Multiple seismic reflectors in Earth’s lowermost mantle

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The modern view of Earth’s lowermost mantle considers a D– region of enhanced (seismologically inferred) heterogeneity bounded by the core–mantle boundary and an interface some 150–300 km above it, with the latter often attributed to the postperovskite phase transition (in MgSiO3). Seismic exploration of Earth’s deep interior suggests, however, that this view needs modification. So-called ScS and SKKS waves, which probe the lowermost mantle from above and below, respectively, reveal multiple reflectors beneath Central America and East Asia, two areas known for subduction of oceanic plates deep into Earth’s mantle. This observation is inconsistent with expectations from a thermal response of a single isochemical postperovskite transition, but some of the newly observed structures can be explained with postperovskite transitions in differentiated slab materials. Our results imply that the lowermost mantle is more complex than hitherto thought and that interfaces and compositional heterogeneity occur beyond the D– region sensu stricto.

The lowermost mantle, extending several hundred kilometers above the ~2,900-km-deep core–mantle boundary (CMB), is of considerable interest because it includes the boundary layer of thermochemical mantle convection across which heat is conducted from the core into the mantle. Almost three decades after the detection of an interface some 150–300 km above the CMB (1), the base of the mantle is still a challenging target for cross-disciplinary research. Both the seismic discontinuity that marks the top of the so-called D– region (1) and the heterogeneity below it (2, 3) have been attributed to a postperovskite (Pv) transition that becomes dominant in the mantle silicate (4–8), and this association has inspired estimation of temperatures above and heat flux across the CMB (9, 10). The D– interface remains enigmatic, however, and recent high pressure–temperature experiments suggest that seismic observations concerning its depth and thickness are inconsistent with those expected for a Pv transition unless the chemical composition of the regions where they occur differs significantly from standard bulk composition models such as pyrolite (11–13). The thickness transition could be reconciled with nonlinearity in the phase fraction profile (11) or lattice preferred orientation in Pv (14), but the depth is a concern because the Pv transition pressure in pyrolite may be too high for it to occur in the lower mantle (13). Candidate compositions for a seismically detectable Pv transition at mantle pressures include midoceanic ridge basalt (MORB) and harzburgite components of subducted and differentiated oceanic lithosphere. Furthermore, silica may transform from modified stishovite to seifertite in Si-rich parts of the lowermost mantle (15). Inspired by these results, we search for multiple interfaces in and above the conventional D– region, using seismic waves that sample the lowermost mantle beneath geographic regions where seismic tomography and plate histories suggest that deep subduction is likely.

Significance

Deep in the Earth’s interior, the region just above the core–mantle boundary exerts control on mantle convection and heat loss from the core. It has long been thought that the so-called D– region is separated from a more uniform mantle above by a single interface, often attributed to a phase transition in Mg perovskite. Systematic deep-mantle exploration with massive seismic waveforms now yields evidence for multiple reflectors up to at least 600 km above the core–mantle boundary. Some of the newly discovered interfaces can be explained by post-perovskite transitions in differentiated oceanic slab materials, transported from Earth’s surface through deep subduction and convection. The lowermost mantle appears more complex than hitherto thought, and this complexity is not confined to the canonical D– region.


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1990–2009] recorded at one or more of a total of 2,700 seismographic stations (Fig. 1, Upper Center). For East Asia, we used 120,000 SKKS traces from 11,000 events recorded at 1,700 stations (Fig. 1, Lower Center). The range of epicentral distances is 0–80° for ScS and 100–180° for SKKS. The data used (~20% of all available traces) passed selection criteria on the basis of signal-to-noise and multichannel cross-correlation values. Source signatures are estimated through principal component analysis and removed from the data through Wiener deconvolution (27, 28). With GRT, we estimate (from scattered energy) elasticity contrasts at nodes of a 3D mesh (10 km vertical, 1° lateral spacing); spatial alignment of such contrasts indicates the presence of an interface. We use a tomographic model (20) to correct for mantle heterogeneity, but the choice of model is not critical for the level of detail discussed here. More information about data selection and processing is provided in SI Text S2.

**Structural Complexity in the Lowermost Mantle**

We illustrate the main structural features in the lowermost mantle beneath the regions under study by means of 2D (vertical) GRT sections through 3D image volumes. The nearly 1,500-km-long Central America section (Fig. 3A) cuts across the main (tomographically inferred) high-wave speed anomaly (Fig. 2, Right). This segment parallels section A–A′ in our previous study (10), but it is shorter because detection of weak structures above D″ requires more stringent imaging conditions and sampling criteria than was necessary for the imaging of CMB and D″. The first of the East Asia sections shown here (B–B′; Fig. 3B) also cuts mainly across seismically fast regions, whereas the other section (C–C′; Fig. 3C) samples slow regions as well. In these profiles, black (red) pulses indicate positive (negative) impedance contrasts with increasing depth; their amplitudes are normalized (with respect to the CMB reflection) and provide only a qualitative indicator of the reflector strength. The background colors depict the tomographic estimates of local variations in shear wave speed used to construct the images.

Beneath both regions, well-aligned black pulses mark the CMB as well as a laterally continuous interface labeled X (fat magenta lines, Fig. 3). Consistent with previous results (8, 10), X occurs some 250–300 km above the CMB beneath Central America (Fig. 3A); however, beneath East Asia, where such an interface was detected previously (2, 29), but where our results establish its large-scale structure, it is positioned closer to the CMB. The data also reveal a weaker, unknown interface (hereinafter Y) some 450–500 km above the CMB. Other structures
exist, but stacking across the sections confirms CMB, X, and Y as the main interfaces (Fig. 3, Right).

Beneath Central America (Fig. 3A), Y is laterally intermittent but is clearly visible over at least a 1,000-km horizontal distance. The data reveal impedance contrasts between CMB and X, as described before (8, 10), and suggest that one or more wave speed drops (red pulses, labeled z) occur between X and Y. Locally, z may look like a side lobe of nearby positive pulses, but in many places it clearly appears as a separate signal. The East Asia section across the fast anomaly (Fig. 3B) is qualitatively similar, with a weak, laterally intermittent interface visible some 200 km above X. In addition, negative pulses appear here between CMB and X and between X and Y (in particular, around 300 km above CMB in the right half of the section). Fig. 3C suggests that the character of these structures changes when moving from (tomographically inferred) high to low wave speeds in the lowermost mantle: interface Y is laterally continuous in the easternmost, 1,200-km, “fast” part of section C–C', intermittent in the center (between 1,000 and 1,500 km horizontal distance), and absent in the seismically “slow” region further west. In the latter, no coherent scattering is visible above interface X, located here ~200 km above CMB.

Phase Transitions in Differentiated Subducted Lithosphere?

Our application of modern imaging techniques to ever-growing data sets confirms the widespread presence of a positive impedance contrast 150–300 km above the CMB beneath Central America and also establishes its large-scale existence beneath East Asia. Consistent with previous results, we interpret this horizon X as the top of the D' region and associate it with the pPv (Pv→pPv) transition. For reference, we also show (black dashes) the hypothetical phase boundary predicted by Sidorin and coworkers (18), who assumed that tomographically inferred wave speed variations have a thermal origin, that a (pressure-induced) mineralogical boundary exists (with a positive pressure–temperature dependence of 6 MPa/K), and that this boundary can be extrapolated globally (18, 19). This prediction is hereafter referred to as the “thermal model.” Along the Central America section (A–A') and in the western part of Asia section C–C', interface X coincides with the hypothetical phase boundary expected from the thermal model. In contrast, it does not correlate with thermal predictions in the high-wave speed parts of the East Asia sections (e.g., B–B').

The large data sets also reveal hitherto unknown structures in the lowermost mantle. The detection of laterally continuous scatter surfaces above the D' interface may still be at the edge of current resolution, but tests with synthetic data demonstrate it cannot be attributed to noise, multiple scattering near sources or receivers (such as depth phases), reverberations within the D' layer (Fig. S2), or multiples of SKKS (i.e., SKS, with n > 2) (17). Our first-order observations (both poor correlation between tomographically inferred wave speed variations and depths to the D' discontinuity and the existence of multiple reflectors) are inconsistent with a single (pressure-induced, temperature-controlled) phase transition in magnesium silicate (MgSiO₃) Pv in a compositionally
homogeneous mantle. Forward (Pv→pPv) and reverse (pPv→Pv) transitions can produce multiple interfaces in the steep thermal gradient near the CMB (7,8,10), but not several hundreds of kilometers above it. If X does indeed mark the top of D∗, our results suggest that interfaces and, by implication, compositional or phase boundaries exist several hundred kilometers above the D∗ region.

With hundreds of millions of years of subduction along the eastern seaboard of Asia and beneath Central America, it is possible that some of the multiple reflectors are a result of the buckling of slabs above the CMB (24), as has been suggested to occur beneath Central America (19, 25). Another explanation involving subduction concerns the presence of material that is chemically distinct from a pyrolytic bulk composition. Variations in mineralogy and iron and aluminum content can influence the propagation speed of seismic waves (30–32), as well as the Ppv transition depth (11, 13, 33) and detectability (13). Of particular interest are predictions from experimental mineral physics that in a heterogeneous mixture of harzburgite, basalt, and bulk mantle, the Pv→pPv transition in the MORB fraction can occur several hundreds of kilometers above the transitions in either the harzburgitic component or pyrolite (Fig. 4) (13). The shallower pPv transition would be weak because MORB contains much less magnesium silicate (30%) than harzburgite or pyrolite (60–80%) (34) and because the pPv transition depth interval in MORB is greater than harzburgite [but still smaller than in pyrolite (13)].

Because of complex mineralogy and phase chemistry, uncertainties in absolute pressure in mineral physics data (35), controversy about seismic velocities of bulk mantle composition (pyrolite) and (mixtures of) recycled materials (MORB, harzburgite) (36, 37), and the fact that weaker impedance contrasts are only now beginning to emerge as robust features, one-to-one mapping of phase transitions and seismic boundaries is still premature. However, combining evidence from mineral physics and seismic imaging (Fig. 4), we suggest that X marks the pPv transition either in average mantle (if effects from lattice preferred orientation and element partitioning decrease pPv transition thickness to within detectable limits) or in the harzburgitic component of differentiated subducted lithosphere, and that Y marks the transition in the subducted MORB component. Negative impedance contrasts (z and closer to CMB in Fig. 3) may reflect transformations in silica (34), local existence of partial melt (38), or reverse pPv transformations in basalt, harzburgite, or bulk mantle. These are exciting targets of future joint research for mineral physics and deep Earth exploration seismology, preferably with both P- and S-type data (39), and may lead to a further revision of the canonical view of a lowermost mantle with compositional and structural heterogeneity restricted to the D∗ layer.

**Methods**

We used a GRT to recover unknown elastic reflectors in the lowermost mantle from the scattered ScS and SKKS wave fields (SI Text S1; Fig. S1). In essence, the GRT maps (as in reverse time migration) the scattered seismic wave field (recorded at the surface) back to subsurface contrasts in elasticity. For each node of a dense 3D mesh, the GRT exploits data redundancy through the integration of waveform data over a wide range of scattering angles and azimuths. In theory, point scatterers can be resolved in the Rayleigh diffraction limit (which depends on frequency), but in practice, spatial resolution depends on how a subsurface point is illuminated (i.e., the wave slownesses and the range of scattering angles over which data are integrated), which depends on source-receiver distribution. To ensure robustness, we require that target points be sampled from a sufficient range of angles (15°) around the stationary point (specular reflection), which enables us to resolve structure at lateral scales of 500 km or larger. GRT imaging does not rely on priors assumptions about the location or shape of geological targets, which facilitates discovery of hitherto unknown structures. More details of the method and data preprocessing are given in SI Text S1 and SI Text S2. For the mineral physics data, the absolute pressure (or depth) is uncertain by at least ±5 GPa (±100 km) (35) and depends on the pressure scale used. We note that the use of the gold pressure scale by Tsuchiya (40), originally used in Grocholski and coworkers (13), gives depths in between those inferred from the gold scales used in Fig. 4B. The relative pressure scale (or difference in pressure or depth) is better constrained

![Fig. 4](https://www.pnas.org/cgi/doi/10.1073/pnas.1312647111)

Fig. 4. (A) Stacks along interface X in sections A–A′, B–B′, and C–C′ (shown in Fig. 3). (B) Depth ranges for pPv transitions in midoceanic ridge basalt (MORB), harzburgite (Harz), and pyrolite (Pyr), after ref. 13, and transitions in silica from modified stishovite to seifellite (Sft), after ref. 15, at ~2,500 K. Plotted are the lower bounds of the pPv transition thickness. The pressures at the phase boundaries are calculated using a shock wave (gold) scale from ref. 41, a scale bar on the left of B, or a static compression (gold) scale from ref. 42, on the right. Pressure was converted into depth using preliminary reference Earth model (PREM; 43). Absolute pressure (depth) is uncertain by at least ±5 GPa (±100 km), but pressure differences are constrained better (±1 GPa). Given these uncertainties, the depth difference between X and Y agrees remarkably well with the difference in pPv transition depth between MORB and Harz.
(≤1 GPa), and all mineral physics data shown in Fig. 4B are constrained, using the same pressure scale (gold). Therefore, depth differences among the phase boundaries in Fig. 4B are more reliable for comparison with seismic data than absolute depths.

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Inverse scattering in seismology refers to estimating \( \theta_c \) (Fig. S1). For a given \( x \) data.

\[ \nu \delta_1 \]

1

\[ \psi = \delta \theta \]

or \( p = \delta \theta \) data and the radial component (SV wave) for \( x = \) can be divided into two parts, \( u(x) = u_d(x) + \delta u(x) \), where \( u_d(x) \) is the direct wave resulting from propagation in \( c_0(x) \), and \( \delta u(x) \) is the scattered wave resulting from \( \delta c(x) \). The background smooth model can be estimated from reference Earth models and seismic tomography. Inverse scattering maps the scattered wave field \( \delta u(x) \) back into images of perturbation of medium properties \( \delta c(x) \).

The scattered wave field \( \delta u(x) \) depends on the position of scattering point \( y \), source \( x \), and receiver \( x' \) (Fig. S1). For a given \( (y,x',x') \), we define the slowness vector \( p(y) \) for the incident ray from source to the scattering point, as well as \( p(y') \) from the receiver side. These two slowness vectors can define migration dip \( \nu = p(y) - p(y') \), with \( p(y) = p(y') + p(y',y') \), scattering angle \( \theta \), and azimuth \( \psi \). The structural reflectivity at \( y \) can be approximated as

\[ I(y) = \int \delta u(y; \nu, \theta, \psi) |p(y')|^2 d\nu d\theta. \]

where \( \delta u(x) \) is integrated over the scattering angle \( \theta \) and azimuth \( \psi \).

In practice, the spatial resolution of GRT depends on how an image point is illuminated (integration range of Eq. 1). The imaging result would be biased by uneven source-receiver distribution (e.g., dominant earthquake direction, poor illumination coverage). To ensure robustness and mitigate inversion artifacts, we conduct two processing steps. First, all image points are required to be sampled from a sufficient range of angles around the stationary point (i.e., the specular reflection). For each vertical profile in Fig. 3 (i.e., the same latitude and longitude with different depths), we check the illumination for an image point at the core–mantle boundary (CMB). Such a point is included in the final image only if the data sampling it contain specular reflections (i.e., migration dip angle is zero) and a wide range of migration dip angles (here we use 15°). Second, the contributions from different earthquakes are balanced through simple ray-count normalization. After trace normalization, the ray count is proportional to the incident wave energy. Before summing over the scattering angle \( \theta \) in Eq. 1, the ray count is calculated for each angle bin (3°), and each partial image is normalized by the ray count.

**SI Text S2 Preprocessing of the Data.** There are several steps in the selection and preprocessing of the data. First, for all data, we remove the instrument response and rotate the data to radial and transverse components. We use the transverse component (SH wave) for ScS data and the radial component (SV wave) for SKKS data.

Second, all data are band-passed between 10 and 50 s with a 4-pole Butterworth filter and then normalized with respect to the reference phase (ScS or SKKS). Third, we discard data on the basis of a simple quality criterion obtained from multichannel cross-correlation (6). For each earthquake, we organize the data in 10° (epicentral) distance bins, and from each seismic record in a bin, we extract a 50-s time window around the theoretical ScS (or SKKS) arrivals. After energy normalization, we cross correlate each trace with all others in the bin, which yields a qualitative estimate of the average correlation coefficient for each trace. Traces with average correlation coefficient lower than a set threshold (we use 0.6) are discarded. Roughly 20% of all data meet this quality criterion. The trace polarity can be corrected according to the sign of average correlation coefficient.

After multichannel cross-correlation, traces are aligned with reference phases. Fourth, principle component analysis is applied to estimate the source signature (7), which is then removed from the data to enhance image resolution by Wiener deconvolution (8). The so-called water level in Wiener deconvolution is selected automatically and adaptively based on the noise spectrum of array data. Finally, travel times are corrected for ellipticity (9) and for 3D wave speed variations, using a tomographic model for mantle shear wave speed (10).

**SI Text S3 Tests with Synthetic Data.** Because the quality of the GRT imaging operator depends on the aperture of angles \( \theta \) and azimuth \( \psi \) (SI Text S1) and the validity of the single-scattering approximation, tests with synthetic data are necessary to examine the effects of uneven sampling, noise, interfering main phases (such as depth phases), and multiple scattering (e.g., near-source or receiver scattering and internal reverberations). In all the tests described here, we computed synthetic waveforms for the real source-receiver geometry (including focal depths) and focal mechanism obtained from the Global Centroid-Moment-Tensor (CMT) catalog. All synthetic seismograms are calculated with the Wentzel-Kramers-Brillouin-Jeffreys (WKBJ) method (11), using a radially stratified wave speed model [ak135 model (12)], on which a 3% S wave velocity jump is superimposed at 250 km above CMB. Before the inversion, all synthetic data are band-pass filtered between 10–50 s. We note that the synthetics merely test numerically whether the available sampling in migration dip is sufficient. In other words, the performance of such synthetics (in linearized inversions) depends mainly on sampling, and not on the strength of the contrast or the noise level. Indeed, a GRT is stable in \( L^2 \) and, hence, effectively attenuates additive random noise, allowing the imaging of weak deterministic reflectors. In concept, this use and limitation as a diagnostic for image quality is similar to the so-called checkerboard test used to verify the effect of sampling on the ability to recover input patterns in tomography.

Fig. S2 shows the effects on the SKKS image gather of random noise (100% noise level), depth phases, and multiple reverberations. Even when the signal cannot be identified in most individual traces, the reflector at 250 km above CMB is clearly recovered after stacking over the scattering angle (Fig. S2A). The depth phase can produce small waveform distortion (Fig. S2B); however, it does not produce significant artifacts in the image. The reverberation in the crust and D* layer can also be a potential source of artifacts because such phases could interfere with the coda wave of SKKS. Applying the GRT to the synthetic data with such signals shows they do not, in general, contaminate the image profile (Fig. S2 C and D). Moreover, such multiples
would generate structures that are (within uncertainty of the background wave speed) parallel to the main interfaces, which is not generally the case in the results presented here. 

SKKS multiples $SKS_n (n > 2)$, which reverberate within the liquid outer core, arrive later and could interfere with the reflections from lowermost mantle interfaces in time domain. In angle gathers, their slownesses differ from those of SKKS (and SKS’SKS), and therefore they can be distinguished by a clear residual move out and suppressed by parabolic Radon transforms (5).
