Investigation of Upper Mantle Discontinuities Near Northwestern Pacific Subduction Zones Using Precursors to $sS$H

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Teleseismic long-period World-Wide Standard Seismograph Network (WWSSN) tangential component recordings of deep and intermediate depth earthquakes are analyzed for the presence of $sS$ precursors, denoted $s_xS$, signifying underside reflections from discontinuities at depth $x$ below the $sS$ reflection point above the source. These $sS$ precursors can be used to place constraints on upper mantle discontinuities in the vicinity of subduction zones. The clearest precursor is usually the $s_mS$ phase, the underside reflection from the Moho, which is observed for a number of events in the northwestern Pacific for paths reflecting under the Sea of Okhotsk and North Korea. The amplitude and timing of $s_mS$ relative to $sS$ are modeled using synthetic seismograms to determine the shear wave impedance contrast at the Moho and the crustal thickness in these regions. The results are compared with previous work on $Pp$ precursors reflected from the Moho for similar paths, with consistent crustal thickness being found. $s_mS$ is strong for continental Moho properties unless broadband data are used. Other precursors, $s_xS$, are much weaker than $s_mS$ and are difficult to identify in individual waveforms. We use slant stacking to enhance the signal to noise and search for precursors with various slownesses, using recordings from 13 deep events. Four $SII$ wave reflectors may exist below the Moho in our study area. The shallowest is the "400-km" discontinuity, with the depth varying from 380 to 400 km, perhaps indicating slight elevation of the olivine-$\beta$ phase transformation near the slabs.

**INTRODUCTION**

Investigation of upper mantle discontinuities is an important step in understanding fine structure, dynamics and chemical evolution of Earth. Of the various remote sensing techniques employed, seismology plays a prominent role in studying mantle structure. By the 1930s, the gross radially symmetric seismic velocity distribution inside the planet was established, showing regions of different seismic character (crust, upper mantle, lower mantle, inner and outer core) corresponding to gross compositional divisions of Earth. With the advent of digital data and refinement of interpretational techniques, modern seismology has revealed that the upper mantle has global layering associated with the Moho, "410-km", "520-km" and "660-km" discontinuities and other regionally varying structures such as the "60-km", "80-km", "220-km" and "330-km" discontinuities.

The seismological Moho, defining the upper boundary of the mantle, has been widely mapped based on the abrupt increase of seismic velocity with depth inferred from earthquake and controlled source seismology, mainly using $PmP$ (reflections from the top of the Moho) or $Pn$ (Moho head waves). In the preliminary reference Earth model (PREM) [Dziewonski and Anderson, 1981], the average shear and compressional impedance contrasts at the Moho are about 29.2% and 32.6%, respectively. Lerner-Lam and Jordan [1987] constructed models EU2 and PA2 for northern Eurasia and the western Pacific Ocean, respectively, using fundamental and higher-mode Rayleigh waves. The shear and compressional impedance contrasts at the Moho are 26.0% and 28.8% in EU2 and 36.8% and 38.4% in PA2, respectively. These models imply that the impedance contrast at the Moho is significantly lower in continental than in oceanic regions. However, there are many regions for which Moho properties, particularly for shear waves, remain undetermined.

The 410- and 660-km discontinuities are global features explored by precursors to $Pp$ [Whitcomb and Anderson, 1970; Husebye et al., 1977; Nakanishi, 1986, 1988], waveform modeling [Burdick and Helmberger, 1978; Grand and Helmberger, 1984a,b; LeFevre and Helmberger, 1989], stacking of reflected and converted phases [Bock and Ha, 1984; Wajeman, 1988; Paulsen, 1988; Bowman and Kennett, 1990; Richards and Wicks, 1990; Shearer, 1990, 1991; Vidale and Benaz, 1992], and migration of mantle reverberations $ScS$ [Revenaugh and Jordan, 1987, 1989, 1991a,b]. Based on these previous studies, average $P$ velocity jumps of 5-6% at 410-km and 4% at 660-km, and $S$ velocity jumps of about 5% at 410-km and 7.8% at 600-km are inferred all with a ± 1-2% uncertainty. The long-wavelength (>50 km) topology for these two discontinuities is on the order of ± 20-30 km. Ringwood [1975] suggested that the 410-km discontinuity is caused by the phase change from olivine to $\beta$-phase for (Mg,Fe)$_2$SiO$_4$. The
670-km discontinuity can be explained by the transition γ-spinel → perovskite + magnesiowüstite [Liu, 1976, 1979; Jackson, 1983; Ito et al., 1984; Ito and Takahashi, 1989]. Garnets also transform to perovskite structure near this depth with a two-phase region several gigapascals in extent (not an efficient reflector at high frequency).

The 520-km discontinuity is potentially a global feature [Whitcomb and Anderson, 1970; Helmberger and Engen, 1974; Jones and Helmberger, 1990; Revenaugh, 1989; Revenaugh and Jordan, 1991b; Shearer, 1990, 1991] but is very weak. The shear impedance contrast at 520-km discontinuity is about 3% [Revenaugh and Jordan, 1991b]. The phase transition of β-phase → γ-spinel is a possible explanation for the 520-km discontinuity [Whitcomb and Anderson, 1970, Ringwood, 1975; Weidner et al., 1984].

Other upper mantle discontinuities have been proposed based on seismological observations, such as the 60-km, 80-km [Revenaugh and Jordan, 1991c], 220-km [Jordan and Frazer, 1975; Hales et al., 1980; Drummond et al., 1982; Bowman and Kennett, 1990; Revenaugh and Jordan, 1991c; Vidale and Benz, 1992], and 330-km [Weidner, 1988; Revenaugh and Jordan, 1991c] discontinuities, but the universality of these features is not yet established nor is their underlying physical explanation.

It is critical to determine the properties of all upper mantle discontinuities and to establish whether they are global or regional structures. Properties that seismology can resolve include the sharpness and size of velocity, density, and impedance contrasts of these discontinuities. This information is needed for mineral physics experiments to determine if the discontinuities can be explained by phase changes or, failing that, must be associated with compositional or other changes such as partial melting or anisotropic fabric. Resolving the nature of these discontinuities will undoubtedly contribute substantially to understanding the dynamics of mantle convection. Of course, no single method or data set can hope to resolve all properties of a given discontinuity, and a variety of methods must be used to fully characterize the upper mantle.

In this paper we focus on analysis of precursors to the transverse component aSH phase (below we write S only for the SH component, which we use exclusively) on long-period recordings for deep earthquakes to analyze the regional mantle structure above the sources. Our procedure is most sensitive to mantle structure above and near to subducting slabs, where strong thermal and chemical perturbations may affect mantle discontinuities. We determine crustal thickness and shear wave impedance contrast at the Moho using nS, the Moho underside reflection, and investigate upper mantle discontinuities using sS, underside reflections from discontinuities at x km depth in the mantle above deep sources in the northwestern Pacific region.

**Data**

Seismic waves that encounter discontinuities in the upper mantle spawn many phases. We use only transverse components (SH waves) in this study since they have particularly simple interactions, with no conversion of S to P energy. Data from 23 intermediate and deep events are analyzed in this study (Table 1). Figure 1 shows the location of these 23 earthquakes in the northwestern Pacific region. The depths of these events range from 115 to 580 km, and the magnitudes vary from 5.6 to 6.0. Table 1 lists the International Seismic Centre (ISC) source parameters, along with focal mechanisms from Gaherty and Lay [1992].

**TABLE 1. Earthquake Parameters (ISC) for Events Used in This Study**

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<th>No.</th>
<th>Date</th>
<th>Time, GMT</th>
<th>Latitude, deg</th>
<th>Longitude, deg</th>
<th>Depth, km</th>
<th>mb</th>
<th>Strike, deg</th>
<th>Dip, deg</th>
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<td>142.58</td>
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<td>5.6</td>
<td>274</td>
<td>88</td>
<td>80</td>
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<td>162</td>
<td>5.9</td>
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<td>38</td>
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<td>51.93</td>
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The sensitivity of a reflected seismic wave to the vertical extent (sharpness) of the velocity or impedance increase across a discontinuity is a function of the signal wavelength [Richards, 1972]. Large wavelength, long-period waves vertically incident on a boundary will be reflected effectively only if the change in material properties is distributed over a depth interval less than \(-1/4\) wavelength. Near vertical incidence reflection coefficients are only sensitive to impedance contrast (Δ(velocity x density)). Thus, if we analyze long-period waves traveling nearly vertically through the upper mantle, we can constrain gross characteristics of upper mantle impedance structure. We examine long-period SH waves propagating upward from deep earthquakes that reflect from discontinuities and travel to teleseismic distances. By using only long-period World Wide Standardized Seismic Network (WWSSN) records in this study (dominant periods close to 20 s and wavelengths near 100 km) we are able to image mantle shear velocity impedance contrasts that are distributed over as much as 25-30 km in depth. The criteria for choosing our data are a high ratio of signal to noise and clear direct S (to use for a source pulse) and surface reflection SS phases. The epicentral distance range in our data is from 30° to 95° to avoid upper mantle triplications and core diffraction.

We define precursors to SS as $s_x S$, the upgoing S wave reflected from the underside of a discontinuity above the source at a depth of $x$ km. These phases arrive at teleseismic distances after direct S but before the surface reflection SS. Figure 2a shows the geometry of ray paths for $s_x S$, where $x$ can be any depth at which there is an impedance contrast above the source. Underside reflections from the Moho are given the special name $s_m S$. For example, if the source depth is 560 km, then precursors may be generated by the Moho, 80-km, 220-km, 330-km, 400-km, or 520-km discontinuities if they exist, and are then labeled as $s_m S$, $s_{80} S$, $s_{220} S$, $s_{330} S$, $s_{400} S$ and $s_{520} S$, respectively. In the teleseismic distance range of our data, the core-mantle boundary reflection ScS may arrive between S and SS phases in the SS interval. Figure 2b is a sketch of a teleseismic waveform corresponding to the ray paths of Figure 2a, suggesting that at some distances each $s_x S$ should be well separated from ScS and thus observable. Figure 2c shows travel time curves for S, SS, ScS, and $s_x S$ phases for a source depth of 560 km.

As an example of our waveform data quality, we show the entire set of waveforms for one representative earthquake, that of August 16, 1979, in Figure 3. The data are globally distributed; thus the waves sample a cone of mantle material above the source. The horizontal components were digitized and rotated to give these tangential components, and polarity reversals due to radiation pattern have been corrected for. The SS phase on all traces are aligned, and we can see the clear, large amplitude arrivals of SS and S phases. The lines give differential travel time curves calculated with the PREM velocity structure, showing the arrival times expected for possible $s_x S$ precursors $s_m S$, $s_{80} S$, $s_{220} S$, and $s_{400} S$ between S and SS. We computed $s_m S$ travel time for a 36-km crustal thickness. Note that these curves are not based on actual arrivals. The
travel time curve for ScS is also shown and clearly has associated arrivals. In the plot we can see a small wiggle about 18 s before the sS phase in most traces, which is the Moho underside reflection smS. Identifying smS in each individual trace is sometimes difficult, but it is even harder to see any other precursors between S and sS. The noise levels are quite low for this event, and the data can be used to place bounds on the size of any upper mantle discontinuities. However, since the mantle discontinuities (above the sources) are expected to be less than half the size of the Moho arrival, we will stack the traces to enhance the signal to noise ratio before drawing any conclusions. Since the Moho produces the strongest and most coherent reflection of the upper mantle discontinuities, we are able to identify smS in the stacked traces readily and to actually model the waveforms to determine crustal thickness and shear wave impedance contrast at the Moho.

As shown in Figure 3, underside Moho reflections smS can be seen in some traces, but it is hard to identify the phase in others. The only significant difference between sS and smS is the path through the crust for sS. If the crust has laterally uniform properties above the source, then all of the arrivals at different azimuths should have similar waveforms. Our events have stable upgoing SH radiation patterns, so the wave shapes are very similar. To increase the signal to noise ratio, we stack the signals. We align the traces for a given event on the sS phase and normalize the peak amplitudes to unity, then stack the arrivals. This is valid because we expect little moveout between smS and sS at teleseismic distances. Figure 4 shows stacked waveforms for all 23 events, some of which have very clear and isolated smS, such as the events on February 1, 1984, April 20, 1984, August 21, 1972, August 16, 1979, September 5, 1970, and September 10, 1973, while some lack a clear smS. Note that alignment on sS causes earlier phases like S and ScS or deeper sS phases to stack incoherently given the large distance range spanned for each event. A total of 451 individual seismograms is included in Figure 4. Since oceanic crust is expected to be less than 10 km thick, the Moho underside reflection smS will overlap the sS arrival in long-period waveforms; therefore, it is very hard to identify smS from oceanic Moho, as for events on March 31, 1969, December 12, 1976, and February 22, 1974, (see Figures 1 and 4).

Other precursors, sS, are much weaker than smS, and they have slightly different slownesses as indicated by the travel time curves in Figure 3, so later we will slant stack the waveforms at various slownesses to seek evidence for any upper mantle discontinuities in the sS precursor window.
ANALYSIS OF $s_0 s$ PHASES TO DETERMINE PROPERTIES OF THE MOHO

We use the September 5, 1970, deep Kurile slab earthquake (Table 1) to develop our modeling method and to estimate the average crustal thickness and shear impedance contrast at the Moho near the $s_0 s$ reflection point under the northern Sea of Okhotsk. We have a total of 58 digitized WWSSN transverse component records available for this event. Twenty traces of $s_0 s$ phases and 34 traces of $s$ phases are chosen for stacking on each phase for this event, at distances where the phases are well isolated. We separately stack the data aligned on the $s_0 s$ and $s$ phases to enhance the ratio of signal to noise. The stacked $s$ pulse is used as a source pulse and the averaged $s$ waveform is modeled with synthetic waveforms. Clearly, this stacking involves lateral averaging over the region sampled by $s_0 s$, as well as the intrinsic averaging of the long wavelength waves. However, the gain in signal stability gives us confidence that the average structure can be determined robustly from the event-averaged data. Later we will consider subdivisions of the data based on reflection point grouping.

In synthetic modeling, the computed seismogram, $S(t)$, is a convolutional product of time series:

$$S(t) = X(t) * E(t) * I(t)$$

(1)

where $X(t)$ is the source time history, $E(t)$ is the Earth transfer function, and $I(t)$ is the instrument impulse response. For a simple rupture, it is sufficient to treat the source as a point source for which the spectrum of upgoing and downgoing radiation from the source is the same, with separate scaling factors due to the difference in radiation pattern for each ray parameter. At teleseismic distances the Earth transfer function is essentially a sequence of arrivals with different geometric spreading and attenuation filters convolved with the receiver function. The isolated downgoing $S$ phase contains the attenuation contribution from the lower mantle and near the receiver, along with any receiver reverberations. Upgoing phases, such as $s_0 s$, have additional attenuation and accompanying multiple arrivals from underside and topside reflections from any discontinuities above the source. To model the $s_0 s$ phase and its precursors, we use the $S$ phase as a source wavelet containing the deep mantle and receiver effects, the source radiation, and the instrument response. The additional factors that have to be accounted for thus include the extra attenuation of the $s_0 s$ phases and the interactions with any velocity discontinuities above the source. We determine precursor characteristics relative to $s_0 s$, so scalar radiation pattern differences of the downgoing phase are not important, being eliminated by normalization. The paths of $s_0 s$ and $s$ in the mantle are very similar, the difference between the two arises near their reflection points. For deep foci, the incident wave at the base of the crust is nearly planar, and a simple plane wave reflectivity method is sufficient for the calculation of synthetic $s_0 s$ and $s$ seismograms. There are many parameters that affect the synthetic modeling, such as ray parameter $p$; $s_0 s$ differential attenuation relative to $S$, $\Delta \alpha s^*$; crustal thickness $h$; and shear velocity and density contrast at the Moho, $\Delta \beta$ and $\Delta \rho$; but which plays an important role in the computation? We discuss this at some length since our procedure differs from previous methods used to study underside Moho reflections (e.g., Revenaugh and Jordan, 1989).

**Parameter Search**

We first test the sensitivity of synthetic waveforms to the five parameters: $\rho$, $\Delta \alpha s^*$, $h$, $\Delta \beta$, and $\Delta \rho$. For each test, only one parameter is varied. For the range of epicentral distance of our $s_0 s$ data (September 5, 1970 event) from 50° to 90°, the parameters $p$ and $\Delta \alpha s^*$ were computed for PREM: $p$ changes from 9.5 to 14.5 s/deg, and $\Delta \alpha s^*$ varies from 1.5 to 3.0 s. The average $p$ and $\Delta \alpha s^*$ for our data samples are 11.81 s/deg and 2.56 s, respectively. For the numerical tests, we prescribe a shear wave surface velocity $\beta_s = 3.2$ km/s and surface density $\rho_s = 2.6 \times 10^3$ kg/m$^3$, uppermost mantle shear velocity $\beta_m = 4.7$ km/s and density $\rho_m = 3.38 \times 10^3$ kg/m$^3$, crustal thickness $h = 25$ km, $\Delta \alpha s^* = 2.56$ s, $p = 11.81$ s/deg, $\Delta \beta = 0.7$ km/s, and $\Delta \rho = 0.4 \times 10^3$ kg/m$^3$, when each particular parameter is not being varied. The crust has a linear velocity and density gradient from the surface to the Moho, approximated by many thin layers in the reflectivity modeling.

In Figure 3a the effect of variable ray parameter $p$ is shown in superimposed waveforms for the range $p = 9.5-14.5$ s/deg; the variation interval is 0.5 s/deg. All other parameters are fixed at the values listed above. These signals show only a small change in waveform as ray parameter varies, indicating the possible smoothing effect of stacking the data. Since the data are actually concentrated in a limited range near 75°, this is a minor bias. Thus, in the following analysis we fix an average value of $p = 11.81$ s/deg, which is an appropriate average for our stacked $s_0 s$ data.

The waveform sensitivity to $\Delta \alpha s^*$ is shown in Figure 5b. $\Delta \alpha s^*$ varies from 1.5 to 3.0 s, with interval of 0.1 s. Although the change is visible, it is quite small. Comparisons with observed waveforms will only be affected if we compare with the entire $s_0 s$ waveform, including its instrumental overshoot.
Fig. 5. Synthetic waveforms of \( s_S \) superimposed waveforms (a) for \( p = 9.5 \) to 14.5 s/deg with an interval of 0.5 s/deg; (b) for \( \Delta t_{ss}^* = 1.5 \) to 3.0 s with an interval of 0.1 s; (c) for \( h = 15 \) to 30 km with a 1-km interval; note that travel time difference between \( s_S \) and \( s_mS \) increases with increasing \( h \); and (d) for \( \Delta \rho = 0.05 \) to 0.75x10^3 kg/m^3 with an interval of 0.05x10^3 kg/m^3; note that \( \Delta \rho \) influences the ratio of \( s_mS/s_S \). Other model parameters are held fixed at the value given in the text.

Therefore we choose an average \( \Delta t_{ss}^* = 2.56 \) s in the modeling below and use a short window which excludes the \( s_S \) waveform overshoot.

Figure 5c shows the waveform sensitivity to crustal thickness \( h \) at the reflection point, which varies from 15 to 30 km. The substantial waveform variation shown in this figure is due to the travel time difference between \( s_S \) and \( s_mS \), which increases with increasing crustal thickness. The ratio of \( s_S \) and \( s_mS \) amplitude also changes due to the variable interference, but this is less pronounced than the time difference effect. Crustal thickness is mainly determined by the difference in travel time between \( s_S \) and \( s_mS \). It is clear that we should avoid stacking waveforms that sample regions with varying crustal thickness. This prevents us from stacking multiple events, as the effective footprint of \( s_mS \) becomes too large.

Figure 5d is an examination of the waveform sensitivity to density variation \( \Delta \rho \) at the Moho reflection point, which ranges from 0.05x10^3 to 0.75x10^3 kg/m^3, with a step of 0.05x10^3 kg/m^3. The surface and Moho densities are fixed at \( \rho_s = 2.6x10^3 \) kg/m^3 and \( \rho_m = 3.38x10^3 \) kg/m^3, respectively, and the crust has a linear gradient from 2.6x10^3 kg/m^3 at the surface to the density at the base of the crust given by \( \rho_m - \Delta \rho \) kg/m^3. The amplitude ratio \( s_mS/s_S \) increases as the density contrast \( \Delta \rho \) increases as shown in the superimposed traces for different \( \Delta \rho \) models. There is no travel time difference between \( s_S \) and \( s_mS \) due to varying average density, since we do not perturb the velocities, although the apparent rise time of the \( s_mS \) phase varies subtly.

We also test \( \Delta \beta \) for waveform sensitivity. If we keep the average crustal velocity constant, identical waveforms can be produced by varying \( \Delta \beta \) and \( \Delta \rho \) since the reflection coefficient at near-vertical incidence is primarily sensitive to the impedance contrast \( \Delta (\rho \beta) \). The \( s_mS/s_S \) amplitude ratio provides our estimate of the impedance contrast.

**Best Model Search**

The tests above show that the parameters \( h, \Delta \beta, \) and \( \Delta \rho \) affect the waveform modeling much more than the parameters \( \Delta t_{ss}^* \) and \( p \). Thus we use average values of \( p = 11.81 \) s/deg and \( \Delta t_{ss}^* = 2.56 \) s in the following data analysis. As \( \Delta \rho \) and \( \Delta \beta \) trade off directly in the modeling, we combine these two to determine the impedance contrast \( \Delta (\rho \beta) \) in the following. Since the velocity and density are likely to be related to each other, we specify the absolute density associated with a given velocity model using the Nafe-Drake relation [Grant and West, 1965]:

\[
\rho_c = 0.173\alpha_c + 1.695
\]

where \( \rho_c \) is density in the crust, and \( \alpha_c \) is the \( P \) wave velocity. We assume \( \alpha_c = \sqrt{3}\beta \), where \( \beta \) is the shear velocity in the crust. We assume a linear crustal velocity gradient throughout the crust and do not have any midcrustal discontinuities. Therefore, the free parameters in the synthetic computations are the surface velocity \( \beta_s \), the Moho velocity \( \beta_m \), \( \Delta t_{ss}^* \), the crustal thickness \( h \), and the impedance contrast \( \Delta (\rho \beta) \) at the Moho.
We allow $\beta_o$, $\beta_m$, and $\Delta t_s^*$ to vary over a limited range compatible with reasonable crustal and mantle velocities, performing complete searches of $h$ and $\Delta(\rho\beta)$ for each case.

First, we vary $h$ over a large range, finding that the correlation coefficients between data and synthetics for crustal thickness less than 14 km or more than 33 km for the September 5, 1970 test event are less than 0.2. For each model suite in the final search $h$ varies from 14 to 33 km at a 1-km interval. For some events we explore thicker crusts. $\Delta(\rho\beta)$ has 20 intervals between 1% and 46%. Our absolute Moho depth determinations thus depend on how well this simple model represents average crustal velocity. For each model search with a specified $\beta_o$, $\beta_m$, and $\Delta t_s^*$, we compute 400 waveforms which we compare with the observed one by computing the cross correlation between the observed stacked $sS$ waveform and the synthetics. We correlate over an interval containing the $s,,S$ arrival and the first half cycle of the $sS$ arrival to reduce the presence of noise.

A contour plot of the correlation coefficients obtained for fixed values of $\rho\beta_o$ and $\rho\beta_m$ and numerous combinations of $h$ and $\Delta(\rho\beta)$ (Figure 6) indicates that the search algorithm does define an optimum choice or range of choices of parameters. There is some coupling between crustal thickness and impedance contrast because we assume a linear velocity gradient in the crust between the surface velocity $\beta_o$ and the velocity at the base of the crust, $\beta_m-\Delta\beta$. This parameterization avoids solutions with unrealistically high mantle velocities. Table 2 shows the best fitting model for the September 5, 1970, event, with velocities $\beta_o=3.1$ km/s and $\beta_m=4.65$ km/s and different $\Delta t_s^*$. For the best model, the crustal thickness is 24 km, and the shear wave impedance contrast at the Moho is 27%. The correlation coefficient is slightly lower for $\Delta t_s^*=1.98$ s than for $\Delta t_s^*=2.56$ s.

**Comparison for Individual Stations**

Since some individual records have clear $s,mS$ arrivals, we can assess the variability in the Moho parameters by fitting single stations, while recognizing that these have a much higher noise level but involve less horizontal averaging of the Moho. We chose a data set from four individual stations with consistently high signal to noise ratios for detailed comparison with synthetics. These four WWSSN stations are located in Europe (KON, TRI, STU) and North American (SCH). Since they have different path effects, we use the corresponding $S$ phase at each station as input signals for the modeling. We use the same search method as described above, i.e., for each station, we compute 400 traces to compare with the observed data at fixed $\rho\beta_o$ and $\rho\beta_m$ and choose the best fitting one. The correlation coefficients differ for various choices of $\rho\beta_o$ and $\rho\beta_m$. Table 3 gives the results and Figure 7 shows the observed and synthetic waveform comparisons for each station. The waveform comparison for the event stack is also shown, with the synthetic parameters corresponding to the best fit found in Figure 6. The results show that $\Delta(\rho\beta)$ varies from 20% to 36%, and $h$ varies from 23 to 25 km for optimal crustal structures. The crustal thickness estimates tend to be more stable because the waveform correlation procedure in the presence of noise is more robust for differential times than for relative amplitudes. As expected, the parameters found for the stacked traces are most reliable, given the higher signal to noise.

**Average Moho Properties Underneath $sS$ Reflection Points**

From the discussion of our method, it is clear that we can determine laterally averaged crustal thickness (for specified average crustal velocity) and shear wave impedance contrast at the Moho from clear and isolated $s,mS$ phases preceding the $sS$ reflection. This is because we use waveform modeling to determine the structure. Figure 8a shows the surface $s$ reflection point locations for the event of September 5, 1970. The actual reflection for $s,mS$ involves a horizontal region with scale defined by the Fresnel zone, which depends on the wavelength and the source depth. The inversion results for $\Delta(\rho\beta)$ and $h$ are thus averages over this zone and not local values at any reflection point. The Fresnel zone radius for our long-period $S$ wave data is about 170 km for a source depth of 560 km and wavelengths of about 100 km. Most of the data

---

**TABLE 2. Parameters for Best Fitting Models for the Event of September 5, 1970**

<table>
<thead>
<tr>
<th>$\beta_o$</th>
<th>$\beta_m$</th>
<th>$\Delta\beta$</th>
<th>$\Delta(\rho\beta)$</th>
<th>Thickness</th>
<th>$\Delta t_s^*$</th>
<th>Correlation Coefficient</th>
</tr>
</thead>
<tbody>
<tr>
<td>3.1</td>
<td>4.65</td>
<td>0.81</td>
<td>27</td>
<td>24.0</td>
<td>2.56</td>
<td>0.993520</td>
</tr>
<tr>
<td>3.1</td>
<td>4.65</td>
<td>0.81</td>
<td>27</td>
<td>24.0</td>
<td>1.98</td>
<td>0.992514</td>
</tr>
</tbody>
</table>
for the event of September 5, 1970, sample the crust north of the source, over an area spanning 4.5°. Only a few traces sample structure south of the event. Thus, stacking the data and the intrinsic Fresnel zone limitations involve smoothing over a region at least 900 km across. Therefore, the estimated crustal thickness and impedance contrast are appropriate averages over the entire reflection point area.

Table 3. Parameters for Best Fitting Single Station Models for the Event of September 5, 1970

<table>
<thead>
<tr>
<th>Name</th>
<th>$\beta_{p}$ km/s</th>
<th>$\beta_{s}$ km/s</th>
<th>$\Delta\beta_{p}$ km/s</th>
<th>$\Delta\beta_{s}$ %</th>
<th>Thickness km</th>
<th>$\Delta\alpha_{s}$ *</th>
<th>Correlation Coefficient</th>
</tr>
</thead>
<tbody>
<tr>
<td>KON</td>
<td>3.4</td>
<td>4.4</td>
<td>0.85</td>
<td>30</td>
<td>25.0</td>
<td>1.98</td>
<td>0.973723</td>
</tr>
<tr>
<td>SCH</td>
<td>3.2</td>
<td>4.7</td>
<td>1.10</td>
<td>36</td>
<td>25.0</td>
<td>1.98</td>
<td>0.969731</td>
</tr>
<tr>
<td>STU</td>
<td>3.9</td>
<td>4.4</td>
<td>0.60</td>
<td>20</td>
<td>24.0</td>
<td>1.98</td>
<td>0.971232</td>
</tr>
<tr>
<td>TRI</td>
<td>3.2</td>
<td>4.7</td>
<td>1.05</td>
<td>34</td>
<td>23.0</td>
<td>1.98</td>
<td>0.983613</td>
</tr>
</tbody>
</table>

To explore for lateral variations on a smaller scale, we separate signals with reflection points into three groups (Figure 8a): group A with reflection points at azimuths to European stations (shaded triangles), group B with reflection points at azimuths to North American stations (open triangles) and group C with reflection points at azimuths to Southeast Asia and Australian stations (solid triangles). Then we separately stack data for each group and perform the waveform modeling. Group C only has four traces all with oceanic reflection points, and cannot be modeled due to the thin crust. Table 4 shows results from modeling of groups A and B. The structural parameters for each case are given in Table 3.

As shown in Figure 3, it is very hard to identify any coherent underside reflection phases other than $s_{n}S$ between $S$ and $s_{n}S$ phases in the raw data, suggesting that only weak impedance contrasts exist above the source. Isolated arrivals observed at only a single station cannot be safely interpreted in terms of mantle layering. $ScS$ arrives between $S$ and $s_{n}S$ phases over much of the distance range, thus we must avoid time intervals in which the strong $ScS$ arrival would obscure any $s_{n}S$ phase. After some editing of the data to accomplish this, we stack all the waveforms on different slownesses. We consider all possible depths for a reflecting boundary.

Because of possible lateral variations in reflector depth, the sharpness and impedance contrast of some upper mantle discon-
Fig. 8. (a) Locations of sS reflection points for the event of September 5, 1970. The asterisk indicates the source location and triangles are reflection points for paths to the labeled stations. We separate reflection points into three groups: group A, reflection points at azimuths to European stations (shaded triangles); group B, reflection points at azimuths to North American stations (open triangles); and group C, reflection points at azimuths to southeast Asia and Australian stations (solid triangles). (b) Similar plot for the North Korean event of September 10, 1973.

Interpretation:

Upper mantle discontinuities may be obscured in a global analysis like that of Shearer [1990, 1991]. Our stacking exploration for sS precursors reveals relatively localized upper mantle properties above the deep source regions near the northwestern Pacific subduction zones. The same Fresnel zone and laterally averaging effects as described for the s,,S analysis apply for this portion of the study, so if the boundary varies over a 1000-km scale we will obtain only locally averaged properties of the structure.

Differing distance ranges and source depths of the earthquakes affect the arrival times for a certain sS phase. We line up all the data for a given event on the sS phase to provide a clear reference phase and to eliminate station statics. We also separate the data into three groups to make the depth influence as small as possible. These three groups are for different focal depth ranges: (1) 115 km < H < 195 km, (2) 348 km < H < 424 km and (3) 503 < H < 566 km. Given possible lateral variations in our study area, as indicated in analysis of s,,S phases, we divide the data into three regions for each depth group; (1) the Sea of Okhotsk, (2) North Korea, and (3) Japan, Ryukyus and Izu.

Upper mantle discontinuities can have different reflection coefficients, but most of them are expected to be very weak. For example, shear impedance contrasts of 2-15% have been suggested for the 80-, 220-, 410-, 520-, and 650-km discontinuities, explaining in part why it is difficult to identify sS precursors in the individual traces. We do know that different phases will have different slowness.

By aligning the traces on sS, we set the corresponding slowness of the sS arrivals to zero. The slowness for different underside reflection phases provides a means for identifying the sS arrivals, if there are any, relative to random or signal-generated noise. Slant stacking on a particular slowness creates interference between the signals, which is destructive except for a phase with the corresponding slowness. Such phases will appear with a maximum amplitude at a certain time on the stacked trace. The confidence that one has in the stacked signal depends on the degree to which noise is suppressed in the overall stack and consistency of the timing of the arrival with the depth expected for a particular slowness. Figure 9 gives an example of slant stacking at different slowness values for the event of August 16, 1979. For slowness Δs = 0.00, which corresponds to a stack with the sS phase lined up, we can see a very strong sS phase and a weak incoherent S phase. As Δs decreases from zero, the sS phase becomes weaker and distorted, while the S phase becomes stronger and more coherent. The energy between the S and sS phases also changes. For Δs = -0.0065 s/km, the S phase is aligned and a single pulse is retrieved. Figure 3 shows that a single differential slowness is not ideal for aligning S, but will be an excellent approximation for any sS phases. Therefore, when slant stacking over a range of slowness, any phase with a particular slowness will have maximum coherence for the corresponding stacks.

Since we exclude traces with ScS between S and sS phases, the traces which can be used for slant stacking are fairly sparse.

### TABLE 4. Parameters for Best Fitting Models for Different Reflection Regions Under the Sea of Okhotsk

<table>
<thead>
<tr>
<th></th>
<th>Sept. 5, 1970</th>
<th>Bθ</th>
<th>Bₘ</th>
<th>ΔB</th>
<th>Thickness</th>
<th>ΔΔ(ρθ)</th>
<th>Correlation Coefficient</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>km/s</td>
<td>km/s</td>
<td>km/s</td>
<td>km</td>
<td>%</td>
<td></td>
</tr>
<tr>
<td>A</td>
<td></td>
<td>3.15</td>
<td>4.8</td>
<td>1.05</td>
<td>23</td>
<td>30.8</td>
<td>0.993747</td>
</tr>
<tr>
<td>B</td>
<td></td>
<td>3.1</td>
<td>4.8</td>
<td>0.9</td>
<td>26</td>
<td>27</td>
<td>0.992433</td>
</tr>
</tbody>
</table>
for each event. Therefore, we combine nearby events that have similar depths into group stacks. This is possible because the events all have simple, similar waveforms. In the slant stacking, we first line up all the traces on the $S_S$ phases, flip any traces that have reversed $S_S$ polarity, and then normalize the traces to the peak of the $S_S$ phase. Given the very weak precursors $S_S$, we plot the logarithm of the envelope of the true slant stacks using the same method as Vidale and Benz [1992]. Very weak variations can be seen in the logarithmic scale. In our slant stacking, we consider relative slowness variation from $-0.015$ to $0.015$ s/km, a range spanning any plausible $s_S$ arrivals.

For shallower focal depth events ($115 \text{ km} \leq H \leq 195 \text{ km}$), it is very difficult to separate precursors from $S$ and $S_S$ in the long-period data. We did tests on shallow events on September 21, 1965, July 4, 1967, May 14, 1968, May 23, 1978. Because of the shallow depth for those events, it is very hard to confidently identify any energy between $S$ and $S_S$ phases, so we only discuss events in the depth range of 348 to 566 km in the investigation of $s_S$ phases.

**RESULTS**

**Results From Analysis of $s_mS$**

Figure 10 shows comparisons between observed (dashed line) and synthetic (solid line) waveforms for all events that have identifiable $s_mS$ phases that could be modeled. Table 5 lists results from the modeling. Table 5 indicates that crustal thickness near North Korea is about 36 km, and the shear wave impedance contrast at the Moho is about 18%. The relatively early $s_mS$ arrivals in this region can be seen in Figure 10. As shown in Figure 4 and Table 1, the duration between $S$ and $S_S$ is less than 50 s for the Kurile Island events on September 21, 1974, December 1, 1967, and the Taiwan event on September 2, 1978, because of their shallow source depths. Since the period of our data is about 20 to 25 s, it is difficult to isolate the $s_mS$ phase between $S$ and $S_S$ for these events. Allowing for substantial uncertainty in the $s_mS$ phase for events on September 21, 1974, and September 2, 1978 (Figure 4), the wave-form modeling indicates crustal thicknesses of 19 and 30 km, respectively, and 21% shear impedance contrasts for these two events. The crust appears to be thicker under Taiwan than under Kamchatka if these numbers are reliable. It should be possible to improve the resolution with broadband data as it becomes more extensive. The depths of events on May 23, 1978, May 14, 1968, and September 21, 1965 along the Ryukyu arc range from 160 to 195 km (Table 1). The modeling (Figure 10 and Table 5) gives crustal thicknesses of 25, 20, and 20 km and shear impedance contrasts of 21%, 29%, and 29%, respectively. Since the $s_mS$ phases are not well isolated, relatively low correlation coefficients are found, as listed in Table 5. Nonetheless, there is some indication of crustal thickening along the arc toward the Japanese islands. Event on March 31, 1969 under the Sea of Japan and events on February 22, 1974, January 31, 1973 and December 12, 1976, under the
TABLE 5. Crustal Thickness and Impedance Contrast for Some Events

<table>
<thead>
<tr>
<th>Date</th>
<th>Latitude, Depth, Thickness, A(p%), Correlation Coefficient</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>(deg) (km) (km) (%)</td>
</tr>
<tr>
<td>Sept. 10, 1973</td>
<td>42.48 552 5.8 36 18 0.995280</td>
</tr>
<tr>
<td>Aug. 16, 1979</td>
<td>41.85 566 5.8 36 18 0.994874</td>
</tr>
<tr>
<td>Sept. 2, 1978</td>
<td>24.81 115 6.0 30 21 0.973122</td>
</tr>
<tr>
<td>Sept. 21, 1965</td>
<td>28.96 195 6.0 20 29 0.975749</td>
</tr>
<tr>
<td>May 14, 1968</td>
<td>29.93 162 5.9 20 29 0.979251</td>
</tr>
<tr>
<td>May 23, 1978</td>
<td>31.07 160 6.2 25 21 0.938021</td>
</tr>
</tbody>
</table>

North Korea

Taiwan

Ryukyu Arc

Sea of Okhotsk

Izu arc, all reflect under oceanic structure as shown in Figure 4 and have unmodelable weak $s_mS$ waveform effects so we cannot quantify the structure there.

The crustal thickness beneath the northern Sea of Okhotsk varies from 21 to 29 km, and the shear wave impedance contrast varies from 26% to 28% for all events except events on December 21, 1975, and March 18, 1964, which give extreme estimates of shear wave impedance contrast of 46% and 12%. Looking at the data for event on December 21, 1975 (Figure 10), we see that the onset of $s_mS$ is unclear and the synthetics give a more impulsive arrival than observed, which might be one reason for the high value of $\Delta(p\beta)$. It is hard to identify $s_mS$ for event on March 18, 1964 (Figures 4 and 10) because the amplitude is small, yielding a low estimate of impedance contrast. This event appears to be unreliable and is the only event from the early days of the WWSSN system when a 30-s pendulum period was being used. These two anomalous results may represent locally complex uppermost mantle structure or simply noise for these events. Figure 11 plots stacked waveforms for different events in the Sea of Okhotsk, showing that clear and strong $s_mS$ phases are observed in the north of this area, for events on September 5, 1970, December 21, 1975, January 29, 1971, April 2, 1984, August 21, 1972 and February 1, 1984, while weak $s_mS$ are observed at events farther to the south: June 21, 1978, and July 10, 1976. We do not report Moho parameters for the latter two events because they are poorly resolved. This variation from north to south probably reflects the crustal variation from quasi-continental to oceanic.

Fig. 11. Stacked $sS$ waveforms for the Sea of Okhotsk events plotted from north to south. The $s_mS$ arrival weakens for the southern events.
structure in this area. We only have six traces for the Kurile event on May 27, 1972, so the stack is less reliable for this event and we did not model it.

Results From Slant Stacking $s_s S$

Plate 1 shows results from slant stacking $s_s S$ phases for source depths varying between 503 and 566 km in three regions. The plot shows the logarithmic amplitude of the stack envelope as a function of relative slowness and travel time. The color scale is a logarithmic amplitude that varies from 0 to 2, which means the true amplitude scale will vary by a factor of 100 relative to the stacked $s_s S$ amplitude. Black indicates less than 1% amplitude relative to $s_s S$. Any arrivals with amplitude 2% to 10% of $s_s S$ can be readily detected. Some clear artifacts of slant stacking are apparent in the streaking. Plates 1a, 1b and 1c are slant stacks of deep events in North Korea, the Sea of Okhotsk, and Izu Japan, respectively. The slowness of any $s_s S$ precursors from horizontal layers will vary between the slownesses of $S$ and $s_s S$ phases (Figure 3). If energy is found between $S$ and $s_s S$ with correct corresponding slowness, it is likely to be an $s_s S$ arrival, but the limited ray parameter resolution must be allowed for. In each case, $s_s S$ is the strongest arrival at a relative slowness of zero. As the relative slowness varies, the $s_s S$ energy will become less coherent in both slowness variation directions. The extent to which the $s_s S$ arrival is localized in time and ray parameter is indicative of how well any arrival can be isolated, and real $s_s S$ arrivals should streak as much as $s_s S$.

A very clear stack is found for two deep events below North Korea (Plate 1a). We can see that relatively strong energy occurs in front of the $s_s S$ phase with very subtle slowness difference from $s_s S$. This corresponds to energy reflected from the Moho and the amplitude ratio of $s_n S$ to $s_s S$ is about 8%. Ahead of $s_n S$, other energy is seen which is consistent with reflection from a 90-km-deep boundary, with about 6% ampli-
Plate 2. Slant stack sections for events with focal depths of $348 \leq H \leq 424$ km below: (a) the Sea of Okhotsk and (b) Izu Japan.

There is no evidence for a $s_{200}S$ phase. In fact, the very low amplitude in the corresponding time window limits any sharp discontinuity to be less than 1% shear impedance contrast. There is a weak, but still distinguishable arrival compatible with a 330-km reflector, with the amplitude ratio of $s_{330}S/sS$ being about 5%. There is a stronger arrival close in time to a reflection from a 400-km discontinuity with the amplitude ratio $s_{400}S/sS$ being about 6%, although the slowness is somewhat anomalous. We convert amplitude ratio to impedance contrast by considering the geometric spreading and attenuation using reference model PREM. The impedance contrasts for the Moho, 90-, 330-, and 400-km discontinuities in North Korea are about 16%, 11.4%, 9%, and 10.8%, respectively. The Moho impedance contrast (16%) inferred from the slant stack is compatible with the 18% value found by modeling the $s_mS$ phases. Strong attenuation in the mantle wedge above the slab could reduce these values, but radiation pattern effects may compensate somewhat. Overall, we estimate a ±3% uncertainty associated with deeper discontinuities and ±1-2% for structures above 100 km for shear impedance contrasts and ±5-10% uncertainty for estimation of discontinuity depths.

The region under the Sea of Okhotsk yields the slant stack shown in Plate 1b. The particular data distribution leads to some strong streaking in the slant stack. We see very strong $sS$ energy, and there is visible energy just ahead of $sS$ associated with $s_mS$ which has an impedance contrast of about 16-26%. This is again consistent with the results of $s_mS$ waveform modeling. Since the crustal structure varies from north to south in the Sea of Okhotsk, the wide energy range associated with $s_mS$ may reflect the transition from continental to oceanic structure. At the time for reflectors near 85-90 km and 160 km depth there are very weak arrivals with 11.4% and 8.4% impedance contrast. But these two arrivals are very narrow, and they may be noise or artifacts. There is again no clear reflection from any structure near 210 km, although a 2% relative amplitude arrival cannot be precluded. There is some energy near the time for a 330-km discontinuity with about 8% impedance contrast, but because of the weak signal we do not have great confidence in this discontinuity. A stronger arrival is seen near 400 km depth with 11.2% impedance contrast. While these do not have very discrete arrivals and lack much ray parameter resolution, these arrivals are consistent with those observed for paths under Korea.

In the area of Izu Japan (Plate 1c), which has an oceanic crustal structure, there is no separation of $s_mS$ energy in our long-period data. The slant stack for this region is in fact very clean. There is weak energy ahead of $sS$ which may come from a depth near 66 km, with about 12.6% impedance contrast. While these do not have very discrete arrivals and lack much ray parameter resolution, these arrivals are consistent with those observed for paths under Korea.
narrow arrival is suggested near a depth of 325 km with 9% impedance contrast, but it too may be noise. Energy from near the 400 km discontinuity is stronger than that of the other precursors, with 5.6% amplitude ratio, but because of focal depth range of events, this is very close to the S phase and may be contaminated by S wave coda.

Plate 2 shows slant stacks for focal depths varying from 348 to 424 km for two regions: Izu Japan (Plate 2a) and the Sea of Okhotsk (Plate 2b). In the slant stack for Izu Japan, we again cannot identify energy from the Moho because of the oceanic crustal structure there. But there is some energy that may come from reflectors around 66 and 210 km with the impedance contrast being 7.6% and 10.4%, respectively. It is hard to determine the amplitude ratio of $s_{33}S/sS$ because the S wave is very close, but there is energy in the window. However, the noise level is high so this identification is very tentative. Plate 2b shows the slant stack for the Sea of Okhotsk, which has strong artifacts. We can still identify some possible arrivals between S and $sS$ phases, which suggest discontinuities near the Moho, 90 and 335 km with impedance contrasts of 8%, 4.7% and 5.5%, respectively, but there is little confidence in these arrivals, other than as supporting the evidence for weak reflectors seen in Plate 1. There is again no evidence for any reflector near 210 km beneath the Sea of Okhotsk.

**DISCUSSION**

From seismic profiling, refraction, and deep seismic sounding, Gnibidenko [1985] pointed out that the principal geomorphic elements of the Sea of Okhotsk are the continental shelf in the north, the continental slope area of the central part of the sea, and the deep-sea basin in the south. They reported that crustal thickness varies from 12 to 35 km. From Figure 11 we see that the $s_{33}S$ waveform varies from north to south, although we were able to determine the crustal thickness and shear wave impedance contrast only in the north where we have clear $s_{33}S$ phases. The crustal thickness of the northern Sea of Okhotsk varies from 21 (August 21, 1972) to 29 km (March 18, 1964) in our study. Our results for the Sea of Okhotsk are consistent with Gnibidenko's [1985] results.

Schenk et al. [1989] analyzed one of the same events as we did (January 29, 1971) under the Sea of Okhotsk, looking for $pP$ Moho precursors. Their model has a crustal thickness of 21-25 km and a P wave impedance contrast at the Moho from 27% to 31%, while we estimate an average crustal thickness of 23 km and an average shear wave impedance contrast of about 28% for this event. The comparison shows very consistent results from study of $pP$ and $sS$ precursors. Future studies will pursue such simultaneous measurements.

Using an event in North Korea, Schenk et al. [1989] find that the crust in Korea appears to thicken from 27 km (in the reflection point region at azimuths to European stations: TOL, MAL, PTO, and STU) to 32 km (in the reflection point region toward Africa stations: AAE and NAI) with the P impedance contrast at the Moho varying from about 23.8 to 17.1%. We have two events that sample the crust slightly north of their $pP$ paths. We combine these two events and sort them into three groups by surface reflection points (Figure 8b). Figure 8b is the locations of $sS$ reflection points for event on September 10, 1973, as an example of these two events. The first, group A, is composed of paths to European stations, group B includes only North American stations, and group C includes the more southern stations including those in Africa. Since these two events are closely located and have almost the same depth (Table 1), we combine them in the stacks for each group and apply the $s_{33}S$ waveform modeling. Unlike Schenk et al. [1989], stations AAE and NAI are not included in the present study's data set, and most group C reflection points lie in the Sea of Japan, giving poor $s_{33}S$ arrivals, so we did not attempt to model group C records. Table 6 summarizes the results of modeling groups A and B. The crustal thickness varies from 35 km in group A to 31 km in group B, thinning slightly toward the coast, and the shear impedance contrast at the Moho ranges from 21% to 31.2%. For the total stacks for August 16, 1979, and September 10, 1973, we determined the average crustal thickness in the area to about 36 km and the shear wave impedance contrast is about 18%. The average crustal thickness is slightly greater than that from regions A and B in part because modeling station CHG (group C) gives a 37-km-thick crust. The CHG reflection point locates near the area where Schenk et al. [1989] found thicker as well. There is some trade off between impedance contrast and crustal thickness, but we tend to prefer the average parameters because the stacks have the highest sig-

<table>
<thead>
<tr>
<th>Aug. 16, 1979/Sept. 10, 1973</th>
<th>$\beta_p$</th>
<th>$\beta_m$</th>
<th>$\Delta\beta$</th>
<th>Thickness</th>
<th>$\Delta(p\beta)$</th>
<th>Correlation Coefficient</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>3.25</td>
<td>4.6</td>
<td>0.69</td>
<td>35</td>
<td>21</td>
<td>0.991637</td>
</tr>
<tr>
<td>B</td>
<td>3.3</td>
<td>4.6</td>
<td>0.99</td>
<td>31</td>
<td>31.2</td>
<td>0.982888</td>
</tr>
</tbody>
</table>
nal to noise. Comparing the results of the $pP$ and $sS$ precursor studies, the estimates of crustal thickness are in reasonable agreement since our events locate slightly north of the event they used. If we compare with other models for continental regions, such as EU2 [Lerner-Lam and Jordan, 1987], the $S$ and $P$ impedance contrasts appear somewhat lower than in other continental regions.

In our study, the average shear impedance contrasts at the Moho are 18% beneath North Korea and 26-28% under the north of the Sea of Okhotsk. We show our preferred crustal models in Figure 12 and Table 7, along with the $P$ results of Schenk et al. [1989]. The shear wave impedance contrast of the continental structure is found to be smaller than that of the quasi-continental to oceanic transition structure under the Sea of Okhotsk (Table 5). This information may contribute to the thermal and petrological modeling of the crustal evolution in different tectonic environments.

Different shallow structure such as oceanic or continental crust should reflect the underlying upper mantle structure in the lithospheric lid. It is generally accepted that the lithosphere structure and thickness varies with different tectonic region. As pointed out by Revenaugh [1989], the base of this lid may be

<table>
<thead>
<tr>
<th>Region</th>
<th>$\beta_0$</th>
<th>$\Delta\beta$</th>
<th>$\beta_{m'}$</th>
<th>$\alpha_0$</th>
<th>$\Delta\alpha$</th>
<th>$\alpha_{m'}$</th>
<th>$h_0$</th>
<th>$h_{m'}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Okhotsk</td>
<td>3.1</td>
<td>0.9</td>
<td>4.8</td>
<td>23</td>
<td>6.2</td>
<td>1.5</td>
<td>8.11</td>
<td>22</td>
</tr>
<tr>
<td>Korea</td>
<td>3.25</td>
<td>0.84</td>
<td>4.6</td>
<td>33</td>
<td>6.5</td>
<td>1.2</td>
<td>8.11</td>
<td>30</td>
</tr>
</tbody>
</table>

Plate 3. Slant stack for events August 16, 1979, and September 10, 1973. (a) Waveform stack at slowness $s_{90}S$, (b) envelope of the stack at slowness $s_{90}S$, (c) slant stack (see detail in Plate 1) with two arrows pointing at slownesses for 90- and 330-km discontinuities, (d) envelope of the stack at slowness $s_{330}S$ and (e) waveform stack at slowness $s_{330}S$. Dash lines point corresponding arrivals.
marked by an abrupt 5-8% decrease in shear wave impedance at an average depth of 67 km in the majority of deep subduction zones and an average depth of 86 km in the western Pacific, which he attributes to the lid low-velocity zone (LVZ) transition. In our study region of the northwestern Pacific, we find evidence for a discontinuity near 66, 80 to 85 and 90 km in the Izu Japan, Sea of Okhotsk, and North Korea regions, respectively, with about 11.4% average impedance contrast. Our basic processing is insensitive to whether this involves a velocity increase or decrease. We can attempt to determine polarity of the reflection. Plate 3 shows details of the slant stack for North Korea. Plate 3c is the same as Plate 1a with two arrows indicating stack slownesses at 90 km (sS0S) and 330 km (sS30S) discontinuities. Plate 3a, 3b, 3d, and 3e are the corresponding stacked waveforms and envelopes. The polarity of sS and sS0S appears to be the same, suggesting reflection from a velocity increase near 90 km rather than from the top of a low-velocity zone. The lid discontinuity is slightly deeper under Korea than under the Sea of Okhotsk, and we again slightly prefer a velocity increase near 85 km in the latter case. Thus the transitional continental margin structure may involve something other than the onset of partial melting. The narrowband data and weak signal make this a tentative conclusion. The Izu Japan region has oceanic crustal structure, and here we find a relatively shallow discontinuity at 66 km. The signal is too weak to make any statement about its polarity, so it may in fact be associated with the lid-LVZ boundary.

In our study of sS precursors, we only found evidence for a 200 to 210 km discontinuity underneath the Izu Japan area, which has oceanic lithosphere. Recently, Vidale and Benz [1992] found evidence for a P wave impedance contrast near 210 km depth near the Bonin, Mariana, and Tonga subduction zones, also beneath oceanic lithosphere. The 200-km discontinuity may exist under both continental and oceanic lithosphere, but the lateral variation must be very large, because we did not find evidence for this structure under North Korea or the Sea of Okhotsk. Therefore our data do not support the hypothesis that the 200-km discontinuity is a global feature as advocated in some studies [Anderson, 1979; Dziewonski and Anderson, 1981; Drummond et al., 1982]. This is supported by the absence of strong reflections from a depth of 200 km in the global stacks of Shearer [1991]. A search for PP precursors in these regions is needed to test whether this is a result of there being no structure or only lack of S wave structure.

A few studies [Revenaugh and Jordan, 1991c; Wajeman, 1988] have found evidence for a 330-km discontinuity. In our study, we found consistent but very weak evidence in all three regions for a reflector near this depth, but we need more data to increase confidence in the identification of this structure. Plates 3c, 3d, and 3e suggest that the polarities of sS30S and sS are the same. Since this weak discontinuity is observed only on profiles sampling regions of active subduction, it may be caused by hydration reactions, occurring in parts of the mantle that have been volatile charged by subducting lithosphere [Revenaugh and Jordan, 1991c]. We will seek more explanations of these discontinuities in future studies that will extend the coverage of subduction zone mantle stratification.

Table 8 summarizes of our results along with estimates of impedance contrasts at x km discontinuities. We include crustal information in Table 8 from the results of sS waveform modeling.

**CONCLUSIONS**

We have established that precursors to sS can be used to determine mantle stratification above subduction zones. For our northwestern Pacific study area, we find oceanic crust beneath Izu Japan, continental crust under North Korea with 36 km crustal thickness and 18% shear wave impedance contrast at the Moho, 20-25 km thick crust under the Ryukyu arc with 21-29% impedance contrast, and beneath the Sea of Okhotsk the crustal structure varies from quasi-continental in the north with 21-29 km thickness to oceanic toward the south. Our study indicates that the shear wave impedance contrast is smaller for the continental Moho than for the transitional structure. Lateral variations also exist in the upper mantle for these regions. Evidence for an 80-km discontinuity was found, with lateral variation in depth. The data suggest a discontinuity structure near 66 km under Izu Japan, 80-85 km under the Sea of Okhotsk and 90 km under North Korea, all with average impedance contrast of about 11.4%. We only found evidence for a 200-km discontinuity under the Izu Japan area, below oceanic lithosphere, with depth near 200 to 210 km. No evidence for this discontinuity is found in the other two regions. Consistent evidence for a 330-km discontinuity is found in all three regions. The depth range varies from 325 to 335 km, and the average impedance contrast is about 9%, but we do not yet have great confidence in this feature. Slant stacking results also show consistence evidence for a 400-km discontinuity in our study area; under North Korea and the Sea of Okhotsk the depth is around 400 km, and under Izu Japan it is 380 km. The average impedance contrast is about 10.8%.

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**Table 8. Results for Upper Mantle Discontinuities**

<table>
<thead>
<tr>
<th>Region</th>
<th>Moho</th>
<th>80 km</th>
<th>220 km</th>
<th>330 km</th>
<th>400 km</th>
</tr>
</thead>
<tbody>
<tr>
<td>Izu-Japan</td>
<td>Oceanic</td>
<td>66</td>
<td>200-210</td>
<td>325</td>
<td>380</td>
</tr>
<tr>
<td>North Korea</td>
<td>36</td>
<td>90</td>
<td>No</td>
<td>330</td>
<td>400</td>
</tr>
<tr>
<td>North Okhotsk sea</td>
<td>22-24</td>
<td>80-85</td>
<td>No</td>
<td>330-335</td>
<td>400</td>
</tr>
<tr>
<td>Possible Impedance Contrast</td>
<td>-8%</td>
<td>-5.7%</td>
<td>-5.5%</td>
<td>-4.5%</td>
<td>-5.4%</td>
</tr>
</tbody>
</table>

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