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INSIDE

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 lithosphere and crust, and variations in the structure of D″ may likewise be a combined result of thermal, chemical, and mineral phase variations. Interpretations of the seismic images, requiring knowledge of the mineral physics of expected mineralogical assemblages at these depths, suggest that the CMB plays a very important role in controlling the dynamics of the core and lower mantle, and therefore of the evolution of the interior of Earth.

Editor's Note:

Each year the David and Lucile Packard Foundation awards 20 Fellowships for Science and Engineering in a national competition to promising young scientists and engineers working in fields that receive relatively less popular attention than high-energy physics, space, and medicine. Each Packard Fellowship provides $100,000 per year for five years to the Fellow's institution, $90,000 of which is for use of the Fellow to support his/her research work. These young researchers are truly among the "best and brightest" in the United States. The science article in this issue is one of several in which Packard Fellows in earth science report on research in their field.

INTRODUCTION

While most geologists, including specialists in the field of seismology, study rocks at Earth's surface, more attention also is being paid to the planet's other major boundary, that between the core and mantle. With a density jump of 4.3 kg/m^3 between the silicate lower mantle and the liquid iron outer core, as well as a temperature increase of possibly 1500 °C between the lower mantle adiabat and outer core, 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 though 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 heterogeneous 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 "continents" at the CMB, speculate about what these maps may mean, 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 represent 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 a function of density, rigidity, and incompressibility, which are complicated functions of temperature, composition, and mineralogical phase. Hope for better understanding exists because different disciplines complement each other in providing constraints about the state of the deep Earth. Long wavelength signals in the geoid are affected by mass variations in the lowermost mantle. Variations in the wobble of Earth's axis of rotation and in the length of days are also a result of mass variations and provide constraints on the topography of the CMB. 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 realistic 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 equations of state, are delineating the kinds 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, laterally swept mantle dregs, 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.

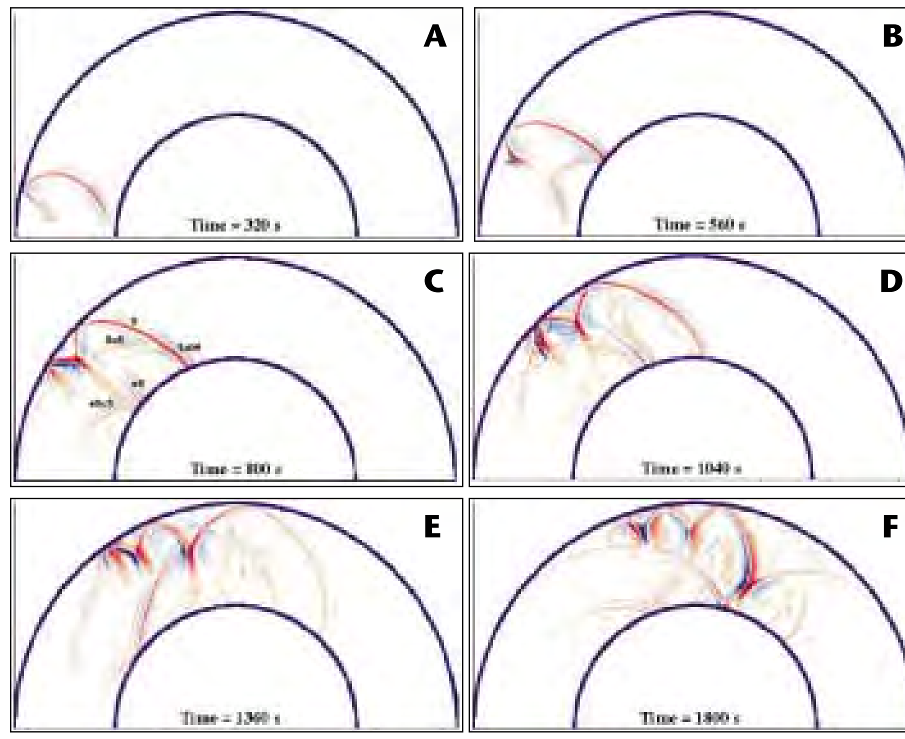

Figure 1. Images from a motion picture showing the propagation of seismic shear energy through the mantle (Wysession and Shore, 1994). The images correctly show the locations of the seismic shear wave fronts at (A) 320, (B) 540, (C) 800, (D) 1040, (E) 1360, and (F) 1800 s after the occurrence of a 600-km-deep earthquake at the lower left of the images. 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 (28,585) 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 I. Lehman 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 D″ 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 mantle adiabat and will have appropriately slower velocities.

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

Core-Mantle *continued on p. 239*

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Core-mantle *continued from p. 237*

importance about D″. Global tomographic inversions 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 (<0.01 Hz) normal modes.

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 1, 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 torsional 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 560 s (Fig. 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 (Fig. 1C) a second wave is reflecting off the CMB, the surface-reflected sScS, 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 (Sdiff, or equivalently, Pdiff), 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 in the distance range of 100°–150° Sdiff and Pdiff arrivals at the surface provide a lot of information about the very base of the mantle. This article provides two examples of CMB studies, one using core-diffracted Sdiff and Pdiff waves, and the other using core-reflected ScS and PcP waves.

CORE-DIFFRACTED WAVES

Sdiff and Pdiff 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 (Pdiff is the first arrival of any kind beyond 100°, and Sdiff 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 allow the picking of clear onset times. The high-frequency decay does not resemble seismic anelastic 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 Sdiff and Pdiff. 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 (Fig. 2) shows six WWSSN Sdiff arrivals (top), modeled by their reflectivity synthetic counterparts (bottom). The slope through the arrivals is the ray parameter and is a direct result of the average velocity structure in D″. Maps of the results for 12 Sdiff and 20 Pdiff profiles (Figs. 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 the tomographic mantle shear velocity models of Su et al. (1994). A possible explanation for this pattern is discussed below.

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 P 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 lowermost mantle. This variation in the Poisson ratio of the lowermost mantle 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 the way surface waves can be used to determine upper-mantle structure. Because all seismic 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) identify the top of D″ where a sharp velocity increase creates an additional seismic precursor (SdS) to ScS. Core-diffracted waves provide additional information about the rest of D″, longer wavelengths sampling more of D″ and 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 the model proposed by Young and Lay (1990), using SdS waves to model a region of D″ nearby to the east. As more high-quality data are obtained from portable arrays of broad-band

Core-Mantle *continued on p. 240*

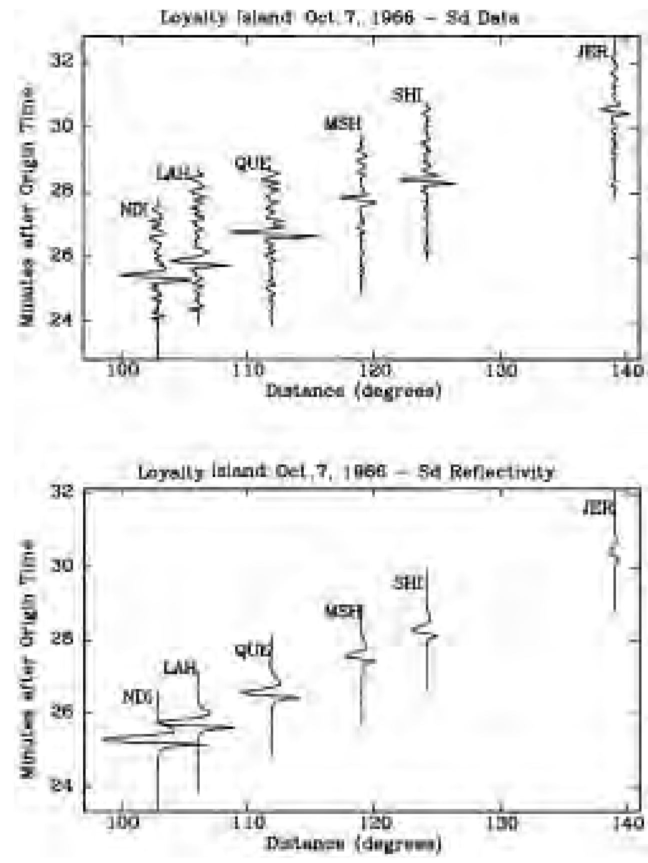

Figure 2. An example of six data (top) and synthetic (bottom) core-diffracted Sdiff seismic arrivals from an earthquake (October 7, 1966) in the Loyalty Islands (from Wysession et al., 1992). All arrivals are along a narrow azimuthal window, so a single patch of the CMB is investigated. Because all of the diffracted waves bottom at the base of the mantle, they share the same ray parameter, represented by the slope through the arrivals. A change in the slope relative to that of the synthetic counterparts implies anomalous velocities in D".

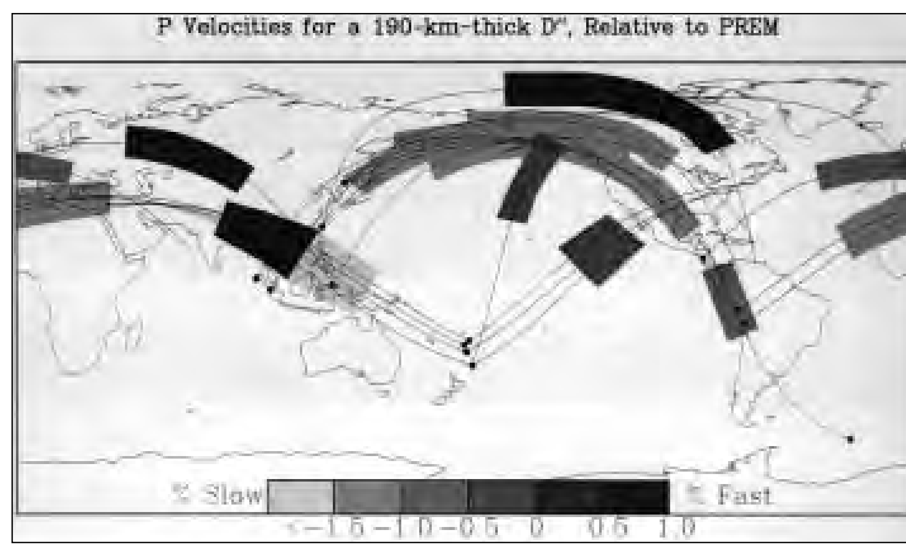

Figure 3. A map from Wysession et al. (1992) showing the very long wavelength average D"P velocity variations, determined from 20 profiles of core-diffracted Pdiff arrivals like those shown in Figure 2. The solid circles are earthquakes used, lines represent average paths for each profile, and the shaded areas are the regions of D" sampled by the waves. Note the unusual transition from slow to fast velocities beneath the western Pacific.

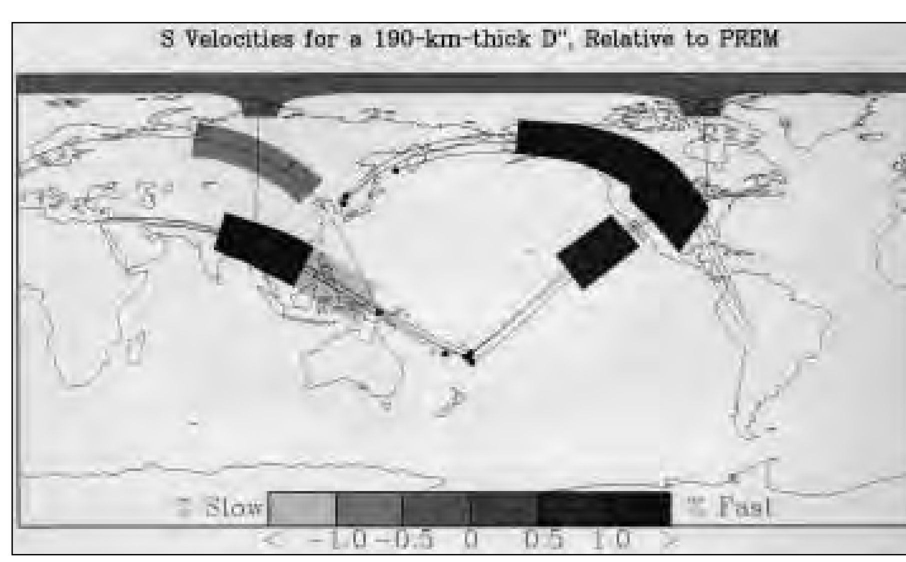

Figure 4. Similar to Figure 3, but for long-wavelength S velocity variations based on 12 independent profiles of Sdiff arrivals (from Wysession et al., 1992).

Core-mantle *continued from p. 239*

seismometers (recording a "broad band" of frequencies) this technique of using Sdiff or Pdiff 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, Wysession et al. (1994, 1995b) further investigated this region using ScS-S and sScS-sS differential traveltimes (the S, ScS, sS, and sScS waves are the first four wave fronts shown propagating away from the earthquake in Fig. 1C). Using 747 differential traveltimes 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 canceled out (Wysession et al., 1995b). After the different phase paths were corrected by ray tracing through a 3-D tomographic mantle S-velocity model (SH8/WM13 of Woodward et al. (1993) to help remove middle and upper-mantle heterogeneity effects, any remaining traveltime residuals were converted into velocity variations along their computed paths through the lowermost 300 km. These velocity variations were superimposed by moving a weighted Gaussian cap with a 300 km radius across them to average the geographical contributions and help simulate the CMB Fresnel zones, or sampling regions, of the ScS and sScS footprints. The resolution of the result (Fig. 6) is on the order of about 300 km, or 5°. (Note that the amplitudes of the original figure in Wysession et al. [1994] were erroneously amplified by a factor of two; this was corrected in Wysession 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 install 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 region. 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 (LVZ) is surrounded on three sides by regions showing fast velocities. The average of these regions is about 2% faster than for PREM but reaches values greater than 3%. The fast velocities to the south and west of the D″ LVZ seem to form one continuous feature that extends from beneath China to beneath eastern Australia. There is a correlation between this fast D″ rock and the location of the paleotrench of the Tethys plate. The fast-velocity region northeast of the LVZ is poorly constrained in its lateral extent and is beneath the northern part of the Pacific Ocean. We have no coverage of what happens to the LVZ east of the study region, but if current tomographic images like those of Su et al. (1994) are an indication, it probably extends a long way eastward as part of a broad low-velocity region beneath the central Pacific.

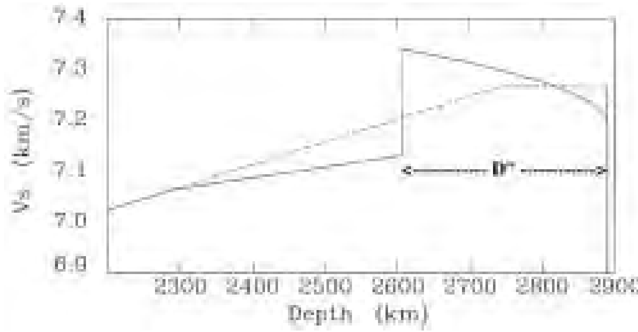

Figure 5. Profile from Valenzuela et al. (1994) showing the S velocity model for a patch of the lowermost mantle beneath eastern Siberia that best modeled the amplitude decay of core-grazing S waves as a function of frequency and distance. The study used five northern California earthquakes recorded in Tibet. The dashed line is the reference model PREM of Dziewonski and Anderson (1981), and the solid line is our best preliminary fit to the amplitude data—a model based on D" structures proposed by studies like that of Young and Lay (1990).

In an attempt to get at what the P velocities might be doing in the same region, Zhu and Wysession (1995) presented a map of D″ P velocities by stacking the differential times of PcP and P for those seismic 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 from combining the very large number of data available—in this case, 78,793. Figure 7 shows the resulting map for D″ P velocities in the same region as that previously shown for S, and determined by the same procedure. The PcP-P residuals were determined relative to the IASP91 reference Earth model of Kennett and Engdahl (1991), and the mean of the entire data set was 0.35% slower 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 large central low-velocity region, but it extends farther west than for the S velocities in Figure 6. An examination and careful analysis of available PcP waveforms for this region will be required before an accurate comparison of P and S velocities in this region can be made.

INTERPRETATION AND SPECULATION

It is clear that there are some interesting and unusual geologic processes at work at the CMB, but it is not clear what they are. Many recent papers have discussed the potential causes and implications of geophysical observations such as the seismic images just shown. Both general and detailed discussions can be found in a variety of papers, which are far too numerous to mention in full (e.g., see Loper and Lay [1995] and Wysession [1995a]). As yet there are more interpretations than hard facts, and each strong argument seems rebutted by an equally strong counter-argument. There are, however, three major categories of possible contributions to the structures seen in D″: thermal variations, chemical variations, and mineralogical phase changes. Each of these in turn presents a variety of geodynamic interpretations. In many places I draw analogies between D″ and the surface's crust and lithosphere. Although there clearly are dangers and limitations with doing so, because of the extreme differences in temperature and pressure, these are Earth's two major boundary layers, and it is likely that we can gain some understanding of CMB geology from processes observed at the surface.

Thermal Variations

Some estimates of the temperature difference between the lower-mantle and outer-core adiabats are as large 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.

As with the thermal lithosphere, we would expect horizontal mass movements to cause lateral variations in the temperature within such a thermal boundary layer. The temperature 50 km below a mid-oceanic 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 up. Three-dimensional tomographic mantle models interpreted as buoyancy forces resulting from temperature variations do a good job of modeling the observed long-wavelength geoid (Forte et al., 1994).

The thermal model of D″ can be taken a step further to incorporate a direct connection with plate tectonics, showing that the CMB is not immune to arguments about the degree of intermixing 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; however, the correlations between D″ seismic variations and paleosubduction would be equally satisfied by thermal coupling between an independent upper 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 "dregs," dense iron alloys, could have formed at the base of the mantle early on in Earth's evolution, or 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 eclogitic crust of sub-

Core-Mantle *continued on p. 256*

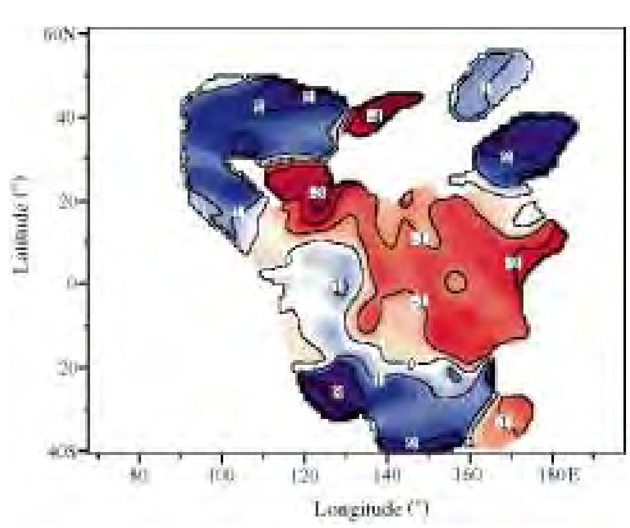

Figure 6. A map from Wysession et al. (1995b) of lateral S velocity variations in D" beneath the western Pacific, computed from ScS-S and sScS-sS differential traveltimes. All ray paths are corrected for mantle path heterogeneities outside of the bottom 300 km of the mantle, and the velocity magnitudes 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 continent-sized long wavelengths.

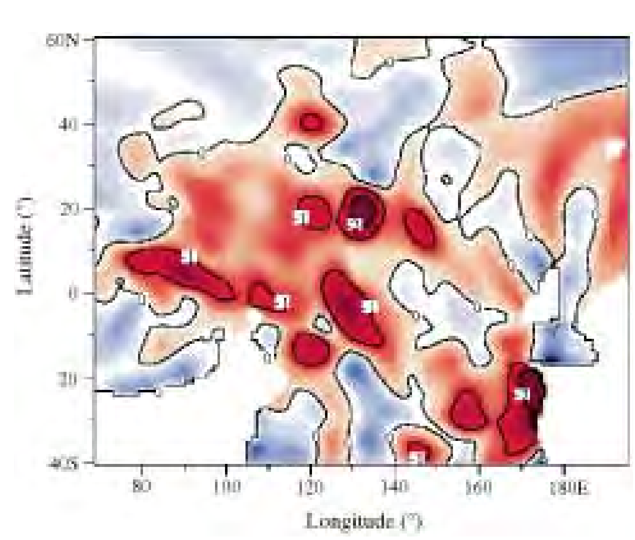

Figure 7. A map from Zhu and Wysession (1995) of the average P velocity variations in a 300-km-thick D" layer for the same region shown in Figure 6 for S velocities, incorporating more than 10,000 PcP-P differential traveltimes reported to the International Seismological Centre. As with Figure 6, the dominant feature is a broad low-velocity region in the center surrounded by slightly faster velocities, although the low velocities here extend farther to the west than in Figure 6.

Core-mantle *continued from p. 240*

ducted slabs could delaminate and form a mineralogical phase denser than the ambient lower mantle. Lighter elements could be settling up and out of the outer core as a more iron-rich inner core freezes at an inner core boundary eutectic point, increasing the outer core percentages of lighter elements. All of these mechanisms have been proposed, but it is not clear which of them, if any, actually occurs.

Chemical variations will cause P and S seismic-velocity variations that are distinct from thermal effects, as shown for the major lower-mantle constituents, perovskite and magnesiowüstite. Figure 8, from Wysession et al. (1992), uses a third-order Birch-Murnaghan equation of state to show the amount of change in temperature, silicate/oxide ratio, and iron/magnesium ratio required to change the P and S velocities in D″ by several percent. This means that seismic observations of P and S velocities varying out of tandem within D″, as was observed for diffracted wave profiles beneath the northern Pacific, may tell us when thermal or chemical effects are dominant. If current tomographic models of D″, such as the S model of Su et al. (1994) and the P model of Pulliam et al. (1993), are combined, they provide a map of the Poisson ratio in D″ that varies from 0.295 to 0.310 (5%). Neither our seismic data nor our thermoelastic constants are quite good enough yet to take a map of D″ Poisson ratios resulting from these models and convert them into temperature and compositional variations, but it is the direction in which we are moving.

Variations in CMB topography are also a kind of lateral chemical variation. The asphericity of the CMB has been detected seismically at both very high and very low wavelengths, as well as through length-of-day variations. Actual determination of CMB topography is very difficult to measure because of trade-offs with velocity heterogeneities, but it is vitally coupled to dynamic CMB processes. The CMB will be depressed beneath regions of lower mantle downwelling because of the isostatic 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 compositionally denser mantle dregs. 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 develops, denser mantle aggregates will be swept laterally to the site of the plume, causing a depression of the CMB. In other words, the convective cycle may cause a temporal transition in CMB topography similar to that modeled by Gurnis (1992) for long-wavelength surface topography during the history of lithospheric subduction.

The electrical conductivity for lowermost mantle phases may vary by 11 orders of magnitude (Jeanloz, 1990). This is important because determinations of core flow using temporal variations in Earth's magnetic field (e.g., Bloxham, 1993) assume that the mantle behaves as a perfect insulator. Because the mantle flows a million times slower than the outer core, D″ structure can also have lasting effects on core convection in more direct ways. CMB topography can serve to channel core flow, like air over mountains, or lateral D″ thermal variations can regionally vary the heat flux out of the core, constraining core convection patterns.

Mineralogical Phase Changes.

The third form of explanation for observed seismic variations is that of pressure-driven changes in the mineralogical assemblages that are present. If there are either thermal or chemical variations across D″, this could cause vertical topography on a mineralogical

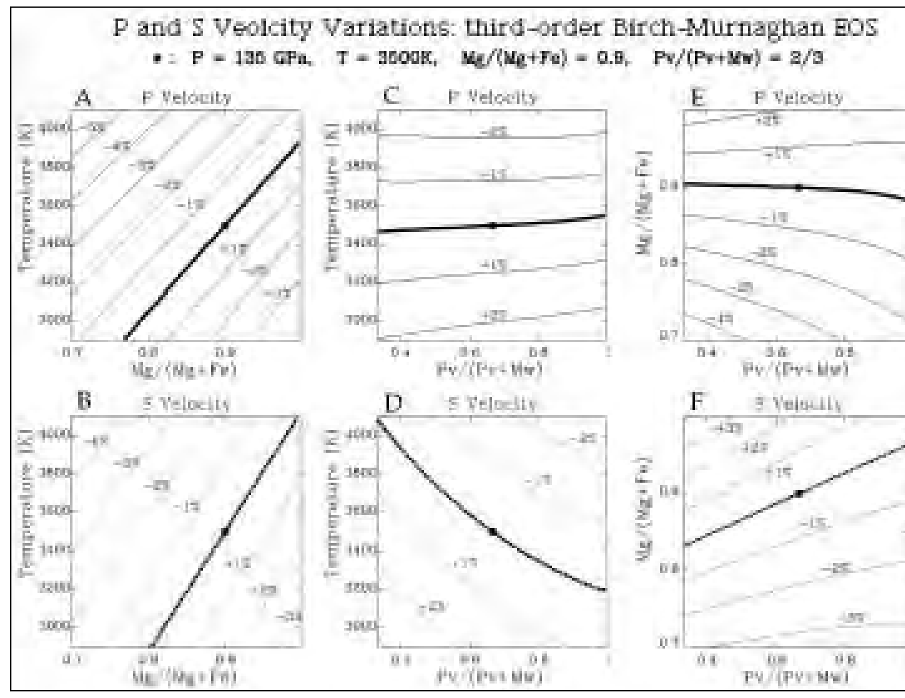

Figure 8. The percentage changes in P and S velocities for rock at the base of the mantle as a result of changes in three parameters: temperature, silicate/oxide ratio, and iron/magnesium ratio. Calculations are done using a third-order Birch-Murnaghan equation of state (EOS) for the iron and magnesium end-members of perovskite and magnesiowüstite, using the best available thermoelastic parameters (from Wysession et al., 1992). The starting material is a pyrolite-type composition at 135 GPa and 3500 K. Note that the P and S velocities do not change in the same way for the different thermal and chemical changes, implying that we may eventually be able to use not only the P and S velocities in D″ but also their relative variations to identify their dominant geophysical causes.

phase boundary, presuming one were present. A possible candidate would be the breakdown of the silicate $(Mg,Fe)SiO_3$ (perovskite) into the oxides $(Mg,Fe)O$ (magnesiowüstite) and SiO_2 (stishovite) (Stixrude and Bukowinski, 1990). Although the phase relations at CMB temperatures and pressures are still poorly constrained, such a phase change would provide the best explanation for the sharp velocity increase seen at the top of D″ in many radial velocity models like Figure 5. It is also possible that high-pressure phases of stripped oceanic eclogitic crust, which would be denser, yet seismically faster, could accumulate at the base of the mantle and provide a chemical basis for the D″ reflector (Christensen and Hofmann, 1994). Both of these mechanisms are shown through thermochemical modeling to be able to provide the necessary 3% seismic velocity jump (Wysession, 1995b). It is unlikely that an inverted temperature change, such as for descended cold rock that has spread out and ponded at the CMB, would have the very steep gradient that is observed seismically.

If such a phase transition does occur, we can also speculate as to the effects of composition and tempera-

Core-Mantle *continued on p. 257*

Book Reviews *continued from p. 249*

overall, Cattermole does a commendable job of summarizing what has been learned. Because it was written so early after publication of the initial results, most of the presented materials are derived from the initial (Saunders and Pettengill, 1991; Saunders, 1992) publications devoted to summary results of the Magellan mission. As such, it is mostly a digested version of those reference works with related bits and pieces and balancing arguments thrown in. However, that is not necessarily bad; after reading it, I found that my grasp of some of the issues and subtopics of secondary interest to me, which I had skimmed in the original references, was somewhat clarified and expanded. However, this book is not "watered-down" science for popular consumption; it rates as a reference book. One wishes that the results of many large-scale scientific studies could have a similar synthesis done in such an easily readable format by one synthesizing author familiar with the science, as Cattermole appears to be.

In addition to the summary of the recent geological story, there are also nicely done summaries of the related atmospheric characteristics, which anyone who uses the term "greenhouse effect" in discussions of environmental science may wish to review. The brief summary of the geometry and overall methods of the synthetic aperture radar imaging technique may be useful to the uninitiated, but one could wish for more details and a discussion of influence of surface properties (roughness and reflectivity) on images and altimetry determinations, especially because that influence serves as a hook for learning about an important remote-sensing technique used in terrestrial environmental studies. Although there is a substantial list of references, relevant references such as Ford et al. (1993) and more recent summaries and data releases are not listed. A discussion of some of the subsidiary image products such as stereo radar image data and digital terrain models would be desirable. At the time the book was written, only preliminary results were in on Magellan-derived gravity data, but there is still room for more summary discussion about the global gravity field, its interpretations, and particularly some of its implications for regional geologic characteristics.

Several matters of production detract from the book. Because much in this book derives from what was largely an imaging mission, the dimensions of the pages are relatively small for image reproduction; the many images and maps that fill the text could have been better presented in a larger format book. Perhaps because of the format, the overall reproduction of the global altimetry and image maps is poor, although the inclusion of several color plates is a welcomed, if fuzzy, addition. Many of the black and white images are too dark and lack contrast. Even though a significant fraction of the Magellan image data are radar dark, it is possible to reproduce the image data in better contrast. Several distracting errors also occur: Figures 3.7, 3.8, and 9.25 are printed upside down (but in fairness, this seems difficult even for some science journals to get right), a number of typographical errors occur in some of the figure captions, at least one map is misattributed, and some symbols are mentioned in captions that do not appear in the figures. The text is generally well written, understandable at the advanced undergraduate and graduate level. However, because of the particular time period during which this text was written, the occasional switch from present to future tense in referring to mission events ensured that the text would sound out of date before it actually made it into print. These are minor distractions, however, not condemnations.

If the geology of Venus is a story, then the "moral" of the story is still being determined. The moral would seem to have something to do with the large-scale effects of subtle differences in starting conditions and environment on subsequent development of two otherwise similar planetary bodies (Earth and Venus), but it is also about some large-scale similarities over which environment has little influence. Perhaps the moral may have been best summarized in the title of a presentation by the mission project scientist Steve Saunders: "Venus and Earth: Twins Separated at Birth." Behavioral scientists found out long ago the value of that experiment. Now we geologists have our chance at a similar experiment.

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Geomorphology of Desert Environments. *Edited by Athol D. Abrahams and Anthony J. Parsons. Chapman & Hall, London, 1994, 674 p., $146.95.*

Desert environments, encompassing about 30% of Earth's land surface, are increasingly being subjected to urban and industrial development. Many of these developed areas have

Book Reviews *continued on p. 259*

Core-mantle *continued from p. 256*

ture. The seismic observations of Kendall and Shearer (1994) and Revenaugh and Jordan (1991) suggest that a correlation may exist between regions of shallow D″, 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 transformation 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 were enriched in magnesium relative to iron, so the phase transition could also be exothermic if rock were significantly depleted in iron relative to its surroundings. Advances in mineral physics will eventually solve this question of the possibility of a D″ phase transition and the form it would take.

FUTURE DIRECTIONS

The only real fact that can be gleaned from the previous section is that the questions still outnumber 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 efforts 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 P velocities in the lowermost mantle (Wysession, 1994). By cross-correlating the Pdiff phases (data and synthetic) and PKP phases separately, we are able to determine the delay of Pdiff relative to PKP and isolate any anomalous behavior of Pdiff during its long path around the CMB. The CMB Fresnel zones of the long-distance Pdiff waves are very large, but the superposition of these cover all parts of the CMB, and an over-determined inversion can be done for the long-wavelength D″ P velocities. Because of the geographical limitations of available earthquakes and seismometers, utilizing all phases that interact with the CMB will help fill in the many gaps that exist in our geographical coverage of the lowermost mantle.

We also need to be constantly developing new techniques to increase our ability to interpret existing seismic data. As an example, Koper and Wysession (1995) have developed a genetic algorithm for simultaneously determining radial velocity structure at both the CMB and inner core boundary using the AB, BC, and DF branches of PKP arrivals. The PKP branches are our best seismic tools for examining the P velocity structure of the core, but difficulties arise because the DF and BCdiff branches interact with both the inner and outer core boundaries. The genetic algorithm is a powerful technique for identifying the different kinds of possible structures at both boundaries that would simultaneously satisfy all of the observed PKP arrivals. These investigations use high-quality seismic array data to examine regional parts of the core, as well as the set of ISC PKP times, to determine a global core model.

An important contribution from seismologists is obtaining new data sets that can be made available to future researchers who will answer the questions that currently puzzle us. In January and February 1995 my colleagues and I, together with Timothy Clarke (University of Illinois) and Karen Fischer (Brown University), installed 18 broad-band seismometers in a straight line, connecting stations CCM (Missouri) and HRV (Massachusetts). This Missouri-to-Massachusetts deployment (MOMA), which will run for one year, is designed to investigate several aspects of deep Earth geology, including core and CMB velocities, the structure of subducting slabs, and the velocity structure of the upper mantle and crust beneath North America. The 20 stations span 16° in the distance range of about 100°–130° from the world's most seismically active regions in the western Pacific, and are therefore ideally set to examine the CMB by means of core-diffracted waves. In fact, as is seen in Figure 9, the northeastern United States is one of two geographical regions that are within the 95°–135° distance range from more large earthquakes than any another parts of the world (the second is southern Africa). A problem of the seismic data set today is that while seismic stations are spaced to cover the globe evenly (although they are limited to mostly continents and islands), this results in a large distance between stations and creates an aliasing problem in our ability to image Earth's interior structure. Regionally dense temporary seismic deployments like MOMA, made possible by using instruments borrowed from IRIS (Incorporated Research Institutions for Seismology), help provide high-resolution windows into the planet. One of the MOMA sites, installed in March 1995 near Lake Newton in eastern Illinois, is shown in Figure 10.

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. SEDI (Studies of Earth's Deep Interior) organizations exist nationally within the American Geophysical Union and the National Science Foundation, and internationally as well. These provide 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.

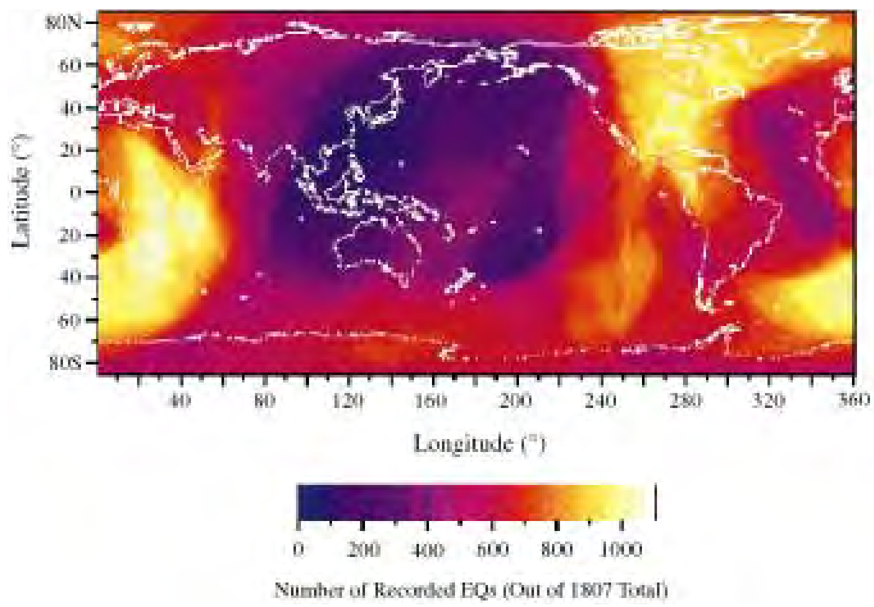

Figure 9. A map showing the number of earthquakes (EQs; mb >5.7) during a ten-year period (1983–1992), which occurred within a distance of 95°–135° from any given location (out of a total of 1807). This distance range is required for examining the CMB using core-diffracted Pdiff and Sdiff waves. The Missouri-to-Massachusetts (MOMA) array of portable broad-band seismometers, 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 seismogenic western Pacific regions. Such temporary arrays, funded through the PASSCAL program of IRIS (Incorporated Research Institutions for Seismology), greatly help in providing new seismic data that fill in the aliasing gaps between permanent seismometers.

Figure 10. Ghassan Al-Eqabi (left) and Patrick Shore (right) completing installation of one of the MOMA (Missouri-to-Massachusetts) portable-array broad-band seismometers near Lake Newton, Illinois.

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