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Bulk sound travel times and implications for mantle composition and outer core heterogeneity

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Abstract. We present a new methodology for constraining mantle composition directly from observed seismic travel times. By measuring travel times of PcP, PcS, ScS, and $PKiKP$ phases generated by the Chinese nuclear explosion of 21 May 1992 and recorded at epicentral distances of $\sim 10°$, we directly determine bulk sound travel times (T^{Φ}) for the mantle and outer core. This eliminates the need to combine independent V_P and V_S profiles, characterizing different paths and frequencies, to obtain a reference V_{Φ} profile. Our observed $PKiKP - PcP$ differential travel time indicates that the outer core model of AK135 is slow by 2.1 s (that of IASP91 by 2.3 s) while that of PREM is slow by only 0.7 s and suggests, through comparison with other data, relative lateral homogeneity of the outer core. Furthermore, we find that T_0^{Φ} for the crust and mantle in this region exceeds the AK135 prediction by 3.5 s (IASP91 by 3.9 s) +1.0 s. Since the lower mantle contribution to this delay depends upon how much arises in the upper mantle, we present two end-member models for such a slower mantle V_{Φ}, in which the delay resides entirely above or below 660 km depth. The latter yields best-fitting lower mantle compositions which are less enriched in Si than those fit to T_0^{Φ} for AK135 (or IASP91) for the former, and neither supports Fe-enrichment of the lower mantle. Applicability of such localized observations to the larger mantle is suggested by tomographic results indicating that lateral heterogeneity in V_{Φ} is very small relative to that in V_S. Our results illustrate the potential utility of T_0^{Φ} in constraining mantle composition.

Introduction

Constraints upon the composition of Earth's interior are frequently derived by comparison of the radial profiles of density and seismic velocity obtained from inversion of seismic data to those obtained from various mineralogical models [*Birch*, 1952; *Jackson*, 1983; *Knittle et al.*, 1986; *Jeanloz and Knittle*, 1989; *Bina and Silver*, 1990; *Stixrude et al.*, 1992; *Hemley et al.*, 1992; *Zhao and Anderson*, 1994]. Direct comparison with seismological P-wave velocity (V_P) and S-wave velocity (V_S) profiles is difficult, since determination of V_P and V_S for mineral assemblages requires knowledge of the poorly constrained behavior of mineral shear moduli at high temperatures and pressures. Therefore comparisons are frequently made using the bulk sound velocity (V_{Φ}), given by

$$V_{\Phi}^2 \equiv V_P^2 - \frac{4}{3}V_S^2 , \qquad (1)$$

in order to avoid the additional uncertainties associated with extrapolation of shear moduli [*Bina and Silver*, 1990]. However, construction of a seismological V_{Φ} profile is complicated by the fact that the V_P and V_S profiles employed in equation (1) generally have been independently determined for different regions [*Bina and Wood*, 1987]. Thus, the component V_P and V_S profiles may be characteristic of different temperature structures, compositional regimes, and frequency ranges, leading to systematic errors in the resulting V_{Φ} profile. Hence attempts to constrain mantle composition, for example, can be as sensitive to uncertainties in reference seismic models as to uncertainties in mineral thermoelastic equations of state [*Bina and Silver*, 1990].

Here we present a new methodology for directly determining bulk sound travel times (T^{Φ}) for the mantle and outer core in a single region, thus circumventing the need to combine independent V_P and V_S profiles characterizing different paths and frequencies. We use these T^{Φ} measurements to test different models of outer core velocity and to evaluate the sensitivity of inversions for lower mantle chemistry to uncertainties in both seismic data analysis and mineral equations of state.

Method

We measure the travel times of core-reflected phases recorded at small source-receiver epicentral distances ($\Delta \sim 10°$). In such experiments, core-reflected P and S waves are recorded on the same instruments, and they traverse nearly identical paths which are essentially independent of Earth model. For a given phase recorded at a distance Δ, we measure the difference (δT_{Δ}) between the observed travel time and that predicted (T_{Δ}) by a reference model such as AK135 [*Kennett et al.*, 1995] (or IASP91 [*Kennett and Engdahl*, 1991]). We correct δT_{Δ} to its value at $\Delta = 0°$ (δT_0) by assuming

$$\frac{\delta T_0}{T_0} \approx \frac{\delta T_{\Delta}}{T_{\Delta}} \qquad (2)$$

for small Δ.

Given any two of the phases PcP, PcS, ScS, and ScP, we can determine two-way T_0^{Φ} through the crust

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and mantle by considering small variations in velocity at a given depth z:

$$(V_\Phi + \delta V_\Phi)^2 = (V_P + \delta V_P)^2 - \frac{4}{3}(V_S + \delta V_S)^2 \ . \quad (3)$$

Upon vertical integration over depth, this yields the perturbation in vertical bulk sound travel time (δT_0^Φ):

$$\delta T_0^\Phi = -2 \int_0^{CMB} \left[\left(1 + \frac{4}{3}R\right)^{\frac{3}{2}} \frac{\delta V_P}{V_P^2} - \frac{4}{3} R^{\frac{3}{2}} \frac{\delta V_S}{V_S^2} \right] dz \ , \quad (4)$$

where $R \equiv {V_S}^2/{V_\Phi}^2$. If we assume that R is independent of depth through the mantle, this becomes:

$$\delta T_0^\Phi \approx \left(1 + \frac{4}{3}R\right)^{\frac{3}{2}} \delta T_0^P - \frac{4}{3} R^{\frac{3}{2}} \delta T_0^S \ , \quad (5)$$

where δT_0^P and δT_0^S are the perturbations in vertical PcP and ScS travel times, respectively. The actual $\pm 5\%$ variations of R about its mean for AK135 (0.4924) induce error in δT_0^Φ of ± 0.05 s.

Given three or more of these phases, we can perform a consistency check by deriving several pairwise estimates of δT_0^Φ, thus providing a simple measure of uncertainty.

Data and Results

We use seismic waves generated by the Chinese nuclear explosion of 21 May 1992, recorded by a PASSCAL instrument deployment on the Tibetan Plateau [*Owens et al.*, 1993] and subjected to a 4-pole Butterworth bandpass filter of 0.75-2.00 Hz (Figure 1). Core-reflected phases arrive late relative to AK135 (and IASP91) predictions. Delays at SANG ($\Delta = 10.80°$, $\phi_{az} = 166.62°$) are 1.6, 1.8, and 1.1 s for PcP, PcS, and ScS, respectively, and delays at USHU ($\Delta = 10.77°$, $\phi_{az} = 140.04°$) are 1.0 and 0.2 s for PcS and ScS, respectively, including corrections for ellipticity and for S-wave splitting due to known receiver-side anisotropy [*McNamara et al.*, 1994] at SANG ($\Delta t = 0.80$ s, $\phi_{az} = 51°$) and USHU ($\Delta t = 0.72s$, $\phi_{az} = 119°$). No ScP phases are observed at these stations.

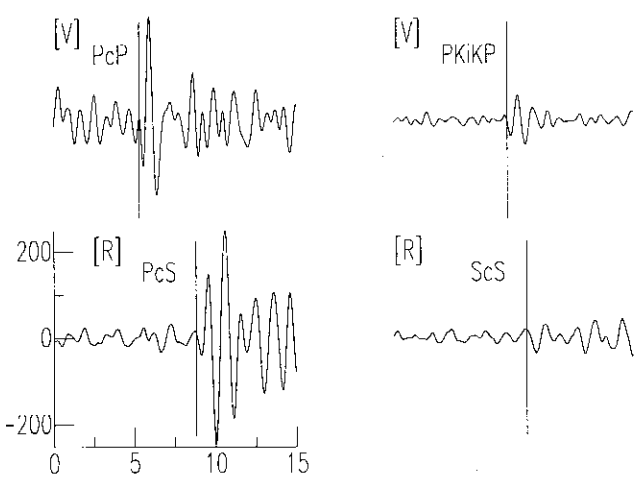

Figure 1. The core phases PcP (vertical), $PKiKP$ (vertical), PcS (radial), and ScS (radial) recorded at SANG ($\Delta = 10.80°$, $\phi_{az} = 166.62°$) and bandpass filtered to 0.75-2.00 Hz. For each window, 15 seconds of data are shown.

Pairwise estimates of δT_0^Φ relative to AK135 at SANG yield 2.9 s from PcP & ScS, 4.6 s from PcS & ScS, and 2.5 s from PcP & PcS. A pairwise estimate at USHU yields 3.9 s from PcS & ScS. Thus, observation of these phases yields T_0^Φ for the crust and mantle which exceeds the AK135 prediction by 3.5 s (IASP91 by 3.9 s) ± 1.0 s, so that mantle V_Φ in this region must be slower than AK135 (similarly for IASP91 and for PREM [*Dziewonski and Anderson*, 1981]). Reassuringly, our delay falls within the broad uncertainties of previous results from long-period multiple-ScS phases [*Sipkin and Revenaugh*, 1994] which indicate that T_0^S for our receiver region is slow by 1.3 ± 1.4 s while that for our source region is fast by 0.8 ± 4.3 s.

Additionally, $PKiKP$ is observed at SANG, constraining V_P ($= V_\Phi$) for the outer core in this region. The observed differential travel time between $PKiKP$ and PcP is 476.8 s, indicating that the AK135 outer core model is slow by 2.1 s (IASP91 by 2.3 s). By contrast, the PREM outer core is slow by only 0.7 s (presumably due in part to differing core model radii).

Discussion and Conclusions

Our $PKiKP - PcP$ differential time observation indicates that the PREM core model provides a better fit to actual core travel times beneath this region than does the AK135 (or IASP91) core model. Furthermore, the relevant differential times used to derive PREM (477.5 s at 10.9° and 477.2 s at 11.73°) were recorded from NTS nuclear explosions in the western United States [*Engdahl et al.*, 1974], thus limiting to 0.7 s the vertically averaged difference in outer core velocities beneath these two widely separated regions and arguing for a laterally homogeneous outer core.

Our PcP, PcS, and ScS observations can be used to constrain allowable lower mantle compositions. The structure of the crust and upper mantle (0-660 km depth) and that of the very top and bottom of the lower mantle (661-764 km, 2726-2889 km) are constrained by features other than T_0^Φ. We therefore begin by fixing densities (ρ) and velocities (V_Φ) for these regions at their AK135 values. By allowing the composition of the bulk of the lower mantle (765-2725 km) to vary in terms of Mg/Fe content and Si content, given by the molar ratios $X_{Mg} \equiv Mg/(Mg+Fe)$ and $X_{Pv} \equiv Si/(Mg+Fe)$, respectively, we solve for the lower mantle compositions whose calculated $\rho(z)$ and $V_\Phi(z)$ profiles along adiabats minimize the misfit to reference models. We follow the method of *Bina and Silver* [1990], except that we minimize the misfit to seismological $\rho(z)$ and T_0^Φ rather than to $\rho(z)$ and $V_\Phi(z)$.

Thus, rather than fit a $V_\Phi(z)$ profile resulting from inversion of seismic data for independent $V_P(z)$ and $V_S(z)$ profiles, we directly solve the forward problem for T_0^Φ for each candidate composition and compare those solutions to T_0^Φ from our observations of core reflected phases. Our approach thus avoids comparison with problematic $V_\Phi(z)$ profiles, and it obviates the necessity of attempting to modify an AK135 $V_\Phi(z)$ profile to yield a larger T_0^Φ value—modification which would result in a multitude of nonunique models, many of which would violate other seismological constraints. Furthermore, our methodology allows us to directly vary the

reference T_0^Φ between different models so as to determine the effect upon mantle compositional constraints of assigning responsibility for our observed delays to the lower mantle or to shallower regions. We present two end-member interpretations in Figure 2: in the bottom panel (AK135+3.5s), all 3.5 s of T_0^Φ delay is assumed to occur in the lower mantle; in the top panel (AK135), it is all mapped into the crust and upper mantle.

While fitting $\rho(z)$ provides no support for an Fe-enriched lower mantle, our measured mantle T_0^Φ of AK135+3.5s results in best-fitting lower mantle compositions which are less enriched in Si than those fit to the T_0^Φ calculated for AK135 (Figure 2). The high-γ_{D0} equations of state for ferromagnesian silicate perovskite [*Bina*, 1995] which had supported a Si-enriched lower mantle are now consistent with less Si-rich compositions, nearer to pyrolite [*Ringwood*, 1975]. The low-γ_{D0} equations of state continue to show no evidence for lower mantle Si-enrichment; indeed, these latter are even consistent with Si-depletion for low estimates of lower mantle temperature.

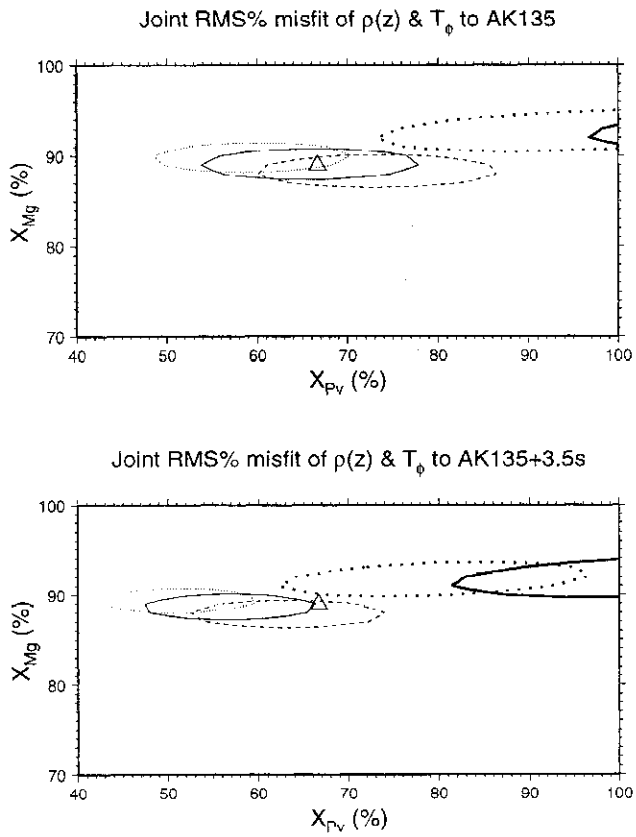

Figure 2. Contours (0.5%) of joint RMS misfit to reference seismic models of calculated $\rho(z)$ and T_0^Φ as functions of candidate lower mantle composition, after the method of *Bina and Silver* [1990]. Compositions are parameterized in terms of the molar ratios $X_{Mg} \equiv Mg/(Mg+Fe)$ and $X_{Pv} \equiv Si/(Mg+Fe)$. Triangle denotes pyrolite composition. Heavy contours are for high-γ_{D0} perovskite equation of state (line 1, Table 1 of *Bina* [1995]); light contours are for low-γ_{D0} equation of state (line 2, loc. cit.). Adiabat temperatures at 660 km are 1800 K (dotted), 2000 K (solid), or 2200 K (dashed). In the top panel, $\rho(z)$ and T_0^Φ are fit to AK135 values. Thus, assuming a low-γ_{D0} perovskite equation of state and a temperature of 2000 K at 660 km, lower mantle compositions which best fit AK135 densities and travel time fall within the solid light contour, overlapping pyrolite. The absence of a heavy dashed contour indicates that for a hot (2200 K at 660 km) lower mantle, even Si-enrichment to 100% perovskite cannot match seismic data for high-γ_{D0} perovskite equations of state. In the bottom panel, $\rho(z)$ is again fit to AK135, but T_0^Φ is fit to AK135+3.5s to match our observed travel times.

While the radial V_S model TIP for the Tibetan Plateau indicates that upper mantle V_S is slow [*Zhao et al.*, 1991], recent 3-D tomographic imaging of Tibetan structure reveals fast V_S and slow V_P in the upper mantle beneath our stations [J. VanDecar, pers. comm.]. Furthermore, $SS-S$ residuals for the Tarim Basin reveal slow V_S in the upper mantle beneath our Lop Nur source [S. Roecker and I. Dricker, pers. comm.]. Thus, while we cannot definitively partition the observed T_0^Φ delay between upper and lower mantle structure, we obtain two models in which either all or none of the observed delay arises within the lower mantle. Our results for these two end-members demonstrate the potential utility of T_0^Φ in constraining mantle composition and illustrate the sensitivity of such inferences to uncertainties in both seismic data analysis and mineral equations of state.

Extension of such results to other regions requires that lateral heterogeneity in V_Φ be relatively small in the lower mantle. Tomographic imaging of mantle seismic structure yields values for $dlnV_S/dlnV_\Phi$ "significantly greater" than unity [*Dziewonski and Su*, 1995] and values for $dlnV_S/dlnV_P$ of about 2 [*Woodhouse and Dziewonski*, 1987], the latter rising from 1.7 to 2.6 with increasing depth in the lower mantle [*Robertson and Woodhouse*, 1996], thus supporting the hypothesis that lateral heterogeneity of V_Φ is significantly less than that of V_S in the lower mantle.

Additional measurements of mantle T_0^Φ from a growing body of seismic data, in conjunction with improving knowledge of the equations of state of mantle minerals, should place tighter constraints upon lower mantle composition and lateral heterogeneity. Furthermore, the possibility of measuring T_0^Φ for phases which interact with transition zone seismic discontinuities may offer a direct probe into the distribution of T_0^Φ delays between upper and lower mantle structure.

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