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

Christine Thomas,[1] Thomas Heesom, and J. Michael Kendall
Department of Earth Sciences, University of Leeds, Leeds, UK

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[1] Seismic array recordings are used to study the heterogeneity of a 15° × 25° region of the lowermost mantle beneath the northern Pacific. We investigate *P* waves from northwestern Pacific events, 68° to 82° from the Yellowknife array in northern Canada. Anomalous arrivals (*PdP*) are observed 2–13.5 s after *P* with a slowness 0.4–1.2 s/deg smaller than *P*, suggesting that they are reflections from a D″ discontinuity. We use vespagrams (slant stacks) and f-k analyses to determine travel times and slowness vectors respectively. The f-k technique simultaneously estimates both the horizontal slowness and backazimuth of arrivals at a receiver, with better resolution than vespagrams. Travel time analysis reveals a mean discontinuity height of 241 km above the core-mantle boundary in the region studied here. However, there appears a systematic variation in this thickness which ranges from 211 to 336 km. The f-k analyses also reveal variations between *P* and *PdP* backazimuths, further implying the existence of lateral variations in D″ in this area. This site is thought to be a transition region from an area of mantle downwelling to a region of upwelling; thus the variations in heterogeneity in this region may be related to its proximity to a site of paleoslab accumulation. *INDEX TERMS:* 7207 Seismology: Core and mantle; 7203 Seismology: Body wave propagation; 8124 Tectonophysics: Earth's interior—composition and state; *KEYWORDS:* lower mantle, seismology, D″ discontinuity, seismic arrays techniques

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1. Introduction

[2] 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 material in others [e.g., *Weber*, 1994; *Scherbaum et al.*, 1997; *Kendall and Shearer*, 1994; *Russell et al.*, 1999].

[3] 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].

[4] 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.

[5] 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 analysis are both used to estimate the arrival

[1]Now at Department of Earth Sciences, University of Liverpool, Liverpool, UK.

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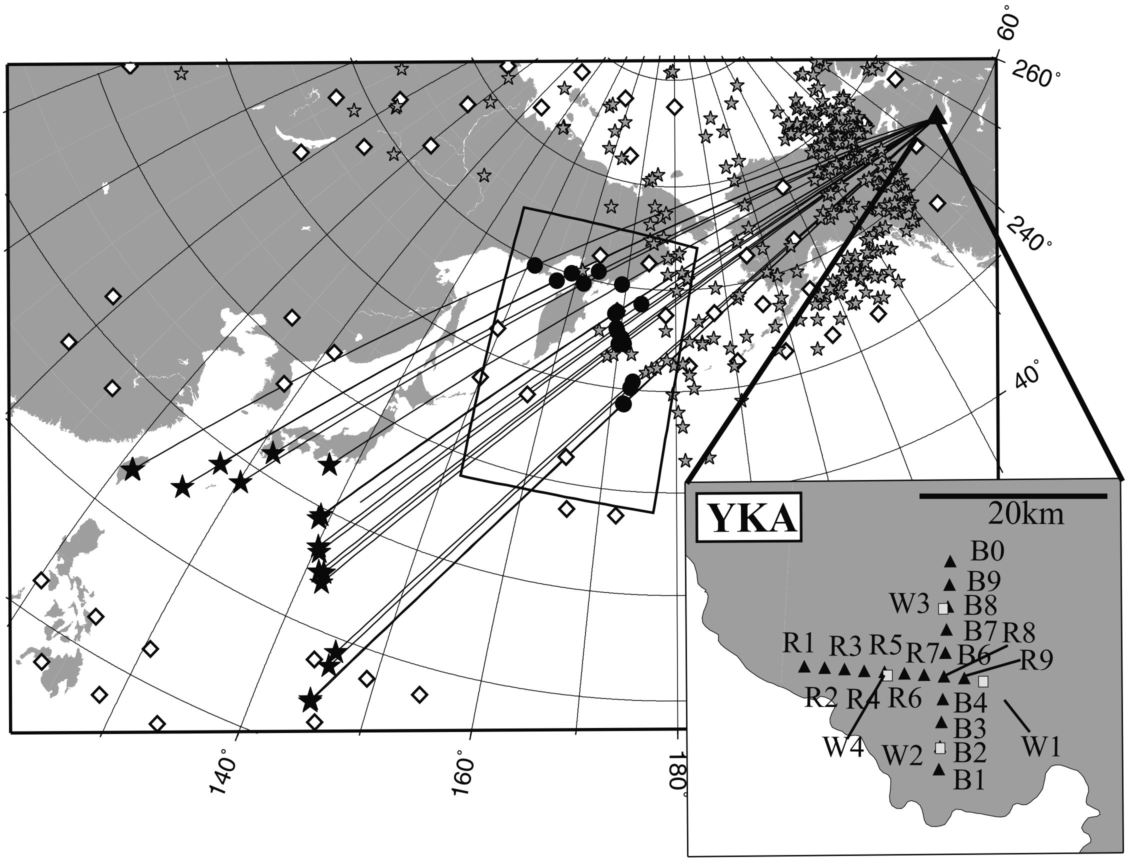

Figure 1. Map showing the sources (solid stars), the YKA array (solid triangle), and reflection points at the D″ discontinuity (solid circles). Our study region is marked by the rectangle. Previous results from *S* wave investigations of the lowermost mantle are shown as open diamonds for *Kendall and Shearer* [1994] and grey stars for *Young and Lay* [1990]. The insert shows locations of short-period stations (solid triangles) and three-component broadband stations (open squares) of the Yellowknife array, YKA, in Canada.

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 $\times$ 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 *N*th root stacking technique [*Muirhead and Datt*, 1976; *Davies et al.*, 1971] that sums the *N*th 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.,

Table 1. Events Used in This Study[a]

Event	Date	Time, UT	Latitude, °N	Longitude, °E	Depth km	m_b	baz	Distance
1	1 Oct. 1995	1706:03.45	29.31	139.04	430	6.1	298	71
2	18 Oct. 1995	2325:58.77	28.20	130.21	27	6.1	304	76
3	16 March 1996	2204:06.24	28.98	138.94	477	6.7	298	72
4	9 June 1996	0112:16.76	17.44	145.46	149	6.0	287	79
5	15 July 1996	1651:22.07	18.74	145.63	177	5.9	287	78
6	18 Oct. 1996	1644:47.91	33.69	137.40	337	5.6	301	68
7	19 Oct. 1996	1444:40.79	31.89	131.47	22	7.0	305	72
8	23 April 1997	1944:28.42	13.99	144.90	100	6.5	286	82
9	13 Aug. 1997	0445:04.86	25.03	125.77	55	6.2	307	80
10	1 Jan. 1998	0611:22.64	23.91	141.91	95	6.6	293	75
11	7 Feb. 1998	0113:36.81	24.79	141.75	525	5.9	293	74
12	7 Feb. 1998	0118:59.50	24.82	141.75	525	6.4	294	74
13	15 May 1998	0558:06.04	14.18	144.88	154	6.1	286	82
14	03 Oct. 1998	1115:42.69	28.51	127.52	227	6.2	307	77
15	18 Oct. 1998	0139:01.35	24.72	141.24	110	5.5	294	75
16	12 Jan. 1999	0232:25.59	26.74	140.17	440	6.0	296	73
17	3 July 1999	0530:10.09	26.32	140.48	430	6.1	295	74
18	22 Sept. 1999	0014:39.15	23.73	121.17	26	6.4	310	82
19	22 Sept. 1999	0049:42.77	23.64	121.14	33	6.2	310	82

[a] The earthquake parameters are from the Preliminary Determination of Earthquakes (PDE) catalogue. Distances and backazimuths (baz) are calculated for the center of the array.

P 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 D″ (*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.

[9] 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 D″ 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

[10] 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-

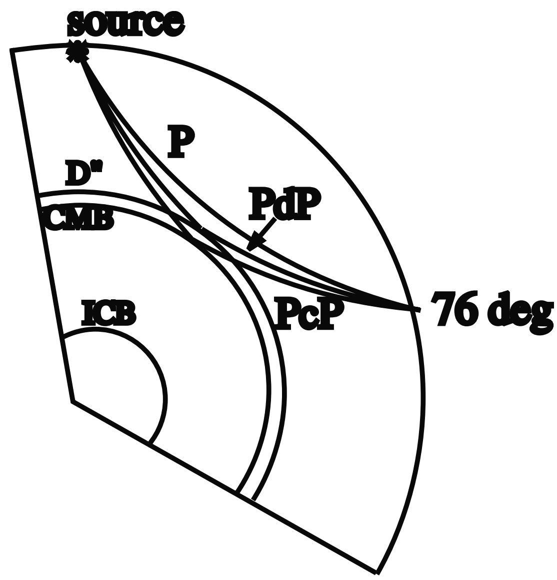

Figure 2. The phases *P*, *PdP*, and *PcP* in the epicentral distance range 68° to 82° are used to investigate the D″ *P* wave discontinuity. See text for further discussion.

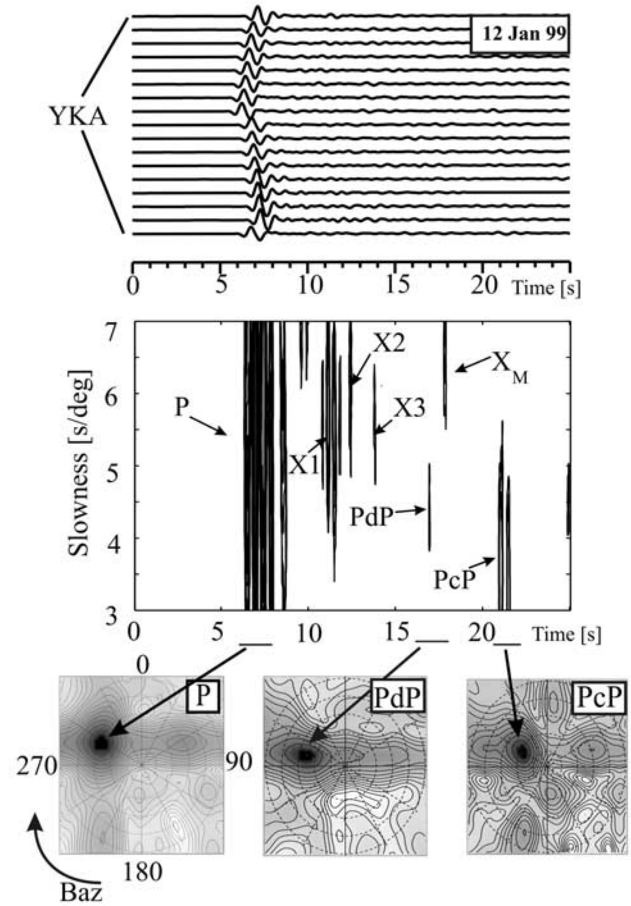

Figure 3. (top) Data, (middle) vespagram (slowness versus time), and (bottom) f-k plots for *P* and time windows around expected *PdP* and *PcP* for event 12 January 1999 (event 16 in Table 1), which shows evidence for a D″ discontinuity. The epicentral distance for this event is 73° and the theoretical 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.

aperture array. Thus it is difficult to confidently interpret these arrivals as *PdP* based on the vespagram alone. Therefore f-k plots are needed to ascertain the coherent nature of the arrivals for both the *P* and *PdP* phases.

[11] Figure 3 shows the result for event 16 in Table 1. It is a typical example of an event showing evidence for *PdP* in the vespagram but little indication of an anomalous arrival following *P* in the raw data. *PcP* is also not observed in the raw data because of the low impedance contrast across the CMB for *P* waves at this distance. However, the *PdP* phase is visible in the vespagram with an expected slowness but 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 additional 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.

[12] 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 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

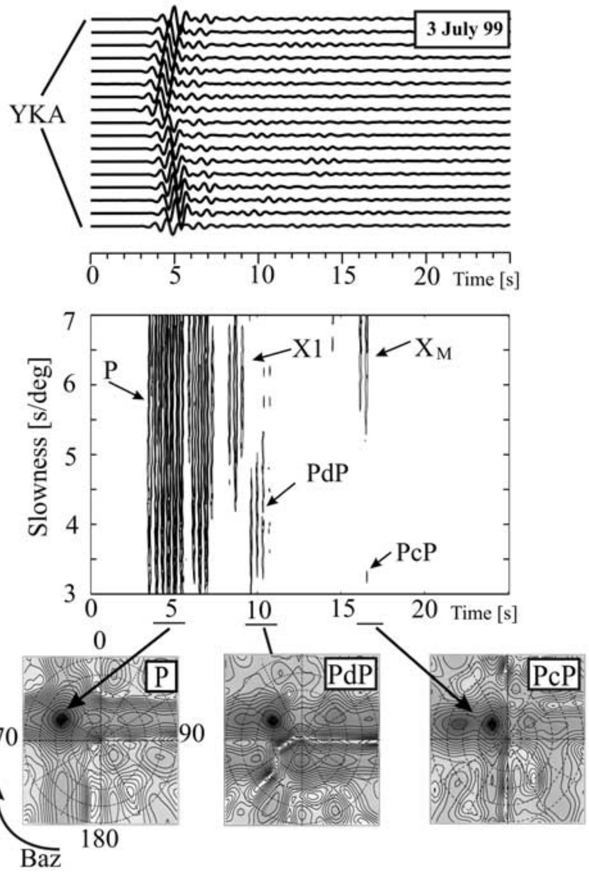

Figure 4. (top) Data, (middle) vespagram, and (bottom) f-k plots for *P* and *PdP* for event 3 July 1999 (event 17 in Table 1). The epicentral distance is 74° and the theoretical backazimuth is 295°. This example illustrates the advantages of the f-k method over the vespa method when phases do not travel along the great circle path. *P*, *PdP*, and *PcP* are clearly visible in the f-k plots but *PcP* seems to arrive with a backazimuth difference of 16°. The slowness of *PdP* in the vespagram seems to be unusually low for *PdP*, but this estimate is affected by the significant deviation from the great circle path (8°). The f-k analysis shows that the phase has in fact a more realistic *PdP* slowness. Two additional phases can be seen, X_M and X_1, which have a slowness comparable to the *P* slowness. X_M is again most likely the Moho reverberation and X_1 again the reflection for the shallow feature. (See also Figure 3 caption).

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.

[15] 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.

[16] 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

[17] 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

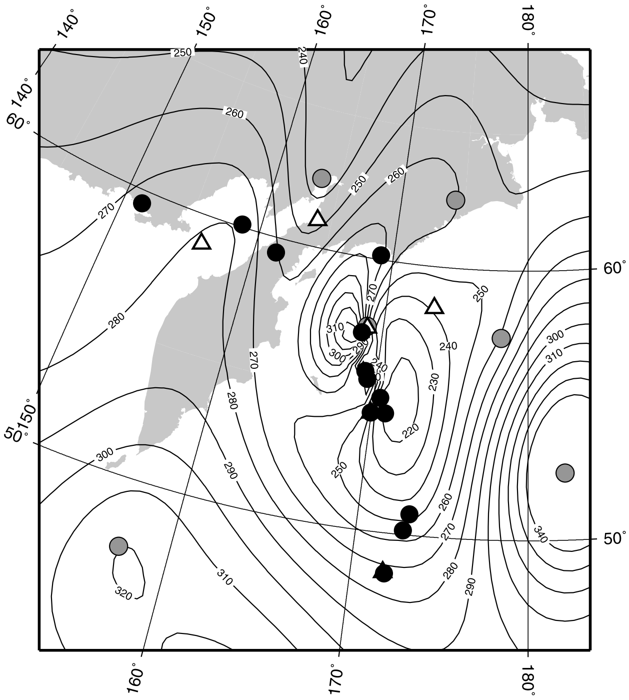

Figure 5. Equal angle projection of the study area summarizing the 21 results (see also Table 2). Solid circles denote the presence of a D″ discontinuity; ambiguous results are denoted by open triangles. The results from *Kendall and Shearer* [1994] are given by the grey circles. The reflector height is shown as contour lines. A trough in the discontinuity can be seen at a latitude of around 55° followed by a peak to the north and increasing height to the south.

Table 2. Summary of Results[a]

Event	Reflector	*u*(*P*), s/deg	baz(*P*), deg	*u*(*PdP*), s/deg	baz(*PdP*), deg	Reflector Height, km
1	?	-	-	-	-	-
2	yes	5.35	304	4.99	306	261
3	yes	5.94	297	5.37	303	334
4	yes	5.55	291	4.93	294	256
5	yes	5.32	292	4.72	293	246
6	yes	5.61	299	4.5	311	273
7	?	-	-	-	-	-
8	yes	5.3	288	5.18	289	273
9	?	-	-	-	-	-
10	yes	5.56	294	5.07	300	218
11	?	-	-	-	-	-
12	yes	5.46	296	3.9	304	238
13	?	-	-	-	-	-
14	yes	5.45	304	4.92	307	278
15	yes	5.52	297	4.81	298	218
16	yes	5.58	297	5.09	283	211
17	yes	5.7	297	4.5	303	289
18	yes	5.03	314	4.8	306	273
19	yes	5.0	316	4.7	318	273

[a] 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.

the top 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].

[18] 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.

[19] 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.

[20] 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.

[21] The f-k analysis has also provided an estimate of the backazimuth of each phase. The results reveal between 1° and 14° difference between the *P* and *PdP* backazimuths (Table 2). 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

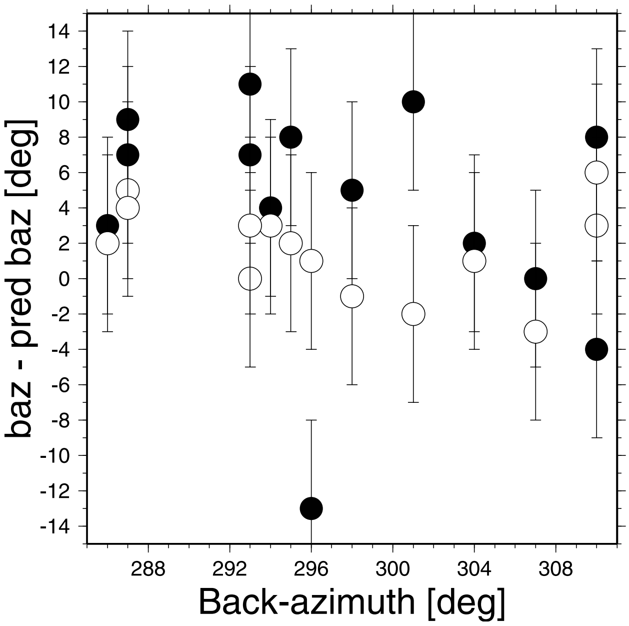

Figure 6. Backazimuth minus predicted backazimuth as a function of backazimuth for *P* (open circles) and *PdP* (solid circles) arrivals with the observational error. Most deviations of *PdP* are toward larger backazimuths (i.e., arrive farther from the north than expected).

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 vespagrams.

[22] Although we are talking about a reflector, the additional phase, *PdP*, could be produced by scatterers [*Scherbaum et al.*, 1997; *Braña and Helffrich*, 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.

[24] **Acknowledgments.** We would like to thank the staff from the Geological Survey of Canada for providing the data from the CNSN. Thanks to Thorne Lay for providing reflection points from previous studies. All maps were produced using the program GMT by *Wessel and Smith* [1991]. C.T. was supported by DFG grant TH763/1-1 and NERC grant GR3/12578. Ed Garnero, Matt Fouch, and Rob van der Hilst are thanked for helpful reviews.

References

Bostock, M. G., Seismic imaging of lithospheric discontinuities and continental evolution, *Lithos*, *48*, 1–16, 1999.

Braña, L. P., and G. Helffrich, A discrete scatterer near the core-mantle boundary under the North Atlantic, *Geophys. J. Int.*, in press, 2002.

Bullen, K. E., Compressibility-pressure hypothesis and the Earth's interior, *Mon. Not. R. Astron. Soc., Geophys. Suppl.*, *5*, 355–368, 1949.

Capon, J., Signal processing and frequency-wavenumber spectrum analysis for a large aperture seismic array, *Methods Comput. Phys.*, *13*, 2–59, 1973.

Christensen, U. R., and A. W. Hofmann, Segregation of subducted oceanic crust in the convecting mantle, *J. Geophys. Res.*, *99*, 19,867–19,884, 1994.

Davies, D., E. J. Kelly, and J. R. Filson, Vespa process for analysis of seismic signals, *Nature Phys. Sci.*, *232*, 8–13, 1971.

Garnero, E. J., Heterogeneity of the lowermost mantle, *Annu. Rev. Earth Planet. Sci.*, *28*, 509–537, 2000.

Grand, S. P., R. D. van der Hilst, and S. Widiyantoro, Global seismic tomography: A snapshot of convection in the Earth, *GSA Today*, *7*, 1–7, 1997.

Gutenberg, B., Über Erdbebenwellen VIIA. Beobachtungen an Registrierungen von Fernbeben in Göttingen und Folgerungen über die Konstitution des Erdkörpers, *Nachr. K. Ges. Wiss. Göttingen Math. Phys. Kl.*, *14*, 125–177, 1914.

Kendall, J.-M., and C. Nangini, Lateral variations in D″ below the Caribbean, *Geophys. Res. Lett.*, *23*, 399–402, 1996.

Kendall, J.-M., and P. M. Shearer, Lateral variations in D″ thickness from long-period shear wave data, *J. Geophys. Res.*, *99*, 11,575–11,590, 1994.

Knittle, E., and R. Jeanloz, The Earth's core-mantle boundary: Results of experiments at high pressures and high temperatures, *Science*, *251*, 1438–1443, 1991.

Lay, T., and D. V. Helmberger, A lower mantle *S*-wave triplication and the shear velocity structure of D″, *Geophys. J. R. Astron. Soc.*, *75*, 799–837, 1983.

Lithgow-Bertelloni, C., and M. A. Richards, The dynamics of Cenozoic and Mesozoic plate motions, *Rev Geophys.*, *36*, 27–78, 1998.

Liu, X.-F., J. Tromp, and A. M. Dziewonski, Is there a first-order discontinuity in the lowermost mantle?, *Earth Planet. Sci. Lett.*, *160*, 343–351, 1998.

Loper, D. E., and T. Lay, The core-mantle boundary region, *J. Geophys. Res.*, *100*, 6397–6420, 1995.

Muirhead, K. J., and R. Datt, The *n*-th root process applied to seismic data, *Geophys. J. R. Astron. Soc.*, *47*, 197–210, 1976.

Ringwood, A. E., *Composition and Petrology of the Earth's Mantle*, McGraw-Hill, New York, 1979.

Rost, S., and C. Thomas, Array seismology: Methods and applications, *Rev. Geophys.*, *40*, doi:10.1029/2000RG000100, in press, 2002.

Rost, S., and M. Weber, A reflector at 200 km depth in the northwestern Pacific, *Geophys. J. Int.*, *147*, 12–28, 2001.

Russell, S. A., T. Lay, and E. J. Garnero, Small-scale lateral shear velocity and anisotropy heterogeneity near the core-mantle boundary beneath the central Pacific imaged using broadband *ScS* waves, *J. Geophys. Res.*, *104*, 13,183–13,199, 1999.

Scherbaum, F., F. Krüger, and M. Weber, Double beam imaging: Mapping lower mantle heterogeneities using combinations of source and receiver arrays, *J. Geophys. Res.*, *102*, 507–522, 1997.

Sidorin, I., M. Gurnis, and D. V. Helmberger, Dynamics of a phase change at the base of the mantle consistent with seismological observations, *J. Geophys. Res.*, *104*, 15,005–15,023, 1999.

Stacey, F. D., and D. E. Loper, The thermal boundary layer interpretation of D″ and its role as a plume source, *Phys. Earth Planet. Inter.*, *33*, 45–55, 1983.

Thomas, C., and M. Weber, *P* velocity heterogeneity in the lower mantle determined with the German Regional Seismic Network: Improvement of previous models and results of 2D modelling, *Phys. Earth Planet. Inter.*, *101*, 105–117, 1997.

Weber, M., *P*- and *S*-wave reflections from anomalies in the lower mantle, *Geophys. J. Int.*, *115*, 183–210, 1993.

Weber, M., Lamellae in D″? An alternative model for lower mantle anomalies, *Geophys. Res. Lett.*, *21*, 2531–2534, 1994.

Weber, M., and J. P. Davis, Evidence of a laterally variable lower mantle structure from *P*- and *S*-waves, *Geophys. J. Int.*, *102*, 231–255, 1990.

Wessel, P., and W. H. F. Smith, Free software helps map and display data, *Eos Trans. AGU*, *72*(441), 445–446, 1991.

Wysession, M. E., Imaging cold rocks at the base of the mantle: The sometimes fate of slabs?, in *Subduction: Top to Bottom*, edited by G. E. Bebout et al., pp. 369–384, AGU, Washington, D. C., 1996.

Wysession, M. E., T. Lay, J. Revenaugh, Q. Williams, E. J. Ganero, R. Jeanloz, and L. H. Kellog, The D″ discontinuity and its implications, in *The Core Mantle boundary Region*, *Geodyn. Ser.*, vol. 28, edited by M. Gurnis et al., pp. 273–297, AGU, Washington, D. C., 1998.

Young, C. J., and T. Lay, Multiple phase analysis of the shear velocity structure in the D″ region beneath Alaska, *J. Geophys. Res.*, *95*, 17,385–17,402, 1990.

T. Heesom and J. M. Kendall, Department of Earth Sciences, University of Leeds, Leeds LS2 9JT, UK. (m.kendall@earth.leeds.ac.uk)

C. Thomas, Department of Earth Sciences, University of Liverpool, 4 Brownlow Street, Liverpool L69 3GP, UK. (tine@liverpool.ac.uk)