

Upper mantle *P* wave velocity structure of the eastern Snake River Plain and its relationship to geodynamic models of the region

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Abstract. Tomographic inversions of ~5000 teleseismic *P* wave travel time residuals image a narrow, deep, low-velocity region centered beneath the eastern Snake River Plain, Idaho. Aligned in the direction of North American plate motion, the eastern Snake River Plain is the locus of time-progressive volcanism leading to the Yellowstone hotspot. The low-velocity anomaly extends to depths of at least 200 km and is flanked by high-velocity mantle to either side. These results are inconsistent with standard mantle plume models which predict a shallow, wide, low-velocity anomaly. By considering the effects of composition, partial melt, and temperature on both *P* wave velocities and upper mantle densities (assuming local isostasy), we conclude that the low-velocity region is partially molten peridotite and the high-velocity regions on either side are depleted residuum. Swell relief at hotspots has generally been attributed to thermal effects and is commonly used to estimate plume thermal buoyancy fluxes. However, we demonstrate that the compositional effects of mantle melting and depletion can account for at least one fifth and possibly all of the swell relief at Yellowstone. These results suggest that melt generation and segregation are dominant processes within the upper mantle of Yellowstone and allow for the possibility that the fundamental origin of the Yellowstone hotspot is something other than a deep-seated mantle plume.

Introduction

In many respects, Yellowstone is a classic hotspot. It is located at the leading edge of a linear, 16 million year old, age-progressive volcanic system [Armstrong *et al.*, 1975]. Heat flow is anomalously high [Brott *et al.*, 1981]. The region surrounding the hotspot track is uplifted in a parabolic or bow-shaped form similar to swells at oceanic hotspots [Crough, 1983; Pierce and Morgan, 1992] which is associated with active faulting and seismicity. And like most other hotspots in the world, it is at the center of a geoid high. Yellowstone might be considered unusual only by virtue of its continental location, which affords it both greater accessibility as well as potentially greater complexity (since the lithosphere through which it is erupting has been previously affected by various orogenies and other large-scale tectonic events). Consequently, Yellowstone, like other hotspots, has been modeled as a thermally buoyant mantle plume impinging on the base of the lithosphere [e.g., Sleep, 1990].

The basic premise of the plume model is that hot buoyant mantle flowing continuously through a conduit will pond at the base of the lithosphere in a disk approximately 100 km thick. Motion of the overriding plate will continuously drag the plume material away from the site of ascent thereby modifying the radial flow to one that is parabolic in shape [Anders and Sleep,

1992; Sleep, 1990; Ribe and Christensen, 1994]. Support for these ideas comes from the observation that the uplifted, tectonically and seismically active region surrounding Yellowstone, the so-called "tectonic parabola" [Anders and Sleep, 1992], coincides with the theoretically expected boundaries of asthenospheric plume flow. The topographically depressed region within the interior of the parabola, the eastern Snake River Plain (SRP), is attributed to crustal thinning from Basin and Range extension and magmatic loading of the crust [Anders and Sleep, 1992].

Although the observations and plume modeling are remarkably consistent and attractive, our tomographic images of the mantle structure beneath the volcanic track leading toward Yellowstone differ significantly from what is predicted by current hotspot models. These models predict a flattened zone of upper mantle that is 100°–200°C hotter than ambient asthenosphere [Ribe and Christensen, 1994; Sleep, 1990] that extends the full width of the swell and has a thickness of less than 100 km. Such a zone should appear as a shallow, wide, low-velocity seismic anomaly, whereas we image a narrow deep anomaly directly below the site of the time-progressive volcanism.

We discuss inversion results for data collected during a 1993 passive seismic experiment as part of the PASSCAL Program (Program for Array Seismic Studies of the Continental Lithosphere) in which seismometers were deployed across the entire width of the tectonic parabola. Previous studies in the vicinity of our experiment [Iyer *et al.*, 1981; Evans, 1982; Dueker and Humphreys, 1990] found a narrow zone of anomalously slow *P* wave velocities extending to depths of at least 150 km; however, crossing rays in the new data set extend to 400 km (almost twice as deep), permitting an improved determination of the depth extent of the anomaly. In addition, the longer array and greater density of seismometers in our study allow

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improved resolution of the lateral extent and magnitude of the low-velocity mantle.

Data Acquisition and Analysis

From May through November 1993, the University of Oregon collected teleseismic earthquake data from a 550 km array of PASSCAL broadband and high-frequency seismometers that were arranged in a line oriented N45°W (Figure 1) with an average station spacing of 10 km. This alignment was particularly advantageous because it was both perpendicular to the track of the hotspot and oriented toward those regions of the globe where large earthquakes are common (Japan and Alaska to the northwest and South America to the southeast) thereby allowing excellent resolution of the upper mantle velocity structure in the vertical cross section directly beneath the array. The array crossed the eastern Snake River Plain 250 km southwest of the current location of the Yellowstone hotspot. It spanned the entire width of the tectonic parabola in a location beneath which the hotspot passed 6 to 8 million years ago. During the 6 month duration of the experiment, 375 events of magnitude (m_b) 4.5 or greater occurred (Figure 2) from which ~5000 usable P wave residuals were derived.

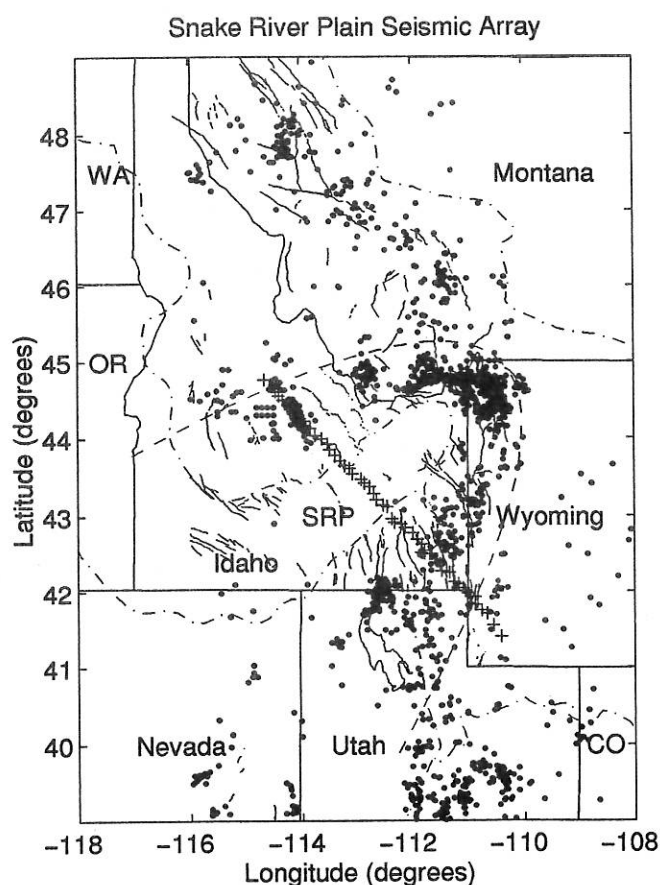


Figure 1. Location map. The eastern Snake River Plain (SRP) is a topographically depressed, 90 km wide swath of silicic calderas that show 16 m.y. of time-progressive volcanism. Aligned parallel to absolute North American plate motion, the eastern SRP defines the track of the Yellowstone hotspot. An array of 55 seismometers (crosses) was arranged in a line perpendicular to this track and spanned the entire width of the parabolic shaped region of active faulting (thin black lines) and earthquakes (dots).

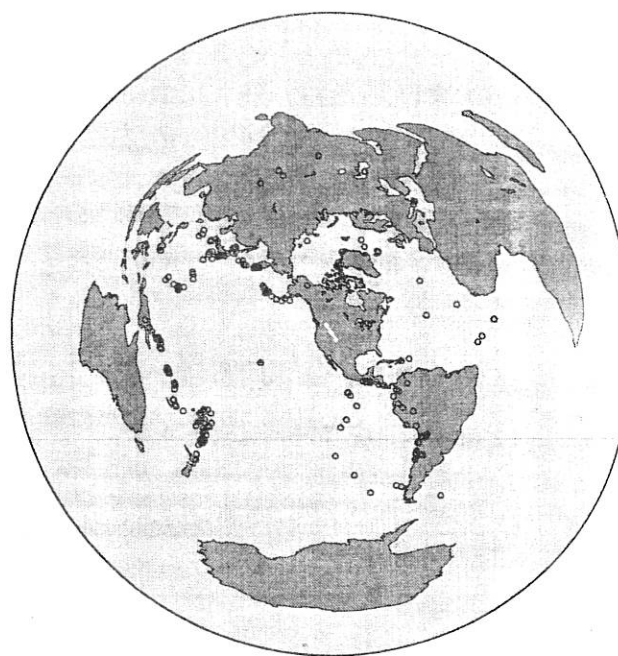


Figure 2. Event location map. During the 6 months of the experiment, 375 events of magnitude $m_b \geq 4.5$ occurred around the globe from which we derived ~5000 travel time residuals. The array of seismometers is shown as a white line centered and oriented correctly but with an exaggerated length.

Both high-frequency and broadband sensors were deployed simultaneously during the experiment. To create a uniform data set from which consistent arrival times could be obtained, we deconvolved the instrument responses from the seismograms before band pass filtering between 0.5 Hz and 2.5 Hz. We made necessary timing adjustments (e.g., leap seconds and "clock jerks") and reversed the polarity of inverted seismograms, on the basis of a careful checking of field notes, log files, data acquisition error files, and waveforms. Relative arrival times were then determined using a cross-correlation technique.

For initial alignment, the seismic traces were windowed in time to 6 s lengths, with several seconds of noise followed by the first P wave arrival. A reference trace was selected for each event, and all other traces were cross-correlated against it. The shifted traces were then visually inspected so that errors due to cycle skipping could be corrected manually and noisy or otherwise unpickable traces could be discarded. The window was then narrowed to less than a second (no more than one cycle) so that the traces could be realigned according to the first arrival rather than the first large peak. The teleseismic waveforms recorded in our study area are very similar so that this travel time picking method worked well for all P wave phases used in this study (direct P , P_{diff} , and core phases). Three different sample rates were used over the course of the experiment, however, the uncertainty in arrival times is estimated at less than the largest sample interval, 0.1 s. We used the IASPEI91 radial earth model to calculate theoretical travel times to each receiver and then took the difference between those theoretical times and the actual arrival times to obtain a travel time residual. The residuals were then demeaned as a first-order correction for differences between sources, as well as to remove the effects of mantle differences outside the region of interest [e.g., Dueker et al., 1993].

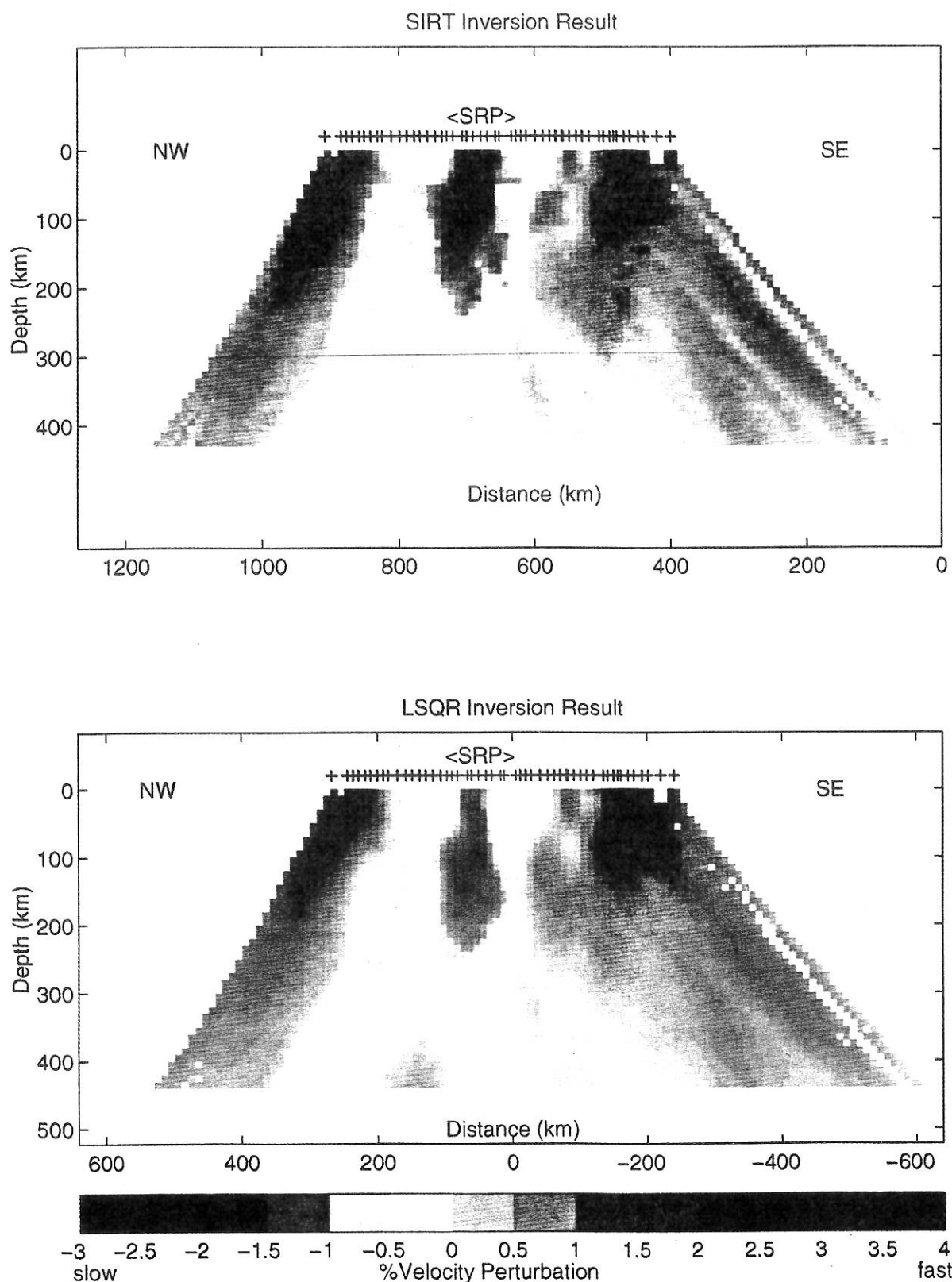


Plate 1. Comparison of SIRT and LSQR inversion results. Both are row-reduction methods for inverting large, sparse matrices. They differ in their weighting of the ray-length matrix elements when determining the magnitude of the velocity perturbation to make in each block. SIRT gives more weight to poorly hit blocks than does LSQR. Despite this difference, both methods produce very similar results, and SIRT can be made to converge to the LSQR solution with appropriate weighting parameters. The similarity of the two methods can be seen in the example shown here, even though no special attempt was made (through choice of damping and smoothing parameters) to make the inversions as similar as possible.

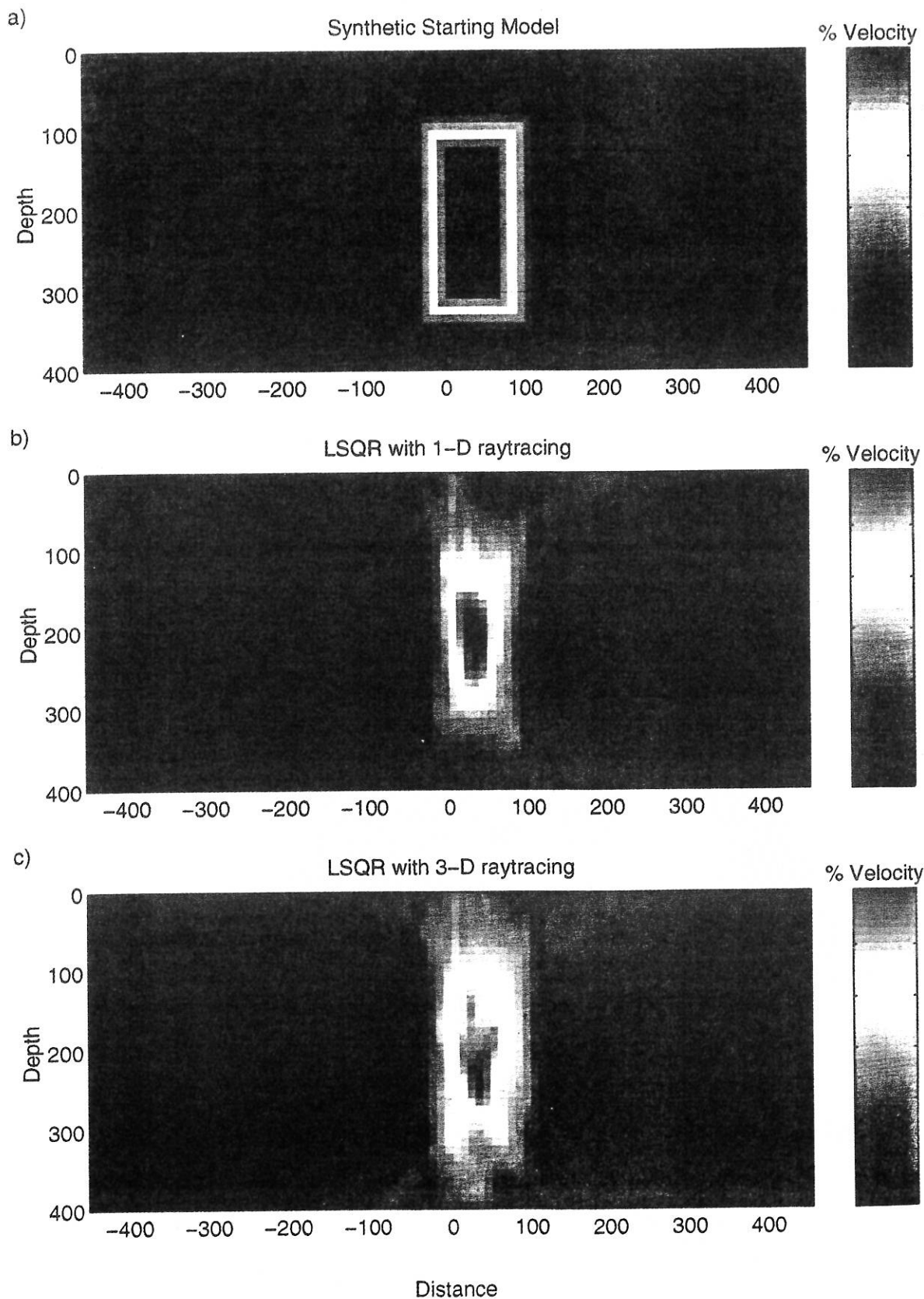


Plate 2. Comparison of ray tracing methods. Synthetic residuals are produced by ray tracing through the synthetic starting model (a) with the three-dimensional (3-D), nonlinear ray tracer. These residuals are then inverted to see how well the image is reconstructed. Inversions using a one-dimensional ray tracer (b) tend to image structure slightly shallower and narrower than is correct while the 3-D ray tracer (c) tends to image the width and depth of the structure more accurately.

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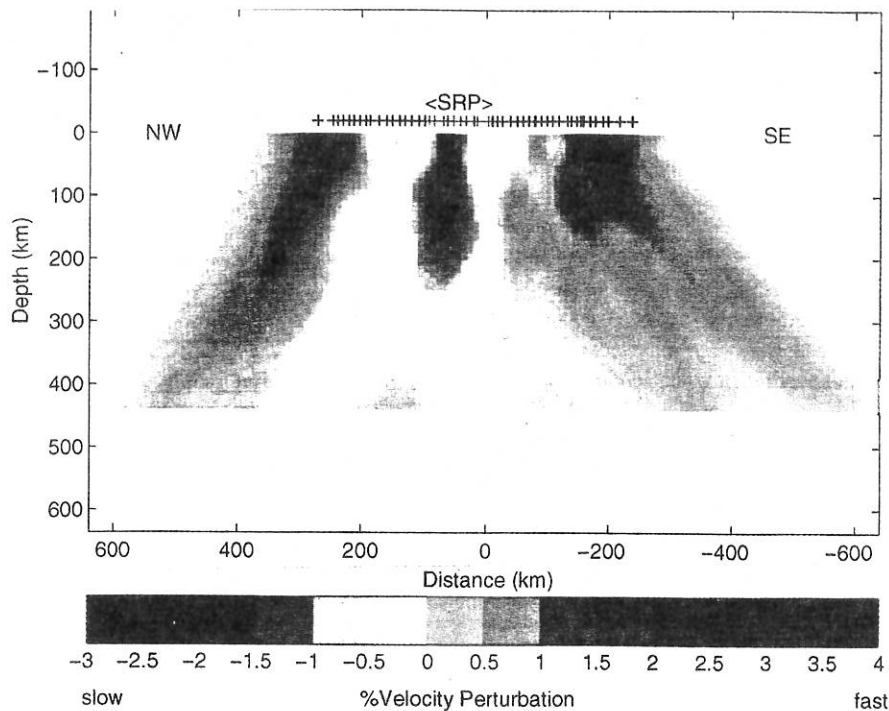


Plate 3. "Typical" tomographic image. Cross section of upper mantle *P* wave seismic velocity structure directly beneath the array of seismometers (crosses). Warm colors (oranges and reds) represent regions that are slower than the one-dimensional starting velocity model, and cool colors (blues and greens) represent regions that are faster. Seismometers located within the eastern SRP volcanic track are plotted in lighter blue. Plate 3 was derived using the nonlinear, 3-D ray tracer and an LSQR (conjugate gradient method) inversion algorithm. Comparison with Plate 2b shows how insignificant the differences are between results obtained using the 3-D ray tracer (Plate 3) and the 1-D ray tracer (Plate 1b).

Inversion

For determining the velocity model that best satisfied the travel time residuals, we used two inversion algorithms, the simultaneous iterative reconstructive technique (SIRT) [Humphreys and Clayton, 1988] and the conjugate gradient method (LSQR) [Paige and Saunders, 1982]. Both methods produced very similar results (Plate 1). The starting velocity structure for the upper 50 km is based upon the crustal refraction profiles of Sparlin *et al.* [1982] and Braile *et al.* [1982] that have been modified by Peng [1996] using receiver function analysis of data recorded by our array. The starting velocity structure for the remainder of the model is the *P* wave structure derived by Walck [1984] for the Gulf of California. We chose this model in particular because it, like western U.S. mantle, tends to be slower on average than the typical whole Earth models. Nonetheless, we also tried inversions with the IASPEI91 radial Earth velocity model and found no discernible differences between the results of the two inversions. The RMS of the data prior to inversion is 0.42 s and after inversion is 0.10 s, corresponding to a 94% variance reduction. Event statics and station statics were estimated during inversion following Humphreys and Dueker [1994]. Variance reduction is 87% without the application of event statics, and the only significant station static is 0.38 s for a station that was located on top of a deep sedimentary basin.

We also explored the effects of ray tracing. We used Snell's law to trace rays through the initial, one-dimensional layered

velocity structure and compared these results with those obtained using a three-dimensional, shortest-path ray tracer based upon the work of Moser [1991], which allows the bending of rays by the imaged structure (modified so that a ray enters the bottom of our inversion domain with the IASPEI91 predicted ray parameter). Synthetic tests of structures with large lateral velocity gradients show that the one-dimensional ray tracer tends to reconstruct low-velocity structures slightly shallower and narrower than is correct, whereas the nonlinear, three-dimensional ray tracer does a better job (Plate 2). The errors in tomographic imaging with one-dimensional ray tracing become progressively pronounced with increasing magnitudes of anomaly and velocity gradient. For our data, however, the effects of ray bending on the inversion are minor (compare Plate 1b with Plate 3), and the trade-off in computational time (15 s versus 7 hours) is severe. In our synthetic test examples, one-dimensional ray tracing misplaces rays horizontally at most by 10 km at the bottom of the model, and typical misplacements in the upper few hundred kilometers (around the low-velocity anomaly) are less than 20 km. We attribute the lack of importance in the particular ray tracing algorithm used to the relatively small lateral velocity gradients actually encountered in the mantle.

Results of Seismic Velocity Modeling

Teleseismic waves travel steeply toward the seismometers and generally do not attain angles greater than 45 degrees from

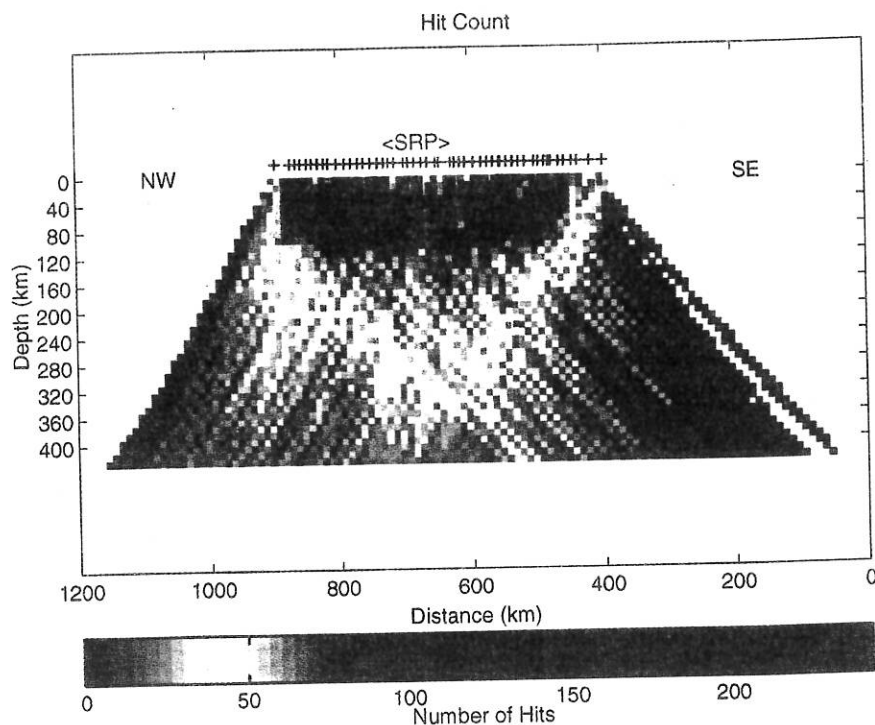


Plate 4. Number of times each block was hit by a ray. The inverted triangle in the center (predominantly green) shows where the rays from the circum-Pacific events cross, hence, where the best resolution is obtained. Outside of this region, where there are far fewer crossing rays, streaking will be greatest and resolution poorest. Of the various resolution tests we examined, the hit count provides the best overall summary of resolution.

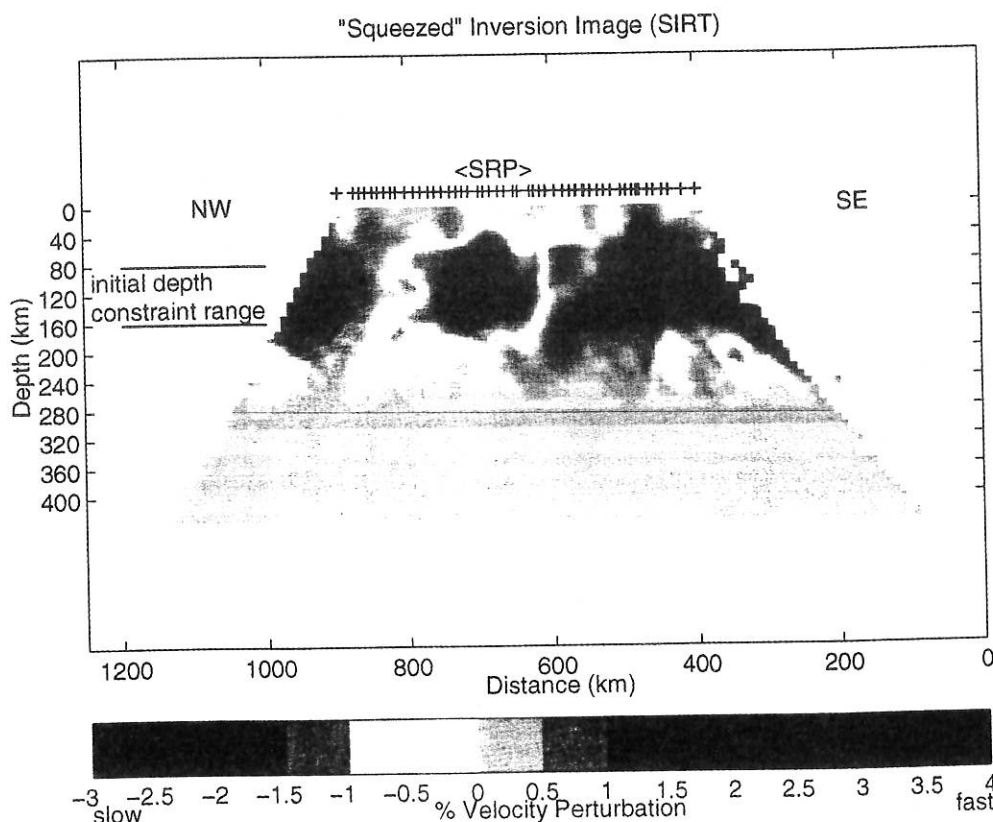


Plate 5. "Squeezed" tomographic image. Cross section of upper mantle *P* wave seismic velocity structure beneath the array of seismometers (crosses) derived with SIRT inversion algorithm. Velocity perturbations are initially limited to the region between 80 km and 160 km depth. In subsequent iterations, that restriction is progressively relaxed to allow perturbations both deeper and shallower. Consequently, structure outside of the initially restricted region will appear only if the model is unable to explain the data with the structure already

vertical in the uppermost mantle that we are modeling. Consequently, lateral resolution of the tomographic images is better than vertical resolution, since there is an inherent tendency for streaking in the directions that the rays travel. This streaking is a result of the minimum model energy constraint that is implicit in standard teleseismic inversion. Tomographic images derived with these standard techniques (Plate 3) show a 125 km wide low-velocity zone below the eastern SRP that extends to depths of ~325 km. The SE boundary of the anomaly is aligned with the SE boundary of the eastern SRP, whereas the NW boundary of the anomaly extends ~40 km from the NW boundary of the eastern SRP. This zone is flanked on either side by relatively high-velocity mantle. These velocities, which are relative in our inversion, can be put in a global reference frame. *Humphreys and Dueker* [1994] compared teleseismic travel time residuals from across the western United States (including the eastern SRP) with the International Seismological Centre (ISC) catalog of station residuals [*Dziewonski and Anderson*, 1983] and found that the relatively high velocities on either side of the Plain are slightly faster than the ISC global averages (~0.25 s) and that the low-velocity region in the center is much slower than the global averages (~1.25 s).

In order to understand which aspects of this imaged structure are resolved, we have tested resolution in a variety of ways. Traditionally, resolution tests focus on the ability to resolve single anomalous blocks, which is assessed with resolution kernels [*Backus and Gilbert*, 1968] or, equivalently, with point spread functions [*Humphreys and Clayton*, 1988]. We ran "checkerboard tests" to evaluate point resolution ability as a function of position within the model domain and found that resolution is very good in the region of the model where there are many crossing rays (i.e., single-block tests reconstruct >70% of the anomalous velocity). The crossing rays form a triangular-shaped region beneath the seismic array. This region can be seen clearly in Plate 4 which shows the "hitcount," the number of times each block is hit by a ray.

A more specific resolution test uses travel time delays calculated by ray tracing through a prescribed structure and then inverting these delays for structure. By comparing the resulting image to the original structure, one can assess the ability of the inversion to resolve the prescribed structure. We performed a number of such tests (e.g., Plate 2), again finding that the triangular region beneath the array is the area best resolved and that broad features extending beyond this region tend to be streaked to greater depth. The low-velocity anomaly imaged in Plate 3 is within this well-resolved region of our model domain.

Standard tomographic techniques (i.e., minimum model energy methods such as were used to obtain Plate 3) produce images of the most dispersed structure that best explain the data in a least squares sense. Since our goal is to compare our seismic results with geodynamical models of a flattened mantle plume, the shallowest, most vertically compact structure satisfying the travel time residuals is of greater interest than the maximally dispersed structure. The most compact structure provides a minimum depth to which the low-velocity anomaly must extend, which, in this case, is the most conservative estimate of the difference between the seismic observations and the geodynamic model predictions.

To determine this minimum possible depth, we ran a set of "squeezing" experiments in which we limited perturbations to an arbitrary range of depths for the first iteration of the SIRT inversion and then progressively relaxed that restriction for sub-

sequent iterations. Conceptually, the goal is to find a "best" model within a spatially restricted domain and then to permit any travel time data that cannot be modeled within this depth range to reconstruct structure elsewhere. This solution differs from the minimum model energy solution (Plate 3) but still satisfies the data equally well in a least squares sense. We found that if we initially restricted velocity perturbations to depths of less than 200 km, the low-velocity zone expanded downward once the depth constraint was released. However, when we constrained the base of the perturbations to depths of 200 km or greater, the structure remained essentially unchanged after the constraint was released (Plate 5). This exercise demonstrates that the data can be satisfied by a "squeezed" velocity model in which the base of the low-velocity zone is ~200 km deep.

In order to test further the limits of this squeezing, we restricted the perturbations to the bottom 10 km of the model and then slowly allowed perturbations to grow upward. For the other end-member, we restricted perturbations to the top 10 km of the model and then slowly relaxed that restriction to allow downward growth of perturbations. While these extreme constraints allowed some energy to be distributed along various parts of the bottom or top of the model space (depending upon the particular starting restrictions), a strong, low-velocity anomaly remained below the eastern Snake River Plain. Regardless of the initial perturbation restrictions, we always found a low-velocity anomaly beneath the Plain extending to depths of at least 200 km. We conclude, as a result of these tests, that the minimum depth to which the low velocity anomaly must extend is ~200 km depth.

The width and horizontal position of the anomaly are far less ambiguous. Synthetic tests reveal that the particular geometry of our source and receiver positions is excellent for resolving lateral velocity variations very accurately. The smearing that is inherent in the vertical direction is virtually nonexistent in the lateral direction. We conclude that there is a low-velocity anomaly beneath the eastern Snake River Plain. It extends to a depth of at least 200 km ("squeezed" solution) and probably no more than 325 km ("typical" minimum model energy solution). The SE boundary of this anomaly is aligned with the SE boundary of the eastern SRP, whereas the NW boundary lies approximately 40 km to the NW of the eastern SRP. The peak-to-peak magnitude of the velocity anomaly is 7% in the "squeezed" solution and 5% in the "typical" solution.

Scaling Seismic Velocities to Physical Properties

Our seismic image of the upper mantle beneath the eastern SRP region provides information on the physical state of the upper mantle near the Yellowstone hotspot. Upper mantle P wave velocity variations (V_P) may be caused by the following: (1) anisotropy, (2) temperature variations, (3) partial melt, and (4) compositional differences. We discuss below how each of these factors might contribute to the low-velocity anomaly we image.

Anisotropy

Alignment of anisotropic crystals, such as olivine, can cause V_P to vary depending upon the direction of wave propagation. Shear wave splitting measurements [*Schutt et al.*, 1995] indicate that anisotropy is present but essentially uniform along the entire length of the array: the fast axis is oriented parallel to the

track of the hotspot and travel time differences between the slow and fast directions are generally about 1.2 s. Hence, all rays for any given event sample the anisotropic structure similarly. Since differences between events are absorbed by event statics computed during the inversion, anisotropic effects on the P wave relative delay times should be minimal.

Temperature Effects

The relationship between changes in temperature T and V_p is somewhat controversial, as reviewed by Sobolev *et al.* [1996]. A temperature-velocity relation of

$$\frac{\partial V_p}{\partial T} = -0.5 \text{ m s}^{-1} \text{ K}^{-1} \quad (1)$$

is suggested by two independent methods: (1) from modeling of wave propagation along subducting slabs [Creager and Jordan, 1986] and (2) from extending the theoretical results of bulk and shear modulus measurements to depth with finite strain calculations [Anderson and Bass, 1984; Duffy and Anderson, 1989]. This value corresponds to a 1% reduction in velocity with every 160°C temperature increase. Therefore an 800°C temperature difference would be required in order to explain a 5% velocity anomaly by temperature alone. Not all workers agree, however, that (1) is appropriate for the upper mantle.

Experimental results of Sato *et al.* [1989] indicate that V_p declines by as much as 6% in mantle peridotite at temperatures within 85% of the solidus (i.e., within 100° to 200°C of solidus at the depths of interest here), permitting the entire low-velocity anomaly we image to be a result of temperature variations alone. Karato and Spetzler [1990] question the validity of these results on the basis that they were obtained at very high frequencies without a sufficient investigation of the frequency dependence of those measurements and that a simple extrapolation to seismic frequencies may not be valid.

The main problem with using (1) is that it is based upon the assumption that seismic waves are propagating through an elastic Earth; hence it does not account for velocity reductions associated with attenuation. If anelastic effects are significant, temperature changes as small as 50°C (for quality factor $Q < 100$) could cause a 1% reduction in V_p [Karato, 1993]. However, when temperature-dependent relaxation is important, waves propagating through hotter, lower velocity material experience greater attenuation. The data for the eastern SRP do not show any obvious effects of greater attenuation (low Q) for rays passing through the low-velocity region.

Partial Melt

The effect of partial melt on seismic velocities depends upon the geometry of the melt, which can be parameterized as

$$\frac{\Delta V_p}{V_p} = A \phi$$

where ϕ is the melt fraction and A is a constant dependent upon the aspect ratio of the melt inclusions [Schmeling, 1985; Mavko, 1980]. Generally, $1 < A < 3$ unless the melt is distributed in very thin films, in which case A can be larger. Therefore the presence of 2% partial melt can give rise to a V_p reduction of 2% to 6% or more, depending upon the melt geometry.

Composition

When upper mantle peridotite is partially melted, the iron-

magnesium-rich residuum (forsterite or Fo) that is less dense but of higher seismic velocity [Jordan, 1979]. A change in peridotite composition from Fo_{88} to Fo_{92} (~10% basaltic melt extraction) would cause a 1% increase in V_p [Jordan, 1979].

Interpretation

We interpret the low-velocity upper mantle to be partially molten peridotite and the neighboring high-velocity upper mantle to be peridotite that has been depleted of a basaltic component because of melt segregation. This interpretation is consistent with both the seismic velocity structure and the local density structure, assuming isostasy.

Excess plume temperatures are generally modeled to be less than 300°C [e.g., Sleep, 1990; Ribe and Christensen, 1994] on the basis of petrological constraints. Therefore, depending upon which value of $\partial V_p / \partial T$ is used, either a maximum of 2% of the velocity anomaly can be explained as the result of mantle temperature variations (using the relation defined in (1)) or virtually the entire anomaly can be explained as thermal variations (using the Sato *et al.* [1989] relation). Anelastic effects modify the temperature derivatives to values partway in-between [Karato, 1993]. However, regardless of which value of $\partial V_p / \partial T$ is chosen, an interpretation that lateral temperature variations are responsible for the seismic velocity variations is, at the same time, an interpretation that the faster regions are colder than the slower regions. Faster, colder regions should also be comparatively dense. To illustrate the magnitude of this effect, Figure 3 shows the results of some isostatic calculations. The gray lines in Figure 3a show the topography that would be produced by assuming the imaged velocity structure were due entirely to lateral temperature variations. The dashed line is derived with (1) whereas the solid gray line is derived from an anelastic $\partial V_p / \partial T$ derivative (i.e., Karato [1993]) assuming a low $Q = 70$. For comparison we have taken the actual topography and subtracted the crustal load using the densities inferred from the crustal refraction profiles [Braile *et al.*, 1982; Sparlin *et al.*, 1982] and the crustal thickness variations determined in the receiver function analysis of our teleseismic dataset [Peng, 1996]. This figure demonstrates that large lateral temperature variations in the mantle would give rise to a topography that is inconsistent with the actual topography (once the density effects of the crust have been removed).

Lending further support to these ideas, Figure 3b shows the similarity between the actual elevation (stippled black line) and the elevation predicted by the crustal density structure (again using the results of the seismic refraction profiles and crustal thickness variations [Braile *et al.*, 1982; Sparlin *et al.*, 1982; Peng, 1996]). Within the tectonic parabola, crustal density variations account for the observed variations in elevations well, although the actual elevation is about 1 km higher than is predicted by the crustal density structure. This result suggests that large mantle density contrasts within the tectonic parabola do not exist, and that instead the mantle within the parabola is uniformly buoyant. Large lateral temperature variations are inconsistent with a uniformly buoyant mantle. Moreover, the crust southeast of the tectonic parabola is thicker than the crust beneath the tectonic parabola and lies at an elevation that is more than 0.5 km lower than predicted. This elevation difference suggests that the mantle beneath the tectonic parabola is more buoyant than mantle southeast of the tectonic parabola by an amount equal to the load of ~1.5 km of crust (as shown

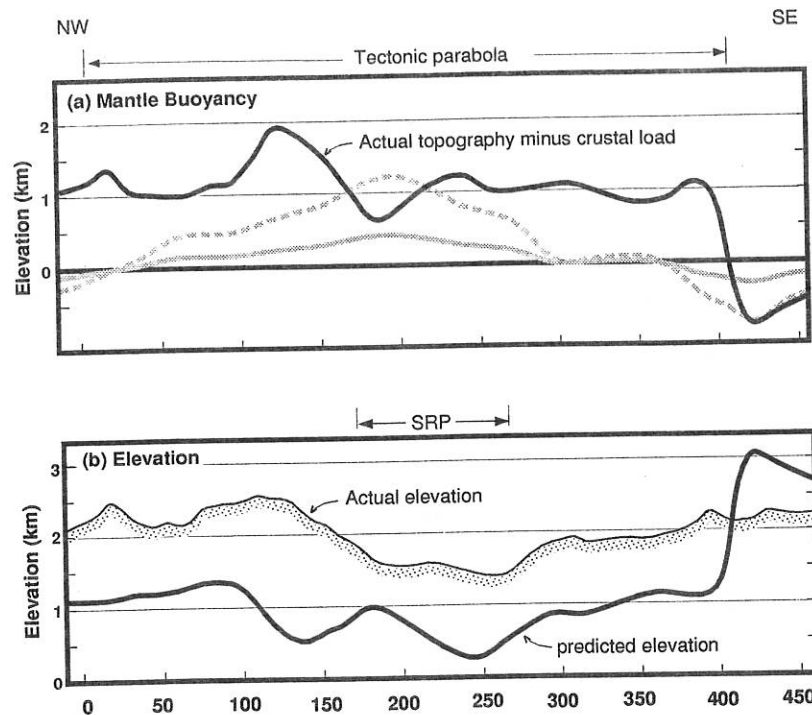


Figure 3. Mantle buoyancy calculations. In (a) we show the predicted variations in topography if the upper mantle velocity anomaly results solely from lateral temperature differences. The solid black line shows the actual topography with the crustal load subtracted. The gray lines show the temperature-predicted topography with two different temperature-velocity derivatives. The dashed gray line is derived from equation (1) and the solid gray line is derived from the anelastic derivative of Karato, [1993]. The large differences between the temperature-predicted topographies (gray) and the actual topography minus the crustal load (black) imply that upper mantle temperature variations are not significant and therefore not the origin of the observed velocity variations. In (b) we compare the actual elevation (stippled black line) with the elevation predicted by the crustal structure (assuming isostasy), with the mantle buoyancy of the United States seaboard along the eastern and southern coastal areas serving as a reference elevation.

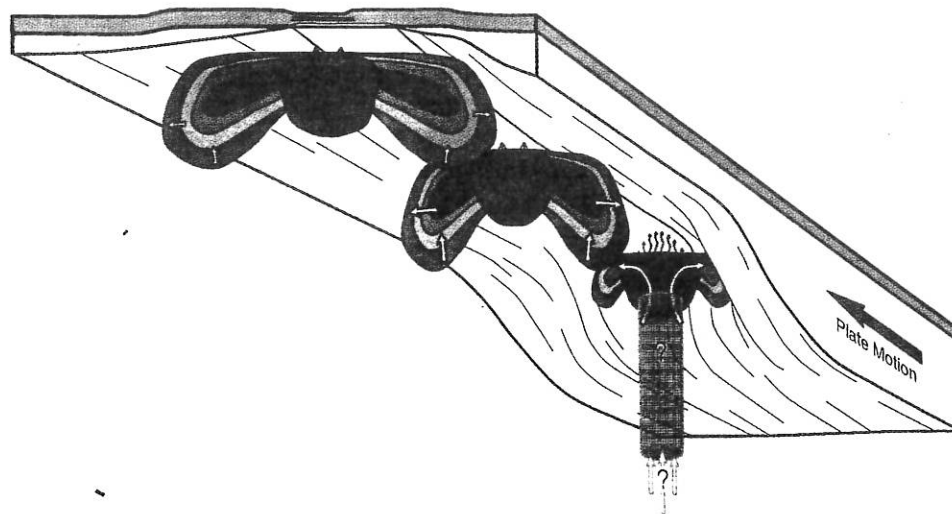


Plate 6. Geodynamic schematic, viewed from below. A partially molten core (red) of approximately constant width and depth is flanked on both sides by depleted residuum (blue) which widens and flattens with distance from the hotspot. Any excess pressure at Yellowstone will drive partially molten mantle to flow in the direction of plate motion, along the existing low velocity channel, similar to a river. The hotspot is currently impinging on the craton, which may enhance this flow.

In sum, we find great seismic velocity variations where there is no indication of lateral buoyancy contrasts (across the width of the tectonic parabola) and we find no significant velocity variations where we infer great buoyancy contrasts (across the southeast margin of the tectonic parabola). Therefore we conclude that the seismic velocity variations we see tomographically are not principally the result of lateral temperature variations.

Absent temperature variations, the only known mechanisms affecting upper mantle seismic velocities are anisotropy, partial melt, and compositional differences. As explained in the section on anisotropy, anisotropy while present, is oriented uniformly parallel to absolute plate motion (so that it should have no effect on travel time arrivals for a given event). Likewise, compositional effects can account for no more than an additional 1% variation. Because partial melt is the only other significant mechanism likely to be operating over the volume being considered, we conclude that it is responsible for the majority of the seismic velocity variations.

While the interpretation of a region of partial melt can explain the seismic velocity variations, it cannot explain the inferred upper mantle density structure. The high-velocity regions in our image lie beneath a normal thickness crust (~40 km) [Sparlin *et al.*, 1982; Braile *et al.*, 1982; Peng, 1996] that stands at great elevation, implying that the upper mantle is unusually buoyant. Relative to typical continental lithosphere, the crust within the tectonic parabola stands at an elevation that is ~1 km higher than would be predicted by the crustal load (see Figure 3b). These elevations are what might be expected for something closer to 10 million year old oceanic lithosphere. The simplest explanation for this high-velocity, buoyant upper mantle is that it is depleted residuum created by mantle melting during passage of the hotspot and subsequent extraction of basalt. Compositional variations are consistent with the tomographic image as well, because depleted residuum has both a slightly higher seismic velocity than fertile mantle peridotite and contains a smaller melt fraction at any given temperature. Therefore we infer that the partial melt variations are modulated by compositional variations which simultaneously explains both the seismic and density structures.

This interpretation is supported by crustal studies [Braile *et al.*, 1982; Sparlin *et al.*, 1982; Peng, 1996] which indicate a high-velocity midcrustal layer beneath the eastern Snake River Plain that is interpreted to be a basalt sill. This layer is approximately 100 km wide and 10 km thick. Formation of such a sill by partial melting of peridotite would have left behind a large volume of depleted mantle residuum (~10 times the volume of the sill). Therefore it is likely that a large volume of depleted upper mantle exists near the eastern SRP. This depleted mantle would not contribute significantly to imaged seismic velocity variations but would have a significant effect on the upper mantle density structure.

On the basis of the above arguments, we conclude that both the seismic image and the topography are consistent with our interpretation that minor temperature variations, at most, exist between the high-velocity and low-velocity regions of our tomographic image and that the velocity contrasts between these regions are due largely to partial melt variations modulated by compositional variations. Indeed, we are hard-pressed to find an alternative explanation that simultaneously accounts for both the topography and seismic structure.

Discussion

The above conclusions are important in several ways. First, the structure predicted by traditional plume models is not what is observed. The most likely explanation for this difference is that plume models solely consider thermal convection effects [e.g., Davies, 1994; Kiefer and Hager, 1991; Ribe and Christensen, 1994], neglecting melt removal and the compositional and geodynamical changes that accompany the creation and removal of partial melt. Nonetheless, the observed structure is plausible in the genetic context of a mantle plume. Our inference of large volumes of depleted upper mantle adjacent to a partially molten core simply suggests that the dynamic processes associated with plume-lithosphere interaction are more complicated than current models allow.

Second, swell relief is not solely thermal in origin. Rather, there is a buoyancy contribution resulting from depletion of a basaltic component from the mantle. A minimum estimate of this contribution can be made by considering the amount of depleted residuum that must have been left behind in the mantle as a result of extracting basaltic melt to create the midcrustal sill. For a region in isostatic equilibrium, the rate at which mass deficiency is created at depth equals the rate of excess mass produced near the surface through uplift. This rate is termed the buoyancy flux [Sleep, 1990] and the compositional contribution to this flux, B_{comp} , can be estimated from the rate of basalt emplacement, as represented by the midcrustal sill:

$$B_{comp} = \frac{\partial \rho}{\partial F} A V \quad (2)$$

for basalt of cross-sectional area A , plate velocity V , and the rate at which mantle density decreases with melt segregation $\partial \rho / \partial F$ (where F is the fraction of mass lost through melting). The value of the derivative $\partial \rho / \partial F$ depends upon the phase assemblage present. Most of our imaged structure is in the garnet stability field (depths ≥ 70 km) where the density difference between depleted and fertile mantle is ~25 kg m⁻³ [Jordan, 1979] yielding $\partial \rho / \partial F = 333$ kg m⁻³. A North American plate velocity $V = 3.5$ cm yr⁻¹ and creation of a sill of cross-sectional area $A = 1000$ km² would produce a compositional buoyancy flux $B_{comp} = 330$ kg s⁻¹ which corresponds to ~220 m of mantle-derived uplift over the width of the tectonic parabola. This uplift estimate neglects the loading effect of the midcrustal sill (because we are interested in the depleted residuum that is horizontally displaced from the sill, as is implied by the tomographic images).

In addition to the midcrustal sill, other amounts of basalt have been extracted from the mantle that are more difficult to estimate. These include the surface basaltic flows that systematically follow the rhyolitic volcanism along the track of the hotspot, probable basaltic underplating of the crust, and any basalt that might have fractionated to form some of the later stages of rhyolitic volcanism, as is argued by Perry *et al.* [1993]. Inclusion of these unknown volumes of basalt would increase the estimate of the compositional buoyancy contribution to net uplift.

When we compare our minimum estimate of the compositional buoyancy flux with the net buoyancy flux of 1500 kg s⁻¹ estimated for Yellowstone [Sleep, 1990], we find that at least 22% of the total buoyancy flux is compositional in origin. Sleep's [1990] estimate is based upon an assumption that

swell-created relief is ~1000 m; however, published estimates of the Yellowstone swell relief vary from 500 m to 1000 m [Smith and Braile, 1994; Pierce and Morgan, 1992; McNutt and Waschbusch, 1994] and still-lower values are possible. If Sleep [1990] had used a 500 m estimate of swell relief instead, then our estimate of the compositional buoyancy would be 45% of the total. Moreover, all of the swell relief can be attributed to depletion of the mantle if one assumes that half of the basalt segregated from the mantle is unaccounted for or that the actual swell relief is just 250 m. That is, within the uncertainties of the data, no thermal buoyancy flux is required.

Previous estimates of global heat flux from hotspots were derived under the assumption that swell relief is entirely thermal in origin [Sleep, 1990]. If swell relief is partly due to compositional changes accompanying mantle melting and depletion, then these estimates of hotspot heat flux may be overstated. Recently, Phipps Morgan et al. [1995] used one-dimensional modeling to argue that depleted residuum contributes to the swell observed at Hawaii, and Manglik and Christensen [1997] included the effects of mantle depletion buoyancy on a model of a plume interacting with a stationary plate. The mantle structure observed beneath the eastern Snake River Plain suggests that future such modeling will be important in improving our understanding of hotspot dynamics.

Dynamical Origin of the Yellowstone-System Upper Mantle Structure

Inversions of our data with teleseismic data from previous experiments [Iyer et al., 1981; Evans, 1982; Dueker and Humphreys, 1990] show that the low-velocity zone maintains a roughly constant width and depth along the axis of the parabola toward the current location of the hotspot, shallowing slightly (~50 km) at the Yellowstone caldera. If the partially molten zone were flanked by "typical" upper mantle, then the low-velocity region would be buoyant relative to the surrounding mantle and hence would be expected to flatten and widen with time and distance from the hotspot source. However, if the central zone of partial melt is surrounded by depleted residuum on either side, then the relative buoyancy between the two regions depends upon the level of depletion of the matrix within which the melt resides. With the passage of time and consequent cooling and loss of melt fraction, the zone of partial melt will become less buoyant than the residuum, assuming the two regions are similarly buoyant at the termination of melt extraction. This net difference in buoyancy will allow the pancaking residuum to confine the partially molten central core thereby preventing it from flattening. Such anticipated behavior can explain the seismic observations well: the eastern SRP-Yellowstone region is surrounded by the tectonic parabola [Anders and Sleep, 1992], which suggests a widening and thinning of hotspot-derived material under its own buoyancy, while the zone of partially molten upper mantle remains roughly constant for a great distance from the Yellowstone caldera.

The geoid high at Yellowstone suggests there is a buoyancy-induced pressure high at depth. If the region of partial melt is prevented from flattening due to the constricting nature of the depleted residuum on either side, then the pressure excess would tend to drive the partially molten mantle to flow along the existing low-velocity channel, similar to a river between levees. The apparent deepening of the low-velocity zone with distance from the caldera suggests such a mass

imbalance in which a lateral pressure gradient drives partially molten mantle "downstream" more rapidly than it is being advected away through simple plate motion. The Yellowstone hotspot currently is abutting the craton which may enhance this effect by inhibiting flow to the northeast, as shown schematically in Plate 6.

What Fundamentally Drives the Yellowstone System?

The traditional view of the Yellowstone hotspot is that it represents a fairly steady state system in which a deep-seated, thermal plume continuously supplies energy. The time-progressive magmatism parallel to North American absolute plate motion, the high He^3/He^4 ratios [Kennedy et al. 1985], and the enriched geochemical signature of erupted lavas [Doe et al., 1982; Leeman, 1982a,b] are key observations supporting this view.

An alternative view has the Yellowstone hotspot as a self-organized system of propagating volcanism that will proceed until lateral density differences become too small to drive the system any further. This idea is embodied in the models of Tackley and Stevenson [1993], who propose that convective instabilities may develop spontaneously in regions of partial melt. Once begun, pressure-release melting causes a decrease in density that will drive upwelling and hence induce more pressure-release melting, resulting in a positive feedback system. When the available melt is released or the depletion buoyancy of the residuum accumulated at the sides of the ascending column balances the melt buoyancy, the instability is effectively choked off. This process is similar to the active flow models of spreading ridges and can produce voluminous volcanism without a continuous supply of energy from a deep source such as a mantle plume. This model predicts that large amounts of depleted residuum will be pushed to the side of the partial melt zone, similar to what we infer from our tomographic images.

In the fully three-dimensional case, melt-driven convection would tend to form elongate convective rolls aligned in the direction of absolute plate motion. This alignment would occur because of both the applied shear field across the asthenosphere [Richter and Parson, 1975] and the advection of melt-depleted residuum downstream, which would draw new, fertile mantle to the evacuated front of the system. The net result would be time-progressive magmatism oriented in the direction of plate motion. Crustal extension above this sort of propagating system would further excite mantle upwelling and greatly enhance melt production. These ideas provide a natural mechanism by which both propagating magmatism and the structure observed beneath our study area can be created by local processes. Similar convective processes may also explain both the Jemez and St. George volcanic trends to the south, which display similarly aligned low-velocity trends in the upper mantle beneath the loci of young magmatism [Humphreys and Dueker, 1994]. This sort of active convective structuring of the upper mantle is possible regardless of whether Yellowstone and other western U.S. volcanic trends are ultimately connected to deep-seated mantle plumes.

A third idea is that lithospheric thickness differences could drive vertical flow [King and Anderson 1994]. At the transition where thick Archean lithosphere joins thinner lithosphere, lateral differences in temperature gradients might induce small-scale convection which, in turn, could produce decompression melting and associated volcanism. Like the self-perpetuating

volcanism model, this model predicts that large amounts of magmatism may occur without a continuous supply of thermal energy from a mantle plume.

All of these models can explain the oriented, propagating magmatism of the Yellowstone hotspot. Neither the self-perpetuating volcanism model nor the lithospheric thickness differences model alone can explain the high He^3/He^4 ratios of Yellowstone volcanics. This geochemical evidence strongly suggests a deep mantle source; however, a plume-like, steady state supply of hot, lower mantle is not required by these arguments. Any of the above suggested origins are consistent with our tomographic findings, and none of these potential processes are exclusive of one another. Fundamental to resolving the issue of "local or plume" will be an improved resolution of the deeper structure (i.e., convincing evidence for the presence or absence of a plume) and comparison of the regional Yellowstone structure with other similar systems.

Conclusions

Tomographic images of the upper mantle velocity structure beneath the track of the Yellowstone hotspot show a low-velocity region that is flanked by relatively high-velocity regions to either side. The shape of the low-velocity anomaly is narrower and deeper than predicted by current mantle plume models. We interpret the low-velocity region to be partially molten peridotite and the high-velocity mantle to be depleted of a basaltic component because of recent melt segregation. This conclusion is derived from combined seismic and isostatic arguments.

Melt generation and segregation is a dominant process in the upper mantle beneath Yellowstone. The forces driving this small-scale convection are derived from melt and compositional buoyancy differences. We cannot resolve whether the ultimate source of the Yellowstone hotspot is a deep-seated mantle plume or simply a more local, convective phenomenon in the upper mantle. Both scenarios are consistent with our tomographic images.

Swell relief surrounding hotspots has generally been attributed to thermal buoyancy, however, a zone of depleted residuum in the mantle can account for at least one fifth and possibly all of the swell relief at Yellowstone. We conclude that hotspot models need to include the effects of compositional changes that accompany melting and depletion of mantle peridotite in order to more correctly predict upper mantle conditions. Without such modifications, estimates of global thermal flux from hotspots may be in error.

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