Investigating the heterogeneity of the D'' region beneath the northern Pacific using a seismic array

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1. Introduction

It has been long known that the core-mantle boundary (CMB) transition zone (D'' [Bullen, 1949]) is a complicated region within the Earth [e.g., Gutenberg, 1914] exhibiting heterogeneities on many length scales [see, e.g., Loper and Lay, 1995; Garnero, 2000]. Velocity models of this region predict variations in excess of ±3% for P and S waves [see Wysession et al., 1998]. Lay and Helmberger [1983] were the first to present compelling evidence for a seismic discontinuity at the top of the D'' layer. Since then a great deal of evidence for a discontinuity has accumulated [see review by Wysession et al. [1998]], but there is also an indication that the discontinuity is not necessarily a global feature [Kendall and Nangini, 1996]. Many possible explanations for the complex structure of D'' have been suggested. These are dependent upon the chemical and thermal processes occurring in the lower mantle, with D'' believed to be a thermochemical boundary layer [e.g., Stacey and Loper, 1983; Ringwood, 1979; Knittle and Jeanloz, 1991]. It has been suggested that these variations are related to the CMB accumulation of ancient subducted slab material in certain regions and the upwelling of slab material in others [e.g., Weber, 1994; Scherbaum et al., 1997; Kendall and Shearer, 1999; Russell et al., 1999].

One evidence for the D'' reflector comes from an additional phase between the P wave and the reflection from the core-mantle boundary. However, although many studies are assuming a reflector as the cause for this nonstandard wave, it could also be produced by scatterers [Scherbaum et al., 1997] or strong 3-D gradients in velocity at different depths [Liu et al., 1998].

Here we investigate the lower mantle in a region beneath the northern Pacific Ocean previously uninvestigated with P waves. The area (Figure 1) marks a transition region between areas of presumed mantle upwelling (mid-Pacific) and downwelling (northern Pacific subduction zones). Previous studies have documented evidence of an S wave discontinuity in this region. In the early work of Lay and coworkers [Lay and Helmberger, 1983; Young and Lay, 1990], long-period seismograms from many stations with a range of epicentral distances were used to estimate the 1-D S wave discontinuity structure. Kendall and Shearer [1994] estimated the discontinuity depth using a phase-stripping method applied to individual long-period seismograms and found considerable lateral variation in the depth of the discontinuity.

In this study, array techniques are applied to short-period data to map detailed variations in a P wave D'' discontinuity in this region. Vespagrams (akin to slant stacks) and f-k analyses are both used to estimate the arrival
times and directions of incidence of seismic phases which sample the lower mantle. Such techniques give improved estimates of the three-dimensional nature of the D″ region. A better image of the lowermost mantle ultimately offers improved insight into mantle dynamics and core-mantle interactions.

2. Data and Processing

[6] Recordings from northwest Pacific events at the Yellowknife array (YKA) in Canada are used to study the lowermost mantle beneath the northern Pacific. The YKA consists of 19 vertical component short-period seismometers which are arranged in a cross configuration of 20 × 20 km. The stations and events are shown in Figure 1 and the event parameters are listed in Table 1. Events with a magnitude >5.5 were selected to ensure a good signal-to-noise ratio. The event depths range from 20 to 525 km, and the epicentral distance range is between 68° and 82°. A variety of phases can be used to study lower mantle structure; in this study we look for P wave reflections from a D″ discontinuity (PdP in Figure 2). The discontinuity produces a triplication in the wave front. The first forward branch is the phase which turns above the discontinuity, the second forward branch is the diving wave which turns beneath the discontinuity (sometimes known as PDP [Weber, 1993]. The reverse branch is the reflection PdP. PDP is difficult to isolate as it merges with PdP in the range of epicentral distances we consider.

[7] The advantages of seismic arrays over single stations are that they can be used to determine the azimuthal direction of the incoming wave and its apparent velocity (slowness), as well as enhance the signal-to-noise ratio through stacking (see Rost and Thomas [2002] for a review on array methods). A D″ reflection, PdP, will arrive as a precursor to the core reflection PcP. A reduced slowness of the phase PdP with respect to P is the decisive element in identifying this arrival as a D″ reflected phase, and it is only through the use of seismic arrays that the slowness of a phase can be determined. The array methods employed in this study are vespagrams (slant stacks) and frequency-wave number (f-k) analysis.

[8] To construct the vespagrams, we use a nonlinear Nth root stacking technique [Muirhead and Datt, 1976; Davies et al., 1971] that sums the Nth root recordings from each station for a slowness and raises the sum to the power of N. Different phases arrive at the array at different angles of incidence, and hence travel across the stations of the array with different apparent velocities (horizontal slowness). By stacking over all possible slownesses, coherent phases (e.g.,
and PcP) are enhanced while the amplitude of the surrounding incoherent noise is diminished. This process is known as beam forming, with a separate beam being generated for each slowness value. The resulting plot of all the beam traces leading from the minimum slowness to the maximum slowness value is called a vespagram. A reflection from D00 (PdP) should lie between the P and PcP phases with respect to both time and slowness. The method is particularly useful for discriminating between low-amplitude arrivals (such as PdP) and noise. A weakness of the method is that it assumes a 2-D medium; phases not travelling in the source-receiver plane will distort the vespagram slowness estimates. However, vespagrams with respect to azimuth can also be constructed. Hence the slowness and azimuth can be found in a multistep process.

The f-k method uses the fact that the time delays required to bring the arrivals at each station into phase provide a simultaneous estimate of the backazimuth and slowness of the arrival. Unlike the vespa method, in which D0 reflections are initially assumed to lie in the source-receiver plane, the f-k procedure searches over all possible backazimuths and slownesses, finding the combinations that result in the greatest amplitudes. The data are transformed into the f-k domain using a 3-D Fourier transform, allowing the slowness and azimuth to be calculated simultaneously [Capon, 1973]. The total energy recorded at the array is a function of the power spectral density and the array response function (ARF). The ARF is controlled by the aperture, configuration and station spacing of the array, and provides a measure of the coherency of the arrivals. This means that if the ARF is visible in the f-k result, the data are coherent. In practice a time window is prescribed around the target phase and the analysis performed. Phases may show the correct slowness and time delays between P and PcP but deviate significantly (>10°) from the backazimuth to the event. Such anomalous signals may be the result of reflections out of the great circle path. Alternatively, they could be the result of a separate but similarly timed event. Backazimuth, slowness, and coherency (as given by the ARF) are the three criteria on which the classification of the arrivals is made. The f-k analysis relies upon correct specification of the time window by the operator. A poorly judged window may significantly affect the f-k analysis result. A sliding window f-k analysis can be used to remove the ambiguity in prescribing the time window selection [Rost and Weber, 2001].

3. Results

Of the 19 events studied, 14 showed clear evidence for a phase that we interpret as a reflection from the D'' discontinuity (PdP) and the remaining proved inconclusive. In some cases the vespagram reveals a distinct arrival (PdP) between P and the theoretical arrival time of PcP, but the slowness of this arrival is often poorly resolved for a small-

![Figure 2. The phases P, PdP, and PcP in the epicentral distance range 68° to 82° are used to investigate the D'' P wave discontinuity. See text for further discussion.](image-url)

Table 1. Events Used in This Study

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<th>Time, UT</th>
<th>Latitude, °N</th>
<th>Longitude, °E</th>
<th>Depth km</th>
<th>mb</th>
<th>baz</th>
<th>Distance</th>
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<td>00:49:42.77</td>
<td>23.64</td>
<td>121.14</td>
<td>33</td>
<td>6.2</td>
<td>310</td>
<td>82</td>
</tr>
</tbody>
</table>

The earthquake parameters are from the Preliminary Determination of Earthquakes (PDE) catalogue. Distances and backazimuths (baz) are calculated for the center of the array.
and there are four additional phases visible. The phase
X_M is interpreted as the reflection from the Moho. The phase X_1 phase shows up in most other
vespagrams and is most likely a reverberation from a shallow feature below the YKA [Bostock, 1999]. The
slowness of these phases is greater than that for P (~6.0 s/deg). The phase X_2 also has a higher slowness than P
and travels on the great circle path, whereas X_3 has a backazimuth more than 20° off the great circle path. It is difficult
to say what is responsible for these phases. It is only with
array analyses that we can isolate the PdP phase from other phases. The error in estimation of the slowness and backazimuth of PdP can be assessed from the contour lines. In this case we estimate an error of 0.5 s/deg in slowness and 5° in azimuth.

Figure 3. (top) Data, (middle) vespagram (slowness versus time), and (bottom) f-k plots for P and time windows
around expected PdP and PcP for event 12 January 1999 (event 16 in Table 1), which shows evidence for a D''
discontinuity. The epicentral distance for this event is 73°
and the theoretical backazimuth is 296°. Note that in this
case the PcP arrival is stronger than the PdP arrival. The
data have been band-pass filtered between 0.03 and 2 Hz.
The f-k analysis confirms that the phases are P, PdP, and
PcP and there are four additional phases visible. The phase
X_M is most likely the Moho reverberation and X_1 possibly a
reverberation from a feature in the crust beneath the array.
The phase X_2 has a higher slowness than P, whereas the phase X_3 arrives from a different azimuth and has a
slowness similar to the P slowness. The f-k plots show
slowness in the radial direction (0 to 10 s/deg) and backazimuth in degrees from north. Isolines are in 1 dB
steps.

with a very small amplitude compared to P and PcP. The f-k
analysis gives values for P of 5.58 s/deg for the slowness
and a backazimuth of 297°, which is close to the theoretical
backazimuth of 296°. PdP is quite clear with a slowness of
5.09 s/deg and a backazimuth of 283°. This indicates that
PdP travels 13° out of the source-receiver plane and may
indicate a dipping reflector. PcP has a slowness of 3.6 s/deg
and a backazimuth of 298°, again close to the theoretical
backazimuth. Also visible in the vespagram are four addi-
tional phases. The phase X_M is interpreted as the reflection
from the Moho. The phase X_1 phase shows up in most other
vespagrams and is most likely a reverberation from a shallow feature below the YKA [Bostock, 1999]. The
slowness of these phases is greater than that for P (~6.0 s/deg). The phase X_2 also has a higher slowness than P
and travels on the great circle path, whereas X_3 has a backazimuth more than 20° off the great circle path. It is difficult
to say what is responsible for these phases. It is only with
array analyses that we can isolate the PdP phase from other phases. The error in estimation of the slowness and backazimuth of PdP can be assessed from the contour lines. In this case we estimate an error of 0.5 s/deg in slowness and 5° in azimuth.

[13] Event 17 in Table 1 (Figure 4) illustrates the 2-D
limitation of the vespa method where the PdP phase travels
out of the source-receiver plane. The vespagram shows an
anomalous arrival in the expected region for PdP, but this
arrival has a low slowness value for PdP (~4.0 s/deg).
However, the f-k plot for the PdP phase shows a difference
in backazimuth of 8° compared to the theoretical backazimuth
(295°) and a more realistic PdP slowness of ~4.5 s/deg,
~1.2 s/deg smaller than that of P. Again, similar errors in
slowness and azimuth are indicated by the f-k contour lines.
Two additional phases are visible in the vespagram. The f-k
analysis indicates that both phases have a slowness very
similar to the P slowness. The X_M phase is again interpreted
as the Moho reverberation and X_1 is again related to a
shallow feature beneath YKA. A phase with a low slowness
is barely visible in the vespagram at the theoretical arrival
time of PcP. This phase can be resolved with f-k analysis
and shows a slowness of about 3 s/deg and a backazimuth
of 311°. If this is the PcP phase, it has travelled 16° out of
the source-receiver plane.

[13] Travel time analysis was performed on the 14 PdP
results to determine the depth to a D'' reflector. Travel time
curves were generated by ray tracing through a range of
models based on modifications of the P wave model PWDK
[Weber and Davis, 1990]. PWDK was developed for the
region beneath northern Siberia and has a 3% P wave
velocity jump located 293 km above the CMB. Previous
investigations of the North Pacific region have been S wave
studies [e.g., Lay and Helmberger, 1983; Young and Lay,
1990; Kendall and Shearer, 1994], and no P velocity model
has been developed for the region.

[14] The P-PdP differential times were measured from
the vespagrams and compared to the generated travel time
curves. New curves were generated using a trial-and-error
approach until the observed and theoretical times agreed to
within the measurement error of ±1 s which corresponds to
±25 km in the discontinuity depth. Errors in the method are
mainly due to inaccuracies in the measuring of the P-PdP
differential times from the vespagrams. The stated errors are
based on the modeling of Weber [1993] for PWDK, who calculates that a 0.2 s pick error results in a 5 km shift in discontinuity height. The range in estimated reflector heights above the CMB is illustrated in Figure 5 as contour lines.

The discontinuity velocity contrast in the model PWDK (3% velocity jump of the P velocity) seems to be too high to explain our data. Vespagrams produced using synthetics calculated for this model predict PdP amplitudes much stronger than those observed in our data. Our data show considerable variability in PdP strength. The inferred D" topography could lead to focusing and defocusing effects making it difficult to constrain the velocity contrast.

Furthermore, as PcP is also a very weak phase, often not observed, it is impossible to use the core reflection to further constrain the D" velocities. In the present study we can therefore only comment on the depth of the reflector.

The results (Table 2) imply a large variation in reflector topography (211–334 km) in the region, with a mean value of 241 km. It is important to keep in mind that the reflector height was calculated with the assumption of travel paths confined to the great circle arc. Deviations from the great circle path cause travel time delays in the PdP phase which would be interpreted as a lower reflector height and misposition the reflector. A difference in backazimuth of PdP to the great circle azimuth can cause a travel time difference of 2 s which corresponds to 50 km difference in height.

4. Discussion and Conclusions

The first P wave investigation of this region using array methods reveals that the D" region south of the Kamchatka peninsula is highly heterogeneous. We show evidence for a P wave discontinuity which has a mean height of 241 km above the CMB. Figure 5 shows the spatial distribution of the reflection points of the 19 events. Positive (solid circles) and ambiguous results (open triangles) overlap, and there is no evidence for negative results (i.e., complete absence of PdP in the vespagram or f-k analysis). Assuming the reflector is a continuous marker of
The average height is slightly lower than values from previous studies in the south (Figure 5) with a peak being visible to the north. The latitude of 55°/C176°/C00° with a peak being visible to the north. The depth of the D'' layer, the layer thins to nearly 210 km near a latitude of 55° and thickens dramatically to the north and south (Figure 5) with a peak being visible to the north. The average height is slightly lower than values from previous studies of the region (e.g., 280 km [Lay and Helmberger, 1983]; 243 km [Young and Lay, 1990]), and mean 296 km [Kendall and Shearer, 1994]), but there is little overlap in image locations between these studies. The depths of the reflector found by Kendall and Shearer [1994] in this area are included in the figure as grey circles. Note that the studies mentioned above are S wave studies; however, comparisons in other areas have shown that predicted discontinuity heights for both P and S waves generally agree [Wysession et al., 1998].

As discussed above, a large deviation from the backazimuth can influence the travel time of the observed phase and therefore give a misleading depth estimate for the reflector. In three cases we found deviations from the great circle path of more than 10°. The estimated depths of the reflector corresponding to those data points may be too large. However, when estimating a higher distance of the reflector for these points, the trend discussed above is still visible.

Due to limited data quality we cannot constrain the magnitude of the discontinuity velocity contrast. This is probably due to D'' reflector topography which causes focusing and defocusing that can turn PdP signals “on and off” [Thomas and Weber, 1997; Wysession et al., 1998]; our ambiguous PdP observations may result from such topography.

An important diagnostic of the D'' reflections comes from slowness estimates. The slowness results quoted in Table 2 are from the f-k analysis. The vespagram results lacked sufficient resolution to enable accurate slowness estimates to be made. One possible explanation is that the PdP phase did not travel within the source-receiver plane as discussed. In general, the accuracy of slowness estimates using the vespagram and f-k method depends greatly upon the geometry and aperture of the array and the dominant wavelength of the incoming signal. The aperture of the Yellowknife array is 20 × 20 km (Figure 1). In comparison, Weber and Davis [1990] achieved better resolution in a similar study using the Gräfenberg array (GRF) in Germany which has an aperture of 60 × 100 km. The effect of the small aperture size at YKA is evident in the relatively wide spacing of the −1 dB contours, revealing larger errors in slowness and backazimuth estimates than those obtained by other workers. Nevertheless, all 14 results showing evidence of PdP show a coherent f-k signature with a reasonable slowness and consistent travel time on the vespagram.

Figure 6 shows a comparison of the variation in measured backazimuth for P and PdP. In most cases the P phase travels close to the great circle path. In contrast, the PdP phase shows deviations from the theoretical backazimuth. Large deviations of PdP from the great circle path can be indicative of variations in the lateral structure in the D'' region. Most PdP deviations are positive indicating that the phase arrives further from the north than predicted. In an earlier study, Scherbaum et al. [1997] noted that PdP phases beneath the Arctic Sea arrived from anomalous backazimuths with no preferred direction of deviation from the great circle path, and suggested that scatterers at a variety of depths (from 2500 to 2750 km) were responsible. Previous studies, in particular S wave investigations [e.g., Young and Lay, 1990] and tomographic modeling [e.g., Grand et al., 1997], have shown the northern Pacific region to be heterogeneous at a variety of length scales. The PdP results of this study indicate that the structure of the region varies

Table 2. Summary of Results

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<th>baz(P), deg</th>
<th>u(PdP), s/deg</th>
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*Ambiguous reflector (Refl) results are denoted by question mark. Slowness (u) and baz values are from f-k analysis. See text for a discussion of slowness and baz errors. Events are earthquakes as given in Table 1.
considerably over a short length scale. This structure will also affect the slowness estimates, and may be a contributing factor to the poor slowness resolution of the vespa.

[22] Although we are talking about a reflector, the additional PdP, could be produced by scatterers [Scherbaum et al., 1997; Braha and Helfrich, 2002] or strong 3-D gradients in velocity [Liu et al., 1998]. At present, however, we are not able to distinguish between these scenarios and more detailed investigations are required.

[23] Many mechanisms have been proposed for a D" reflector (see review by Wysession et al. [1998]). The North Pacific and Alaska regions are areas of predicted paleoslab ponding at the CMB [Lithgow-Bertelloni and Richards, 1998], the southern extent of which coincides with the area covered by this study. It is therefore tempting to interpret PdP as reflections from the top of buckled slab material that has accumulated on the CMB [Christensen and Hofmann, 1994], from the top of displaced D" material [Wysession, 1996] or from a phase change in a region of subduction [Sidorin et al., 1999]. Thus slab material may be contorted and mixed with D" material leading to the observed complex morphology of the D" discontinuity.

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References


THOMAS ET AL.: D" HETEROGENEITY BENEATH THE NORTHERN PACIFIC

ESE 3 - 7

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