Mantle structure beneath the western edge of the Colorado Plateau


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1. Introduction

[1] Teleseismic traveltime data are inverted for mantle Vp and Vs variations beneath a 1400 km long line of broadband seismometers extending from eastern New Mexico to western Utah. The model spans 600 km beneath the moho with resolution of ~50 km. Inversions show a sharp, large-magnitude velocity contrast across the Colorado Plateau-Great Basin transition extending ~200 km below the crust. Also imaged is a fast anomaly 300 to 600 km beneath the NW portion of the array. Very slow velocities beneath the Great Basin imply partial melting and/or anomalously wet mantle. We propose that the sharp contrast in mantle velocities across the western edge of the Plateau corresponds to differential lithospheric modification, during and following Farallon subduction, across a boundary defining the western extent of unmodified Proterozoic mantle lithosphere. The deep fast anomaly corresponds to thickened Farallon plate or detached continental lithosphere at transition zone depths.


2. Methods and Model

[4] Instrument responses were deconvolved from La RISTA 1.5 P and S wave recordings. P waves were bandpassed from 0.2–1.5 Hz and S waves from 0.03–0.25 Hz. Traveltime residuals with respect to the IASPEI model [Kennett and Engdahl, 1991] were then measured by a cross-correlation technique for P and S waves indepen-
dently. Only data with back azimuths within 20° from the
trend of the seismic line were used to limit inclusion of off-
strike structure. Two stations in the La RISTRA 1.5 array
reoccupied La RISTRA 1.0 sites and were used to combine
the residuals from both deployments into a single dataset.
Residuals were corrected for topography and crustal thick-
ness variations using the models of Wilson et al. [2005] and
Wilson et al. [in preparation, 2008] so that final residuals
reflect mantle velocity variations. La RISTRA 1.5 produced
767 P wave residuals and 559 S wave residuals. These data
were added to the identically processed 5007 P wave and
2164 S wave residuals from La RISTRA 1.0 [Gao et al.,
2004]. The mantle beneath the seismic line was parame-
terized into 1464 blocks 25 km wide and 25 km deep, and
the raypath corresponding to each time residual was back-
projected through the mantle assuming the IASPEI model.
The traveltime residuals were inverted using the LSQR
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algorithm [Paige and Saunders, 1982] to give a regular-
ized least squares solution for slowness perturbations
within the blocks relative to the IASPEI model following
the approach of Gao et al. [2004]. To examine model
resolution, synthetic traveltime residuals were computed
using the raypaths of the measured data for a checkerboard
model with 50 by 50 km blocks of alternating velocity
(Figure 2a) and were inverted using identical methodology.
Figures 2b and 2c show the recovered S and P waves
models for this resolution test. The centers and shallower
depths of the recovered models are well resolved, but the
model edges, lacking crossing raypaths, are significantly
less well resolved. Note that most blocks show recovery
of the sign of the checkerboard anomalies, but that
spreading of information between adjacent blocks gener-
ally reduces recovered anomaly amplitude relative to the
test model. We would expect the sharpness and the mag-
nitude of the anomalies to increase with 3-D ray tracing or finite
frequency kernel considerations. This would not change our
conclusions.

[5] The final Vp and Vs models from eastern New
Mexico into the GB are shown in cross section in
Figure 2d with areas of interest labeled A–E. Notable
features are the very slow P and S wave speeds in the
upper 200 km beneath the GB (A) and the very sharp
transition in seismic wave speed at the western edge of the
CP (B) where Vp and Vs change ~6% and ~12% respect-
tively over a horizontal distance of less than 100 km. The
eastern edge of the CP is also marked by a strong lateral
gradient from fast velocity beneath the CP to slow velocity
beneath the Jemez lineament (D) associated with the RGR
[Gao et al., 2004] although velocities beneath the Rift are
not as slow as beneath the GB. Interestingly, the core of the
CP is underlain by shallow mantle velocities that are slower
than its edges. Under the central CP the “4-corners anom-
aly” zone (in the region of the Navajo volcanic field) of
slightly lower than average velocity exists from 200 to
500 km depth (C). At depths below 400 km there is no
indication of a strong anomaly beneath the GB or the RGR.
Finally, the dominant deep anomaly is a zone of fast mantle from 300 to 600 km beneath the eastern GB (E).

3. Discussion

[6] The sharpness and magnitude of velocity change beneath the GB-CP transition indicate an abrupt change in mantle properties similar to what Zandt et al. [1995] found across the GB-CP boundary south of our line. Assuming that the velocity contrast is due to a sharp lateral temperature gradient and using the temperature derivatives given by Cammarano et al. [2003] (assuming a Qs from 80 to 140 over depths from 150 to 250 km), our model implies an average temperature contrast across the western edge of the CP of ~600°C averaged over the upper 200 km. However, temperature estimates from xenoliths [Riter and Smith, 1996] and gravity modeling [Wilson et al., in preparation, 2008] suggest a temperature change of only ~200–400°C.

Mantle compositional differences across the western CP boundary are likely insufficient to contribute substantially to the observed anomaly [Schutt and Lesher, 2006]. Hydrous minerals slow seismic velocities significantly but are stable at relatively low upper mantle temperatures (e.g. hornblende is stable up to 1000°C). Hacker et al. [2003] find that typical upper mantle depleted harzburgite contains stable hydrous phases only at temperatures less that 800°C. Smith [2000] found temperatures in the mantle 45 km beneath the GB to be >1000°C, thus ruling out hydrous mineral induced velocity reduction as the primary cause of the deeper slow velocities.

[7] Supersolidus conditions in the mantle result in melt between mantle grains which reduces the elastic moduli of the mantle and thereby the seismic velocity. Mavko [1980] and Hammond and Humphreys [2000] model realistic crystal-melt configurations and predict a large range in velocity reduction with a Vp reduction of 1 to 3.6% and a Vs reduction of 2 to 7.9% respectively for a 1% melt fraction indicating that melt in the mantle beneath the GB could contribute to the observed velocity contrast. Hydrogen in nominally anhydrous mantle minerals can also greatly reduce seismic velocities. Starting from a typical mantle (0.01 wt.% hydrogen and Q = 100), the average S wave velocity reduction is 8% for a 0.1 wt.% water content and 19% for a 1 wt.% water content [Karato, 2006]. Hydration of nominally anhydrous mantle minerals localized beneath the GB could thus also partly explain the observed velocity contrast across the GB-CP boundary. Magnetotelluric imaging of this region [Wannamaker et al., 2001] also shows very low mantle resistivity beneath the GB margin at depths below 100 km.

[8] We conclude that elevated temperatures combined with partial melting and/or high levels of hydrogen in nominally anhydrous minerals cause the extremely slow mantle velocities beneath the GB. Wyllie [1979] estimates a peridotite solidus drop of ~500°C at a depth 100 km for an addition of ~0.4% water. Considering the inferred history of flat-slab subduction in the western U.S. hydration of the shallow mantle beneath the GB would not be surprising. After slab rollback and asthenospheric upwelling [Humphreys et al., 2003], water-assisted melting of fertile lithosphere beneath southwestern North America appears inevitable. We propose the retention of ~1 to 2% melt and a temperature increase of ~300 to 500°C beneath the GB produces the strong gradient in seismic velocity across the GB-CP boundary.

[9] The contrast in seismic velocity across this boundary is not only large but is also very sharp, and our limited resolution inversion results will underestimate the true sharpness. Our smoothed images show the entire change in S wave velocity (~12%) occurring over less than 100 km laterally. Some of the contrast could be due to thinning of the lithosphere with subsequent asthenospheric upwelling under the highly extended GB, but the contrast in velocity extends over almost 200 km in depth. Even with strains approaching 100% for the GB, thinning due to extension can not explain the entire seismic anomaly across the boundary.

[10] O’Reilly et al. [2001] and Poudjom Djomani et al. [2001] (comparing compositions, ages, and strengths of 16,000 global mantle xenocrysts) found lithospheric buoyancy and refractoriness to increase with age. Therefore, under certain conditions Proterozoic lithosphere may resist disintegration while partial melting of fertile Paleozoic lithosphere leads to enhanced heat advection through melt migration and ultimately to pervasive lithospheric weakening. The response of juxtaposed Paleozoic and Proterozoic mantle lithosphere to heat and water input could therefore result in the present-day contrast beneath the GB-CP transition. We conclude that our image reveals the western extent of unadulterated Proterozoic mantle lithosphere which is maintaining the stability of the western edge of the CP and that the factors controlling lithospheric stability in this area were established by preexisting differences between the Proterozoic and Paleozoic mantle lithospheres that meet at our inferred boundary. While Precambrian crust, possibly underlain by a thinning sliver of Paleoproterozoic upper mantle lithosphere, clearly extends west of our boundary to the SR 7076 line [Burchfiel et al., 1992], we believe the deeper mantle lithospheres separated by our inferred boundary are distinct. The narrow high velocity zone bounding the edge of the CP may also be a manifestation of edge driven convection as proposed by van Wijk et al. [in preparation, 2008].

[11] The fast anomaly imaged from 300 to 600 km beneath the NW portion of the array (E) has Vp and Vs anomalies at 400 km of up to 2% and 4% faster than the surrounding mantle respectively, implying a temperature anomaly of ~400°C [Cammarano et al., 2003]. Though checkerboard tests indicate that the model amplitude and precise shape of this anomaly are not reliable, its existence as a strong and large (250 by 250 km) feature is not in doubt. Van der Lee and Nollet [1997] also find transition zone seismic anomalies in this region and interpret them as arising from the trailing edge of the Farallon plate. Schmid et al. [2002] models the present-day thermal signature of the Farallon plate and predicts that fragments remaining in the upper mantle retain a thermal anomaly of 200–400°C, consistent with anomaly (E) being the Farallon plate. The volume of the anomaly, however, is too large to be a simple descending oceanic plate of relatively young age. Internal plate deformation due to resistance at the endothermic
~660 km phase change [Schmid et al., 2002] could distort the slab to produce the volume of material necessary to give the observed anomaly.

[12] Humphreys [1995] observes two migrating areas of magmatism in western North America converging to southern Nevada around 20 Ma and correlates these fronts with the edges of the foundering Farallon plate. In this model the slab buckles in a concave downwards fashion forming a locus of downwelling plate. This area of convergence is spatially associated with the large deep anomaly imaged in this study. Continental mantle lithosphere delaminated from beneath the GB could also contribute to the deep anomaly. Paleozoic mantle lithosphere is gravitationally unstable under normal mantle temperatures and geotherms [O’Reilly et al., 2001; Poudjom Djomani et al., 2001]. Therefore, gravitational instability enhanced by lithospheric cooling during flat-slab subduction might have resulted in delamination of GB mantle lithosphere as low-density upwelling asthenosphere replaced the foundering Farallon plate.

[13] Figure 3 shows the present-day mantle state interpreted from our tomograms. The sharp velocity contrast across the GB-CP transition corresponds to a boundary between altered GB Paleozoic mantle lithosphere, or possibly juvenile mantle that replaces older delaminated lithosphere, juxtaposed with largely unaltered CP Proterozoic mantle lithosphere. A thinning sliver of Precambrian upper mantle lithosphere may extend west of our inferred boundary. Beneath the eastern GB the Farallon plate or delaminated mantle lithosphere is sinking in the transition zone. Counter-flow associated with the sinking fast anomaly may be coming up beneath the CP and perhaps feeding hot mantle to the GB to the west and the RGR to the east.

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