Teleseismic P-Wave Tomogram of the Yellowstone Plume

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[1] Inversion of a new data set of telesismic P-wave travel-times from three PASSCAL seismic deployments around the Yellowstone hotspot reveals a 100 km diameter upper mantle plume that extends from the Yellowstone volcanic caldera to 500 km depth and dips 20° to the northwest. A monotonic decrease in the velocity perturbation of the plume from −3.2% at 100 km to −0.9% at 450 km is consistent with a uniform thermal anomaly of 180°C. Where the plume crosses the 410 km discontinuity, previous research shows a depression in the 410 km discontinuity consistent with a warm plume (Fee and Dueker, 2004). Additionally, a region of high velocities extends to 250 km beneath the Wind River basin in NW Wyoming that may represent a convective downwelling of the lithosphere. 

[2] From 17–14 Ma, a large-scale magmatic event in the back-arc of the Juan de Fuca subduction zone created the Columbia River basalt group and the eastern Oregon volcanic plateau along with extension and lesser volcanism within the western Snake River Plain and northern Nevada rift [Smith and Braile, 1994; Humphreys et al., 2000]. By 14 Ma, two sets of silicic calderas and domes began propagating outward from SE Oregon: the NE propagating Yellowstone hotspot track (YHT) [Armstrong et al., 1975] and the NW propagating Oregon High Lava Plain trend [Jordan et al., 2004]. Two models for the origin of the YHT remain under consideration. The first model suggests that a modest-sized plume head impacted and spread beneath SE Oregon with a plume tail subsequently migrating to its present location beneath the Yellowstone caldera (YC) [Iyer et al., 1981; Parsons et al., 1994; Smith and Braile, 1994]. A second model suggests that the 17–14 Ma magmatic and extensional events are manifestations of back-arc spreading and that the YHT is not the track of a plume, but instead is a rift along an ancient lithospheric flaw [Christiansen et al., 2002]. Fundamental to testing these differing models is assessment of whether or not a plume tail extends beneath the YC. To date, previous tomographic imaging has lacked sufficient seismic coverage to unequivocally answer this question. This study greatly augments the previous seismic coverage with the addition of data from 76 new Program for Array Studies of the Continental Lithosphere (PASSCAL) stations.

2. Data and Methods

[3] Our seismic waveforms are provided primarily by three PASSCAL seismic deployments (Figure 1): the Yellowstone Array (YISA; June 2000 to May 2001), the Billings array (Aug. 1999 to Aug. 2000), and the Snake River Plain array (SRP; May 1993 to Oct. 1993). Events with magnitude >5.2 Mb and epicentral distance >30°are selected. A multi-channel cross correlation technique is used to measure 14,327 relative travel-time residuals from P, PP and PKP_{df} phases. Given the lack of temporal overlap of the three arrays, a baseline array of nine permanent stations that operated concurrently with all three PASSCAL deployments was constructed from National Seismic Network and University of Utah network stations (Figure 1). For each of our PASSCAL arrays, the mean travel-time residual of the baseline stations with respect to each array’s residuals was calculated. These time shifts are consistent between baseline stations (Figure S2) and are consistent with the known upper mantle velocity variations: the largest array (YISA) has the smallest static (0.03 s) and the Billings array, which resides over high velocity mantle has a negative static (−0.29 s); the SRP array, which dominantly resides over low velocity mantle, has a positive static (+0.21 s). Timing variations induced by crustal heterogeneity are evaluated using P-wave receiver function moho times that were mapped to depth using a 3-D shear wave image derived from Rayleigh wave analysis (D. L. Schutt and K. G. Dueker, in review) and assuming a V_p/V_s ratio of 1.76 (Figure S2). The peak-to-peak amplitude of the crustal timing corrections is 0.35 s, much smaller than the 2.05 s variation of the data set.

[4] The tomographic model space is 1600 × 1600 × 700 km in the longitude, latitude and depth directions and is parameterized into 20 × 20 × 25 km constant slowness blocks. Our choice of a 700 km deep model was guided by analysis of the tradeoff between model depth and variance reduction. This analysis shows that model depths >700 km do not provide significant increases in variance reduction (Figure S3). Our one-dimensional background velocity model is the AK135 model [Kennett et al., 1995] with the upper 200 km set to the average shear wave velocity found beneath the YISA and Billings arrays by D. L. Schutt and K. G. Dueker (in review). To account for sphericity, an earth flattening transformation was used. To assess the effects of 1-D versus 3-D ray tracing on our final model, comparison of ray paths through the most anomalous portions of our model was conducted. This shows that ray path variations are <15 km at 200 km depth and <40 km at 500 km depth; these variations are small with respect to the lateral resolu-
3. Results and Resolution

Inspection of the P-tomography images above 250 km depth (Figure 2) reveals two primary features: a 100 km wide 2–3% low velocity swath under the east Snake River Plain (ESRP) and Yellowstone Caldera (YC), and a 2–3% curtain of high velocity under the Wind River basin in NW Wyoming. The ESRP anomaly lies directly below the volcanic trough formed by the Yellowstone hotspot track. The NW Wyoming high velocity curtain lies below the region of maximum Laramide thrust fault shortening between the Wind River, Owl Creek and Bighorn faults (Figure 1). Below 250 km depth, this ‘near surface’ velocity pattern disappears and the dominate structure is a low velocity pipe that extends from beneath the YC to ~500 km (Figure 2g). This ~100 km diameter pipe is tilted to the NW at 20° and decreases in amplitude from 3.2% at 100 km to ~0.9% at 450 km. This amplitude decrease is consistent with the depth dependence of the anelastic velocity derivatives [Cammarano et al., 2003] and requires a 180°C thermal anomaly. Given these observations, we name this feature the Yellowstone plume. In addition, we note that an ~80 km diameter ~0.7% vertical low velocity pipe is imaged in the transition zone about 100 km NW of the YC where the 660 km discontinuity topography indicates warmer than normal mantle (Figures 2f, 2h, and 2i).

This low velocity pipe translates into a 120 km wide 2–3% low velocity pipe that extends from beneath the YC to 300 km depth. Further support for this conclusion derives from the observation of a localized 12 km depression in the 410 km discontinuity (Figure 2) where the Yellowstone plume crosses this phase transition (beneath

4. Discussion

Our most important result is the resolution of a tilted low velocity plume that extends from beneath the YC to 500 km depth. Further support for this conclusion derives from the observation of a localized 12 km depression in the depth of the 410 km discontinuity (Figure 2) where the Yellowstone plume crosses this phase transition (beneath

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**Figure 1.** Station and topography map. The blue circle is the current location of the Yellowstone caldera. Station associations are shown in legend. The white line outlines the eastern Snake River Plain. Fault labels denote: WR, Wind River; OC, Owl Creak; and BH, Bighorn thrust faults.
Figure 2. Tomographic velocity image. (a)–(f): Map views at indicated depths and the average transition zone velocity structure. Blue crosses are the stations. The white ‘circle’ beneath SW Montana contours the 12 km downwarp of the 410 km discontinuity [Fee and Dueker, 2004]. Cross sections are presented along A-A’ , B-B’ and C-C’ in (g)–(i). The x-offset of the cross-sections have a common origin where the three cross-sections intersect at YC. The black lines at 411 and 656 km depth are the mean depths of the 410 and 660 km discontinuities and the white lines are the topography with a factor of two vertical exaggeration. Note that the Yellowstone plume low velocity pipe crosses the 410 km discontinuity where it is downwarped, but the 660 km topography does not show any upwarp associated with a crossing plume.

Figure 3. Resolution tests. (a)–(d) Checker board test. The red and blue contours outline the 67% amplitude of each input ‘spike’. The spike is a Gaussian function with a 40 km half-width. (a)–(b): Map view at 200 and 400 km depth. (c)–(d): Cross sections along D-D’ and E-E’ shown in (a)–(b). (e)–(g) Images from inversion of three synthetic data sets created by integrating travel-time anomalies through the first 200 km (e), 400 km (f) and 700 km (g) of the true velocity model (Figure 2). The A-A’ cross section is shown in Figure 2. These three cross sections show that the bottom of the Yellowstone plume can be well resolved with <50 km of downward smearing.
Dillon, Montana). Both the 410 topography and the velocity anomaly of the plume are consistent with a ‘plume-like’ 150–200°C thermal anomaly. The observation of a high velocity anomaly at 400–500 km beneath Yellowstone Park (Figure 2d) is problematic because this anomaly is not observed in the 410 km discontinuity topography. However, Figures S5–6 suggest that this anomaly may be a velocity artifact created by shallower structure. Regarding a dynamical understanding of the 20° tilt of the plume, calculations of plume conduit advection due to large scale mantle flow show a variety of trajectories for an upper mantle plume [Steinberger, 2000]. While it is noteworthy that most of the Yellowstone plume trajectories tilt to the west, model uncertainties are too large to permit any firm conclusions.

[9] To speculate on the dynamical significance of a plume that stops at 500 km depth, we are guided by two constraints. First, it seems reasonable that a modern-day lower mantle extension of the Yellowstone plume does not exist [Montelli et al., 2004]. Second, two recent global tomographic images resolve a 0.8–1.1% low velocities ‘pond’ between 700–1000 km depth beneath much of the western U.S. [Grand, 2002; Ritsema and Allen, 2003]. This low velocity pond would be consistent with a ~200° thermal anomaly. Thus, it is possible that the Yellowstone plume originated as a thermal upwelling from this warm pond. A rough estimate of the volume of warm mantle required to make the Yellowstone hotspot track is provided by assuming that the plume impacted the lithosphere around 17 Ma in SE Oregon and that the plume tail has since migrated 700 km to the YC. A proxy for the volume of plume material emplaced below the lithosphere is provided by multiplying the length of the YHT (700 km) by the cross-sectional area of low velocity mantle beneath the ESRP (100 km wide by 100 km deep). Adding this uppermost mantle volume to the volume of the imaged plume conduit (100 km in diameter by 400 km in depth) results in a volume equivalent to a 332 km diameter sphere. Convection modeling shows that the transient release of similar sized warm volumes across the 660 km discontinuity is dynamically plausible [Cserepes and Yuen, 2000].

[10] Finally, we suggest that the curtain of high velocity anomalies extending to 250 km beneath NW Wyoming plausibly represent convective downwelling of the lower lithosphere. These velocity anomalies require 100–150° temperature reduction consistent with lithospheric downwelling models [Schott et al., 2000]. The location of this high velocity curtain adjacent to the Yellowstone plume suggests that this convective downwelling of lithosphere may be balancing the upward flux of plume material. In addition, we note that this region did experience the greatest (15%) Laramide compressive strain [Bird, 1998] between the Wind Rivers, Owl Creek Mountains, and Bighorn thrust faults (Figure 1) which also may have promoted subsequent convective destabilization of the lithosphere here.

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References


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