high amplitude arrivals between 11 and 14 km range are products of the reflections from the Moho discontinuity. This interpretation is supported by the raytracing and travel time curve. The multiple of the mantle reflection appears as a high-amplitude second-arrival in the 22 to 28 km range. At greater ranges the mantle refractor and its multiple dominate the arrivals.

The Troodos Ophiolite

It has been suggested that the Troodos ophiolite could be representative of crust generated either in a back-arc basin environment [Miyashiro, 1973; Upadhyay and Neale, 1981] or at a more typical oceanic spreading center [Vine and Moores, 1972]. The compressional ($V_p$) velocity-depth function for Troodos (Figure 34) shows that this ophiolite is slightly thicker than the Vourinos section. The synthetic
Figure 32 The P-velocity depth function of the Vourinos ophiolite with the synthetic seismograms generated from the model. The synthetic seismograms in (c) are plotted with each trace scaled to the largest amplitude of the first trace, and adjusted for range using a range scaling factor of $(r/r_d)^2$ where $r_0 = 2.5 \text{ km}$ for the first trace. Both seismograms are reduced over 6.5 km/sec. The distinct coherent high amplitude first arrivals in (b) at a range of 30 to 120 km are mantle refraction arrivals followed by mantle reflected refractions.
VOURINOS OPHIOLITE, GREECE

(a)

(b)

(c)
Figure 33  The P-velocity depth function of the Vourinos ophiolite with the travel time curve and raypaths for the ophiolite model. The crosses (x) on the travel time curve in (b) display the time of arrival associated with each raypath shown below in (c). The dashed line represents the mantle refractor. The travel time section is reduced over 6.5 km/sec.
VOURINOS OPHIOLITE, GREECE

(b)

RV = 6.50

(a)

(c)
seismogram for the Troodos model has distinctly different seismic characteristics than that for Vourinos. The uppermost seismic velocity gradient of the velocity-depth function for this ophiolite has very little effect on the amplitude of arrivals at ranges greater than 7 km in the seismogram. This suggests a need for near receiver, high density shot spacing to resolve upper crustal structure detail. The discontinuous increase in P-velocity from 5.3 km/sec to a 6.8 km/sec refractor produces a triplication and corresponding increase in amplitude at approximately 8 km range (Figure 35). The existence of this 6.8 km/sec refracting horizon (associated with the amphibolite sheeted dikes and the thin plutonic sequence) suggests that the layer 2-layer 3 boundary is a metamorphic facies boundary rather than a major petrologic boundary [Cann, 1974; DeWit and Stern, 1976]. The (PmP) mantle reflection branch dominates the second-arrivals in the remainder of the synthetic seismogram at ranges greater than 10 km. The synthetics show that it is impossible to resolve either the low-velocity trondhjemite layers in the upper plutonic section of the ophiolite, or the thin basal P-high-velocity cumulates given the frequencies and wavelengths used in these computations. Because of the limited bandwidth the P-velocity gradient between the crust and mantle produces signals characteristic of a first order discontinuity. Again, multiple reflected refractions are apparent as distinctive second arrivals.
Figure 34 The P-velocity depth function of the Troodos ophiolite with the synthetic seismograms generated from the model. Amplitudes for (b) and (c) are scaled in the same manner as in Figure 32. The continuous high amplitude second-arrivals at a range of 30 to 120 km represent reflected mantle arrivals. The steeply dipping low amplitude first arrivals at the same range are mantle refractions.
TROODOS OPHIOLITE, CYPRUS

![Graph (a)](image)

![Graph (b)](image)

![Graph (c)](image)
Figure 35  The P-velocity depth function of the Troodos ophiolite with the travel time curve and raypaths for the ophiolite model. The diagrams are plotted in the same manner as in Figure 33.
TROODOS OPHIOLITE, CYPRUS

(a) Depth (km)

(b) Range (km)

(c) Velocity (km/sec)
The Papuan Ophiolite

The P-velocity depth function of the thick Papuan section is, at most, a simplification of the actual velocity structure. The ophiolite section is composed of three homogeneous layers. These are: (1) a thin ultramafic layer at the base of the crustal section with a P-velocity of 7.5 km/sec; (2) the overlying layered gabbros (7.2 km/sec); and (3) the greenschist spilites (6.5 km/sec) and thick zeolitic basaltic sequence at the top of the crust [Davies, 1971, 1977]. Figures 36 and 37 show the synthetic seismogram and travel time curve for the Papuan ophiolite velocity-depth model, respectively. The synthetic seismogram exhibits a more complex seismic signature than the two ophiolite synthetic seismograms previously described. If interpretation of this seismic data were based on first arrivals only, (e.g., the Tau-P inversion method) the seismic velocity structure for the Papuan ophiolite structure would be resolved. To obtain a correct inversion from seismic refraction data both amplitude and second-arrival information has to be incorporated in the inversion technique. The need for this resolution for proper inversion of refraction data is exemplified in the Papuan synthetics. Much of the information necessary for proper inversion (including second-arrivals) is produced at distances greater than refraction lines are normally extended. The Papuan synthetic seismograms show the need to extend seismic refraction shot lines to ranges greater than 80 km. Lewis [1978] pointed out similar problems in ocean refraction experiments.
Figure 36 The P-velocity depth function of the Papua ophiolite with the synthetic seismograms generated from the model. Amplitudes for (b) and (c) are scaled in the same manner as in Figure 32. The synthetic seismograms are characterized by numerous reflection branches seen as coherent high amplitude second-arrivals at ranges greater than 70 km. Reflected refractions are easily seen in (c) at a range of 70 to 80 km with an arrival time of approximately 5.0 seconds.
PAPUA OPHIOLITE, NEW GUINEA

(a)

(b)

(c)
Figure 37 The P-velocity depth function of the Papua ophiolite with the travel time curve and raypaths for the ophiolite model. The diagrams are plotted in the same manner as in Figure 33.
PAPUA
OPHIOLITE,
NEW GUINEA

(b)

(c)
Although the Papuan ophiolite is too thick to be a reasonable example for mature oceanic crust, many features found in the synthetics for the model are similar to some ocean basin seismic refraction data. The first major P-velocity boundary, marked by an increase in velocity from 5.5 km/sec to 6.5 km/sec at the zeolite-greenschist transition, focuses energy in the 10 to 12 km range as seen in the synthetic seismogram in Figure 36 and the travel-time curve (Figure 37). The homogeneous nature of the thick greenschist basalt layer results in small low-amplitude first-arrivals from 15 to 20 km range. Reflection branches from both the 6.5 and 7.2 km/sec refracting horizons dominate the second-arrivals out to 120 km range and defines the strong gradients or first order discontinuities within this crustal model. The 7.5 km/sec layer at the base of the crust generates a small refraction branch. This refractor is difficult to identify in the synthetic seismogram and would not be resolvable in actual seismic refraction data. The extremely small differences in the arrival time for these reflected arrivals generally results in an interpretation consisting of a single (PmP) mantle reflection branch; the result is a misinterpreted crustal section. Similar results have been found for synthetic seismograms generated for the Bay of Islands ophiolite [Kempner and Gettrust, 1981a].
Discussion

Resolution of Data

It is apparent that seismic refraction data cannot discern the exact details of the ocean crustal structure. For example, interpretations of synthetic seismograms for ophiolites, based on first arrival information, tend to smooth the rapid or discontinuous seismic velocity transitions into seismic velocity gradients. Our results show a need for longer seismic refraction lines that delineate the reflection and refraction branches in the data to constrain interpretations of the velocity transitions within the crust. Control of the relative amplitude of second arrivals is also required if these arrivals are to be used to constrain the inversion of seismic data. Synthetic seismograms will play a more important role in constraining interpretations of seismic data if the field experiments are designed to properly sample second arrivals and multiple reflected refractions. [Helmberger, 1976; Kempner and Gettrust 1981a, b].

The Ophiolites

Thin petrologic sections for both the Troodos and Vourinos ophiolites produce very distinctive synthetic seismograms. Using the "infinite onion" theory of Cann [1974] and the plutonic/intrusive arguments of Kusznir [1980], the petrologic sections of both ophiolites suggest slow spreading rates. Seismic refraction data from the crustal section near a ridge-transform junction in the Marianas back-arc basin,
Hussong and Sinton, 1980, per. comm.] show seismic characteristics similar to those found in the Vourinos model. On the other hand, the synthetic seismogram for the Troodos model is not similar to available marine refraction data.

If we imply that the Vourinos or Troodos ophiolites are crustal sections for very young crustal sections generated at slow spreading centers, it is then necessary to have an aging process for the crust whereby the crustal sections thicken more than 2 km with age. Nichols et al. [1980], modeling the Point Sal ophiolite of the California Coast Range, addressed this problem by suggesting continuous serpentinization of mantle material to greater depths with time thereby creating a low velocity zone at the base of the crust. For mature crustal sections this P-low-velocity zone should have a distinct seismic signature. Their synthetic model is not conclusive. Even though Lewis and Syndsman [1977; 1979] interpret a thick P-low-velocity zone at the base of the crust on the Cocos plate their overlying crustal sections are generally greater in thickness than either the Troodos or Vourinos sections. Kempner and Gettrust [1981a, b] have shown that a thick P-low-velocity zone at the base of the crust is not required to fit synthetic seismograms to marine seismic data for mature ocean crust formed in moderate to fast spreading environments. Synthetic seismograms for Troodos with an appropriately thick P-low-velocity zone (on the order of 2 km) cannot produce second arrivals comparable to oceanic refraction [Lewis, pers. comm.]. This suggested that a thick (greater than 1.0 km) serpentinized basal section is not necessary and leads to further
questions concerning the origin of thin ophiolite sections. Our data show that neither ophiolite, as presently described, represents a "normal" oceanic section.

Although the overall structure of the synthetic seismogram for the Papuan ophiolite is not characteristic of marine refraction data for oceanic crust, the synthetic seismograms do show some salient features found in refraction data for mature ocean basin crust, particularly in the "layer 3" seismic structure. The ophiolite section is greater in thickness than suggested for "normal" ocean crust. If the Papua ophiolite has plateau affinities the 7.2 km/sec horizon must be thicker [Gettrust et al., 1980], and the basal high velocity 7.5 km/sec layer needs to be thick enough to generate a distinct arrival as seen in some plateaus [Carlson et al., 1979]. These features are not apparent on the Papua synthetic seismogram. The Papua synthetics show seismic characteristics of oceanic crust but an excessive outpouring of basalt and a complex thermal overprinting now controls much of the seismic signature.

Conclusions

Although these ophiolite models are based on estimated velocity structures of complex ophiolitic sections, which may be tectonically dismembered, the synthetic seismograms represent valid first-order approximations of the seismic structure of the ophiolites. As a result of modeling these more variable ophiolite sections, it has become
necessary to question the assumptions that crustal processes, associated with different spreading rates and regional tectonic environments, produce remarkably similar crustal sections throughout geologic history, independent of basin age, dynamics, and size. The results from the synthetics seismograms for these three ophiolites show that not all ophiolites are capable of producing seismic signatures equivalent to marine refraction observations for "normal" oceanic crust.
In this project we evaluate the hypothesis that ophiolites are representative sections of oceanic crust. The method employed to test this hypothesis has been to generate synthetic seismograms for five different ophiolite sequences and to compare synthetic seismograms for the most stratigraphically complete sections with marine seismic refraction data. We have considered the details of spreading rates, crust and ophiolite ages, stratigraphic continuity of ophiolite sections, lateral heterogeneity within both crustal and ophiolite sections, and the nature of the serpentinization found within basal sections of many ophiolites. Assumptions have been made concerning the seismic velocity structure and spreading rates of the ophiolite suites. We also assume that the seismic velocity structure determined for ophiolites represents the velocity-depth function of oceanic crust which has been "frozen" into the ophiolite section at the time of emplacement. Although post-emplacement serpentinization of the ultramafics is evident in most ophiolites, the seismic velocity structure used in the modeling was initially corrected to exclude the effects of the serpentinization. Approximate spreading rates for the ophiolites are based on plutonic/intrusive relations within the ophiolites. The validity of these assumptions and techniques is supported by the excellent agreement between OBS refraction data for different age crustal sections and synthetic seismograms for ophiolite sections of appropriate ages.
(Chapter II and III). In particular, synthetic seismograms for the BOI ophiolite (45 m.y. age) are indistinguishable from marine seismic refraction data from crust of 60 m.y. age (Chapter II). Similarly, synthetic seismograms for the (5-8 m.y. age Samail ophiolite are in excellent agreement with refraction observations from 4.5 m.y. crust obtained during the ROSE Experiment. These results, together with petrologic and geochemical data for the ophiolites, allow us to propose a model (Chapter III) whereby young crust with slowly varying seismic velocity gradients is altered to a crustal section composed of seismically discrete layers (Figure 38). This has many implications for the structure of the ocean crust and some ophiolites. These are:

(1) Near ridge hydrothermal convection initially alters the rocks of the upper crust and thereby modifies the seismic velocity in layer 2 and the upper part of layer 3.

(2) Although water may penetrate to 5 km depth within ophiolites, as has been proposed for the Samail ophiolite [Gregory and Taylor, 1981], there appears to be little effect on the velocity of the lower structure of the crust at less than 5 m.y. age. Comparison of ROSE seismic data with synthetic seismograms for the Samail ophiolite does not support a P-low-velocity zone (serpentinization) at the base of the young crust.

(3) As the crust ages, the heat transfer mechanism becomes dominately conductive and two different processes may occur which can alter the seismic velocity structure of the mid- and lower crust. If seawater is present at 5-7 km depth in the crust, then simple cooling will be
Figure 38 An aging model for the oceanic crust. The model accounts for both petrologic and seismic variation of the oceanic crust through time as inferred from the results of the comparison of marine seismic refraction data and synthetic seismograms from ophiolites.
(after Dewit & Stern, 1976)
sufficient to lower crustal and upper mantle P-velocities. This alteration with time should follow simple isotherm contours. If seawater is not present, then a slow diffusion of seawater downward is required to alter the mid-crustal seismic velocity structure and, eventually, the basal section of the crust. This serpentinization process does not appear to occur until after 12 m.y. age for crust formed at moderate to fast spreading centers (5 cm/yr half-rate). These conclusions are consistent with petrologic controls from ophiolite complexes and the marine seismic refraction observations.

The primary focus of this dissertation has been to substantiate the hypothesis that ophiolites represent oceanic crustal sections. However, the seismic modeling of the Vourinos ophiolite, the Troodos complex, and the Papua ophiolite, which cannot represent "normal" oceanic crust, may be more important to the overall understanding of the origin of the ophiolite sections. The velocity-depth functions for these ophiolites do not generate synthetic seismograms that fit typical marine seismic refraction data. Thus, to impose a simple genetic model for all ophiolites is not valid.

Given the results from the comparison of the Samail ophiolite and the BOI ophiolite with marine refraction data, it has been shown that carefully selected ophiolites are seismically indistinguishable from oceanic crust. Thus some some ophiolites do accurately represent sections of normal oceanic crust.
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