Seismic Images of the Core-Mantle Boundary

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ABSTRACT

Seismology presents several ways of providing images of the geologic structures that exist in the lowermost mantle just above the core-mantle boundary (CMB). An understanding of the possibly complex geophysical processes occurring at this major discontinuity requires the combined efforts of many fields, but it is the role of seismology to geographically map out this largely uncharted territory. Seismic phases that reflect, diffract, and refract across the CMB can all be used to provide different information in different ways. Profiles of core-diffracted and core-reflected waves are especially powerful when used as differential traveltimes in relation to direct phases. The resulting seismic maps show long-wavelength lateral heterogeneity in the lowermost few hundred kilometers of the mantle (a region called D") with a magnitude of at least 6%, which is comparable only to Earth's upper few hundred kilometers. The maps of lateral seismic variations show significant continent-sized features that are most likely a result of the convective dynamics occurring at the base of the mantle. The geophysics of the CMB and lowermost mantle probably has many analogies with that of Earth's other boundary layers and core, and variations in the structure of D" may likewise be a combined result of thermal, chemical, and mineral phase changes. In the variations described here, the density jump of 4.3 kg/m³ between the core and mantle. With a planet's other major boundary, that of the core, we produce, which represent the densities of materials we might expect to occur in the lower mantle. Mineral physicists, conducted across it, and a thermal wave, in the core, have noted the effects of CMB topography and the thermal variations of the lowermost mantle create observable variations in the geomagnetic field by affecting core flow. Geodynamic modeling, both experimental and numerical, is providing real-time histories of the patterns of convection that might occur in the lower mantle. Mineral physicists, through both high-pressure diamond anvil experiments and theoretical equation of state, are delineating the kind of mate- rials we might expect to occur at these great pressures and temperatures. The stories emerging about the CMB are quite exciting, involving rising hot plumes, sinking cold mantle, lateral swept mantle dredges, core-mantle chemical reactions, and core-mantle dynamic coupling, but because they are compatible with evidence across many independent disciplines, they are not quite as speculative as they may seem.

INTRODUCTION

While most geologists, including specialists in the field of seismology, are familiar with the Earth's core, its geology is less well known. The core consists of an outer core of liquid iron, as well as a temperature increase of possibly 1500 °C between the lower mantle adiabat and outer core. Within the core-mantle boundary (CMB) may well be Earth's most significant and dramatic discontinuity. Our increasing knowledge of this highly variable and heterogeneous region has come through the combined efforts of geoscientists in a wide array of fields, and an important part of this effort has been the use of seismology to map out the structures that exist there. Because of the limitations of imaging a surface nearly 3000 km beneath us through a heterogenous mantle, our images lack clear resolution. In a sense we are like the seafaring explorers of 500 years ago who had mapped out the outlines of the world's continents but still knew little of what lay within them. In this article I discuss a few attempts to get clearer maps of the continental structures in the mantle and describe some of the directions that may be taken to develop a sharper image. The red-and-blue seismic maps that we produce, which reflect the velocities with which P and S waves propagate through a given region, do not mean very much by themselves. However, these two velocities are functions of density, rigidity, and incompressibility, which are complicated functions of temperature, composition, and mineralogical phase. If we cannot be sure of the densities and mineralogy in this region, we cannot be confident in our interpretations of how the Earth is deforming. Geodynamic modeling, both experimental and numerical, is providing real-time histories of the patterns of convection that might occur in the lower mantle. Mineral physicists, through both high-pressure diamond anvil experiments and theoretical equation of state, are delineating the kind of materials we might expect to occur at these great pressures and temperatures. The stories emerging about the CMB are quite exciting, involving rising hot plumes, sinking cold mantle, lateral swept mantle dredges, core-mantle chemical reactions, and core-mantle dynamic coupling, but because they are compatible with evidence across many independent disciplines, they are not quite as speculative as they may seem.

Figure 1. Images from Mohorovicic picture showing the propagation of seismic wave energy through the mantle (Wysession and Shoemaker, 1994). The images correctly show the locations of the seismic wave fronts at (A) 320, (B) 540, (C) 800, (D) 1040, (E) 1360, and (F) 1800 s after CMB. The occurrence of a 600-km-deep earthquake at the lower left of the image. Red is out of the page; blue into the page. Amplitudes are normalized and raised to a power of 0.8 to enhance smaller features. Images were made by interpolating between a grid of 72,846 synthetic seismograms calculated by the superposition of all torsional normal modes (28,385) with periods greater than 12 s.

The examination of the CMB using the seismic waves from large earthquakes has a long history. R. Oldham first identified the core in 1906, and L. Lehmann discovered the inner core in 1936. By the 1940s, scientists like K. Bullen had not only determined reasonable radial models of Earth's seismic velocities, but had even noted the unusual behavior of the then-named D" layer at the bottom of the mantle. As late as the 1980s most seismologists observed a decrease in P velocities relative to the rest of the mantle, which made sense thermodynamically; if the CMB is a chemical boundary between rock and iron, then heat must be conducted across it, and a thermal boundary layer will likely form at the bottom of the mantle. This thermal boundary layer will have temperatures hotter than the rest of the lower mant-"e adiabat and will have approximately slower velocities.

The 1980s, however, brought two seismological findings of primary importance.

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Editor's Note: Each year the David and Lucile Packard Foundation awards 20 Fellows in earth science, including some in Alas-
SEISMIC TOOLS FOR STUDYING THE CMB

There are three categories of waves that can be used to examine the CMB: reflected, refracted, and diffracted. These are usually demonstrated by ray-tracing, where the wave paths are represented by the straight lines of particle paths. This simple ray-tracing, however, is inadequate for describing the true nature of the interactions; the waves that leave the earthquake do not behave like particles, but travel as three-dimensional wave fronts. For this reason, a better picture of the waves that interact with the CMB can be seen in Figure 1A, which is an accurate representation of the horizontal shear (SH) waves that would propagate through the mantle from an earthquake, in this case at a depth of 600 km. The images represent the displacement of the waves in slices through the mantle at different times after the earthquake. The images are part of a movie created through the summation of different normal modes of Earth’s oscillations (Wysession and Shore, 1994).

In Figure 1A, 320 s after the earthquake, the initial wave front (Scs) is still quite simple, having only just reflected off the surface, but as time passes the waves become more and more complex because of their continued interactions with the surface, CMB, internal mantle discontinuities, and a velocity structure that increases with depth. By 640 s (Figure 1B) the Scs wave can be seen leaving the CMB and heading back to the surface. This core-reflected phase is easily recorded at the surface at distances of up to 85° away from the earthquake and has provided the majority of information about the seismic shear structure above the CMB. By 800 s (Figure 1C) a second wave is reflecting off the CMB, the surface-reflected Scs, but by this time the bottom part of the initial wave front is no longer “reflecting” off the core; the wave has turned the corner around the core and is now diffracting along the CMB. The diffracted waves (Sidf, or equivalently, Pdif), which are recorded at distances of greater than about 100° from the earthquake, theoretically continue indefinitely around the core, but in reality quickly lose their energy and are rarely observed beyond about 150°. This means, however, that at any distance range of 100°–150° Sidf and Pdif arrivals at the surface provide a lot of information about the structure of the lowermost mantle.

Important about D". Global tomographic images of huge seismic data sets began to show coherent patterns of very long wavelength variations at magnitudes comparable only to those of Earth’s surface. In addition, mounting evidence supported the findings of Lay and Helmberger (1983) that in many if not most regions of the CMB, the top of D" is characterized, surprisingly, by a significant increase in velocity. The current state of CMB seismology is very active. Many seismologists are now studying the CMB with both global and regional approaches and with types of data that range from high-frequency (1–10 Hz) CMB-scattered P waves to very low frequency (<0.1 Hz) normal modes.

CMB-DIFFRACTED WAVES

Sidf and Pdif are excellent waves for looking at the structure of the base of the mantle because they can spend up to one-third of their total traveltime within D". They are also the first arrivals of their kinds (Sidf is the first arrival of any kind beyond 100°, and Pdif is the first shear arrival), which often makes them easy to detect. Some complications with these phases have prevented their widespread incorporation into seismic studies. High-frequency energy dissipates very quickly during diffraction, so the very long period arrivals do not make the picking of clear onset times. The high-frequency decay does not re semble seismic analytic attenuation and cannot be easily corrected. In addition, because the diffracted waves also travel a great distance through the heterogeneous mantle and crust on their way to and from the CMB, it is difficult to distinguish D" heterogeneities from those present elsewhere.

The studies of Wysession et al. (1992) used a stringent set of requirements and corrections to map out D" variations from profiles of Sidf and Pdif. The ray parameters, or slownesses, were determined for many arrivals traveling a long distance along a narrow swatch of the CMB. Combined with mantle path corrections using three-dimensional (3-D) tomographic models as well as synthetic modeling, this technique reduces contamination from source mislocation, slab diffraction, upper-mantle path heterogeneity, ellipticity, and high-frequency energy dissipation.

An example (Figure 2) shows six WSSN Sidf arrivals (top), modeled by their reflectivity synthetic counterparts (bottom). The slope through the data is the ray parameter and is a direct result of the average velocity structure in D". Maps of the results for 12 Sidf and 20 Pdif profiles (Figures 3 and 4) show the windows onto the core where enough diffracted arrivals meet our criteria. The total variation for both P and S velocities, determined at very long wavelengths, was about 4%. The most striking feature in both maps is a region of D" beneath the western Pacific islands where the seismic velocities were 3% slower than for the radial Earth model PREM (Dziewonski and Anderson, 1981). Just to the west, the inferred P and S velocities were found to be about 1% faster than PREM. This pattern correlates well with the results of other seismic studies done using totally independent data sets, such as tomographic mantle shear velocity models of Su et al. (1994).

Another interesting pattern was found for the D" velocities beneath the northern Pacific, which were sampled by paths from earthquakes near Japan to stations in North and South America. Here it was found that the shear waves were consistently faster than average, whereas the average radial D" velocities were slower than average. This suggests that the P and S velocities may not always vary in the same manner, an observation that has also been made in tomographic models of the lowermantle structure. The variation of the Poisson ratio of the lowestmantle may be real, just as the Poisson ratio in Earth’s crust is seen to vary regionally.

Core-diffracted waves also provide information about the poorly known vertical velocity structure in D". In much way surface waves can be used to determine upper-mantle structure. Because core phases that sample the CMB must pass vertically across D" and back, there is difficulty in resolving the layer's vertical structure. It is hard to tell whether the heterogeneities are at the top or bottom of the layer. Studies similar to that by Lay and Helmberger (1983) identified the top of D" where a sharp velocity increase creates an additional seismic precursor (PS) to Scs. Core-diffracted waves provide additional information about the rest of D", longer wavelengths traveling shorter wavelengths staying closer to the CMB. Valenzuela et al. (1994) showed preliminary results using core-grazing S waves from five northern California earthquakes recorded at the Tibetan Plateau PASSCAL array. The data were forward-modeled by synthetic seismograms for a wide array of seismic models, and the structure in Figure 5 was found to be the best fit, a sudden increase in velocity 290 km above the CMB, with a rapid decrease at the bottom of the layer. This structure is very similar to those obtained from portable arrays of broad-band
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seismometers (recording a "broad band" of frequencies) at the center of using Sdfit or Pilfit amplitudes as a function of distance and frequency should play an important role in helping resolve vertical structure for limited regions within D*.

**CORE-REFLECTED WAVES**

Because of the unusual nature of the D* region beneath the western Pacific, Wyssen et al. (1994, 1995b) further investigated this region using Scs and Scs-s6 differential travel times (the Scs-s5, s5, and Scs waves are the first four wave fronts shown propagating away from the earthquake in Fig. 1C). Using 747 differential travel times between direct and core-reflected shear waves, we attained a higher resolution map of the lateral variations in D* shear velocities for this region (Fig. 6). The use of differential times of seismic phases from the same earthquake is a powerful tool for examining Earth structure, because source and receiver effects are minimized, and the Sd velocity model (SHS/WM13 of Woodward et al. [1993] to help remove mid-lithosphere effects on the travel time. Any remaining traveltime residuals were converted into velocity variations by moving a weighted Gaussian cap with a 300 km radius across them to average the geophysical contributions and help simulate the CMB flexural zones, or sampling regions, of the Scs and Scs-s5 footprints. The resolution of the result (Fig. 6) is on the order of about 300 km, or S°. (Note that the amplitudes of the original figure in Wyssen et al. [1994] were erroneously amplified by a factor of two; this was corrected in Wyssen et al. [1995a].)

The variations in seismic velocity found for this part of the lowermost mantle range over about ±3%, with several notable features. Not all of the region is sampled because of our inability to image permanent seismometers in the oceans and because of the uneven distribution of earthquakes across Earth. In the middle of the region where we do have coverage, corresponding to D* beneath Micronesia, we find a broad low-velocity zone. The average velocity is 1.5% slower than for PREM, but reaches values up to 3%, especially in the slow-velocity arm that extends toward the west. This broad low-velocity zone (LVD) is surrounded on three sides by regions showing fast velocities. The average of these regions is about 25% faster than for PREM but reaches values greater than ±3%. The fast velocities to the south and west of the D* LVD seem to form one continuous feature that extends from beneath China to beneath the Australian region. This is a correlation between this fast D* "rock" and the location of the paleocean of the Jethys plate. The fast-velocity region northeast of the LVD is poorly constrained in its lateral extent and is beneath the southern part of the Pacific Ocean. We have no coverage of what happens to the LVD east of the study region, but it could represent large-scale inhomogeneous regions with amplitudes like those of Su et al. (1994) are an indication, it probably extends a linear way eastwards as part of a broad low-velocity region beneath the central Pacific.

In an attempt to get at what the P velocities might be doing in the same region, Zhu and Wyssen (1995) presented a map of D* P velocities by stacking the differential times of PcP-P for those stations that reported arrivals of both to the International Seismological Centre during the time 1964–1987. While these times, especially for the secondary and often much smaller PcP arrivals, are not as reliable as times determined through personal analyses, there is statistical significance in the picture obtained. The mean of the entire data set was 0.35% larger than for IASP91. This could partly be an indication that IASP91 is on average too fast for the lowermost mantle or for the regions that had the greatest coverage, but it may also be the result of a systematic bias in picking the PcP arrivals too late. It is interesting to note that Figure 7 shows a map of lateral velocity variations in D* beneath the western Pacific, compiled from Scs and Scs-s5 differential travel times. All ray paths are corrected for mantle path heterogeneities outside of the bottom 300 km of the mantle, and the velocity magnitudes are computed assuming that the traveltime residuals are the result of heterogeneities only within this bottom layer. The data are robust, containing little scatter, and the resulting image shows coherent velocity variations at continents-sized wavelengths.

**Thermal Variations**

Some estimates of the temperature differences between the lower-mantle and outer-core adiabats are as large as or larger than 1500 °C (Boehler, 1994). As little or no mass seems to be transported across the CMB, this heat must pass into the mantle via conduction, and although there are some reports that a very high thermal conductivity in D* could lessen the effect, the result will be a thermal boundary layer. This would be analogous to the thermal lithosphere at the surface, where heat brought near to the surface by convection must be conducted across the lithosphere boundary before radiating into space. With the thermal lithosphere, we would expect horizontal mass movements to result in lateral variations in the temperature within such a thermal boundary layer. The temperature 50 km below a mid-ocean ridge is much hotter than the temperature 50 km beneath an oceanic abyssal plain, and this can be observed as an increase in seismic velocities as waves move away from ridges. Something analogous is probably happening in D*, and some of the lateral seismic variation seen there probably has a thermal component. If the vertical change in temperature across D* is 1500 °C, then it is possible to have lateral temperature variations approaching this amount. Seismic variations would then be representative of vertical mass movements associated with lower mantle convection. Fast regions are cold and sinking, or recently sunk. Slow regions are hot, buoyant, and on their way back toward the core. Dimensional tomographic models interpreted as buoyancy forces resulting from these mass variations do a good job of modeling the observed long-wavelength geoid (Fort et al., 1994).

The thermal model of D* can be taken a step further to incorporate a direct correlation with plate tectonics, showing that the CMB is not immune to arguments about the degree of mixing between the upper and lower mantles. A correlation has long been identified between the location of major subduction zones and bands of fast seismic shear velocities in D* (as in Fig. 6), and likewise between slow D* velocities and regions that have a high density of hotspots, like the central Pacific and the western African plates. It is exciting to think of a mantle-wide circulation system bringing subducted plates all the way to the CMB where they heat up and eventually rise to the surface as hotspot mantle plumes, but the correlations between D* seismic velocities and paleosubduction would be equally satisfied by thermal coupling between an independent upper mantle and lower mantle. It is doubtful that a solution to the whole-mantle vs. layered-mantle convection argument will be found at the CMB.

**Chemical Variations**

Earth's surface has not only thermal variations but also compositional variations, and it is possible that a chemical boundary layer analogous to the crust exists in D*. Chemical "drugs," dense iron alloys, could have formed at the base of the mantle early on or at subduction zones and could be continually settling out of the lower mantle during convection. Core-mantle reaction byproducts could be stripped away from the CMB by horizontal convection to form laminar aggregates. The eclogite crust of sub-

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**Figure 5.** Profile from Valenzuela et al. (1994) showing the 5-velocity model for a patch of the lowest mantle northeast of the northern part of the core-mantle boundary (CMB) geology from processes associated with lower mantle convection. Fast regions are cold and sinking, or recently sunk. Slow regions are hot, buoyant, and on their way back toward the core. Dimensional tomographic models interpreted as buoyancy forces resulting from these mass variations do a good job of modeling the observed long-wavelength geoid (Fort et al., 1994).

**Figure 6.** A map from Zhu and Wysession (1995) of lateral S velocity variations in D* beneath the western Pacific, compiled from Scs and Scs-s5 differential travel times. All ray paths are corrected for mantle path heterogeneities outside of the bottom 300 km of the mantle, and the velocity magnitudes are computed assuming that the traveltime residuals are the result of heterogeneities only within this bottom layer. The data are robust, containing little scatter, and the resulting image shows coherent velocity variations at continents-sized wavelengths.
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ducted slabs could delaminate and form an anomalous phase, as opposed to the ambient lower mantle. Lighter elements could be setting up and out of the outer core as a more iron-rich inner core freezes at an inner core boundary of isostatic equilibrium, increasing the iron content of the outer core. All of these mechanisms have been proposed, but it is not clear which one will dominate any time.

Chemical variations will cause $P$ and $S$ seismic-velocity variations that are not derived from thermal effects. For example, the major lower-mantle constituents, perovskite and magnesium silicates, have different densities and reflectivity. Figure 8, from Wysession et al. (1993), uses a third-order Birch-Murnaghan equation of state to show the amount of change in temperature, density, and elastic properties. The temperature dependence of these equations is very difficult to measure because of trade-offs with velocity hetero-
geneities, but it is vitally coupled to dynamic CMB processes. The CMB will be depressed in the regions of lower mantle downwelling because of the iso-
static weight of the colder rock as well as the dynamic force of the convection. However, we would also expect the CMB to be depressed beneath regions of compositional density manto.

A possible scenario is that the CMB topography undergoes a cyclic transition during the cycle of mantle mass transport. During the initial stages of the birth of a mantle plume, the increased temperature will cause an elevated CMB, but as the plume devlops, denser mantle aggregates will be swept largely to the site of the plume, causing a depression of the CMB. In other words, the cyclic convective cycle may cause a temporal variation in CMB topography similar to that modeled by Gunn (1992) for long-wavelength sur-
face topography during the history of lithospheric subduction.

The electrical conductivity for lower-
mantle phases may vary by 11 orders of magnitude (Jeanloz, 1990). This is important because determina-
tions of CMB topography are sensitive to electrical conductivity. If such a phase transition does not occur at the bottom of the core, then the “moral” of the story is still standable at the advanced undergradu-
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down science for popular consump-
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ture. The seismic observations of Kendall and Shearer (1994) and Beavan and Jordan (1991) suggest that a correlation may exist between regions of shallow D Challenger, as defined by the height of the discontinuous velocity increase and fast seismic velocities, as determined from tomographic models. Fast velocities mean colder temperatures, suggesting that the phase transition would be endothermic just as with the 660 km discontinuity, colder temperatures would depress the phase boundary. However, mineral physics experiments suggest that this transformation would occur at higher pressures (greater depths) if the rock are enriched in magnesium relative to iron, so the phase transition could also be exothermic if rock were significantly depleted in iron relative to its surrounding. Advances in mineral physics will eventually solve this question of the possibility of a D9 phase transition and the form it would take.

FUTURE DIRECTIONS

The only real fact that can be gleaned from the previous section is that we still do not have the answers. Several very good scenarios have been identified, but much more work needs to be done to discern among them. The future directions for seismology in mapping the CMB and lowermost mantle include: improving in using new phases, developing new techniques, and obtaining new data sets. An example of using new phases is the utilization of differential Pdiff and PKP phases (which refract through the core) for looking at long-wavelength DEP and BCdiff seismic waves, which will consist of 18 stations linearly connecting CCM and HRV and will be recording until March 1996, is ideally suited to record core-grazing and diffracting waves from the seismic wave-front Pacific regions. Such temporary arrays, funded through the NASGAL program of IRS (Research Institutions for Seismology), greatly help in providing new seismic data that fill in the aliasing gaps between permanent seismometers.

The most important aspect of the future of seismology in imaging the CMB and lowermost mantle is the continued communication between seismologists and scientists from other fields. SED (Studies of Earth's Deep Interior) organizations exist nationally within the American Geophysical Union and the National Science Foundation and are organized internationally as well. This provides many opportunities for seismologists to share both observations and insights. Input from these interactions gives us an understanding of what the important questions are and where to concentrate our efforts. The recent successes in understanding the CMB have come about through interdisciplinary cooperation and will continue to happen in this way.

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