Reconciling the Post-Perovskite Phase With Seismological Observations of Lowermost Mantle Structure

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The discovery of post-perovskite (pPv), the high pressure polymorph of $(Mg_x,Fe_{1-x})SiO_3$ perovskite (Pv), may have profound implications for the thermal, chemical and dynamical structure of the deep Earth, *if* it can be demonstrated that pPv currently exists in the lower mantle. The requisite basis for such a demonstration is seismological observation of elastic velocity and density structures diagnostic of or at least compatible with the expected properties of pPv occurrence in a realistic lower mantle chemical and thermal environment. A critical assessment of seismological observations versus predictions for a lower mantle model with pPv is undertaken, with the limitations and robust attributes of the seismological data being summarized. The existence of a seismic velocity discontinuity several hundred kilometers above the core-mantle boundary is a primary line of evidence for the presence of a Pv-pPv phase change. However, some attributes of the discontinuity, such as localized *P* wave reflections from a large *P* velocity increase and the sharpness of observed *P* and *S* velocity increases, reveal inconsistencies with expected properties of a Pv-pPv phase transition. Lateral variations in temperature can produce complex phase boundary structure that explains variable *S* wave observations, but such elastic heterogeneity intrinsically complicates testing the pPv hypothesis. The combination of rapidly expanding seismological data sets and new high resolution data analysis procedures reveal multiple seismic discontinuities near the base of the mantle; in some cases these may be consistent with forward and reverse Pv-pPv transformations, bounding a pPv layer or lens in D″ above the CMB. Other phase changes or compositional contrasts could also be involved. At present, existence of pPv in the deep Earth is quite plausible, but not yet conclusively demonstrated. Future seismological and mineral physics research directions for further testing the hypothesis that pPv is present in the lower mantle are suggested.

1. INTRODUCTION

Geophysical remote sensing of Earth's deep interior is a challenging undertaking, and even when some aspects of the structure, such as its elasticity, can be characterized with confidence, there is always ambiguity associated with ensuing thermal, chemical and dynamical interpretations. Given that

Post-Perovskite: The Last Mantle Phase Transition
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uncertainties remain regarding the precise mineralogical and compositional contrasts even between crust and upper mantle rocks, it should come as no surprise that interpretations of lowermost mantle observations are fraught with large uncertainties. Nonetheless, motivated by recent advances in seismology and mineral physics, here we will explore whether one can actually make, or hope to make in the near future, high-confidence statements about composition and temperature nearly 3000 km deep into the Earth, with no realistic prospect of ever having a hand-sample to provide ground truth.

One of the key approaches to reducing uncertainty of deep thermal and chemical interpretations has been to associate laboratory-calibrated phase transitions in expected abundant Earth minerals with seismically determined *P*- and *S*-wave velocity contrasts in the interior. With mineral physics experiments and theory providing guidance as to chemical and thermal controls on phase transitions and predictions of observable properties such as elastic velocity and density contrasts, consistency between predicted phase transition properties and seismic observations can lead to surprisingly high-confidence deductions about deep Earth thermal and chemical state. This is exemplified by the transition zone seismic velocity discontinuities near 410-km and 660-km depth, which are generally (but not universally) attributed to the occurrence of phase transitions from olivine to wadsleyite and from ringwoodite to ferro-periclase plus magnesium-perovskite, respectively [e.g., *Ito and Yamada*, 1982; *Ito and Takahashi*, 1989; *Jackson and Rigden*, 1998]. The plausibility of significant $(Mg,Fe)_2SiO_4$ mineralogy being present in the upper mantle coupled with extensive experimental and theoretical quantification of properties of the various transition zone polymorphs of this composition, enable comparisons with seismic observations to serve as a powerful probe of absolute temperatures, bulk composition, kinetic effects on the transitions, and even large-scale dynamical flow. These phenomena are tied to specific pressures by the seismically imaged discontinuity depths. There has not previously been such a probe for the lowermost mantle; the situation may now have changed.

With all primary upper mantle mineral forms occurring as $(Mg_x,Fe_{1-x})SiO_3$ perovskite (Pv) at lower mantle pressures and temperatures, and this mineral found to be remarkably stable over large pressure-temperature-composition (P-T-X) domains (including coexistence with other likely lower mantle components such as ferro-periclase and Ca-perovskite), it is generally accepted that Pv is the most abundant lower mantle mineral [e.g., *Knittle and Jeanloz*, 1987; *Wentzcovitch et al*., 1993; *Serghiou et al*., 1998; *Fiquet et al*., 2000; *Shim et al*., 2001; *Wentzcovitch et al*., 2004; *Gong et al*., 2004]. The recent discovery that $MgSiO_3$ perovskite (Mg-Pv) undergoes a transition to a new phase, called post-perovskite (pPv), at pressures and temperatures likely to exist within a few hundred kilometers distance above the core-mantle boundary (CMB) [*Murakami et al*., 2004; *Iitaka et al*., 2004; *Oganov and Ono*, 2004], has elicited great excitement: mineral physics theory predicts seismically observable attributes of the pPv phase transition [e.g., *Tsuchiya et al*., 2004a; *Stackhouse et al*., 2005a; *Wookey et al.*, 2005a; *Shieh et al*., 2006] generally consistent with long-standing unexplained seismic observations of *P*-wave [e.g., *Wright, 1973*] and *S*-wave [*Lay and Helmberger*, 1983a] D″ discontinuities. Thus, for the first time, a joint mineral-physics and seismological probe of composition, temperature, and possibly dynamics, may be in-hand for the deep mantle [*Lay et al*., 2005; *Hirose*, 2006], yielding key information fortuitously close to the thermo-chemical boundary layer believed to exist above the CMB [e.g., *Lay and Garnero*, 2004].

While it is enticing to immediately and boldly interpret past and present seismological observations in the new context of the pPv discovery [e.g., *Helmberger et al*., 2005; *Wookey et al.*, 2005a; *Hutko et al*., 2006], our purpose here is to cautiously assess the validity of doing so. We will specifically assess the supporting arguments for and possible errors in such interpretations, and define key areas for mineral physics and seismological efforts that will ultimately place interpretations of lower mantle state and processes on at least as firm of a basis as exists for the transition zone. We proceed by considering the implications of laboratory and theoretical predictions for seismically observable attributes of pPv presence jointly with a survey of actual seismological observations. We do not seek to be comprehensive with regard to the latter, given that there are several recent thorough reviews of the topic [e.g., *Garnero*, 2000; *Lay et al.*, 2004b; *Lay and Garnero*, 2004]; instead, we focus on the salient observations relevant to the possible existence of pPv in the lower mantle. With deep mantle seismological research having been conducted for decades with little opportunity to test specific hypotheses, we exploit the opportunity presented by the discovery of pPv to re-assess seismological observations in the new and exciting context.

2. PREDICTED POST-PEROVSKITE PROPERTIES

This volume includes summaries of current experimental and theoretical constraints on the properties of pPv (see Sections 1 and 2). As is true for other mineralogical phases, it is important to know the P-T-X dependence of the pPv polymorph, the stability and element partitioning when coexisting phases are present, the phase transition kinetics, and the transport properties of the phase (Figure 1). Given that there is large uncertainty in lowermost mantle bulk composition and thermal structure, it is necessary to consider a wide range of mineral physics parameters when making comparisons

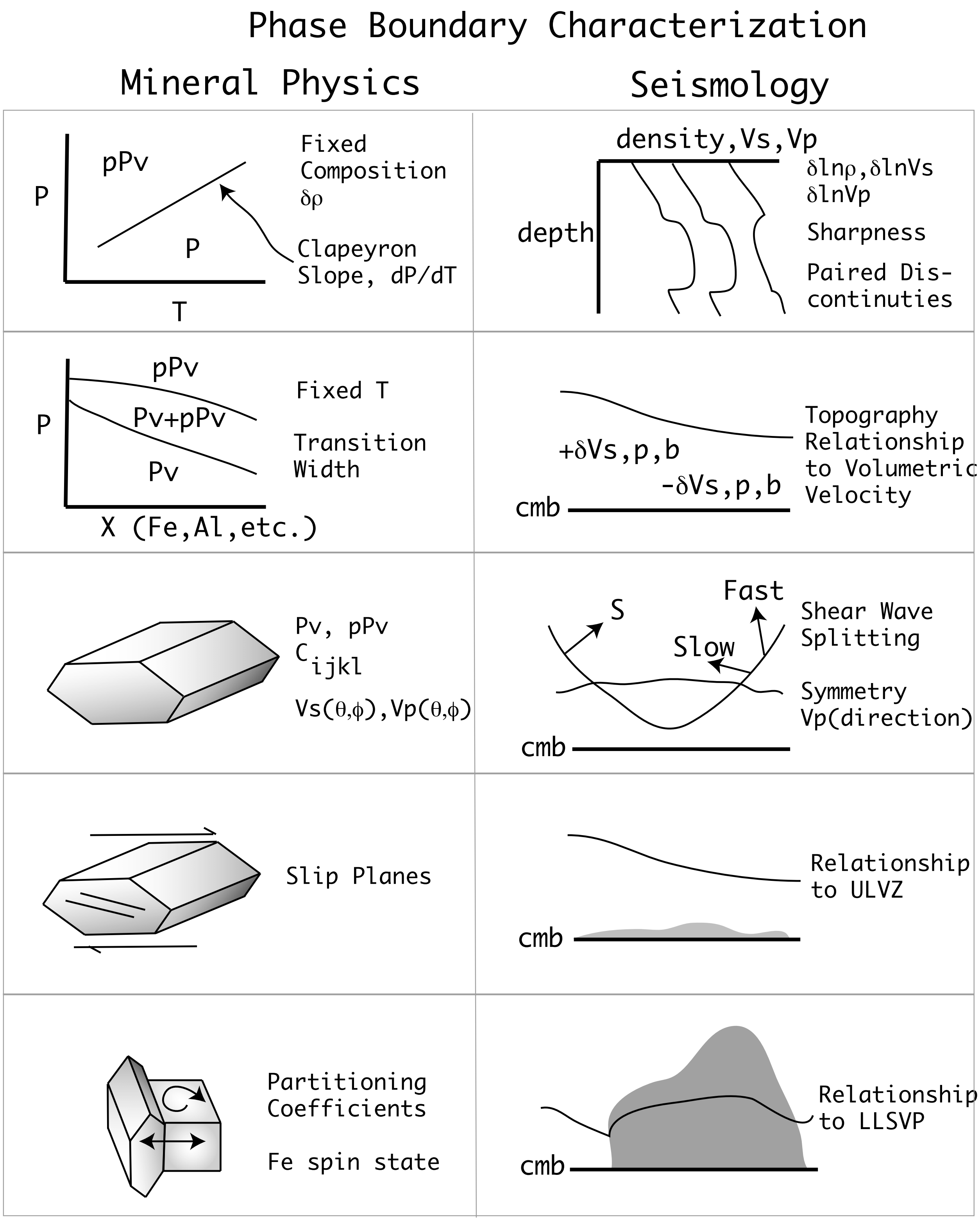

Figure 1. A schematic of key types of information contributed by mineral physics and seismology disciplines to the characterization of phase changes inside the Earth. Experimental and theoretical mineral physics provide the recognition of phase changes within minerals of given composition, the P-T position and slope of the phase boundary, the change in density ($\delta\rho$) across the phase boundary, the width of the two-phase domain and the phase boundary dependence on composition (X), the elasticity tensor, c_{ijkl} and associated anisotropic P and S velocities $V_p(\theta,\phi)$, $V_s(\theta,\phi)$, the slip planes activated under prescribed shear stresses, and element partitioning, Fe spin-state and associated transport properties. Observational seismology determines the existence, depth and sharpness of velocity and density discontinuities in the Earth, with the associated changes in material properties, the patterns of topographic variation of the reflectors and their relationship to volumetric V_s, V_p and V_b (bulk sound velocity) variations, the magnitude and orientations of split shear waves, directionality of V_p velocity, and relationship of reflectors to features such as ULVZ structure right at the core-mantle boundary (cmb) and to LLSVP structures that may indicate temperature and chemical effects.

with seismological observations. Mg-pPv is experimentally unquenchable and in situ measurement of many properties at high P-T conditions is difficult; some properties must be predicted by theory or inferred from low P-T analogues [e.g., *Hirose et al*., 2005b; *Tateno et al*., 2006]. Due to the short time elapsed since the discovery of pPv, the full-suite of parameters for the phase has not yet been explored; we can thus anticipate limitations in assessment of pPv occurrence in the deep mantle at the present time due to inadequate knowledge of pPv properties at lower mantle conditions.

While the detailed properties of pPv are not yet fully elucidated, enough information is available to initiate assessment of possible pPv existence in the deep Earth, and we highlight key findings here. The initial work on the Mg-Pv end-member established the existence and change in volume of the pPv phase, and provided constraints on the atomic lattice that enabled theoretical modeling of the precise crystal structure of the mineral [*Iitaka et al.*, 2004; *Oganov and Ono*, 2004; *Tsuchiya et al.*, 2004b; *Stackhouse et al.*, 2005b; *Wentcovich et al*., 2006]. Theoretical calculations provided the first prediction of a large positive Clapeyron (P-T) slope of the phase boundary (~7.5 MPa/K) [*Tsuchiya et al.*, 2004b; *Hirose and Fujita*, 2005]; recent experiments suggest that the slope may be as large as 11.5 MPa/K [*Hirose et al.*, 2006]. The Mg-Pv to pPv transition was found to involve discontinuous increases in rigidity and density, a decrease in incompressibility, and a change in the anisotropic crystal properties [e.g., *Stackhouse et al.*, 2005b; *Oganov et al*., 2005; *Merkel et al*., 2006; *Wentzcovich et al*., 2006]. Corresponding seismic observables would be an increase of 2-4% in *S*-wave velocity ($\delta \ln V_s$) [*Tsuchiya et al*., 2004a; *Wookey et al*., 2005a], a much smaller increase or decrease in *P*-wave velocity ($\delta \ln V_p$) [*Wentzcovich et al.*, 2006; *Wookey et al*., 2005a], a density increase of 1-2%, and a change in shear wave splitting between regimes of Pv and pPv [*Wookey et al*., 2005a; *Wentzcovich et al*., 2006]. These abrupt changes in elastic properties are expected to occur at greater depth in warmer regions and shallower depth in cooler regions due to the positive Clapeyron slope. The precise depth at which the Pv to pPv transition occurs has significant uncertainty due to issues involving pressure calibration for diamond cell experiments [*Hirose et al*., 2006], and, at face value, some estimates indicate that the Pv-pPv transition does not even occur within the mantle pressure range [e.g., *Shim et al*., 2004].

Compositional effects of iron (Fe) and aluminum (Al) have been explored experimentally and theoretically for $MgSiO_3$-$FeSiO_3$ and $MgSiO_3$-Al_2O_3, respectively [*Mao et al*., 2004; *Akber-Knutson et al*., 2005; *Tateno et al*., 2005; *Caracas and Cohen*, 2005; *Stackhouse et al*., 2005a, 2006; *Mao et al*., 2006; *Sinmyo et al*., 2006; *Tsuchiya and Tsuchiya*, 2006]. Lower mantle silicates probably contain 5-10% iron substitution for magnesium. Initial experiments indicated that the presence of iron in the Pv mineral reduces the pressure of the phase transition, such that it may occur hundreds of kilometers shallower in the mantle than for pure Mg-Pv [*Mao et al*., 2004]. These results have been contested by experiments that find less pressure effect upon Fe inclusion as a result of using more uniform pressure standards [*Hirose et al*., 2006]. The range in pressure estimates, taken at face value, is ±5 GPa [Hirose, 2006]. Theoretical predictions of the effects of Al substitution for both Mg and Si in the Pv crystal lattice suggest that there may be a significant depth range (> 10 GPa; a few hundred kilometers) over which Pv and pPv can coexist, thus distributing the otherwise abrupt seismic velocity change and weakening any seismic wave reflections from the Pv-pPv transition [*Akber-Knutson et al*., 2005; *Tateno et al*., 2005].

Experiments on more complex MORB and pyrolitic compositions have also commenced [*Murakami et al.*, 2005; *Ono and Oganov*, 2005; *Hirose et al*., 2005a; *Ohta et al*., 2006], with the Pv-pPv transition occurring over about 5 GPa range with a Clapeyron slope of about 8.6 MPa/K for both compositions. The pPv transition in MORB occurs at pressures 3-4 GPa lower than in pyrolite for the same temperature, and is accompanied by a phase transition in SiO_2 [*Ohta et al*., 2006], with both resulting in shear velocity reductions rather than increases due to the presence of Al and Fe [*Tsuchiya and Tsuchiya*, 2006]. Additional recent experimental and theoretical constraints on pPv properties are described elsewhere in this volume.

The properties of ferro-periclase (Mg,Fe)O are also important, especially the partitioning coefficient of iron between Pv and ferro-periclase [e.g., *Kobayashi et al*., 2005; *Murakami et al*., 2005; *Spera et al.*, 2006]. Recent experimental work [*Badro et al*., 2003, 2004; *Lin et al.*, 2005] indicates that at high pressure, Fe, in its high-spin state (for Fe^{3+} and Fe^{2+}) in most of the lower mantle, will prefer to be in a low-spin state in the lowermost mantle, which will favor iron partitioning into ferro-periclase rather than Pv or pPv. This Fe spin-transition may occur at depths similar to the pPv phase boundary, so iron partitioning may affect the pPv composition [*Sturhahn et al*., 2005]. Thermal, electrical and mechanical transport properties of lower mantle rocks will be strongly influenced by iron distribution, thus future work on realistic assemblages under high P-T conditions is very important for assessing the effects of the precise composition of any pPv in Earth.

3. SEISMOLOGICAL OBSERVABLES

The basic attributes of lowermost mantle structure that can be determined from seismic wave analysis and brought to bear on assessment of the possible presence of pPv are briefly considered here. Seismic wave travel time analysis can reveal three-dimensional *P*- and *S*-wave velocity variations over scales of several hundred kilometers or larger, average velocity gradients just above the CMB can be resolved over a several hundred kilometer depth range, and reflectivity associated with velocity discontinuities can be resolved for sharp velocity jumps of ~0.5% and larger (Figure 1). Frequency-dependence of reflections can reveal the depth range of a transition zone, but broadband signals over a suitable distance range are needed for this. The distribution of seismic sources and recording stations imposes severe limits on the lateral extent over which high resolution structural determinations can be made. Patterns of large-scale volumetric velocity heterogeneity have been mapped on a global scale, but fine-scale seismic stratigraphy can be resolved in only a handful of locations, as extensive stacking of signals from many events recorded by arrays of stations is required. Structural attributes that can only be resolved with good azimuthal sampling by many raypaths, such as full characterization of seismic anisotropy or discontinuity topography, are almost inaccessible; even the best-sampled regions have very restricted azimuthal coverage.

With these limitations kept in mind, the key diagnostic features that we will consider for assessing whether pPv occurs in the lowermost mantle (Figure 1) are: (1) abrupt increases or decreases in seismic velocity ($\delta \ln V_s$, $\delta \ln V_p$) that may correspond to Pv-pPv phase transition boundaries, (2) sharpness (depth-extent) of $\delta \ln V_s$ and $\delta \ln V_p$ contrasts that may reflect the pressure range of the two-phase domain, (3) relative $\delta \ln V_p$ and $\delta \ln V_s$ volumetric changes that may correspond to the predicted rigidity increase and incompressibility decrease for pPv when Al and Fe are not present, (4) topographic variations of seismic velocity contrasts that may be consistent with the strong phase boundary topography expected for a large positive Clapeyron slope Pv-pPV transition near a heterogeneous thermal boundary layer, (5) pairing of a velocity increase and an underlying decrease that could correspond to forward and reverse Pv-pPv transformations in a steep thermal gradient like that expected just above the CMB, (6) density contrasts and correlation with velocity changes that might indicate presence of the dense pPv phase, and (7) changes in anisotropic properties coupled to velocity changes that might correspond to differences in anisotropy for Pv and pPv. Detailed aspects of anisotropic observations and predictions for pPv are discussed in greater detail by Kendall and Wookey (this volume). Other seismological characteristics of the lowermost mantle, such as seismic attenuation, scattering from small-scale structure, and topography of the CMB will not be considered in detail since they provide no straightforward diagnostic or compatibility tests of the presence of pPv. As the seismological features are discussed, some consideration will be given to the underlying methodologies and the confidence levels that can be assigned to each result.

4. SEISMOLOGICAL OBSERVATIONS VERSUS PREDICTIONS FOR PPV

This section considers various seismological observations of deep mantle structure and whether they can be reconciled with the hypothesis that pPv is present in the lowermost mantle.

4.1. Abrupt Increases or Decreases in Seismic Velocity With Increasing Depth

Seismic velocity discontinuities in the mantle give rise to reflections and, for velocity increases, triplications that have critical angle amplifications of reflected and refracted energy. Seismic velocity discontinuities in the transition zone are usually interpreted as being caused by phase transitions, so it is natural to seek lower mantle velocity discontinuities as indicators of the Pv-pPv transition. *Wysession et al*. [1998] review seismic studies conducted prior to the discovery of pPv that indicate the presence of *P* and *S* wave velocity discontinuities several hundred kilometers above the CMB. Taken at face value, the reviewed studies suggest that in many regions there is at least one abrupt $\delta \ln V_p$ and/or $\delta \ln V_s$ increase (or decrease) of from 0.5 to 3% at depths from 100 to 300 km above the CMB, but with significant lateral variation in properties (Figure 2).

Before considering the details, it is important to recognize intrinsic limitations of the seismic observations. Detection of reflections from small deep mantle velocity increases is strongly dependent on distance range, with amplifications for triplications occurring at epicentral distances from 70 to 85°. At closer ranges, and at all ranges for velocity decreases, precritical reflections are expected, and these can be factors of 3 or so weaker than critical refractions; there is thus a strong likelihood of observational bias toward detection of velocity increases as a result of the critical angle amplifications [e.g., *Flores and Lay*, 2005]. Without careful modeling, data having amplitudes varying with range may easily be misinterpreted as requiring lateral variations when none are present. Inspection of individual waveforms for reflected arrivals, which will occur as precursors to *PcP* and *ScS* reflections from the CMB, must allow for signal-generated-noise in the coda of direct *P* and *S*, originating near the source and near the receiver. Together with typical ambient noise levels, this makes it difficult to detect individual arrivals, even for triplications, when velocity contrasts less than ~1% are involved. The most reliable detections of arrivals are made when seismic arrays or networks are used to stack the arrivals for variable travel-time move-out, establishing the ray parameter and back-azimuth of the extra arrivals. Stacking methods also improve signal-to-noise ratios, allowing sharp velocity increases as small as ~0.5% to be detected [e.g., *Kruger et al*.,

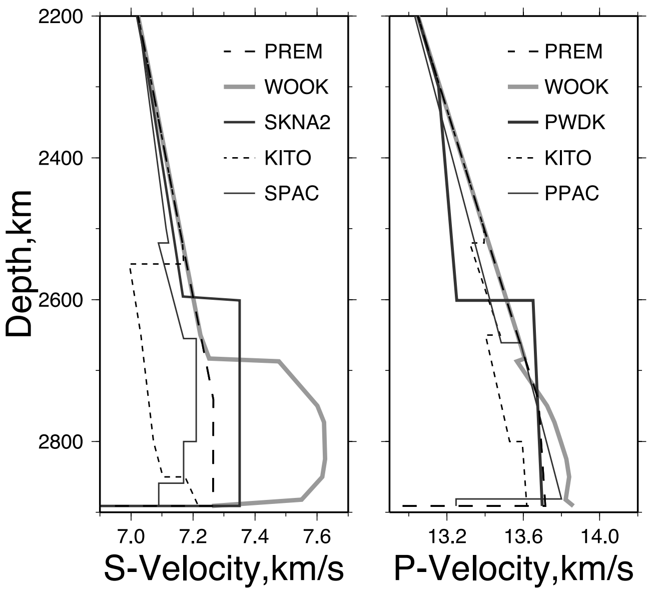

Figure 2. *S*-wave and *P*-wave velocity models for the lowermost mantle. PREM is the global average reference model from *Dziewonski and Anderson* [1981], WOOK is model based on theoretical elasticity calculations for high P-T post-perovskite [*Wookey et al.* [2005a] adjusted by us relative to PREM, SKNA2 is for the region below the Caribbean [*Kendall and Nangini*, 1996], PWDK is for the region below northern Siberia [*Weber and Davis*, 1990], KITO is for the western portion of the Pacific LLSVP [*Kito et al*., 2004], SPAC is for the central Pacific (region B of *Lay et al*. [2006]) and PPAC is for the central Pacific [*Russell et al.*, 2001]. These models are representative of the range of seismic velocity structures inferred from recent studies, and details are discussed in the text.

1995; *Reasoner and Revenaugh*, 1999; *Avants et al.*, 2006b]. But even in this case, source radiation nodes and nulls in reflection coefficients between pre-critical and post-critical arrivals of opposite sign must be recognized. Thus, the lack of observed reflections does not necessarily preclude the presence of a discontinuity. Absence of high frequency pre-critical reflected energy may be due to existence of a transition zone rather than a sharp discontinuity, so this is also neutral with respect to presence of phase boundary structure. Sharpness of velocity gradients can only be resolved by having broad bandwidth, including high frequency energy, over a diagnostic range of distances. The weakness of pre-critical reflections makes near-vertical reflections very difficult to detect, so effects of angle of incidence must always be considered. Sensitivity of the data to structure must be understood; wide-angle reflections are sensitive to the velocity contrasts for the associated wave type, but not to density contrast, whereas pre-critical reflections are sensitive to impedance contrasts.

4.1.1. P wave discontinuity observations. The initial indications that a velocity discontinuity exists in D″ were based on rather subtle *P* wave slowness analyses [e.g., *Wright*, 1973; *Wright and Lyons*, 1975, 1979, 1981] that did not compel broad acceptance; the nature of the *P* velocity structure remains enigmatic. Ironically, this may be a factor compatible with pPv occurrence, as discussed below. While *P* wave data are more abundant than *S* wave data, and less signal processing is needed for their analysis, there are two key observational challenges specific to *P* waves: (1) *PcP* is usually a very weak and difficult to observe arrival (especially compared to *ScS* at lower mantle triplication distances), so it is less useful as a reference phase to control precursor timing and amplitude for mapping of D″ reflectors relative to the CMB and less useful for assessing whether the *P*-wave energy has favorable seismic radiation directed at the deep mantle, and (2) the relatively high D″ velocities of *P* waves (compared to *S* waves) cause lower mantle phases to arrive close together in time (<10 s), making them less readily isolated as discrete arrivals, particularly in broadband data. At large distances where triplication arrivals for lowermost mantle velocity increases are strong, the lack of clear *PcP* energy requires that only the direct *P* arrival be used as a reference phase even though it may have a turning point 500 to 700 km above the CMB. This is less desirable than having two reference phases (as in the case of *S* and *ScS*), which allows the effects of volumetric heterogeneity to be better accounted for, suppressing bias in the estimates of discontinuity depths. Despite these issues, there are some high quality data that provide confident detection of D″ *P* velocity discontinuities.

Perhaps the most compelling observations favoring the existence of a lower mantle *P* velocity discontinuity are the array observations for paths beneath northwest Siberia and the Kara Sea in *Davis and Weber* [1990], *Weber and Davis* [1990], *Weber* [1993], *Houard and Nataf* [1992; 1993], *Thomas and Weber* [1997], *Freybourger et al*. [1999], and *Thomas et al.* [2002]. These studies show clearly visible secondary arrivals 3 to 5 s after *P* in array beams, individual seismograms, and vespagrams. Modeling provides fairly robust estimates of a positive velocity increase with $\delta \ln V_p$ = 2.3-3%, 250-316 km above the CMB in this region. The arrivals' polarity clearly requires a velocity increase and systematically changing phase-shifts indicate that they are post-critical reflections. Model PWDK [*Weber and Davis*, 1990] in Figure 2 is representative of the 1-D models produced for the region under Northern Siberia. The mantle below the Nansen Basin in the Arctic north of Russia also yields relatively consistent observations of $\delta \ln V_p$ = 2-3% increases at depths from 139 to 389 km above the CMB [*Weber and Davis*, 1990; *Houard and Nataf*, 1992; *Weber*, 1993; *Krüger et al.*, 1995; *Thomas and Weber*, 1997;

Scherbaum et al., 1997]. This swath from northern Eurasia to the North Pole stands out as the most-consistent, and best-studied region for lower mantle *P* wave reflectivity [*Wysession et al.*, 1998]. However, the *P* velocity structure found in this region is certainly not globally representative.

Precursors to *PcP* have been detected intermittently in other regions, such as under the western Pacific where there are varying reports of a $\delta \ln V_p$ = 1-2% increase 200-300 km above the CMB [*Shibutani et al.*, 1993, 1995], a laterally varying discontinuity from 170 to 280 km above the CMB [*Yamada and Nakanishi*, 1998], a $\delta \ln V_p$ < -1% decrease 239 to 369 km above the CMB [*Kito and Krüger*, 2001; *Kito et al.*, 2004] and $\delta \ln V_p$ = 0.5% increase 89 km above the CMB [*Kito et al.*, 2004]. While advanced double array stacking and migration methods were employed in these studies, the raw observations appear to have much greater scatter in waveforms and weaker arrivals compared to the data used in constructing PWDK. Subtle or weak reflections in scattered data are less resolved, and warrant follow-up work (especially considering that studies beneath the western Pacific utilized data in the distance range 65°-74°, which lacks the strong amplitudes found at post-critical distances). These models (e.g., KITO in Figure 2) are very different than for those derived from high quality data that samples under Eurasia, and have much more uncertainty.

The region under the central Pacific has also been studied using array analysis, with *Kohler et al.* [1997] reporting $\delta \ln V_p$ = 1-2% increase 140 km above the CMB, *Reasoner and Revenaugh* [1999] estimating a $\delta \ln V_p$ = 0.4-0.55% 190 km above the CMB, and *Russell et al.* [2001] finding $\delta \ln V_p$ = 0.75% 230 km above the CMB. These studies sample the same D″ region, and the differences in estimates of the discontinuity depth are indicative of uncertainties arising due to differences in modeling procedure and reference model choice that plague all seismic models; however, it appears that any $\delta \ln V_p$ increase is less than ~1% in this region. This is important, as such a weak velocity jump (increase or decrease) will not produce secondary arrivals likely to be picked in routine seismic bulletin data based on isolated waveforms. Thus, efforts to use bulletins as indicators of presence or absence of a *P* discontinuity [e.g., *Weber and Kornig*, 1992], are, in fact, useful only as reconnaissance indicators of presence of exceptionally large reflections such as under northern Eurasia, and not reliable for the very weak V_p discontinuities reported in other locations. Only a few other regions have been reported to have such large *P* velocity increases, for example, *Wright et al.* [1985] suggest $\delta \ln V_p$ = 2.8% 160 km above the CMB beneath southeast Asia. In other regions, detected arrivals are significantly weaker than expected for such a strong contrast, as is the case below the northwestern Pacific [*Thomas et al.*, 2002], below the Arctic where *Vidale and Benz* [1993] estimate $\delta \ln V_p$ = 1.5% 130 km above the CMB, below the Indian Ocean where a 0.2-0.4% discontinuity is proposed [*Wright and Kuo*, 2006], and below the Cocos plate where an upper bound of 1% on any *P* velocity increase is stated based on lack of any observed arrivals [*Ding and Helmberger*, 1997] and a 0.5-0.6% *P* velocity increase is estimated based on stacked signals [*Reasoner and Revenaugh*, 1999].

The significance of reported "non-observations" of *P* reflections, as compiled in Plate 1 of *Wysession et al.* [1998] is very difficult to judge; our judgement is to accept only inferences from carefully stacked high quality data that are shown to have detected *PcP* arrivals (to ensure suitable radiation pattern). We find in work in progress that reanalysis of some data sets using *PcP* detectability as a data selection criterion significantly changes even extensive stacking results like those of *Reasoner and Revenaugh* [1999], so one must be carefully skeptical and critical of past literature addressing *P* discontinuities. As discussed in *Wysession et al.* [1998], array analyses sometimes indicate out-of-plane scattering for *PcP* precursors, which suggests that the extra arrivals are caused by scattering from either volumetric heterogeneities or an undulating discontinuity. In either case, estimates of the velocity contrasts and discontinuity depth can be in error if the scattering is not correctly modeled. Modeling data with a simple velocity discontinuity may not be justified. *Weber* [1994] suggests that reflections from a $\delta \ln V_p$ = 3% high velocity lamella 27 km thick could match data as well as model PWDK, while *Thomas et al.* [1998] model some reflections with an $\delta \ln V_p$ = -10% low velocity lamella 8 km thick. This non-uniqueness results from the difficulty of modeling very small arrivals in *P* wave coda for paths that are always thousands of kilometers long through a heterogeneous mantle, with limited geographic sampling of isolated corridors and little or no crossing raypath coverage.

We use the reflectivity method to compute *P* wave synthetic seismograms for standard Earth model PREM [*Dziewonski and Anderson*, 1981], model WOOK based on the isotropic version of theoretical elasticity for the Mg-Pv to pPv transition from *Wookey et al.* [2005a], model PWDK for observations sampling below Northern Siberia [*Weber and Davis*, 1990], model KITO for observations sampling below the western Pacific [*Kito et al.*, 2004], and model PPAC for observations sampling below the central Pacific [*Russell et al.*, 2001] (Figure 2). The synthetics are uniformly processed, and include periods down to 2 s.

Distance profiles for models PWDK and WOOK are shown in Figure 3, aligned in time on *PcP* arrivals and amplitude-normalized on direct *P*. Isotropic radiation is assumed, so the relative amplitudes within a given trace are the result of differential geometric spreading and reflection coefficients between the phases. The distance range is 65-85°, which spans the triplication range for any positive lower

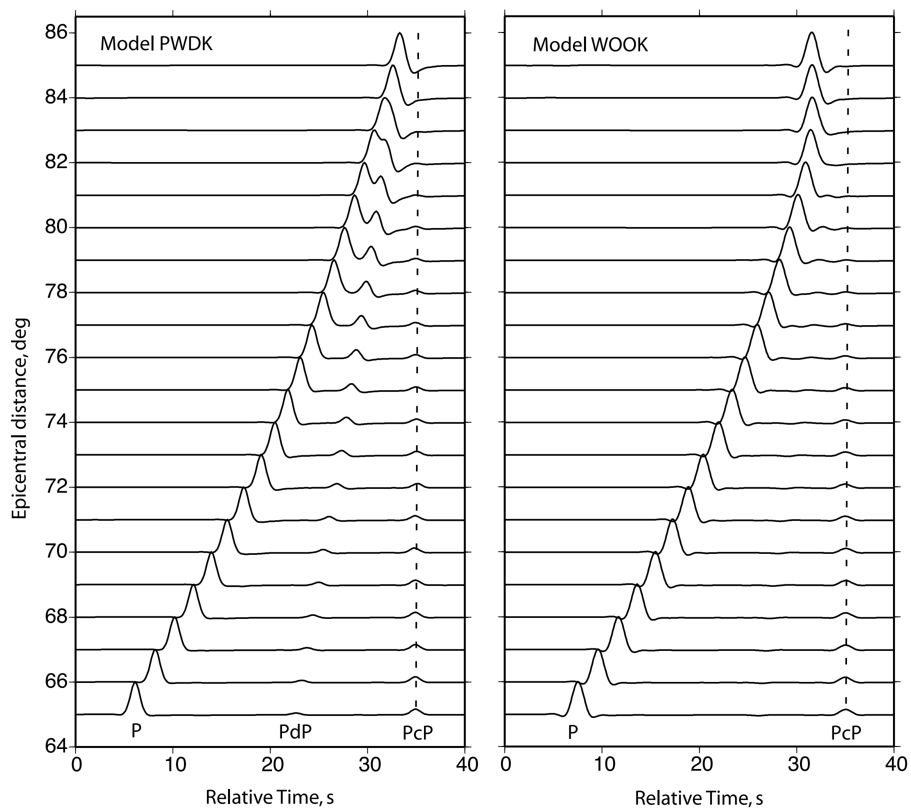

Figure 3. Profiles of reflectivity displacement synthetics for vertical component *P* waves for models PWDK and WOOK in Figure 2, with the traces aligned on the *PcP* arrival for a 400 km deep isotropic source with an impulsive source. A Butterworth low-pass filter with a corner at 2 s has been applied. Crustal reverberations at the receiver were suppressed to keep the signals as easy to interpret as possible. PdP is the generic label for the arrivals produced by interaction with a discontinuity in the deep mantle. For PWDK the 3% jump in V_p produces a triplication, with large amplitudes for the *PdP* phase at post-critical distances beyond about 75°. Model WOOK produces little reflectivity; the isotropic version of the Pv-pPv phase change does not predict strong *P* arrivals from the phase boundary.

mantle velocity discontinuities. Figure 3 demonstrates that *PcP* is expected to be a weak arrival in this range, and it is seldom observed above the typical noise levels on individual traces. This motivates stacking procedures for detection of even the core reflection. Model PWDK has a triplication arrival, labeled *PdP*, which arrives between *P* and *PcP*, with critical-angle amplification causing it to be a stronger arrival than *PcP* beyond 72°. Casual inspection of waveforms at closer distances will likely fail to detect the *PdP* phase, even for this case of noise-free synthetics for a 3% velocity increase, so, again, stacking procedures are imperative. From about 76° to 82°, the triplication arrivals for model PWDK are large enough to observe in individual waveforms, as found in the data for northern Eurasia. Model WOOK has weak reflectivity, so the isotropic version of a Mg-Pv to pPv phase boundary may be undetectable in raw waveforms at all distances.

Comparisons of synthetics for all of the *P* wave models in Figure 2 at a distance of 78° are shown in Figure 4. The peaks of arrivals in the synthetics associated with first-order discontinuities in the various velocity models are indicated by arrows with corresponding polarity. Model WOOK predicts a small negative arrival of very low amplitude after *P*, which is in strong contrast to the large positive arrival for the large positive velocity increase in model PWDK, or even the much weaker increase in model PPAC. Model KITO, for the western Pacific does have a small velocity decrease that could be consistent with model WOOK, but the overall velocity structure is low for model KITO, so *PcP* arrives relatively late. It is also clear that the features in model KITO are very subtle; with arrivals smaller than the weak *PcP* arrival being predicted. With real data typically having noise larger than the *PcP* amplitudes in any one trace, some form of stacking is essential to have any confidence in the small features and even then the interpretation involves faith in the degree of signal-to-noise enhancement.

Figure 5 shows the results of double-array stacking applied to synthetics for the distance range 65-74° for each model.

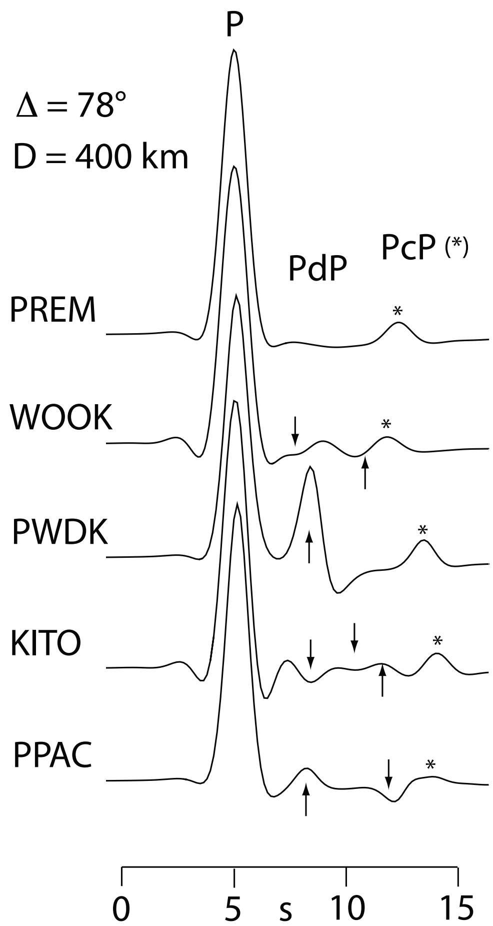

Figure 4. Reflectivity displacement synthetics at 78° epicentral distances (Δ) for vertical component *P* waves for the *P*-wave models in Figure 2, for an impulsive isotropic source at a depth (D) of 400 km. A Butterworth low-pass filter with a corner at 2 s has been applied. Crustal reverberations at the receiver were suppressed to keep the signals as easy to interpret as possible. The traces are aligned on direct *P*, and the *PcP* reflection from the CMB is indicated by an asterisk for each trace. *PdP* arrivals associated with velocity discontinuities in the models are indicated with arrows, with upward arrows corresponding to energy from velocity increases with depth and downward arrows corresponding to energy from velocity decreases with depth. The *PcP-P* time varies between the traces due to the differences in average V_p in the lowermost mantle for the models. The *PdP* arrival for PWDK has an upward pulse followed by negative overshoot; the latter is a result of phase shift of the post-critical reflection from the positive velocity discontinuity in the model. The upward pulse involves both reflected energy from the discontinuity and energy diving below the discontinuity and turning within the D″ region.

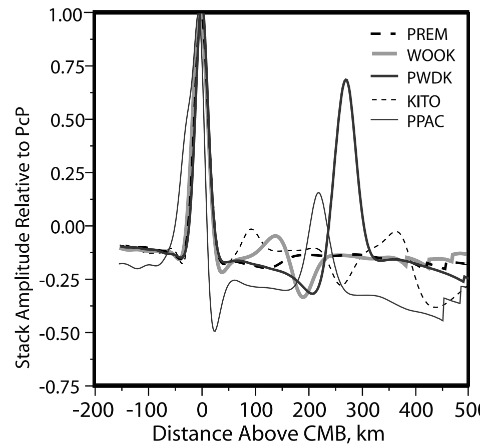

Figure 5. Double-array stacks for synthetics in the distance range 65-75° for each of the *P* wave models in Figure 2. The synthetics have the same parameters as described in Figures 3 and 4 for each model. The stacks represent the peak energy in a stack of traces shifted for predicted move-out of reflections from a target depth relative to a reference phase (in this case, *PcP*) for a reference velocity model (in this case, PREM). Thus, the amplitude of the stack at each depth relative to the CMB (where *PcP* arrivals are perfectly aligned and give a stack amplitude of unity) indicates the strength and polarity of reflections from various depths. The *PdP/PcP* amplitude ratio differs from that in the individual seismograms in Figure 4 due to the range of distances involved in the stacks. The inferred depth will be biased by the extent to which the true model velocities differ from those of the assumed reference model. With careful alignment and stacking of arrivals over at least a several degree distance aperture, small arrivals can be resolved. For real data, the signal-to-noise ratio for *PcP* itself is often very low at these distances, so it is very hard to detect arrivals from sharp discontinuities with less than 0.5% increase, or even larger (1%) decrease.

Double-array stacking involves shifting and summing traces in accord with predicted relative arrival times for a specified 1D reference velocity model, assuming reflections at various target depths in the lower mantle. With our traces initially being aligned on *PcP* (not usually possible for actual data), the depths are given relative to the CMB. This stacking provides images of reflectivity in the lower mantle that correspond to the depths and strength of discontinuities in the seismic models, although the depths are biased by up to tens of kilometers by use of a reference model for the stacking (here we use PREM) that departs from the true structure. Even small features, as in model WOOK, can potentially be detected by stacking, as long as *PcP* is well-resolved. In practice, *PcP* is usually observed with a signal-to-noise ratio of no more than about 5-10 (and often only 2-3), so it is intrinsically very difficult to resolve features weaker than the 0.75% positive discontinuity in model PPAC.

Overall, there appears to be little basis at the present time for confidence in associating published *P* wave discontinuity models with the pPv phase boundary. Qualitatively, the weak expected *P* wave reflectivity for pPv is consistent with the limited extent over which strong *P* reflections are observed.

However, the only way to reconcile the phase boundary with the full range of *P* wave observations is to invoke variable anisotropic properties, such that for some regions there is a strong *P* velocity increase between the Pv above the phase boundary and the pPv below it, while in other regions, the contrast is weak, or even negative. Acquiring such lattice-preferred-orientation (LPO) anisotropy requires favorable patterns of deformation within the boundary layer. The accompanying article by *Kendall and Wookey* [this volume] considers the anisotropic properties in greater detail. Here, it suffices to say that it will be difficult to make a compelling argument for the presence of post-perovskite in the mantle based on *P* wave reflectivity observations alone unless one can characterize the full anisotropic properties of a given region of the lower mantle. At best, one can make some general consistency arguments that pit the pPv hypothesis against other possible explanations such as chemical heterogeneity.

4.1.2. S wave discontinuity observations. General acceptance of the existence of a lower mantle D″ discontinuity first emerged based on *S* wave observations, which prove to have more extensively observable triplication arrivals. *Lay and Helmberger* [1983a] first showed that individual seismograms had clear arrivals between *S* and *ScS* on transverse components that shift systematically with distance and source depth in a fashion that requires a lower mantle structure. Modeling suggested the presence of a 2.75% shear velocity increase from 250-320 km above the CMB beneath Alaska, the Caribbean, and Eurasia. Subsequent work indicated similar structure beneath India [*Young and Lay*, 1987], and extended the models for the initial study regions [e.g., *Young and Lay*, 1990; *Weber and Davis*, 1990; *Gaherty and Lay*, 1992; *Weber*, 1993; *Kendall and Shearer*, 1994; *Kendall and Nangini*, 1996; *Lay and Young*, 1996; *Lay et al*., 1997; *Ding and Helmberger*, 1997; *Bilek and Lay*, 1998; *Garnero and Lay*, 2003; *Lay et al*., 2004a; *Thomas et al*., 2004a,b; *Wang et al*., 2006; *Chambers and Woodhouse*, 2006a,b; *van der Hilst et al*., 2007]. *Wysession et al*. [1998] tabulate basic properties of the *S* wave discontinuity models from many of these studies, but the basic style of structure is represented by the Caribbean region model SKNA2 [*Kendall and Nangini*, 1996] in Figure 2, which has a 2.5% shear velocity increase about 290 km above the CMB. This type of velocity model produces a triplication in the *S* wave field at distances beyond 70° similar to that seen for model PWDK in Figure 3, with the difference that *ScS* is a strong arrival out to about 80°, typically half the amplitude of direct *S*. Analysis of diffracted waves has yielded models with strong SH discontinuities of up to 4.6% under northeastern Siberia and 3.4% under the east central Pacific [*Valenzuela and Wysession*, 1998], although resolution provided by diffracted waves is limited due to the signals having long propagation paths in structure that appears to vary over shorter scales; thus it is difficult to assess or define resolution of such strong jumps mapped by diffracted waves, especially lacking evidence for triplication arrivals of such strength in the same region.

It is important to note that strong *S* velocity increases can be found both in regions of strong *P* velocity increases, as is the case for northern Siberia, and in regions of weak, or no *P* velocity increase, as under the Cocos Plate or Alaska. *Weber* [1993] finds that under the Nansen Basin under the Arctic there is a strong *P* reflector with no apparent *S* reflector. The estimated depths of the *P* and *S* discontinuities in a given region can be consistent, as under northern Siberia, or different by as much as 50 km, but little significance should be attached to this unless a specific effort has been made to calibrate the regional background *P* and *S* velocity structure.

It appears that a model like SKNA2 does not apply everywhere, as some areas appear to lack strong triplication arrivals [e.g., *Schlittenhardt et al*., 1985], and regions that do have discontinuities sometimes appear to grade into regions without a strong reflector [e.g., *Kendall and Nangini*, 1996; *Garnero and Lay*, 2003; *Chambers and Woodhouse*, 2006a,b]. One of the most intensely studied regions is below the central Pacific, where a diverse array of laterally varying discontinuity models has been proposed [e.g., *Garnero et al.*, 1988; *Revenaugh and Jordan*, 1991; *Garnero et al*., 1993a; *Russell et al*., 2001; *Avants et al.*, 2006a,b]. The variation in the models is itself suggestive of the heterogeneity in the region, and no single 1D model is truly representative. The most extensive localized seismogram stacking analysis published to date [*Lay et al*., 2006], yields models like SPAC in Figure 2, which involve small (~1%) increases several hundred kilometers above the CMB, with additional shallower and deeper velocity decreases. The overall velocity in the lower 300 km is low relative to PREM, as this region is part of the large low shear velocity province (LLSVP) in the deep mantle beneath the Pacific. The velocity increase appears to be significantly smaller than for model SKNA2, but estimation of this value has large uncertainties due to the presence of poorly constrained three-dimensional structure. The data for this region intermittently show clear secondary arrivals, but stacking is required to achieve sufficient signal-to-noise ratio for confident identification. This raises questions about whether regions lacking clear triplication arrivals require vanishing of the discontinuity, significant short scale topographical variations that disrupt coherency of the reflected wavefield, or simply reduction to the 1% level below which individual arrivals are not identifiable.

The western region of the Pacific LLSVP is represented by model KITO (Figure 2) from *Kito et al*. [2004], with the shear velocity having a sharp drop of 2.5% 239 km above the CMB. This is not at the same depth as either of the two drops in the regional *P* velocity structure, but this may not be significant. *Russell et al.* [2001] were able to find compatible depths for *P* and *S* velocity discontinuities under the central Pacific that had been separately modeled at depths 40 km apart, simply by modifying the reference structures. The LLSVP beneath southern Africa, the South Atlantic and the southern Indian Ocean has also been modeled as having a sharp drop of from 1 to 3% a few hundred kilometers above the CMB [e.g., *Wen et al*., 2001; *Wen*, 2001; *Wang and Wen*, 2004, 2006a; *Ni and Helmberger*, 2003a,b; *Ni et al*., 2002; *Ni et al.*, 2005], although detailed stacking and resolution of the negative discontinuity has not been performed in this region. The northwestern Indian Ocean has relatively high lowermost mantle shear velocities, and a SH reflection from the top of D″ is observed in this region [*Sun et al*., 2007]. In the western Pacific, the transition from the LLSVP to surrounding mantle may be accompanied by a change in the sign of the D″ *S* wave discontinuity, as *He et al*. [2006] report a positive discontinuity westward of the LLSVP and *Kendall and Shearer* [1994, 1995] report detections of isolated arrivals in the coda of *SH* that may be due to reflections from D″ over a broad region of southeast Asia. We feel that little confidence can be placed in individual observations that cannot establish the systematic travel time behavior expected for a lower mantle phase.

We consider the range of *S* velocity models included in Figure 2 to be representative of the diversity of structures obtained by detailed seismological studies, and compare them with model WOOK, predicted for the Mg-Pv to pPv phase boundary by *Wookey et al.* [2005a]. The high T theoretical calculations predict a large, 4%, dlnVs and associated high shear velocities throughout D″ down to the base of the mantle where post-perovskite converts back to perovskite in a steep thermal gradient just above the CMB. Such a double-crossing of the phase boundary was postulated by *Hernlund et al*. [2005], and is discussed further below. Synthetic *SH* seismograms for the *S* wave models in Figure 2 are shown in Figure 6. Synthetics are shown with two lowpass corner frequencies; 0.12 Hz and 0.3 Hz. Typical deep focus earthquakes of magnitudes 6.0-6.3 produce broadband teleseismic signals with dominant frequencies roughly matched by the 0.12 Hz filter. It is clear that for some models this passband limits observability of secondary arrivals produced by *S* wave discontinuities, and this limitation increases for larger distances. Bandwidth extension procedures, such as source wavelet deconvolution [e.g., *Lay et al*., 2004a; *Avants et al.*, 2006b] can extend the bandwidth to about 0.3 Hz, improving the temporal isolation of secondary arrivals, but at the cost of adding some deconvolution noise and slight Gibbs effect. The 0.3 Hz filtered synthetics are thus representative of the most optimistic potential resolution of carefully processed shear wave signals.

The various models based on seismic data have about 10 s fluctuations in *ScS-S* differential times, consistent with global observations, but the large amplitude of *ScS* allows its ready identification in all cases. The synthetics for model SKNA2 have a strong *SdS* triplication phase, but this is best observed when bandwidth extends to 0.3 Hz, because of the short *SdS-S* differential time. For many observations in the strong triplication range from 78-82°, raw broadband data do not reveal the secondary arrival clearly, and one observes only a broadening of the *S* pulse relative to that for *ScS*. Source wavelet deconvolution is then essential for revealing what may be strong secondary arrivals [e.g., *Lay et al*., 2004a], and deconvolution always has attendant concerns. The velocity decrease in model KITO is as strong as the velocity increase in SKNA2, but the associated *SdS* phase is less than one third as large; this represents the amplification that favors observations of triplicated arrivals. Again, the arrival is obscured by direct *S* in the 0.12 Hz synthetics (and these synthetics lack receiver crust so the coda is minimal). For model SPAC, the weak positive discontinuity (0.6%) produces an *SdS* arrival only 25% as large as *ScS*, and even weaker arrivals are produced by the comparable velocity drops. This represents the lowest plausible level of discontinuity detection, and only with hundreds of source-wavelet deconvolved seismograms being stacked is there any chance of reliably detecting such small features [*Avants et al.*, 2006a; *Lay et al*., 2006].

The predicted waveforms for model WOOK show the expected large *SdS* triplication arrival, which exceeds *ScS* in amplitude. *ScS* is very early for this model due to the high velocities, and it is fair to say that no data have been observed with such short *ScS-S* times at this distance. The triplication arrival is similar to that for SKNA2, which is representative of many regions, as noted above. While the thermo-elastic calculations of *Wookey et al*. [2005a] give too large of a velocity increase and too high of absolute velocities in D″ to match any data, it is the prediction of a strong *S* reflection from the phase boundary that provides a viable explanation for the triplicated arrival in SKNA2. In this case, anisotropic properties of the phase boundary contrast can be invoked to modulate the increase, while chemical variations can weaken the velocity contrast or distribute it over a less-reflective transition zone. This situation is quite different from that for *P* waves, where the largest observed arrivals are not predicted by the isotropic average of the phase boundary model. *Ohta et al*. [2006] note that Pv-pPv transition in MORB material can involve a shear velocity decrease, so one possible explanation for the KITO type model for an LLSVP is that chemical differences of the dense pile bring about a negative amplitude *S*-wave reflection from a Pv-pPv phase boundary. This suggests that assessing the presence of pPv in the lower mantle will require a full mapping of compositional effects

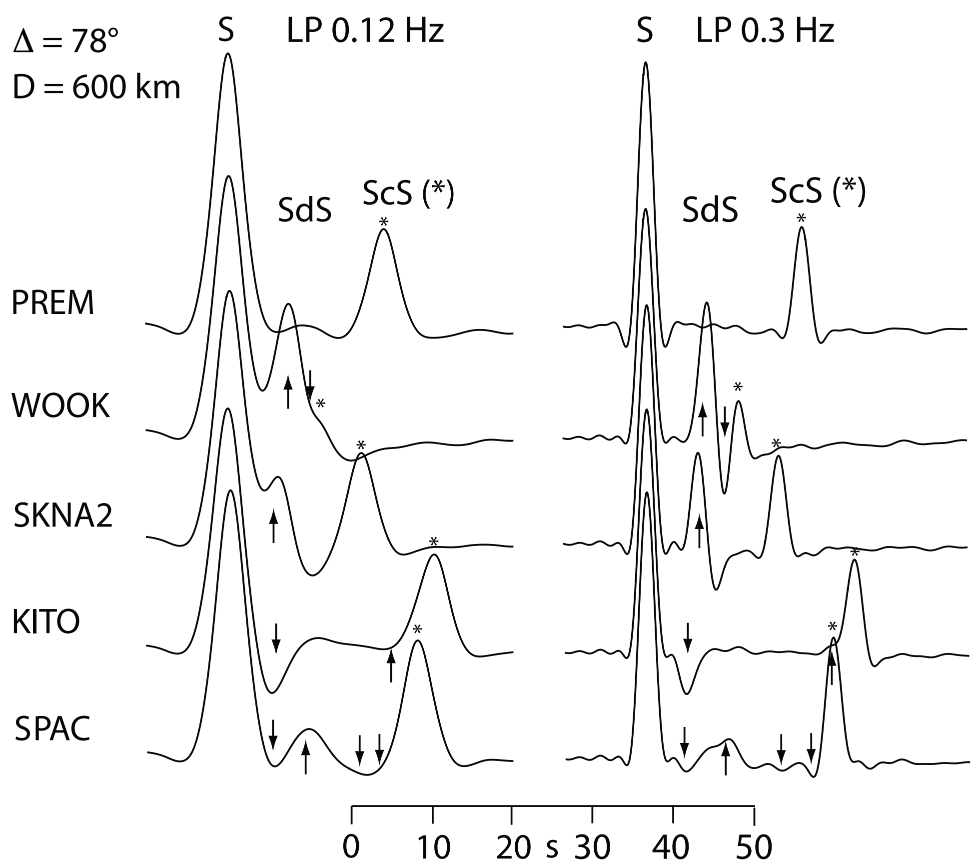

Figure 6. Reflectivity displacement synthetics at 78° epicentral distances (Δ) for transverse component *SH* waves for the *S*-wave models in Figure 2, for an impulsive shear source at a depth (D) of 600 km. A Butterworth low-pass filter with a corner at 0.12 Hz has been applied to the traces on the left, and a low-pass filter with a corner at 0.3 Hz has been applied on the right. Crustal reverberations at the receiver were suppressed to keep the signals as easy to interpret as possible. The traces are aligned on direct *S*, and the *ScS* reflection from the CMB is indicated by an asterisk for each trace. *ScS* arrivals associated with velocity discontinuities in the models are indicated with arrows, with upward arrows corresponding to energy from velocity increases with depth and downward arrows corresponding to energy from velocity decreases with depth. The *ScS-S* time varies between the traces due to the differences in average V_s in the lowermost mantle for the models. The *SdS* arrival for SKNA2 has an upward pulse followed by negative overshoot; the latter is a result of phase shift of the post-critical reflection from the positive velocity discontinuity in the model. The upward pulse involves both reflected energy from the discontinuity and energy diving below the discontinuity and turning within the D″ region.

on shear velocity reflectivity. The associated challenge is that the compositional effects may themselves provide reflectivity without the presence of a phase change, so this is an exceedingly difficult issue to resolve. Overall, the shear velocity increase in model SKNA2 and similar structures represents a plausible manifestation of pPv, and this will be considered in relationship to volumetric heterogeneity in D″ below.

4.1.3. ULVZ observations. The strongest velocity discontinuities in the lower mantle are associated with large velocity drops in the intermittent ultra-low velocity zones (ULVZs) found just above the CMB [e.g., *Garnero et al., 1993*b; *Mori and Helmberger*, 1995; *Garnero and Helmberger*, 1995, 1998; *Revenaugh and Meyer*, 1997; *Garnero et al.*, 1998; *Wen and Helmberger*, 1998; *Rost and Revenaugh*, 2003; *Rondenay et al*., 2003; *Thorne and Garnero*, 2004; *Rost et al*., 2005, 2006a,b]. ULVZ structures involve thin (<40 km) layers or mounds with dlnVp and dlnVs of -5% or more. While this strong velocity reduction is generally attributed to partial melting, possibly of locally chemically distinctive material right at the base of the mantle [e.g., *Williams and Garnero*, 1996; *Lay et al*., 2004b; *Rost et al*., 2005, 2006a], a possible connection to the presence of pPv has been raised by *Mao et al*. [2006]. This is the experimentally-based suggestion that large amounts of iron can be accommodated in the pPv structure, and as the Fe enrichment grows, very strong subsolidus reductions in elastic velocities can result. Thus, a thin layer of very Fe-rich pPv could possibly account for ULVZs. If plumes rise from ULVZs, they could then bring up Fe-enriched material. Penetration of iron into the lowermost mantle in regions of CMB topography has been invoked as one way to enhance Fe content of the base of the mantle, but chemical reactions at the core-mantle interactions tend to lead to depletion of iron on the mantle side [e.g., *Sakai et al*., 2006]. As a test of the presence of pPv, this is problematic, in that partial melting is hard to preclude as an alternative, and the stability of pPv in the hottest part of the mantle (right above the CMB) is highly questionable. The experiments supporting large iron accommodation by pPv have been for temperatures well-below the expected CMB temperatures, which may be 1000° hotter [e.g., *Williams*, 1998] than the 2500K range associated with the Pv-pPv phase transition a few hundred kilometers above the CMB. Issues of partitioning of Fe between pPv and ferropericlase also need to be considered under appropriate conditions. At this time, the occurrence of ULVZs does not provide clear evidence favoring the presence of pPv in the deep mantle.

4.2. Sharpness (Depth-Extent) of D″ Velocity Contrasts

Compositional contrasts tend to be very abrupt, and give rise to velocity discontinuities like the Moho and the CMB, whereas phase changes tend to occur over a finite pressure range, distributing the velocity change over a transition zone across which there is varying proportion of the two phases. Thus, establishing the depth-extent of a D″ velocity contrast (either positive or negative) would provide one test of the phase transition explanation. Unfortunately, for lower mantle discontinuities this is observationally challenging, even in the regions with large velocity contrasts, and definitive results are not presently available. For a velocity increase, as the phase transition zone thickness increases relative to the observing wavelength, the range over which triplication arrivals are strong narrows, but the peak amplitudes (at distances beyond ~78°) are unaffected. This means that the sensitivity to sharpness is unfortunately at closer, pre-critical ranges where the expected reflection amplitudes are small to begin with. Distinguishing between a sharp jump and a distributed transition zone requires determination of the amplitude of reflections as a function of range for varying frequencies. Individual observations are too variable to do this reliably, and combining different events presents difficulties, particularly for stacked signals, as needed for quantifying weak arrivals. At best, upper bounds on plausible transition zone thicknesses compatible with seismic observations have been explored. This should not be misinterpreted as favoring the existence of sharp boundaries; this is usually just a default modeling assumption in seismic studies. If one begins with smooth tomographic models, the question can be alternatively approached by finding the smoothest models that account for observed waveform features [e.g., *Liu et al*., 1998]. But, in almost all cases either the gradients or the overall strengths of heterogeneity in tomographic models need to be increased significantly to produce triplication-like features that can match the data [e.g., *Ni et al*., 2000].

Lay and Helmberger [1983a] first explored this issue by modeling short-period and long-period *SH* observations, finding that the broadband data could be modeled with transition zone thicknesses of up to 50 km, but a sharp discontinuity could do as well. Long-period *SH* triplication data alone cannot constrain the width of the transition zone, as even a 100 km thick transition of several percent can match the wide-angle data [e.g., *Young and Lay*, 1987; *Gaherty and Lay*, 1992; *Garnero et al*., 1993a]. Modeling broadband *SH* triplication signals under Alaska, *Lay and Young* [1990] found good fits only for transition zone thicknesses less than 30 km. *Weber* [1993] and *Weber et al*. [1996] explored the effect of variable thickness of transition zones for regions with strong *P* and *S* wave velocity increases, finding an upper bound on the thickness of ~75 km. *Revenaugh and Jordan* [1991] analyzed long-period multiple *ScS* reflections, and suggest that a transition zone thickness of less than 50 km is required. The small velocity contrasts of lower mantle structures make near vertical reflections very weak, but there are some reports of *P* and *SH* reflections that suggest transition zone thicknesses of 8 to <50 km [*Schimmel and Paulssen*, 1996; *Wysession et al*., 1998], but the observations do not appear very robust relative to wide-angle observations. This is a topic worthy of further seismological effort, as the extent of any transition zone thickness may constrain the possible presence of Al, which tends to increase the width of the two-phase regime.

4.3. Volumetric P-wave and S-wave Velocity Changes

While observation of reflections from a velocity contrast provide the most direct evidence for a phase transition being present, relevant observations for the lowermost mantle are spatially restricted, as discussed above. However, the volumetric velocity effect associated with a phase transition may be more extensively detected by travel time tomography, and relative *P* and *S* wave velocity behavior can be tested against the predictions of a phase transition even when reflectivity is not resolvable. Any interpretation of volumetric velocities must evaluate contributions from thermal and chemical heterogeneity. The predicted elastic properties for $MgSiO_3$ are that V_s should be several percent faster for pPv than for Pv, while V_p should be relatively unchanged. Thus, the seismic tomography expression that would result is high $\delta \ln V_s/\delta \ln V_p$ ratios in regions containing pPv. The pPv layer should be thicker in relatively cool regions, which should make the overall velocities high as well.

The existence of large-scale patterns of seismic velocity heterogeneity in the lowermost mantle has been recognized for several decades [e.g., *Dziewonski et al*., 1977; *Dziewonski*, 1984]. Recent V_s models (Plate 1) are characterized by relatively high velocities beneath the circum-Pacific, with relatively low shear velocities in two large provinces under the central Pacific and south Atlantic/Africa [e.g., *Ritsema et al.*, 1998; *Mégnin and Romanowicz*, 2000; *Masters et al.*, 2000; *Castle et al.*, 2000; *Kuo et al*., 2000; *Gu et al*., 2001; *Grand*, 2002; *Becker and Boschi*, 2002; *Antolik et al.*, 2003; *Ishii and Tromp*, 2004]. The recent models have $\delta \ln V_s$ fluctuations ranging over ±5% within D″ and there is general consistency of the large-scale features between different models. $\delta \ln V_p$ variations (Plate 1) are characterized as having a range of ±1.5% fluctuations, with low velocity areas beneath the southern Pacific and south Atlantic/Africa and high velocities under eastern Eurasia [e.g., *Bijwaard, et al.*, 1998; *Vasco and Johnson*, 1998; *Boschi and Dziewonski*, 1999, 2000; *Zhao*, 2001; *Fukao et al*., 2001; *Kárason and van der Hilst*, 2001; *Antolik et al*., 2003]. The strength of estimated V_p heterogeneity is stronger by factors of 3 to 10 in models that seek to invert for CMB topography in combination with a 20 to 300 km thick boundary layer [e.g., *Doornbos and Hilton*, 1989; *Garcia and Souriau*, 2000; *Sze and van der Hilst*, 2003], so the large-scale tomographic models may be heavily smoothed images that underestimate even large scale velocity fluctuations. In general, the consistency amongst V_p models is lower than between V_s models in D″, possibly due to differences between the bulletin *P* wave arrival time and *S* waveform data sets, the weaker heterogeneity for *P* waves and/or the additional spatial sampling provided by *ScS* and *SKnS* phases that helps to resolve the V_s structure.

Some studies do find good correlation between V_p and V_s structures [e.g., *Masters et al*., 2000], however this correlation appears to break down in certain regions, such as beneath the northern Pacific, where V_p anomalies tend to be positive and V_s anomalies tend to be negative (Plate 1). A south-to-north increase in the $\delta \ln V_s/\delta \ln V_p$ ratio has also been found using diffracted waves traversing D″ below the northern Pacific [*Wysession et al.*, 1999]. V_s models show markedly stronger increases in RMS velocity heterogeneity in the lowermost 300 km of the mantle than do V_p models, although the various models differ in the extent to which shear velocity heterogeneity is concentrated toward the CMB.

Simultaneous inversions of *P* and *S* data have been performed in attempts to isolate bulk sound velocity variations from shear velocity variations [e.g., *Robertson and Woodhouse*, 1996; *Su and Dziewonski*, 1997; *Kennett et al*., 1998; *Masters et al.*, 2000; *Antolik et al*., 2003], but there are significant discrepancies between these models, perhaps as a consequence of incompatible resolution of V_p and V_s structures on a global basis. Direct comparisons of *P* and *S* travel time anomalies on specific paths support the possible decorrelation of these elastic velocities for localized regions within D″ [e.g., *Saltzer et al*., 2001; *Wen et al.*, 2001], as well as confirming the greater variations in δlnVs than δlnVp [e.g., *Tkalcic and Romanowicz*, 2001; *Simmons and Grand*, 2002]. In circum-Pacific regions there are strong correlations between P and S velocity anomalies for localized regions [e.g., *X. Sun et al*., 2007], which are not evident in the global models, so the degree of correlation is still unclear. Global tomography inversions lack detailed resolution of radial velocity gradients, however, there is no question that large-scale patterns of velocity variations dominate in the boundary layer.

Under the assumption that pPv is most likely to occur in lower temperature regions, and using higher velocities as indications of the combined effects of lower temperatures and the increase in V_s expected for pPv; the maps in Plate 1 suggest a circum-Pacific band of a relatively thick layer of pPv [e.g., *Helmberger, et al*., 2005]. While V_p actually tends to be high in the circum-Pacific, the velocity anomalies are smaller, so the expected δlnVs/δlnVp behavior can be reconciled with the presence of pPv in these regions in general, although not necessarily in the regions with strong V_p discontinuities like beneath northern Eurasia. There is uncertainty in the reference level for all tomographic models, so shifts of the baseline can affect the relative size and, of great significance, the sign of the ratio of perturbations, so this is a weak line of consistency. However, the same argument would suggest that low shear velocity regions are warm and have a thin or no layer of pPv, in which case these regions would have δlnVs/δlnVp behavior appropriate for warm Pv, unless there is a chemical contribution. The anticorrelation of bulk sound velocity and shear velocity found by *Masters et al*. [2000] is dominated by the patterns in LLSVPs beneath the Pacific and Africa, and this is commonly invoked as evidence for chemical heterogeneity in these regions. The lateral boundaries of the LLSVPs do appear to be quite sharp, possibly over tens of kilometers or less [e.g., *Luo et al*., 2001; *Ni et al.*, 2002; 2005; *Toh et al*., 2005; *Ford et al*., 2006; *Garnero et al*. 2006], which is generally attributed to a chemical change. Assessing whether pPv exists in these regions must allow for the effects of chemistry on the depth and velocity contrasts in the presence of the compositional differences. As noted above, reflectivity in the LLSVPs can be associated with pPv in iron-rich or MORB-enriched piles, but the problem of solving for the thermal,

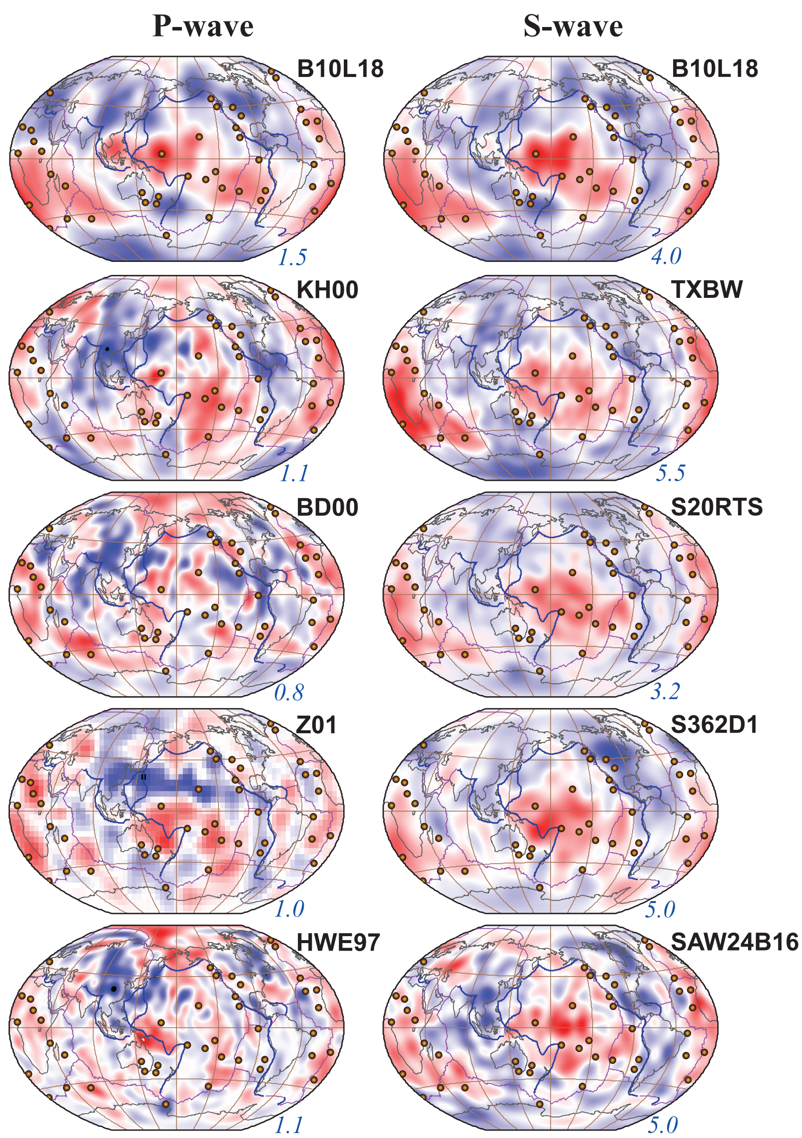

Plate 1. Tomographically derived global *P*-wave (left column) and *S*-wave (right column) velocity perturbations at the base of the mantle. Red and blue colors indicate lower and higher velocities than global averages, respectively. Color scales are not uniform for the different models; the peak-to-peak value is indicated in the lower right of each map (blue number). Model names are given in the upper right, and correspond to the following studies: B10L18 [*Masters et al.*, 2000], KH00 [*Kárason and van der Hilst*, 2001], TXBW [*Grand*, 2002], BD00 [*Becker and Boschi*, 2002], S20RTS [*Ritsema and van Heijst*, 2000], Z01 [*Zhao*, 2001], S362D1 [*Gu et al.*, 2001], HWE97 [*van der Hilst et al.*, 1997], and SAW24B16 [*Mégnin and Romanowicz*, 2000]. Plate boundaries are magenta lines, but convergent boundaries are shown in blue.

compositional and phase boundary effects is rather ill-constrained [see *Reif*, this volume]. Overall, this line of evidence is somewhat supportive of presence of pPv, but the evidence for chemical heterogeneity complicates the argument and the variability in current V_p models weakens confidence in any interpretation.

4.4. Topographic Undulations of Seismic Velocity Contrasts

Sidorin et al. [1999a, b] used the regional variations in timing of *S* reflections from strong D″ velocity increases relative to direct *S* phases to predict that a large positive Clapeyron slope is required for an unspecified phase boundary model to match the data, which is consistent with the subsequently discovered pPv phase transition. At face value this seems like a compelling line of support, but there are several important caveats. The differential times are influenced by mantle heterogeneity, and more recent models have much stronger lowermost mantle heterogeneity than in the models considered by *Sidorin et al*. [1999a, b], so an up-dated evaluation of all observations is worth pursuing as demonstrated by *Sun et al*. [2007]. There is also strong small-scale fluctuation in seismic wave differential times that suggest significant structural variations within a given region [e.g., *Lay et al*., 1997] are as large as the fluctuations on a larger scale that support the general inference of a positive Clapeyron slope. For example, *Hutko et al*. [2006] image changes in depth of a strong SH reflector of about 100 km over lateral distances of 200-300 km. Accounting for such large undulations, even with a large positive Clapeyron slope, requires strong thermal gradients of 500-1000° on such scale-lengths. Special dynamical conditions appear to be needed to produce such thermal heterogeneity, and *Hutko et al.* [2006] advocate folding and piling of cold slab material to produce strong thermal gradient that modulate the phase boundary as one possible way to reconcile this. This notion is also advocated by imaging and modeling studies on a broader scale below Central America [*Sun et al.*, 2007; *van der Hilst et al*., 2007]. Finally, regions like the central Pacific, for which *Sidorin et al*. [1999a, b] invoke and/or predict a very deep or no shear velocity discontinuity, actually do appear to have discontinuities that are in some places no deeper than in circum-Pacific regions. This weakens the original argument, and while chemical heterogeneity may account for this even with a pPv phase change being present, it is not clear that repeated analysis with current data would favor the phase change model or a positive Clapeyron slope. Finally, the original argument is heavily dependent on specific lower mantle flow patterns (with slabs reaching the lowermost mantle), and this is itself contentious. Our assessment is that while it is possible to develop consistency between observations of reflector undulations and a pPv scenario, it is less diagnostic than originally hoped.

4.5. Pairing of Seismic Velocity Increases and Decreases

One of the unique aspects of the pPv phase change scenario is that it may occur in close proximity to the huge compositional contrast at the CMB, and thus in close proximity to a major thermal boundary layer. Rapid increase in temperature above the CMB by 1000 to 1500° appears to be needed if there is not a transition zone or mid-mantle thermal boundary layer [e.g., *Williams*, 1998], and this gives a situation where the Pv-pPv phase boundary could be re-intersected right above the CMB because temperatures increase beyond the stability regime for pPv [*Hernlund, et al*., 2005]. This scenario predicts a second velocity discontinuity, with opposite sign of velocity changes at a distance above the CMB controlled by the thermal gradient and the Clapeyron slope.

The double phase boundary crossing scenario is testable when detailed reflectivity is resolved, but this is very challenging due to the intrinsic differences in detectability of velocity increases versus decreases discussed above [see *Flores and Lay*, 2005]. Nonetheless, the appeal of this test is strong; if paired discontinuities are observed anywhere, one can infer that the CMB temperature is hotter than the stability for pPv of corresponding composition everywhere (since the CMB itself is close to isothermal), so any double-crossing should be widely observed if high quality data are available. In addition, having two proximate P-T tie-points can constrain the thermal gradient, critical to estimation of heat flux through the boundary layer. The observations invoked by *Hernlund et al*. [2005] in the initial discussion of this scenario are drawn from two studies [*Thomas et al*., 2004a, b] suggesting the presence of a velocity increase overlying a velocity decrease, but the observations appear ambiguous. A velocity model for beneath the Cocos plate [*Thomas et al*., 2004a] obtained by migration has a velocity increase above a velocity decrease, but the modeled energy attributed to a second crossing of the pPv phase boundary does not appear to be from a horizontally extensive feature [*Flores et al*., 2005], and is consistent with an origin from isolated D″ scatterers [*Hutko et al*., 2006]. A separate analysis gave a velocity model for Eurasia based on *S* wave migration [*Thomas et al*., 2004b] with a V_s increase overlying a much stronger V_s decrease; the strength of the deeper negative discontinuity is more consistent with a ULVZ origin, that may not involve a reverse transformation or pPv to Pv.

In more recent work, paired velocity increases and decreases have been suggested in a region beneath central America [*Sun et al*., 2006; *van der Hilst et al*., 2007] and below the central Pacific [*Avants et al*., 2006b; *Lay et al*., 2006]. The latter region exhibits systematic changes in the depths of the paired shallower increase and deeper decrease in V_s that are compatible with lateral variations in temperature intersecting a fixed phase boundary with positive Clapeyron slope [*Lay et al*., 2006]. This is a rather specific behavior that is diagnostic of phase boundary behavior in a laterally varying thermal environment where double-intersection of the phase transition occurs (alternatively, chemical heterogeneity could be contrived to behave as required to match the data). Overall, this line of evidence, while subtle due to the intrinsically weak reflections from velocity decreases, is supportive of the presence of pPv. If this is the case, pPv exists both in circum-Pacific regions and within LLSVPs. Further stacking of data in other regions to detect weak paired reflectors is strongly motivated as a test of pPv presence. *Hernlund and Labrosse* [2007] document the constraints on the phase transition Clapeyron slope that ensue from observation of a double-crossing.

4.6. Density Contrasts and Correlation with Velocity Changes

The 1-2% density increase of pPv relative to Pv should accompany a shear velocity increase in regions of pyrolitic composition. Unfortunately, the wide-angle triplication arrivals that reveal a shear velocity increase are insensitive to density contrast, and weak pre-critical reflections that can constrain the impedance contrast are seldom observable. Multiple ScS reflections under the western Pacific do prefer a 1.7% density increase a shear velocity increase of 2.75% is assumed [*Revenaugh and Jordan*, 1991], but disentangling lower mantle and upper mantle reverberations is difficult, so this is, at best, a weak constraint. The large-scale patterns of density heterogeneity inferred from some normal mode analyses [*Ishii and Tromp*, 1999; 2004; *Trampert et al*., 2004] are not supportive of this, in that the LLSVP regions appear to have high density relative to circum-Pacific regions. This is at odds with the notion that high shear velocity circum-Pacific regions have low temperatures and thick layers of pPv. But the density models are not well determined [e.g., *Romanowicz*, 2001; *Kuo and Romanowicz*, 2002] and chemical contributions may simply overwhelm the systematics expected within uniform composition material. Some combination of a buoyant upwelling above a chemical pile can also be reconciled with the data [e.g., *Simmons et al*., 2007]. Thus, at this time, it is not possible to form firm conclusions from testing the pPv hypothesis using seismically determined lowermost mantle densities. Stacking to detect small pre-critical reflections and modeling of the results should help to resolve impedance contrasts, though ideally this will be done for regions where grazing ray paths independently constrain the velocity contrasts so that density effects can be isolated.

4.7. Changes in Anisotropic Properties Coupled to Velocity Changes

The contrast in anisotropic parameters across the Pv-pPv phase boundary, combined with differences in the mid-mantle and thermal boundary layer flow regimes may provide a test of pPv presence. In contrast to most of the lower mantle, the lowermost mantle has extensive regions where shear wave splitting occurs [see reviews by *Lay et al.*, 1998a,b; *Kendall*, 2000; *Moore et al*. 2004]. Splitting of *ScS* phases has long been observed [e.g., *Mitchell and Helmberger*, 1973; *Lay and Helmberger*, 1983b; *Rokosky et al.*, 2004, 2006], and observations of diffracted waves demonstrate that anisotropy is present in D″ [e.g., *Vinnik et al.*, 1989, 1995, 1998; *Lay and Young*, 1991; *Kendall and Silver*, 1996; *Matzel et al.*, 1996; *Garnero and Lay*, 1997; *Ritsema et al.*, 1998; *Russell et al.*, 1998, 1999; *Ritsema*, 2000; *Thomas and Kendall* 2002; *Panning and Romanowicz*, 2004; *Garnero et al.*, 2004a,b; *Maupin et al*., 2005; *Usui et al*., 2005; *Ford et al.*, 2006; *Thomas et al*., 2007]. D″ anisotropy estimates are subject to large uncertainties because of limitations of corrections for upper mantle anisotropy and the possibility of significant near-source anisotropy, even for deep focus earthquakes [e.g., *Wookey et al*., 2005b; *Rokosky et al*., 2006].

The large majority of observations involve horizontally polarized shear wave components (SH) traveling faster through D″ than vertically polarized shear waves (SV) for grazing incidence or wide-angle reflections, with relatively large-scale regions of D″ displaying 0.5-1.5% anisotropy [e.g., *Garnero and Lay*, 2003; *Thomas and Kendall*, 2002; *Fouch et al.*, 2001; *Vinnik et al.*, 1998; *Kendall and Silver*, 1996; *Ritsema* 2000; *Rokosky et al.*, 2004; *Usui et al.*, 2005]. None of the studies has significant azimuthal raypath sampling, but the observations are at least compatible with vertical transverse isotropy (VTI), as may be caused by hexagonally symmetric material with a vertical symmetry axis or by fine-scale horizontal layering. The onset of shear wave splitting appears to be linked to the *S*-wave velocity increase at the top of D″ in high velocity regions [e.g., *Matzel et al.*, 1996; *Garnero and Lay*, 1997], but resolution of the depth distribution of anisotropy in the boundary layer remains very limited [*Moore et al.*, 2004].

Localized observations of shear wave splitting in D″ with SV velocities being higher than SH velocities have been reported, with localized upwellings in the boundary layer being invoked as one possible way to modify the symmetry axis for shape-preferred orientation (SPO) or LPO anisotropy [e.g., *Pulliam and Sen*, 1998; *Russell et al.*, 1998, 1999; *Rokosky et al*. 2004]. *Garnero et al*. [2004a], *Maupin et al*. [2005], *Wookey et al.* [2005b] and *Rokosky et al*. [2006] present clear observations of azimuthal anisotropy, with fast and slow waves mixed on the SH and SV components. There appears to be complex varying azimuthal anisotropy near the border of the western region of the Pacific LLSVP [*Wang and Wen*, 2006b]. Thus far it has only been possible to characterize the horizontal component of the symmetry axis in any of these regions.

The observations of D″ anisotropy, including the apparent increase in macroscopic anisotropic fabric below strong V_s discontinuities are, at least, consistent with the notion that the pPv transition favors the development of LPO in a horizontally sheared boundary layer. *Stackhouse et al*. [2005a] and *Stackhouse and Brodholt* [this volume] compute the anisotropic properties of post-perovskite for high P-T and consider the propensity for developing LPO compatible with seismic observations. It appears that pPv has more favorable elasticity than Pv for acquiring LPO in D″. Experiments on pPv analogs suggest that the preferred slip plane activated in the presence of simple shear, will tend to yield LPO with $V_{sh} > V_{sv}$ [*Yamazaki et al*., 2006], with some azimuthal variability, although there are inconsistencies between deformation experiments [*Merkel et al*., 2006]. As yet, there is no unique fingerprint of pPv anisotropy that would preclude alternative explanations for the seismic observations, such as LPO in ferro-periclase [e.g., *Karato*, 1998; *Kendall*, 2000; *Mainprice et al.*, 2000; *Yamazaki and Karato*, 2002] or SPO involving low velocity lamellae comprised of partially melted crust [e.g., *Kendall and Silver*, 1996] or other melt or chemical heterogeneities [e.g., *Russell et al.*, 1998]. Since the development of LPO is dependent upon the elasticity, the activation of preferred slip planes, the thermal structure and grain size increase subsequent to phase transition, there are many factors to consider, and realistically, seismic anisotropy observations will only yield compatibility tests, not unique constraints. This issue is considered in more detail by *Wookey and Kendall* [this volume].

5. DISCUSSION

The foregoing review indicates the complexity of seismological observations for the lowermost mantle and the challenges of interpreting the structures given the likely presence of thermal, chemical, and dynamical heterogeneity. Whereas the 410-km and 660-km transition zone discontinuities are globally observed, and can be sampled with many seismic wave interactions, lowermost mantle reflectivity is only sparsely resolved and there are first-order uncertainties in the thermal and chemical environment. The latter uncertainties combine with the currently limited characterization of the properties of pPv to preclude definitive conclusions regarding the presence of this phase in the mantle. However, the circumstantial evidence is still substantial and justifies the excitement regarding the discovery of pPv.

Most compelling is the observation of a rapid increase in V_s at depths compatible with the Pv-pPv transition expected for pyrolitic mantle, the generally weaker and more variable nature of any associated V_p change, the tendency for the V_s increase to be prominent in regions of volumetrically high velocities consistent with relatively low temperatures, and the indications that the V_s increase is accompanied by an increased propensity for the rock to acquire anisotropic properties manifested in shear wave splitting. Topographic variations in the depth of the V_s increase are plausibly attributable to thermal variations in a complex boundary layer, and the lateral weakening/vanishing of the reflector indicated by some studies can be reconciled with a thinning or even pinching-out of the pPv layer due to lateral thermal gradients. A few observations of paired velocity increases and decreases that jointly vary systematically in depth provide a strong line of support for a phase change explanation.

Problematic observations include (a) indications that low V_s regions tend to have relatively high density and high bulk sound velocity, (b) the presence of V_s increases several hundred kilometers above the CMB in volumetrically low velocity regions that might have been thought to be too warm to host pPv or where only a very thin layer might have been expected, (c) localized observations of very strong P wave reflections from the same depth of the V_s increase, and (d) observations of shear wave splitting in both regions of high and low velocity. There is, of course, no requirement that pPv behavior be reconciled with all of these observations; it may only be present in local regions that are cool enough for the phase to be stable. However, the likelihood that some of the problematic observations require chemical heterogeneity, particularly in LLSVPs, may actually hold the key to reconciling these disparate observations with extensive pPv occurrence. For example, compositional changes can affect the phase transition pressure for Pv-pPv, breaking down the correlation between depth of the transition and volumetric velocity expected for uniform composition with lateral temperature variations. Composition may also affect the strength and even polarity of the V_s change, allowing some negative V_s discontinuity observations and regions of volumetric low V_s velocity to be associated with the pPv phase transition. Plate 2 summarizes the spatial distribution of seismic observations that have been used to assess pPv occurrence on large scale. Unfortunately, the large number of degrees of freedom allows one to invoke a variety of combinations of thermal, chemical, and phase change heterogeneity that can plausibly reconcile the observations with either small or large volumes of pPv in the lower mantle.

Clearly more work needs to be done to achieve robust conclusions, and the potential offered by having a well-calibrated deep mantle thermometer is strong motivation. With expanding global coverage by seismic networks, it is important to apply waveform stacking approaches to *P* and *S* waves to increase the coverage and the reliability of characterizations of lower mantle reflectivity. Techniques for this analysis are well-established; the main need is for careful, data-intensive endeavors that apply sensible data selection criteria (for example, making sure that PcP is detectable in stacks ensures that suitable radiation illuminated the deep mantle). Seeking pre-critical reflections from

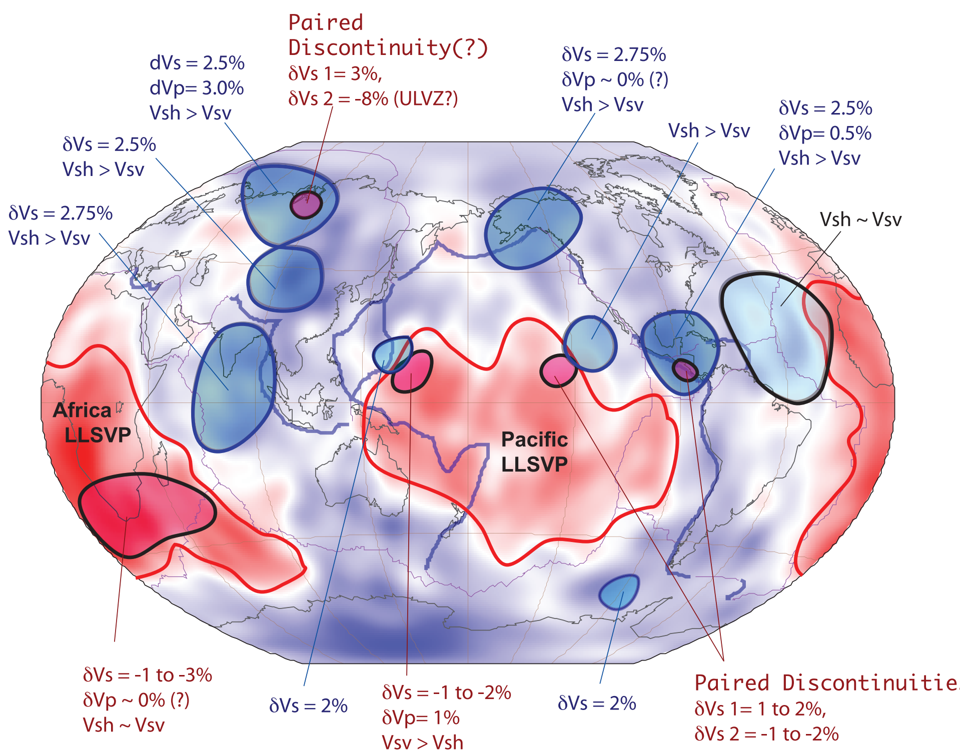

Plate 2. Summary of spatial distribution of key seismological observations of structure in the lowermost mantle relevant to assessing whether pPv exists in the boundary layer. The background pattern is the shear velocity variation in the lowermost layer of the global tomography model of *Grand* [2002], which has a peak-to-peak range of variations of 5.5%. Blue regions are faster than average, near-average regions are white, and slower than average regions are red. The locations of the Pacific and African LLSVPs are indicated. The results of regional studies discussed in the text that determine the *S* wave velocity jump ($\delta V_s = \delta \ln V_s$) and/or *P* velocity jump ($\delta V_p = \delta \ln V_p$) of any predominant D″ reflector, and average shear wave splitting behavior (Vsh denotes the horizontally polarized shear wave velocity; Vsv denotes the vertically polarized shear wave velocity) are also indicated. Many additional localized results have been published, but these are the best-sampled and most consistent results. The blue areas are candidates for presence of pPv in pyrolitic mantle material, which should produce a fairly strong *S* reflector and weak *P* reflector, with higher than average volumetric *S* velocity. The LLSVP regions appear to have stronger V_s reductions than V_p reductions (see Plate 1) and corresponding bulk sound velocity variations anticorrelated with V_s. These regions may have negative Vs discontinuities, and strong lateral gradients. A few regions show paired velocity increases and decreases consistent with Pv-pPv forward and reverse transitions. If chemical anomalies in LLSVPs reduce the pPv transition depth (as expected for MORB composition or for Fe-enriched pyrolite), pPv may exist in the LLSVP regions. Alternatively, it may be too hot in such regions for pPv, and chemical effects are responsible for the reflectivity rather than phase change boundaries.

regions sampled by wide-angle reflections is very desirable for quantifying the density contrast at any discontinuities. This type of effort needs to be conducted with locally dense sampling, given the observations of rapid variability of wavefields in both LLSVP and circum-Pacific regions. For regions with strong reflectors for either *P* or *S*, a special effort should be made to characterize the reflection over a range of ray parameters and frequencies to provide better constraint on the sharpness of the velocity change. Combining the reflectivity analysis with travel time analysis is needed for better establishing the absolute depths of reflectivity and for quantifying the relationship between reflectivity and volumetric velocity. Simultaneous investigation of *P*- and *S*-waves sampling the same region is highly desirable, as there are currently only a handful of regions where similarly estimated structures can be directly compared. Waveform stacking and migration methods must be applied, both to address the issue of model parameterization and finite-frequency effects, but modeling with 2D and 3D synthetics that are similarly processed should be conducted to quantify the amplitudes rather than simply demonstrating detections of arrivals. There has been rapid progress in detection of deep mantle anisotropy, but only a few studies have convincingly isolated the structure to the deep mantle and there are clear indications that azimuthal anisotropy must be allowed for, not just radial anisotropy. This latter area is very challenging, and may benefit from hypothesis testing approaches that draw upon laboratory and mineral physics constrained parameterizations to guide modeling of the data.

The priority areas for mineral physics research can also be enumerated. The elasticity for pPv of varying compositions needs to be determined, including that associated with transformations in rather complex compositions such as MORB. With indications being that either positive or negative shear velocity discontinuities might be expected, more extensive characterization of chemical effects is essential. Experimental results are beginning to directly constrain the Clapeyron slope for end-member compositions, complementing theoretical predictions, and this is of importance for mapping topographic variations of reflectors into thermal estimates. It is also key for the interpretation of paired discontinuity observations, if they result from forward and reverse transformations of pPv. Chemical effects on the magnitude and depth range expected for velocity changes also need further investigation. Predictions of anisotropic properties, as suggested above, requires high T deformation experiments to characterize the slip planes likely to be activated in boundary layer flow, and the full elasticity parameters for variable compositions need to be mapped out.

6. CONCLUSIONS

The discovery of post-perovskite, the high P-T polymorph of magnesium perovskite, has created a sea-change in deep mantle research, as it provides a potential breakthrough for interpreting puzzling seismological attributes of the lowermost mantle and for establishing a thermometer that may constrain absolute temperatures and temperature gradients more directly than any other approach. A review of the numerous recent seismological observations indicates both supporting evidence and possibly contradictory evidence of the presence of pPv in the lower mantle. On balance, there is good support for the possibility that pPv comprises a significant volume of the D″ layer, and some evidence points toward a very extensive occurrence of the phase in the deep mantle. However, the seismic data are complex and admit other possibilities and complications, thus caution is warranted in interpreting resulting seismic models in terms of quantitative properties of pPv. Clear directions for future seismological and mineral physics research effort to reduce the non-uniqueness and to fully exploit the potential impact of pPv presence in the mantle are apparent. These directions are at the forefront of analysis techniques and data collection in the respective discipline, but progress is viable even with existing capabilities. The potential return for our understanding of thermal and chemical processes in the mantle is enormous.

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