THE CORE-MANTLE BOUNDARY

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INTRODUCTION

The largest compositional discontinuity within the Earth is the core-mantle boundary (CMB), at a depth of 2889 km. This boundary and the adjacent transition zones in the lowermost mantle and outermost core play a critical role in the Earth’s thermal and chemical evolution. Estimates of the heat flux out of the core indicate that the 200-km-thick $D''$ region at the base of the mantle is a major thermal boundary layer. The $D''$ region, like its counterpart boundary layer in the lithosphere, is laterally heterogeneous and possibly chemically stratified. This complexity obscures the exact dynamic behavior of the $D''$ region. However, it is clear that the large density contrast (4.3 g cm$^{-3}$) across the CMB provides favorable conditions for density stratification on both sides of the boundary. The $D''$ region may be the refuse pile for “heavy” mantle heterogeneities, while the outermost core may have a concentration of “light” components. Boundary-layer instabilities in $D''$ have been invoked as sources of both localized mantle plumes and major upwelling currents in whole-mantle convection systems. Dynamically supported topography on the CMB and lateral temperature gradients in $D''$ may produce coupling between the core and mantle, influencing the excitation of the geodynamo. The outermost core is generally believed to be radially and laterally homogeneous, although chemical stratification has not been ruled out. Seismological, geodynamical, and geomagnetic investigations are unveiling the three-dimensional complexity of the deep Earth in the vicinity of the CMB.

SEISMOLOGICAL CONSTRAINTS

Several geophysical disciplines provide information about the CMB; however, seismological investigations continue to provide the best resolution...
of the region. A variety of seismic phases have been used to investigate the density and elastic velocity structure, anelastic properties, and the lateral heterogeneity of the lowermost mantle and outermost core. However, the history of seismological models, particularly for the $D''$ region, has been a checkered one, with many incompatible results being published. Present-day quantitative modeling procedures are yielding somewhat more consistent models, although there is still no consensus on the detailed characteristics of the $D''$ region. A new generation of three-dimensional models that explicitly include lateral heterogeneity may reconcile the outstanding inconsistencies. In this review, contemporary seismological models for the CMB are considered first and then integrated with constraints from other disciplines. Particular emphasis is placed on the evidence for a thermal or chemical boundary layer at the base of the mantle.

**Radial Models Based on Body-Wave Travel Times and Free Oscillations**

While recognizing the importance of lateral heterogeneity near the CMB, we still find it useful to consider radially symmetric “average” models for the region. The average properties, such as velocity gradients in $D''$, are important boundary conditions for thermal modeling of the CMB. The classic procedure for determining radially symmetric velocity models is inversion of large data sets of body-wave travel times. As early as 1939, seismic velocity models (Gutenberg & Richter 1939, Jeffreys 1939) indicated that the $P$- and $S$-wave velocity gradients in the lowermost mantle are anomalously diminished relative to the overlying mantle. The primary evidence for this is changes in the slopes of the travel-time curves at a distance of about 95°. In an important review, Cleary (1974) compared the early velocity models for $D''$ based on travel-time inversions and showed that most of the models suggest decreases in the velocity gradients about 200 km above the CMB. Some models, notably that of Gutenberg, actually predict decreasing velocities with depth. The most recent Earth model, PREM (Dziewonski & Anderson 1981), based on global travel-time and free oscillation measurements, shows similar smooth decreases in velocity gradients above the core (Figure 1). These decreases in velocity gradients above the CMB led Bullen (1949) to define the $D''$ region, which he concluded was an inhomogeneous zone. Such inhomogeneity could result from either compositional changes or nonadiabatic thermal structure within $D''$. Nonadiabatic conditions could arise as a result of the presence of a thermal boundary layer, which requires a large heat flux from the core into the mantle. Alternatively, compositional stratification or a phase change could account for the seismic velocity behavior, although a smooth model such as PREM would require gradational changes.
Figure 1  Recent S-wave (left) and P-wave (right) velocity models proposed for the $D''$ region; PREM (Dziewonski & Anderson 1981); PEMC-L01 (Doornbos & Mondt 1979b); SYLI (Young & Lay 1986); WL (Wright & Lyons 1981); POLAR1 (Ruff & Helmberger 1982). The dashed lines are the extremal velocity bounds from Lee & Johnson (1984a,b); bounding velocity values are consistent with travel-time inversions for radially symmetric models.

Models for the $D''$ region based on travel-time inversions alone have limited resolution for several reasons. Foremost among these is that the travel-time slope, $dT/d\Delta$, is nearly constant at distances beyond about 92°. A constant slope is expected for either a diffracted wave (traveling parallel to the CMB) or for a velocity structure with a critical velocity gradient, $dv/dr = \nu/r$. In either case travel-time inversions are unreliable. Frequency-dependent effects due to diffraction are not accounted for in ray-theory inversions, leading to potential biases in the resulting structure. A critical velocity gradient over a 100-km thickness of $D''$ requires a $P$ velocity decrease of 0.4 km s$^{-1}$ or an $S$ velocity decrease of 0.2 km s$^{-1}$. In regions of the Earth where seismic velocities decrease with depth, travel-time inversions tend to be unstable. Direct measurement of the travel-time slopes using array techniques should improve the accuracy of the travel-time inversions; however, typical array measurements (e.g. Chinnery 1969, Johnson 1969, Corbishley 1970, Wright & Cleary 1972) show substantial scatter at distances beyond 85°, which limits their usefulness in global inversions [see Cleary (1974) for a summary of such measurements]. An indication of the limitations of travel-time inversions for resolving $D''$
velocity structure is given by the extremal bounds in Figure 1, taken from a study of 108,000 P-wave travel times and 75,000 S-wave travel times by Lee & Johnson (1984a,b). These bounds indicate the range of radially symmetric Earth models consistent with the travel-time observations. The average bound widths in the central mantle are $0.13-0.14 \text{ km s}^{-1}$. The splaying of the bounds in $D''$ indicates both increased variation in the data and decreased resolution of the inversions. These bounds do not necessarily bracket lateral velocity variations. Part of the increase in the S-wave velocity bound widths can be attributed to the fact that at distances greater than $80^\circ$ the core-traversing phase SKS arrives ahead of direct S and contaminates the travel-time measurements. As is discussed later, relaxing the requirement of a radially symmetric model allows further information about $D''$ to be obtained from large travel-time data sets.

The seismic properties of the outermost core are also difficult to extract from travel-time measurements. The principal reason is that the core behaves as a liquid for seismic wave frequencies and thus does not transmit shear waves, while the P-wave velocities are much lower than in the lower mantle, producing a major low-velocity zone. The resulting downward deflection of P waves crossing the CMB prevents PKP-type phases from bottoming in the outermost core; therefore, these phases only provide integral constraints on the structure. The P velocities in the outer core are fortunately higher than the fastest S-wave velocities in the mantle, so that phases such as SKS and SKKS do turn in the outermost core. However, these phases are often difficult to time accurately, and most travel-time studies extrapolate the outermost core velocities from the directly determined velocities at greater depths. None of the travel-time studies have resulted in particularly complex outermost core P velocity structures. The P velocity just below the CMB has been estimated by SKS travel-time studies to lie between 7.9 (Hales & Roberts 1971) and 8.1 km s$^{-1}$ (Jeffreys 1939), a range that spans the extremal bounds on outer core velocities found by Johnson & Lee (1985). The low-velocity estimate of Hales & Roberts (1971) results in a steep outermost core velocity gradient, suggestive of a transition zone mirroring the $D''$ region.

In addition to travel-time data, observations of free oscillations, particularly higher modes, have been used to develop radial models for the CMB region. While intrinsically insensitive to fine velocity characteristics, free oscillations do provide uniform global averaging that is difficult to attain even with large International Seismological Centre travel-time data sets. Utilization of the normal modes to determine deep Earth structure is complicated by strong trade-offs between density and shear velocity structure and depth of the CMB (Masters 1979). However, joint analyses of travel-time and free oscillation measurements have helped to constrain
$D''$ properties. Differential travel times $P_{c}P - P$ and $S_{c}S - S$ (Engdahl & Johnson 1974, Jordan & Anderson 1974) are particularly useful in conjunction with normal-mode measurements, since they place relatively tight constraints on the depth to the CMB. The currently preferred depth is $2889 \pm 4$ km (Dziewonski & Anderson 1981), corresponding to an outer core radius of $3482 \pm 4$ km, although this value is dependent upon the accuracy of the mantle velocity models [See Dziewonski & Haddon (1974) for a review of this problem.] The velocity models for $D''$ obtained by joint inversion of travel-time and free oscillation measurements tend to result in positive velocity gradients that are usually slightly diminished relative to the shallower mantle (Dziewonski & Gilbert 1972, Jordan & Anderson 1974, Dziewonski et al 1975, Gilbert & Dziewonski 1975, Anderson & Hart 1976, Dziewonski & Anderson 1981). The limited resolution of the combined data sets is indicated by the fact that models with strong negative velocity gradients in a thin zone above the CMB cannot be ruled out (Jordan & Anderson 1974). It is not possible to unambiguously distinguish between thermal and compositional explanations for $D''$ inhomogeneity using normal-mode data (Masters 1979). The velocity models of the outermost core derived by the same joint inversions tend to agree well with the travel-time models, having $P$ velocities just below the CMB of $7.98 - 8.06$ km s$^{-1}$. The PREM model has a steep velocity gradient of $0.17$ km s$^{-1}$ per 100 km at the top of the outer core, compared with a gradient of about $0.12$ km s$^{-1}$ per 100 km at a depth 600 km below the CMB. This decrease in gradient, albeit only marginally resolved, is evidence for inhomogeneity of the outermost core. As is discussed later, a region of thermally stratified, stable outermost core has been suggested on other grounds, and this possibility cannot be ruled out by the travel-time and normal-mode data (Masters 1979). However, Choy (1977) performed detailed waveform modeling of $S_{n}KS$ phases, which is the most accurate procedure for modeling the outermost core, and he found that a relatively low gradient, like that in Jeffreys’ model, fits the data best. More studies of this type are needed to resolve this issue. In order to place tighter constraints on the velocity structure above the CMB, detailed analyses of a variety of body-wave amplitudes and waveforms have been performed.

**Radial Models of $D''$ Based on Body-Wave Analyses**

Many of the early attempts to constrain the properties of the CMB used core-reflected phases $P_{c}P$ and $S_{c}S$. In theory, by determining the reflection coefficients of $P_{c}P$ and $S_{c}S$, bounds on the density and $P$-wave velocity contrast, as well as the outer core viscosity, can be determined (Kanamori 1967a,b). However, in practice, substantial unexplained amplitude vari-
ations of the short-period phases have precluded definitive results. In
general, the efforts to interpret the unstable short-period $PcP$ observations
led to complex transition zones with strong decreases in velocities above
the core (Buchbinder 1968a,b, Ibrahim 1971, 1973, Berzon et al 1972,
Buchbinder & Poupinet 1973). Spectral studies of $ScS$ have suggested finite
outer core rigidity (Sato & Espinosa 1967, Suzuki & Sato 1970). However,
the scatter of the data and the intrinsic instability of the spectral techniques
applied in the early studies have weakened confidence in these analyses
(Frasier & Chowdhury 1974). Quantitative waveform modeling of more
stable long-period phases also gives mixed results, with $PcP$ data sug-
gestig a mild negative gradient in $D''$ (Müller et al 1977) and $ScS$ data
requiring at least some regions either to have a strong positive gradient or
to be anisotropic (Mitchell & Helberger 1973, Lay & Helberger 1983b).
The core reflections also intrinsically sample restricted regions, so it is
difficult to develop radially averaged models.

Perhaps the most straightforward analysis of $D''$ velocity structure
involves body waves that are diffracted along the CMB and observed at
distances in the core shadow ($>95^\text{o}$). These phases usually have long
pathlengths in $D''$, providing laterally averaged (but not necessarily glob-
ally representative) structures. Many of the studies of these phases have
involved simply measuring the apparent velocity of the diffracted signals.
In conjunction with an estimate of the core radius, a geometric optics
interpretation of the waves yields a direct estimate of the velocity at the
base of the mantle. This procedure typically results in quite low velocities
at the CMB for both $P$ waves (Sacks 1967, Bolt 1970, 1972) and $S$ waves
(Cleary 1969, Bolt et al 1970), requiring negative velocity gradients in $D''$.
However, this simple analysis may lead to incorrect models for two
reasons. First, diffraction is a frequency-dependent phenomenon that can-
ot be rigorously described by geometrical ray theory. Second, the presence
of a low-velocity zone at the base of the mantle could simply refract energy
to large distances, where it may mistakenly be interpreted as having been
diffracted. In either case, the signals are sensitive to velocity structure
throughout the $D''$ region, rather than only at the CMB.

Frequency-dependent effects of the diffracted signals are most readily
accounted for by synthesizing waveforms for a given Earth model. This
allows the frequency-dependent amplitude effects of diffraction as well
as the travel-time measurements to be exploited in determining the $D''$
structure. Since the pioneering study by Alexander & Phinney (1966), it
has been standard practice to analyze the amplitude decay coefficients.
Diffraction theory [see Chapman & Phinney (1972) for a review] predicts
that the amplitude decay into the shadow zone is directly related to the
velocity gradients in $D''$. Positive velocity gradients produce faster rates
of amplitude decay, and high-frequency signals decay much more rapidly than long periods. While long-period signals are less sensitive to $D''$ structure, they are less influenced by other propagation effects, so most quantitative studies have analyzed the amplitude decay of long-period (8–64 s) $P$ waves (Alexander & Phinney 1966, Sacks 1966, Phinney & Cathles 1969, Chapman & Phinney 1972). More recent studies of larger data sets of both long-period $P$ and $SH$ waves by Mondt (1977), Doornbos & Mondt (1979a,b), Mula & Müller (1980), Mula (1981), and Doornbos (1983) have presented radially symmetric models for $D''$, but despite similarities in the procedures, these studies have inconsistent results. The models preferred by Doornbos & Mondt (1979b) are shown in Figure 1, where the PEMC model (Dziewonski et al. 1975) has been modified to have negative velocity gradients for both $P$ and $S$ waves in the lowermost 75 km of the $D''$ region. Mula (1981) prefers a model with slightly positive gradients that is essentially the same as PREM. These studies have shown that the $S$-wave decay coefficients differ substantially from the $P$-wave coefficients, and that in the long-period band it is principally the $S$-wave observations that require decreased or negative velocity gradients. The major explanation for the inconsistencies in the models is that the $S$-wave decay coefficients, as well as the $P$-wave decay coefficients at the shortest periods, show substantial scatter. The various authors have chosen different ways of weighting and linearly combining the scattered values in determining a "best" model. The very presence of the large scatter in the observations casts doubt on the validity of the resulting models as global averages. These studies explored a limited class of smoothly varying structures, and the evidence for radial discontinuities discussed below indicates that more complex models may need to be considered.

The recent quantitative modeling of diffracted phases has shown that the diffracted ray parameters are more sensitive to average velocity levels in $D''$ than to velocity gradients (Okal & Geller 1979, Mula & Müller 1980). Synthetic models for long-period body waves also indicate that measured apparent velocities can be lower than the true CMB velocities by as much as 0.4 km s$^{-1}$ for $P$ waves and 0.12 km s$^{-1}$ for $S$ waves as a result of the effects of dispersion. Thus, even the PREM model may be consistent with the early measurements of low apparent velocities of diffracted waves. However, the evidence for strong dispersive effects in the diffracted waves has been disputed by Doornbos (1983) and Bolt & Niazi (1984). Doornbos (1983) found that for $SH$ waves, low-frequency dispersion is close to zero, indicating a near critical velocity gradient. Bolt & Niazi (1984) argue that impulsive, high-frequency onsets of the diffracted waves are usually picked in determining the travel-time slopes, and the high frequencies should "see" the deepest part of $D''$. Both arguments
have merit and suggest that short-period diffracted signals hold the key to determining $D''$ velocity structure.

The amplitude decay of diffracted short-period $P$-wave signals has been studied by several workers (Sacks 1966, Carpenter et al 1967, Bolt 1972, Ansell 1974, Booth et al 1974). These studies had limited success because of the large amount of scatter in the $P$-wave data. This scatter can be only partially attributed to the $D''$ region, since receiver and source effects are known to be very strong. Ruff & Helmberger (1982) devised an experiment to suppress the shallow effects by taking the ratio of amplitude patterns at North American stations for events at several Soviet test sites. They found an abrupt onset of short-period amplitude decay near $95^\circ$ and a subsequent, rather steep falloff with distance that they attribute to a localized positive velocity gradient at the base of $D''$. One of the proposed models (POLAR1) that fits these data is shown in Figure 1. Doornbos (1983) and Bolt (1972) have presented observations of high-frequency $P$-wave arrivals at large distances into the shadow zone that would not be consistent with a positive gradient of this type, which suggests that such a model is only appropriate for a limited portion of $D''$. Recently, Ruff & Lettvin (1984) have observed very different amplitude behavior for short-period $P$ waves sampling other regions of $D''$, which indicates that lateral variations are in fact significant.

A fundamentally different class of models based on short-period $P$ waves has been suggested in a series of detailed array studies (Wright 1973, Wright & Lyons 1975, 1981, Wright et al 1985). In these studies, $P$-wave apparent velocity measurements between 75 and $95^\circ$ have been interpreted as resulting from a lower-mantle triplication produced by a sharp 1.5–3% $P$-wave velocity increase about 180 km above the CMB. Figure 1 shows one of the proposed models (WL; Wright & Lyons 1981), which has a strong negative velocity gradient below the 1.5% discontinuity. This model differs dramatically from the other $P$-wave models proposed for $D''$, but a 1.5% discontinuity could easily go undetected by travel-time, long-period diffracted wave, and free oscillation studies. The detection of such a discontinuity is very complicated because the associated changes in ray parameter around $85^\circ$ are small, and the later branches of the triplication are difficult to recognize because of strong interference with the first arrivals. The possibility of crustal contamination of the array measurements has prevented general acceptance of the $P$-wave velocity discontinuity model, and an array study by Schlittenhardt (1984) has failed to find supporting evidence for this model.

The evidence for a velocity discontinuity in $D''$ has been strongly bolstered by the studies of Lay & Helmberger (1983a), Zhang & Lay (1984), and Young & Lay (1986). These investigations have established that in
four separate regions of the lowermost mantle (beneath Eurasia, Alaska, Central America, and India), a 2.75% shear velocity discontinuity exists about 280 km above the core. Unlike the evidence for the $P$-wave discontinuity, the $S$-wave structures (an example of which is shown in Figure 1) are based on long-period observations in the distance range 70 to 95°. The observations show a systematic arrival preceding the core reflection, $ScS$, throughout this range. This arrival results from a triplication produced by the abrupt velocity increase, the size of which was constrained by detailed waveform modeling. The relatively low $S$-wave velocities in $D''$ separate the triplication arrivals more than for the $P$ waves, enabling the use of stable long-period signals in the analysis. The long-period data that were modeled are not particularly sensitive to the velocity gradient below the discontinuity (Lay & Helmberger 1983b), but analysis of diffracted waves in the presence of such a discontinuity strongly favors the presence of at least a mild negative velocity gradient (Lay 1985). The discontinuities in the proposed $P$- and $S$-wave models are separated in depth by about 100 km, but it may be possible to reconcile the models if they are constrained to have similar velocity gradients below the discontinuities. It appears that a global $P$-wave velocity discontinuity of greater than 1.5% is incompatible with some diffracted $P$ observations (Schlittenhardt et al 1985); however, a smaller discontinuity cannot be detected with these phases. Alternate interpretations of the $S$-wave triplication arrivals as results of $SKS$ scattering, receiver reverberations, or source multipathing have been ruled out by detailed analysis (Lay & Young, 1986, Lay 1986). The existence of a globally stratified $D''$ would have profound impact upon the thermal and compositional models for the CMB, so intensive efforts are needed to establish the validity and extent of these structures. In particular, it is important to appraise the significance of these models in the light of the evidence for strong lateral heterogeneity in $D''$.

**Lateral Heterogeneity of $D''$ and the Outermost Core**

Radial characterizations of the average velocity structure near the CMB are required for thermal, chemical, and dynamical analyses of the region. However, the previous discussion and the perplexing array of recent radial seismic models shown in Figure 1 do not provide a unified model. In fact, the significance of all of these models is placed in question because of clear evidence for lateral heterogeneity in $D''$.

Travel-time studies provided the first indication of large-scale lateral heterogeneity in $D''$ (e.g. Chinnery 1969, Julian & Sengupta 1973). With the accumulation of large travel-time data sets, it has been possible to invert directly for the three-dimensional configuration of the lower-mantle
velocity variations (Sengupta & Toksöz 1976, Dziewonski et al 1977, Sengupta et al 1981, Clayton & Comer 1983, Dziewonski 1984). Each of these global inversions has indicated the presence of greater heterogeneity in $D''$ than in the overlying mantle. These large travel-time data sets cannot resolve detailed velocity layering but do extract the long-wavelength component of the velocity variations in $D''$. In a recent analysis of 500,000 travel-time residuals, Dziewonski (1984) found that the low-order (degrees 2 to 6) spherical harmonic components of the heterogeneity have 1.0–1.5% velocity variations in $D''$, three to four times greater than variations in the central mantle. This is considered to be a lower bound on the actual range of variations, given the smoothing effect of the low-order expansion. In the higher-resolution models (Clayton & Comer 1983), some of the heterogeneities extend upward from the CMB into the mantle in a manner suggestive of rising plume structures.

While the global inversions indicate the presence of very long wavelength velocity variations in $D''$, many scales of heterogeneities are indicated by different seismic waves. Long-period diffracted signals indicate coherent large-scale variations in the velocity gradients in $D''$ (Alexander & Phinney 1966, Bolt & Niazi 1984) that may account for the scatter found in diffracted S-wave decay coefficients. Diffracted short-period $P$ waves indicate similar lateral variations (Ruff & Lettvin 1984). In numerous studies, short-period precursors to $PKP$ phases have been attributed to lateral heterogeneities in $D''$ with scale lengths of 10 to 150 km and mean velocity variations of about 1% (Cleary & Haddon 1972, Doornbos & Vlaar 1973, Haddon & Cleary 1974, King et al 1974, Wright 1975, Husebye et al 1976). Comparison of the $PKP_{ab}$ and $PKP_{df}$ branches led Sacks et al (1979) to conclude that various portions of $D''$ produce different amounts of focusing and defocusing. Strongly scattering regions have scale lengths of 150 km with 1% mean velocity changes, whereas other regions have much smaller lateral variations. Haddon (1982) suggests that the heterogeneity has both short (10–20 km) and longer (500–1000 km) scale lengths, with the large-scale features concentrated in preferred directions.

Some of the scattering attributed to $D''$ heterogeneity may instead be due to topography on the CMB itself. Doornbos (1978) showed that the $PKP$ precursors may be explained by topography of a few hundred meters. Precursors to $PKKP$ phases and off-azimuth anomalies of $PKKP$ have been attributed to backscattering from the underside of a rough CMB (Doornbos 1974, 1980, Chang & Cleary 1978, 1981). The variability of the $PKKP$ precursors is indicative of considerable lateral variation in the topography on the CMB. Greater relief of up to 5 km has been suggested by recent three-dimensional models of the outer core obtained by inverting large $PKP$ travel-time data sets (Creager & Jordan 1986, Morelli & Dzie-
wonski 1986). The amount of inferred topography trades off with the degree of heterogeneity attributed to the $D''$ region; however, comparable excess ellipticity of the CMB may also be required to account for anomalous splitting of normal modes that are sensitive to the outermost core (Ritzwoller et al 1985). Any topography on the CMB can be sustained only by dynamic processes because of the large density contrast between the core and mantle, and hence such topography may be crucial to understanding these processes. It may also affect the interpretation of core reflections and diffracted waves.

Lateral variations in the outermost core are difficult to detect for the same reasons that radial models have limited resolution. In general, however, the outer core has usually been considered to be laterally homogeneous, principally on the basis of the small scatter of travel times of $PmKP$ phases (Engdahl 1968, Buchbinder 1972). In a recent global travel-time inversion, Creager & Jordan (1986) suggested substantial lateral heterogeneity of the outermost core, but this is a preliminary result. As a result of the low viscosity of the outer core, it is unlikely that lateral heterogeneity of sufficient magnitude to be observed seismically can be sustained, even dynamically.

**Seismic Constraints on Anelastic Properties Near the CMB**

Anelastic processes in the Earth are thermally activated; thus, a detailed model of the attenuating properties near the CMB would help to resolve whether thermal or compositional effects dominate in this region. Many studies have suggested that anomalously high attenuation of seismic waves occurs in the $D''$ region, while the outer core appears to transmit seismic body waves with almost no anelastic loss. Radial models of the quality factor $Q$ (which is inversely proportional to the amount of anelastic loss) based on normal-mode data tend to have low-$Q$ values in $D''$ relative to the rest of the deep mantle (Anderson & Hart 1978a,b, Anderson & Given 1982); however, it is not clear that the mode data alone require this low-$Q$ zone (Sailor & Dziewonski 1978, Masters & Gilbert 1983). Anderson & Given (1982) presented a frequency-dependent absorption-band model for $Q$ variations in the Earth in which the absorption band in the $D''$ region shifts to higher frequencies in order to match relatively high $Q$ values for the low-order spheroidal modes $\nu S_2$ to $\nu S_4$. This model predicts low-$Q$ values in the range 100 to 150 for periods less than 500 s, with increasing $Q$ for longer periods.

The question of whether body-wave observations are consistent with the presence of a low-$Q$ zone in $D''$ has been strongly contested. Many early studies indicated the presence of a low-$Q$ zone at the base of the mantle by spectral ratio measurements of body waves (Mikumo & Kurita...
1968, Teng 1968, Shore 1984); however, these studies have not convincingly accounted for frequency-dependent effects due to diffraction. Separating the effects of $Q$ and diffraction is difficult, because both strongly modify the spectral content, and accurate correction for the diffraction effects requires detailed knowledge of the $D''$ velocity structure. Doornbos & Mondt (1979a) considered this problem in detail and concluded that the observed $S$-wave spectral decay coefficients are inconsistent with a thick low-$Q$ zone at the base of the mantle, although a thin zone cannot be excluded. Mula (1981) explored this issue further, finding that $Q$ values of less than about 250 lead to $P$-wave amplitude decay predictions that are inconsistent with the data. Velocity models with positive velocity gradients can match the observations well with no attenuation at all (i.e. infinite $Q$). If the $Q$ structure in $D''$ is similar to that in the overlying mantle ($Q$ values near 300–500), models with slightly negative velocity gradients in $D''$ will fit the data best. Recent high-frequency body-wave studies have also failed to detect any low-$Q$ zone in $D''$ (Ruff & Helmberger 1982, Choy & Cormier 1986).

Body-wave studies of the outer core indicate high-$Q$ values of 5000 to 10,000 (Qamar & Eisenberg 1974), 3000 to 10,000 (Sacks 1969), 4000 (Buchbinder 1971), and 10,000 (Cormier & Richards 1976). $Q$ values of several hundred were proposed by Suzuki & Sato (1970), but these results may be in error as a result of the spectral procedure employed and the assumption of frequency independence of the reflection transmission coefficients of SKS. There has been no clear indication of radial variation of $Q$ in the outermost core.

GEODYNAMIC, GEOMAGNETIC, AND COMPOSITIONAL CONSTRAINTS ON THE CMB

While seismology directly probes the velocity structure near the CMB, other geophysical studies place general constraints on the region. It is important to recognize that many of these studies explicitly utilize inferences from seismology; therefore, the ambiguity in the current seismic models must affect the reliability of the conclusions drawn from other disciplines.

Evidence for a Thermal Boundary Layer

The general decrease in velocity gradients in the $D''$ region has often been attributed to the presence of a thermal boundary layer. A thermal boundary layer would exist if there is a strong contrast in temperature
between the lower mantle and the outer core, which would lead to heat flux out of the core. The existence of such a thermal boundary layer is of paramount importance to models of lower-mantle dynamics because of the difference in convection geometries resulting from heating from below versus internal heating only. Systems with heating from below tend to have much stronger upwelling thermal plumes arising from boundary-layer instabilities. Unfortunately, estimating the heat flux out of the core is fraught with uncertainty, as is estimating the temperature at the base of the mantle (Jeanloz & Richter 1979). The calculations require assumptions about whether whole-mantle convection occurs, as well as whether the $D''$ region and the outermost core are stratified, all of which are unresolved problems.

Several calculations do predict a strong temperature contrast of 650 to 1300° across the CMB, which would produce a major thermal boundary layer (Verhoogen 1973, Jones 1977, Jeanloz & Richter 1979, Stacey & Loper 1983, Zharkov 1985). These estimates indicate boundary-layer thicknesses of 75 to 100 km over which seismic velocity gradients would be expected to decrease and possibly become negative (Doornbos 1983). If the $D''$ region is dynamically separate from the overlying mantle, a second thermal boundary layer may exist at the top of $D''$, although all of the estimates of deep Earth temperature contrasts may be in error if this is the case (Jeanloz & Richter 1979). If the $D''$ region is not dynamically stratified, the boundary layer should have reduced viscosities at the base, giving rise to strong horizontal flow and thermal plumes of relatively small dimensions (on the order 20 km in diameter) that would rise into the mantle (Loper & Stacey 1983, Stacey & Loper 1983, Loper 1984, Zharkov et al 1985). The time scale for the growth of these instabilities is on the order of 1 Myr (Yuen & Peltier 1980). The localized thermal plumes may be entrained in a larger-scale lower-mantle circulation (Boss & Sacks 1985). If viscosity in the Earth increases with depth, the surrounding large-scale circulation may overturn slowly, while the small-scale plumes will ascend rapidly. If $D''$ includes patches of material that are too dense to rise, the convective overturn in $D''$ may be irregularly distributed around them. In this case the $D''$ region may resemble the lithosphere, where oceans and continents have very different participation in the upper-mantle convective system (Jordan 1979, Doornbos et al 1986). The strong horizontal shear flow at the base of $D''$ predicted by the thermal boundary layer calculations may result in observable seismic anisotropy (Lay & Helmberger 1983b, Doornbos et al 1986).

The three-dimensional heterogeneity structures for the lower mantle obtained by global travel-time inversions indicate a complex configuration of velocity and, presumably, density variations. Over the long time scales
operating in the Earth, any density heterogeneity must drive viscous flow. The resulting large-scale flow in the deep mantle should produce dynamically supported topography on the CMB with a total excursion of about 3 km (Hager et al. 1985). Larger relief may exist in regions with stronger heterogeneity.

**Geomagnetic Constraints on the Outermost Core Structure**

The Earth's magnetic field is produced by convective motion in the outer core; hence, it is plausible that some characteristics of these motions should be reflected in measurements of the magnetic field at the surface of the Earth. Several efforts have been made to extract information about the actual radial convection currents in the outermost core by analyzing the secular variation of the magnetic field (Roberts & Scott 1965, Backus 1968, Benton et al. 1979, Whaler 1980, 1982, Gubbins 1982, 1983, Le Mouël et al. 1985, Gire et al. 1986); however, Backus (1982) has applied singular perturbation theory to demonstrate the nonuniqueness and limited resolution of this technique. Even if the outermost core is actively convecting, the very low viscosities should not allow significant lateral heterogeneity that would be seismically observable to persist. The core convection will also maintain the CMB at a nearly constant temperature.

A thermally stratified outermost core, which would inhibit large-scale radial motions, has been proposed on the basis of various thermal and chemical arguments (Higgins & Kennedy 1971, Gubbins et al. 1982). This stable layer may serve as a reservoir for light materials segregated out as the inner core grows. As discussed previously, the seismological evidence for such a stabilized zone is marginal, and efforts to directly observe inertia-gravity waves due to such a region have yielded ambiguous results (Masters 1979, Yukutake 1981, Crossley 1984).

Correlation of changes in the geomagnetic and gravitational fields with changes in the length of day and minima in the Earth's rotation rate suggests that there is a coupling between the outer core and mantle. The coupling is thought to be due to either of two causes: electromagnetic coupling caused by motions of the convective systems on either side of the CMB, or topographic coupling caused by irregularities on the CMB (Hide 1969, Anufriyev & Braginski 1975, Moffatt & Dillon 1976, Moffatt 1978, Le Mouël & Courtillot 1981, Le Mouël et al. 1981, Hassan & Eltayeb 1982). Topography of several kilometers would be sufficient to couple the mantle and core. This level of topography is consistent with that suggested in the recent global travel-time inversions and the geodynamic calculations. As the seismic models improve, it will be possible to test the competing coupling hypotheses.
Compositional Models for the CMB Region

The composition of the deep mantle is not precisely known, as only rough constraints are provided by seismic velocities and densities. In general, it is believed to consist primarily of magnesiowustite \((\text{Mg,Fe})\text{O}\) and silicates \((\text{Mg,Fe})\text{SiO}_3\). Iron enrichment of the \(D''\) region, brought about by exsolution from magnesiowustite and resulting in a general decrease in seismic velocity, has been suggested by Anderson & Jordan (1970) and Jeanloz & Ahrens (1980). The seismic discontinuity models for \(D''\) are suggestive of a compositional boundary, but it is not clear whether this is of global extent, nor is it known what compositions are involved. Jeanloz & Ahrens (1980) suggested that a phase change in \((\text{Mg,Fe})\text{O}\) could explain some of the anomalous properties of \(D''\). No other candidate phase changes that might account for the discontinuity models have been proposed.

The composition of the outer core is principally iron, but both density and bulk modulus constraints require a light alloying component. The cosmochemically plausible light constituents are \(\text{Si, C, S, K, and O}\) (Ahrens 1982). The possibility of \(K\) being a significant component is particularly important because of the large heat release resulting from radioactive decay. This heat source could increase the temperature contrast across the CMB, enhancing the likelihood of a thermal boundary layer (Elsasser et al 1979). Ruff & Anderson (1980) have suggested that a heterogeneous layer in the outermost core involving an uneven distribution of \(\text{Al, Ca, U, and Th}\) could have accumulated as the core evolved. Uneven heating in this layer could drive the geodynamo.

CONCEPTUAL MODELS OF DYNAMICS NEAR THE CMB

The CMB clearly represents a compositional and geodynamical discontinuity in the Earth. The exact role of the \(D''\) region in lower-mantle dynamics has not yet been established by seismological or thermal modeling. Several plausible scenarios are schematically portrayed in Figure 2. Perhaps the simplest model of the CMB is that it is a major thermal boundary layer at the base of a convecting lower-mantle system, with no significant stratification. In this case (Figure 2a), a thin thermal boundary layer will exist from which small-scale plumes may rise as a result of boundary-layer instabilities. The mass flux will be balanced by slow downward flow from the surrounding mantle, and large-scale upwelling may also be present. The density heterogeneity resulting from temperature
variations will produce viscous flow resulting in dynamically supported relief of several kilometers on the CMB. Some of the plumes originating in the $D''$ region may penetrate to the surface, producing hotspots.

An alternative model, which is also consistent with seismological observations, is one in which the $D''$ layer is a stably stratified, compositionally distinct layer in the lower mantle. This model is supported by the evidence for widespread, if not global, velocity discontinuities at the top of $D''$.

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**Figure 2** A schematic diagram of three possible dynamic models for the CMB region. (a) A model with a single-layer convective system and a thermal boundary layer (TBL) above the CMB. A plume is shown emanating from the TBL. (b) A compositionally stratified model with a two-layer convective system. ($D''$ is separated from the overlying mantle.) Note the second TBL at the top of $D''$ and the different scale lengths of the two convective layers. (c) A mixed model with a single-layer convective system in a laterally heterogeneous mantle. The heterogeneities (shaded) are concentrated in $D''$ as a result of an increase in viscosity with depth.
These discontinuities are presumably embedded within or on top of the laterally varying $D''$ zone, with the lateral variations arising from thermal and compositional gradients associated with small-scale convection in the layer. This possibility has not received much consideration in recent thermal modeling, but it would presumably result in a double boundary-layer system (Figure 2b), with a layered convection system of very different scale lengths. While dynamically supported topography may exist on the CMB, the actual sign of the deflections depends on the rheology of $D''$ (Hager et al. 1985).

A third conceptual model, similar to one proposed by Davies (1984), lies between these two extremes. This model is one in which significant compositional heterogeneity exists in the lowermost mantle, although the $D''$ region is not strictly a compositionally distinct layer (Figure 2c). The chemical heterogeneities exist on many scales, some of which are stably stratified and do not participate in the $D''$ dynamics, while others are entrained in narrow thermal plumes as well as any larger-scale upwelling currents rising out of $D''$. Some of the heterogeneities may represent material subducted from the upper mantle, including oceanic crust (Hofmann & White 1982). In this model, dynamically supported topography of the CMB would exist as a result of the viscous flow driven by the mantle heterogeneities. If viscosities are sufficiently high, the residence time for these heterogeneities may be on the order of billions of years, providing the reservoir of primordial material needed to satisfy geochemical isotopic observations. Strong concentrations of heterogeneities could give rise to reflections that have been interpreted as seismic discontinuities in $D''$. Though complex, this model seems to be the most readily reconciled with the current constraints on the CMB.

CONCLUDING REMARKS

Seismological investigations have not yet reached a consensus on the velocity structure near the CMB, apparently because of the strong lateral heterogeneity of the region. The $D''$ region is particularly complex and may be both a chemical and a thermal boundary layer. While the majority of the seismic analyses indicate the presence of a general decrease in velocity gradients in $D''$, it is not clear whether or not the velocities actually decrease with depth. There is evidence for stratification of $D''$ as well as lateral variations in the velocity gradients, which indicates that the use of a single smoothed velocity model in geothermal analysis is unjustified. Topography on the CMB is likely to result from convection in the overlying mantle. This relief is dynamically supported and provides coupling between the mantle and core. The possibility of a chemically or thermally stratified
zone in the outermost core with a concentration of light components cannot be ruled out. Better resolution of the seismic structure of the CMB region is needed before confidence can be placed in either the chemical or dynamical studies that have been published.

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