

Upper-mantle origin of the Yellowstone hotspot

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ABSTRACT

Fundamental features of the geology and tectonic setting of the northeast-propagating Yellowstone hotspot are not explained by a simple deep-mantle plume hypothesis and, within that framework, must be attributed to coincidence or be explained by auxiliary hypotheses. These features include the persistence of basaltic magmatism along the hotspot track, the origin of the hotspot during a regional middle Miocene tectonic reorganization, a similar and coeval zone of northwestward magmatic propagation, the occurrence of both zones of magmatic propagation along a first-order tectonic boundary, and control of the hotspot track by preexisting structures. Seismic imaging provides no evidence for, and several contraindications of, a vertically extensive plume-like structure beneath Yellowstone or a broad trailing plume head beneath the eastern Snake River Plain. The high helium isotope ratios observed at Yellowstone and other hotspots are commonly assumed to arise from the lower mantle, but upper-mantle processes can explain the observations. The available evidence thus renders an upper-mantle origin for the Yellowstone system the preferred model; there is no evidence that the system extends deeper than ~200 km, and some evidence that it does not. A model whereby the Yellowstone system reflects feedback between upper-mantle convection and regional lithospheric tectonics is able to explain the observations better than a deep-mantle plume hypothesis.

Keywords: helium, hot spots, mantle plumes, tomography, upper mantle, Yellowstone National Park.

INTRODUCTION

The Yellowstone Plateau volcanic field is the youngest part of a magmatic system that has propagated northeastward along the path of the eastern Snake River Plain since at least the middle Miocene. The volcanic field has been cited frequently as an archetype of a continental hotspot associated with a convective plume from the deep mantle since its first such attribution by Morgan (1972a).

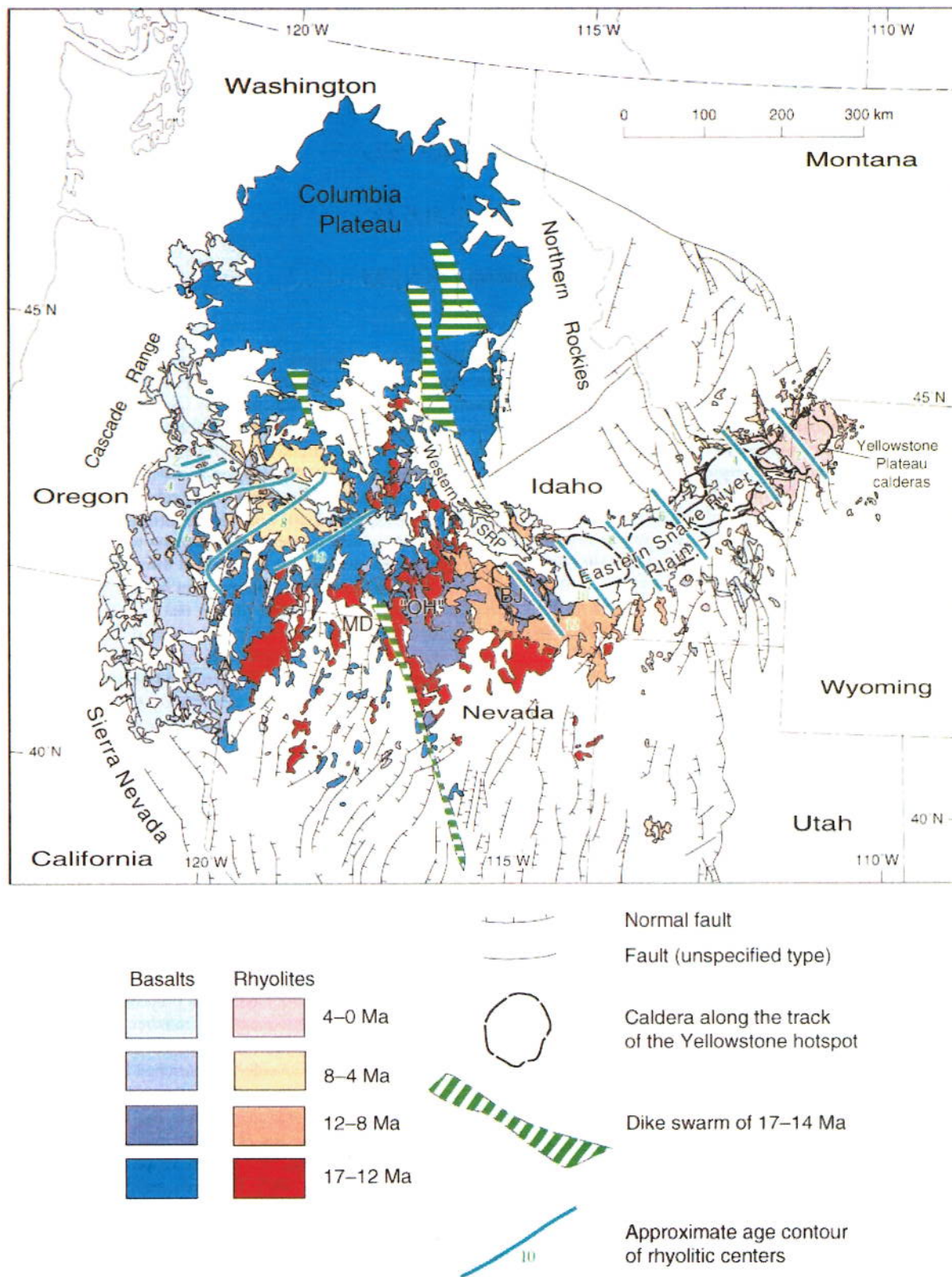
It is widely assumed that hotspots are global features of sublithospheric origin that do not share the motion of the tectonic plates but are points of reference with respect to which plate motions may be

measured. The term "hotspot" was introduced by Wilson (1963) for linear chains of oceanic volcanoes that appear to record the motions of the overlying plates. Subsequently, this terminology became nearly universal jargon. Various investigators have enumerated and catalogued hotspots differently (e.g., Burke and Wilson, 1976; Crough and Jurdy, 1980; Richards et al., 1988; Duncan and Richards, 1991), but all agree that most hotspots are long-lived centers of enhanced magma production, commonly associated with voluminous surface volcanism that has propagated linearly with respect to the plates on which they occur.

Such generalizations are little debated, but there is considerable dispute over an extension of the hypothesis that regards hotspots as near-surface expressions of convective plumes that rise through the entire mantle from a thermal boundary layer at the base of the mantle. Indeed, many investigators regard "hotspot" and "deep-mantle plume" as virtual synonyms (e.g., Morgan, 1971; Pierce and Morgan, 1992; Steinberger and O'Connell, 2000). The term "hotspot" itself is taken by many to imply that the observed features result from the addition of excess heat to the upper mantle from the deep mantle. In response to that common assumption, Shaw and Jackson (1974) made a point of using the term "melting anomaly" rather than "hotspot," emphasizing a specifically recognized upper-mantle process rather than a lower-mantle process that is presumed with few or no directly supporting data. As the term "hotspot" is so well entrenched in the literature, we use it here but without any specific implication of significant excess heat or temperature in the mantle. Where a term that particularly lacks a genetic connotation seems more appropriate, we use "melting anomaly" more or less synonymously.

Morgan (1971, 1972a, 1972b) published the seminal papers in which the deep-mantle plume model was proposed as a global explanation for hotspots. Subsequently, a number of characteristics have come to be widely regarded as diagnostic of their deep-mantle origin. In particular, some consider the enriched lithophile trace element and related isotopic compositions of associated basaltic magmas in many of these systems to represent the advection of more primitive material from the lower mantle into a largely depleted upper mantle (e.g., Schilling, 1973; Hart et al., 1992; Hanan and Schilling, 1997; Hoffman, 1997). One frequently cited criterion of a lower-mantle origin for hotspot systems (e.g., Craig and Lupton, 1981) is a high ratio of ^3He to ^4He compared to values observed away from hotspots, such as in continental crust or mid-ocean ridge basalt (MORB).

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THE YELLOWSTONE HOTSPOT

The current manifestation of the Yellowstone hotspot is the Yellowstone Plateau volcanic field, a voluminous, compositionally bimodal but predominantly rhyolitic system that has erupted as much as 6,000 km³ of rhyolitic and basaltic magma in about the past 2 m.y. (Christiansen, 1984, 2001). The field has evolved through three cycles of activity, each climaxed by the rapid eruption of hundreds to thousands of cubic kilometers of rhyolitic magma and the formation of a large caldera. The volcanic field is interpreted to overlie a batholithic system of partially molten to solidified rhyolitic magma chambers, aggregating an order of magnitude greater in volume than the volcanic products, and a complex network of basaltic intrusions. Because they are significantly denser than rhyolitic magmas, basaltic magmas are trapped beneath a large rhyolitic magma chamber, feeding surface volcanism around the margins of the active rhyolitic area but not within it. The Yellowstone Plateau volcanic field is topographically high and lies at the center of a major regional geoid high (Milbert, 1991).

A series of voluminous volcanic systems analogous to the Yellowstone Plateau was active sequentially from southwest to northeast across the North American plate along the eastern Snake River Plain, at least after ca. 12.5 Ma (Fig. 1). The resulting rhyolitic volcanic fields are preserved along the margins of the plain, the subsided axis of which is flooded by younger basalts. Pierce and Morgan (1992) have given the fullest published depiction of these rhyolitic, caldera-related fields and have outlined exposed remnants of the rhyolitic outflow sheets and interpreted locations of source calderas now largely buried by younger basalts (Fig. 1). Zones along the flanks of the eastern Snake River Plain form a topographic swell that gradually declines southwestward with distance from Yellowstone.

Morgan (1972a) cited the Yellowstone hotspot as an example of a deep-mantle plume in a continental intraplate setting, and numerous authors subsequently adopted that model (e.g., Matthews and Anderson, 1973; Anders et al., 1989; Pierce and Morgan, 1992; Parsons et al., 1994, 1998; Smith and Braile, 1994; Thompson, 1998; Pierce et al., 2000). Pierce and Morgan (1992) considered the northeastward progression of rhyolitic volcanism to have begun with the McDermitt volcanic field at ca. 16–15 Ma and to include an “Owyhee-Humboldt volcanic field” of ca. 14 Ma, farther northeast (Fig. 1). Christiansen and Yeats (1992), however, noted that silicic as well as basaltic volcanism of 17–14 Ma was regionally widespread in the Pacific Northwest and that a well-defined northeastward progression along the eastern Snake River Plain can be recognized with certainty only beginning with the ca. 12.5 Ma Bruneau-Jarbridge volcanic field (Fig. 1). Several investigators have elaborated the original deep-mantle plume model in order to account for otherwise anomalous aspects of the regional eastern Snake River Plain–Yellowstone tectonic setting (e.g., Duncan, 1982; Draper, 1991; Pierce and Morgan, 1992; Geist and Richards, 1993; Pierce et al., 2000).

The deep-mantle plume explanation for the Yellowstone hotspot was never universally accepted. An early model regarded the Yellowstone

Plateau as the tip of a propagating rift, now represented by the eastern Snake River Plain (Hamilton and Myers, 1966; Smith et al., 1974). That concept continued to find a place in some later discussions but is generally considered inadequate to explain the great magmatic productivity of the eastern Snake River Plain–Yellowstone system. Explanations related to lithospheric plate motions were suggested by Christiansen and McKee (1978) and by Dickinson (1997). Notably, the recent summary by Humphreys et al. (2000) of teleseismic observations and regional geophysics concludes that a deep-mantle source is not required by any of the data examined and is disfavored by many of them. Favela and Anderson (2000) have pointed out spatial and temporal coincidences required by plume explanations of the eastern Snake River Plain–Yellowstone system and have suggested that lithospheric extension and the focusing of stress and magma are important. In this paper we critically examine geologic and seismic data that conflict with a deep-mantle plume model for Yellowstone and discuss implications of noble-gas geochemical observations. We emphasize here that we do not argue against the importance of mantle convection in the origin of the Yellowstone hotspot. We do, however, consider that convection entirely within the upper mantle, and perhaps only the shallowest upper mantle, is sufficient to account for the characteristics of the hotspot. Significant flow of material from the lower mantle can be ruled out. Local convection triggered by the juxtaposition of contrasting lithospheric blocks probably plays a major role (King and Anderson, 1998).

The logic of the following sections is as follows: (1) Certain geologic features conflict with the deep-mantle plume hypothesis, (2) seismic images that should test for the existence of a deep-mantle plume demonstrate its absence, and (3) helium isotope data that have been considered definitive of a deep-mantle plume can be explained by upper-mantle models. Therefore, an upper-mantle origin for the Yellowstone hotspot is preferred.

GEOLOGIC CONSTRAINTS

Certain geologic aspects of the Yellowstone Plateau volcanic field as well as the regional tectonic evolution are either inconsistent with a simple deep-mantle plume hypothesis or must be attributed to coincidence (Christiansen and Lipman, 1972; Christiansen and McKee, 1978; Christiansen and Yeats, 1992; Favela and Anderson, 2000; Christiansen, 2001). Relevant arguments from the geology of Yellowstone, the eastern Snake River Plain, and regional tectonics can be summarized in the following five major points.

Temporal Persistence of Magmatism

The basic model of a deep-mantle convective plume for the Yellowstone hotspot does not explain the continuing voluminous basaltic volcanism that occurs along the eastern Snake River Plain for hundreds of kilometers away from the Yellowstone Plateau. Indeed, a nearly

Figure 1. Map of the northwestern United States showing basin-range faults and basalts and rhyolites of 17 Ma and younger. The track of the Yellowstone hotspot is indicated by approximate age contours (blue-green heavy lines) of rhyolitic volcanic centers (ca. 12, 10, 8, 6, 4, and 2 Ma) across the northeast-trending eastern Snake River Plain; a contemporaneous trend of oppositely propagating rhyolitic volcanism that trends northwest across central Oregon is indicated by similar contours. Locations of calderas along the Yellowstone hotspot track from Pierce and Morgan (1992) and Christiansen (2001). SRP—Snake River Plain, MD—McDermitt volcanic field, OH—Owyhee-Humboldt volcanic field of Pierce and Morgan (1992), BJ—Bruneau-Jarbridge volcanic field.

continuous zone of young basalts extends westward across the entire region to the Cascades arc (Fig. 1).

Regional Patterns of Middle Miocene Magmatism and Extension

Volcanism associated with the Yellowstone hotspot began in the region of northern Nevada, southwestern Idaho, eastern Oregon, and southeastern Washington during a middle Miocene time of major tectonic reorganization throughout much of a region 2000 km wide (Christiansen and McKee, 1978). Beginning at ca. 17 Ma, basin-range extension and fundamentally basaltic volcanism became regionally distributed across this entire region. Flood basalts erupted in the Columbia Plateau and Oregon Plateaus north of the subsequent hotspot track, but only moderate- to small-scale basaltic and bimodal volcanism occurred to the south.

Northwest-Propagating Magmatism

The Yellowstone melting anomaly is one of a pair of complementary—and, in many ways, similar—linearly propagating systems that evolved simultaneously from the south edge of widespread basaltic and bimodal magmatism of 17–14 Ma in the region of southeastern Washington, eastern Oregon, western Idaho, and northern Nevada (Fig. 1). The Newberry melting anomaly, like the Yellowstone system, produced bimodal rhyolite-basalt magmatism, but rhyolitic volcanism propagated northwestward across the High Lava Plains of eastern and central Oregon during the same period and at a similar rate as the northeastward propagation of eastern Snake River Plain–Yellowstone rhyolitic volcanism (MacLeod et al., 1976). As with the Yellowstone hotspot track, basaltic volcanism continues along the track of the Newberry melting anomaly.

Regional Tectonic Boundary

The basin-range region south of the two zones of oppositely propagating rhyolitic volcanism has undergone large amounts of tectonic extension, probably widening by ~250 km during the past 17 m.y. (Wernicke and Snow, 1998). In contrast, the region to the north has extended an order of magnitude less (a few tens of kilometers), and extension becomes virtually nil within several tens of kilometers north of the propagating systems (Lawrence, 1976; Christiansen and McKee, 1978; Christiansen and Yeats, 1992). The eastern part of this boundary along the eastern Snake River Plain marks a profound change in lithospheric structure between the thin, hot, extending lithosphere of the basin-range region to the south and the ~40-km-thick cold lithosphere of the North American craton to the north. Although not as strongly differentiated, the western part of the boundary is also marked by mantle regions of lower wave speeds on the south and higher wave speeds on the north.

Preexisting Structural Control

Volcanism associated with the Yellowstone hotspot first became evident at the west edge of the Archean continental crust of the Wyoming province (Reed, 1993). Eaton et al. (1975) and Mabey et al. (1978) showed that a regional aeromagnetic anomaly pattern that marks the axis of the eastern Snake River Plain continues both southwestward into Nevada and northeastward across Montana to the Canadian border, interrupted only by a magnetic low associated with the high crustal temperatures of the active Yellowstone Plateau. Northeast of Yellow-

stone this anomaly pattern crosses Laramide structures of Late Cretaceous to early Tertiary age; this region has been practically undeformed in late Cenozoic time and the anomaly pattern parallels major regional Precambrian structural trends. Iyer and Healy (1972) interpreted teleseismic residuals at the LASA (Large Aperture Seismic Array), located along this trend in northeastern Montana, as reflecting a northeast-trending zone of thickened crust, with the Moho depressed by ~15 km along the trend. Thus, the Yellowstone hotspot magmatism began at the edge of a cold thick Archean craton and appears to follow a regional structural zone of Precambrian origin across that craton.

SEISMIC IMAGING

The crust and upper mantle beneath the Yellowstone Plateau and the eastern Snake River Plain have been studied by several seismic experiments. In the period 1973–1975, Iyer et al. (1981) gathered data on 57 short-period vertical seismometers that formed a 430×250 km network traversing northwest-southeast across the Yellowstone caldera. They inverted 1,382 teleseismic P-wave arrival-time residuals from 221 teleseisms to study the structure of the crust and upper mantle to a depth of ~190 km. They found a low-wave-speed body beneath Yellowstone with an anomaly of 15%–20% in the upper crust, decreasing to 5%–10% in the lower crust and upper mantle. That the low-wave-speed body does not extend deeper than ~200–250 km was unambiguously evident in the raw data, as the region of late arrivals centered on Yellowstone was limited to the central part of the long axis of the network (cf. Figs. 15 and 16 of Iyer et al., 1981). The data of Iyer et al. (1981) have been reprocessed and incorporated into other studies by independent workers using different inversion methods (e.g., Dueker and Humphreys, 1990; Humphreys and Dueker, 1994a, 1994b). In all studies the low-wave-speed body underlying Yellowstone is found to extend no deeper than ~240 km.

In order to examine this result critically, we selectively reinverted these relative traveltimes residuals for three-dimensional wave-speed structure, by using more modern techniques and the most homogeneous ray set available. Our objective was to investigate in the greatest detail possible the robustness of the finding that the anomaly beneath Yellowstone appears to be confined to shallow depths. We selected 51 stations and 93 events, yielding a total of 1,192 rays sampling the Yellowstone region. We inverted these data with the ACH algorithm (Aki et al., 1977). To suppress parameterization artifacts and maximize the resolution of small features such as deep narrow conduits, we used the layer thinning and threefold offset-and-average methods described by Evans and Achauer (1993). The base model to which the thinning and averaging are applied has resolution-matrix diagonal elements ranging from 0.75 to >0.9 at all depths in the central part of the model and reduces the data variance by 95%. Volume metrics (Evans and Achauer, 1993) provide a more complete description of the size, shape, and directions to which individual blocks of the inversion model are likely to smear their wave-speed anomalies, partly obscuring real structures or creating artifacts. For the layer-thinned models, these volume metrics indicate that features as small as ~50 km vertically and 35 km horizontally can be resolved in the upper half of the model, to a depth of ~200 km. The corresponding limits of resolution in the lower half of the model are about twice those values.

The principal feature of the resulting model (Fig. 2) is a low-wave-speed anomaly beneath the caldera, having a strength that decreases with depth to its weakest value of ~2% at ~175 km. Beneath this depth, there is no evidence of low wave speeds either directly beneath the caldera or beneath its northwest edge, where the shallower anom-

alies beneath both Yellowstone and the eastern Snake River Plain are centered (Evans, 1982; Evans and Iyer, 1991). Ray coverage and resolution are good to at least 300 km depth beneath Yellowstone. Because traveltime residuals of 0.05–0.10 s can be resolved at such depths, bodies with wave-speed anomalies stronger than about 1% can be ruled out if they are comparable in diameter to the shallower mantle anomaly (50–100 km; Fig. 2A). Similarly, if they are comparable in wave-speed perturbation, they must be significantly narrower than the shallow feature. Some smearing along the main P-wave ray bundles is evident, as expected, and is exemplified by the discontinuous yellow ring in Figure 2A and the weak yellow-green streaks in the deepest two-thirds of Figure 2B. These features are the expected smearing of the strong low-wave-speed feature beneath Yellowstone, as predicted by the volume metrics. This sensitivity also suggests that the numerical inversion is unlikely to have suppressed any weak, deep anomaly. Furthermore, the volume metrics of selected blocks show that there is a tendency for the inversion to smear anomalies downward, artificially exaggerating the vertical extent of the real anomaly rather than concealing any downward continuation of the low-velocity body.

The asthenospheric low-wave-speed layer (LVZ—low-velocity zone) beneath the Western United States extends down to a maximum depth of 150–200 km (Helmberger, 1972). Our results show to a high degree of certainty that the anomaly beneath Yellowstone is largely or completely contained within and above this LVZ.

Mantle structure beneath the eastern Snake River Plain and its flanking topographic swell has been studied by using data gathered on transverse profiles up to 500 km long at distances of 100 km and 200 km southwest of Yellowstone. Teleseismic tomography images a relatively narrow (~130 km wide, northwest-southeast), low-wave-speed anomaly beneath the plain and its northwest margin. That zone has an anomaly as strong as about 5%, extends from a depth of ~50 km down to ~200 km, and is flanked by high-wave-speed bodies up to 200 km wide to the northwest and southeast, underlying the swell (Evans, 1982; Dueker and Humphreys, 1990; Humphreys and Dueker, 1994b; Saltzer and Humphreys, 1997). This structure is not what is expected if the eastern Snake River Plain and swell represent a trailing plume head, because that model requires a low-wave-speed body that extends from near the surface to ~100 km depth and everywhere underlies the plain and surrounding swell (e.g., Sleep, 1990). Instead, the low-wave-speed body underlying the plain is comparable in size and depth extent to that which underlies Yellowstone except that the top 50 km is missing.

The presence of any thermal anomaly in the 400–650 km depth range, as allowed in one of the models by Humphreys et al. (2000), would be expected to be accompanied by thinning of the transition zone through depression of the 400 km discontinuity and elevation of the 650 km discontinuity (Bina and Helffrich, 1994). The depths to these discontinuities have been studied by using receiver functions for earthquakes recorded on a 500-km-long profile traversing the eastern Snake River Plain and flanking swell ~200 km southwest of Yellowstone (Dueker and Sheehan, 1997). The results provide no evidence for a coherent thermal anomaly extending throughout the transition zone under the plain. These results are confirmed by reanalysis of the data by using different techniques (Beucler et al., 1999). Beneath the profile, transition-zone thickness was found to vary by 30–35 km, but the topographies of the two discontinuities are uncorrelated. The most significant feature is a deepening of the 400 km discontinuity by 20 km from the northwest margin of the plain to the eastern edge of the basin-range region. If interpreted as thermal anomalies, the discontinuity topographies imply uncorrelated lateral temperature variations of

up to 250 K across the plain and flanking swell, with maximum temperatures at 400 km depth at a location 150 km southeast of the plain. Possible nonthermal explanations for the observed topography of the mantle seismic discontinuities include the combined effects of garnet-pyroxene phase transformations, chemical layering, and variations in mantle hydration (Dueker and Sheehan, 1997).

Mantle structure on a broad scale is revealed by whole-mantle tomographic images. A southwest-northeast cross section (Fig. 3) through model S20RTS (Ritsema et al., 1999) passing through Yellowstone illustrates clearly the location of Yellowstone at a position of profound lithospheric structural change where the basin-range region to the southwest abuts the North American craton to the northeast. A major high-wave-speed body apparently continuous with the craton occupies most of the thickness of the upper mantle beneath Yellowstone. There is no evidence for a significant low-wave-speed anomaly beneath Yellowstone either in the lower mantle or the lower part of the upper mantle, although bodies smaller than a few hundred kilometers across cannot be ruled out.

HELIUM GEOCHEMISTRY

Of all geochemical tracers, the ratio $^3\text{He}/^4\text{He}$ (R) is considered by many to be the most robust indicator of a lower-mantle component. The average value of R for the global spreading-ridge system is $(9.1 \pm 3.6)R_a$ (1σ), where R_a is the atmospheric value of 1.38×10^{-6} ; for MORB, excluding samples that are subaerial or from backarc basins or seamounts, it is $(8.58 \pm 1.81)R_a$ (Anderson, 2000b), equivalent to a $\pm 2\sigma$ range of $4.96R_a$ to $12.20R_a$. Hotspots such as Hawaii, Iceland, and Yellowstone are commonly associated with a much wider spread of values, including a few that are significantly higher than the MORB value—up to $42.9R_a$ (Hilton et al., 1999; Breddam and Kurz, 2001). These very high R observations are commonly interpreted as indicating a component from a high- R reservoir, commonly assumed to be the lower mantle, which is deeper and less degassed than the upper-mantle MORB-source reservoir. Observed hotspot values that are less than the maximums are most often interpreted as having been reduced from initial high values by degassing and contamination with crustal ^4He or air.

The noble-gas geochemistry of thermal waters at Yellowstone has been summarized by Fournier (1989). R ranges from $<3R_a$ immediately outside the caldera to values of $16R_a$ in the Mud Volcano area and $13.4R_a$ in the Gibbon Geyser Basin area (Craig et al., 1978; Welhan, 1981; Kennedy et al., 1985), higher than the mean for MORB at the 2σ confidence level (Fig. 4). Values of $10.5R_a$ and $9.5R_a$ (Hearn et al., 1990) for Crater Hills and Shoshone Geyser Basin (Fig. 4) are within the 2σ variation for MORB.

The maximum values observed, which are unusually high for a continental region, have been interpreted as indicating a convectively rising plume in the lower mantle beneath Yellowstone (e.g., Dodson et al., 1997), and this conclusion has been used to set limits on interpretations of other data (e.g., Leat et al., 1991). However, the model by which high $^3\text{He}/^4\text{He}$ ratios are attributed to the lower mantle is flawed, and there are viable processes by which such values may be generated in the upper mantle. An upper-mantle origin for high $^3\text{He}/^4\text{He}$ ratios would render these observations compatible with the geologic and seismic results.

DISCUSSION

Within a deep-mantle plume framework, certain major features of the geology and tectonic setting of the eastern Snake River Plain—

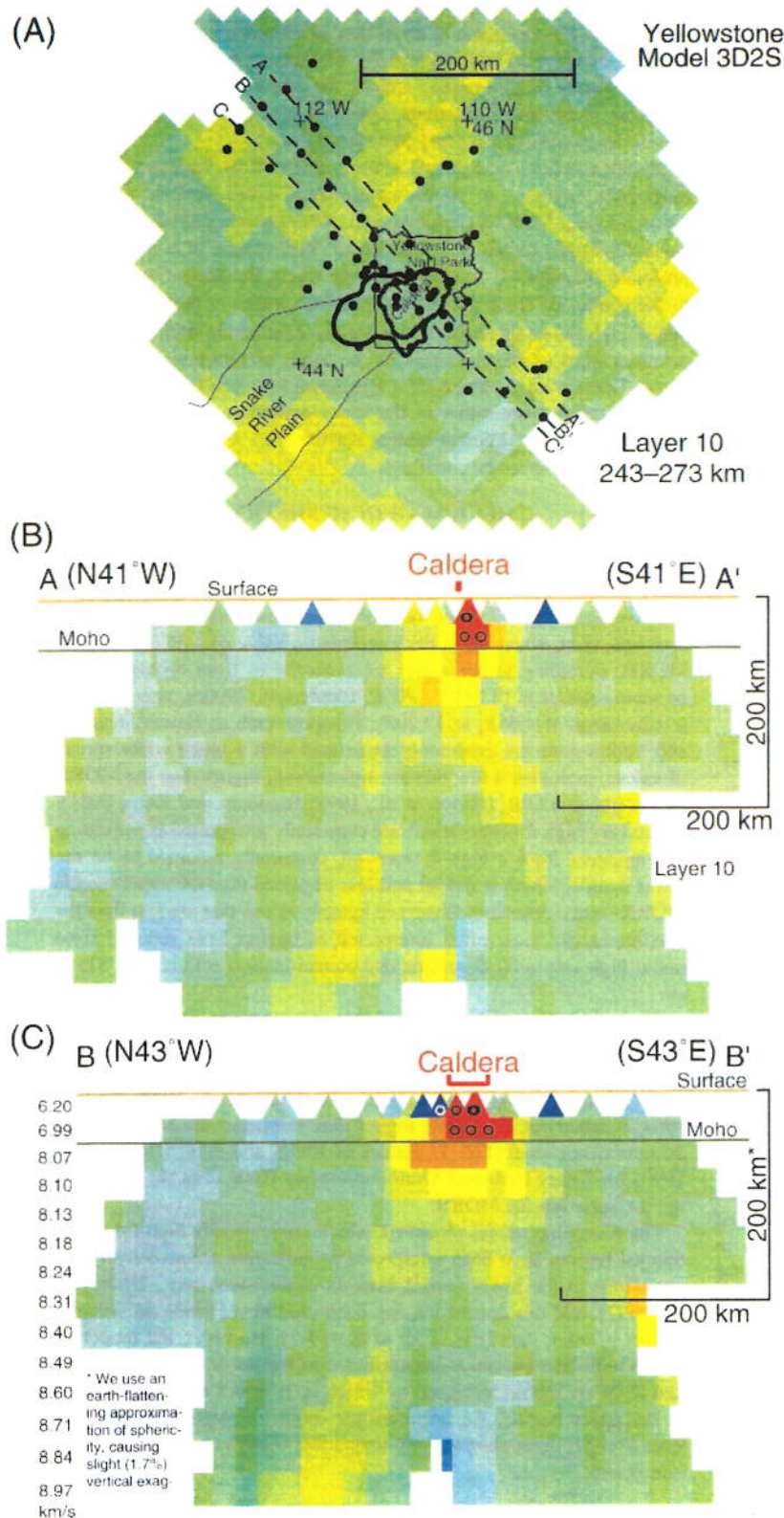


Figure 2. Teleseismic tomographic structure beneath Yellowstone obtained with the techniques of Evans and Achauer (1993) on a subset of the data collected by Iyer et al. (1981). Data were selected for uniform coverage in event back-azimuth and distance, optimizing the ray set. The color scale shown in D applies to all parts of the figure; distance and depth scaling are also constant; small circles in some of the blocks indicate that the velocity anomalies exceed the range of the color scale. (A) Dots are the stations used; the boundary of Yellowstone National Park and the edges of the eastern Snake River Plain are shown; lines of cross section shown in B, C, and D are indicated. Colors indicate wave-speed variations in the layer in the depth interval 243–273 km, where a deep plume-like structure would be imaged if one exists. Irregularly shaped, closed rings outline calderas of the Yellowstone Plateau volcanic field (Christiansen, 2001). (B) Cross section through the model at the north-east edge of the caldera, presumed by some investigators to be the center of the hotspot. (C) and (D) Cross sections farther southwest through the caldera.

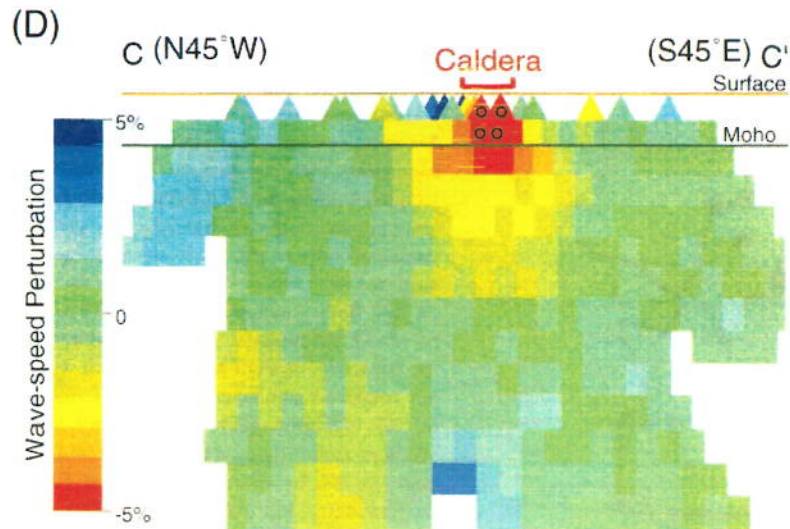
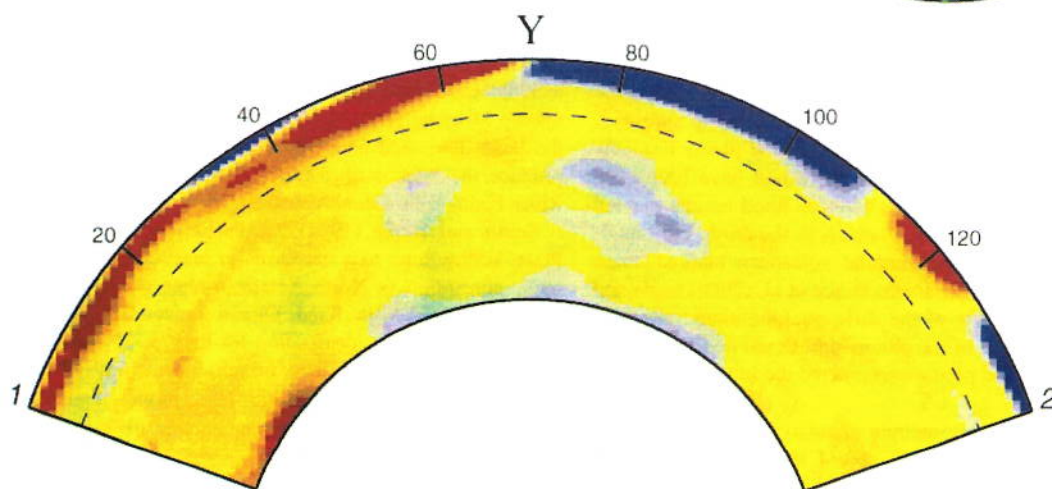
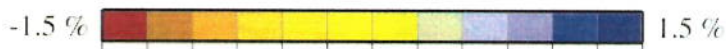


Figure 2. (Continued.)

Figure 3. Southwest-northeast cross section through model S20RTS (Ritsema et al., 1999) collinear with the eastern Snake River Plain and passing through Yellowstone, which is located at 70° along the line of section at "Y." Top panel shows the line of section; bottom panel shows structure from the surface to the core-mantle boundary. Dashed line indicates the 650 km discontinuity. Points 1 and 2 are cross-reference marks between the map and section, analogous to A and A' in Figure 2. Resolution of the image is ~500 km.



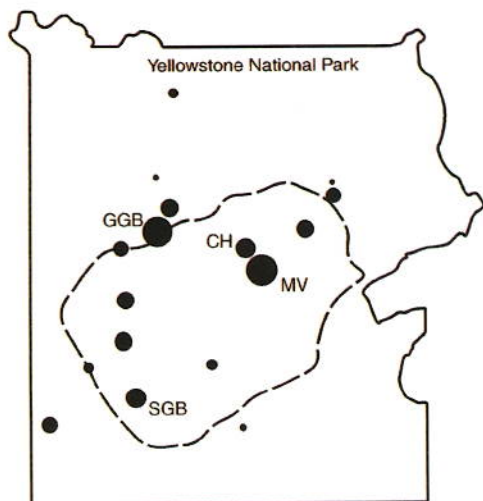


Figure 4. Map of Yellowstone showing variations in helium isotope ratios. Each black dot is proportional in area to the maximum $^3\text{He}/^4\text{He}$ ratio observed locally in the hydrothermal fluids. The largest dot corresponds to $^3\text{He}/^4\text{He} = 16R_a$. Solid outline shows boundary of Yellowstone National Park; dashed line shows Yellowstone caldera. CH—Crater Hills, GGB—Gibson Geyser Basin, MV—Mud Volcano, SGB—Shoshone Geyser Basin (adapted from Kennedy et al., 1985).

Yellowstone Plateau system can only be attributed to coincidence or to special auxiliary hypotheses, many of them mutually incompatible. Nothing inherent in the model explains the continuing basaltic volcanism along the hotspot track at great distances from Yellowstone. Morgan (1972b) briefly noted “flood basalts” of the Snake River Plain (presumably intending to include the voluminous 17–14 Ma basalts of the Columbia River and Oregon Plateaus) as analogous to the Deccan Traps of the early Reunion hotspot but proposed no explanation for persistent basaltic volcanism along both the Snake River Plain and the eastern Oregon High Lava Plains. Similarly, the model fails to account for initial development of hotspot volcanism at the edge of the only Archean craton in the western United States or for structural control of the hotspot track across that craton.

In order to explain the arrival of a deep-mantle plume at the base of the lithosphere precisely during a middle Miocene time of major tectonic and magmatic reorganization of a 2000-km-wide region, Draper (1991), Pierce and Morgan (1992), Parsons et al. (1994), and some others have proposed that the regional 17–14 Ma extension and magmatism were direct consequences of a newly arrived plume head. Its effects, however, according to the proposals, would have been very different to the north and to the south. Whereas flood basalts erupted in the Columbia Plateau and Oregon Plateaus to the north, only moderate- to small-scale basaltic and bimodal volcanism marked linear zones that reach >600 km to the south. Pierce et al. (2000) explained this discrepancy by the effects of the different lithospheric structure north and south of the center of the plume but offered no reason other than coincidence for why the plume encountered the lithosphere at that tectonic boundary.

Propagation of the Newberry melting anomaly in a symmetrically opposite direction seems unlikely to record the same deep-mantle plume as presumed to be responsible for the Yellowstone melting anomaly. Two lines of reasoning have been offered to explain this

relationship in the context of a deep-mantle plume. Draper (1991) suggested that the Newberry system represents partial entrainment of the spreading Yellowstone plume head in mantle counterflow related to Cascadia subduction. Pierce and Morgan (1992) compared the post-14 Ma northwestward-propagating Newberry system to the entire post-17 Ma history that they attribute to the Yellowstone hotspot, emphasizing significant differences between the two systems, and proposed different and unrelated origins. The Newberry system, however, would be better compared only to the distinctly northeastward-propagating part of the eastern Snake River Plain–Yellowstone system, which also postdates 14 Ma; both propagating systems arose within the same wide region of 17–14 Ma bimodal volcanism. Although magmatic productivity of the Newberry system has been much lower, that system nevertheless has produced significant volumes of rhyolitic magma, including 9.3–6.5 Ma ash-flow fields, each having volumes of hundreds of km³ (Walker, 1979; Streck and Grunder, 1999). Newberry’s magmatic productivity has, however, decreased with time. Despite these differences, the symmetry and overall similarity of these two coeval propagating magmatic systems suggests related origins. Furthermore, the two systems together bound a large region of great extension, thin lithosphere, and high heat-flow on the south from cooler, thicker, much less extended lithosphere to the north.

Several investigators have elaborated the deep-mantle plume model to account for aspects of the regional tectonic setting. Duncan (1982) hypothesized that a Yellowstone mantle plume beneath the eastern Pacific during the Eocene produced oceanic-island basaltic systems that later accreted to North America in the Oregon and Washington Coast Ranges, that its surficial expression was suppressed as the Cascades subduction system migrated over the plume, that it reemerged farther east to produce the Columbia River flood-basalt province of 17–14 Ma, and that the plume then shifted abruptly southward to form the eastern Snake River Plain–Yellowstone magmatic progression. Draper (1991) proposed that the plume head first arrived at the base of the lithosphere at ca. 17 Ma, spread laterally, and became entrained into sublithospheric mantle counterflow above the subducting plate. Geist and Richards (1993) proposed that the Cascades subduction system temporarily deflected a relatively narrow plume from its vertical ascent path and carried it along with the northward motion of the subducting plate, allowing it to emerge in the northern part of the region between 17 and 14 Ma to produce the Columbia River basaltic volcanism but then finally burn through the subducting plate, return to vertical, and propagate northeastward relative to North America along the eastern Snake River Plain. Pierce and Morgan (1992) proposed that, after initial arrival of a plume head perhaps 1,000 km across to produce basin-range tectonic extension, the Columbia Plateau flood basalts, and a broad middle Miocene pulse of regional volcanism in the Pacific Northwest, the head dissipated and only a chimney-like plume tail followed to produce the later northeastward progression of silicic eastern Snake River Plain–Yellowstone volcanism.

Smith and Braile (1994) calculated that the eastern Snake River Plain–Yellowstone axis conforms at least approximately to the kinematic constraints of North American plate motion, if the Hawaii hotspot is assumed to be fixed. Others, however, have noted significant deviations from this idealization for the pre-10 Ma part of the track, even allowing for the effects of subsequent local extension (Rodgers et al., 1990). Pierce et al. (2000) have proposed that the plume was deflected westward beneath the subducting plate to account for a deviation between 17 and 10 Ma.

Given the apparent necessity for such elaborate variations of the deep-mantle plume model, it is reasonable to question whether other

observations require a plume in the lower mantle beneath Yellowstone, provide no support for one, or contraindicate such a model.

The primary seismological observations from Yellowstone and the eastern Snake River Plain are inconsistent with or do not support a plume model. Such a model predicts a low-wave-speed body beneath Yellowstone that extends continuously from the surface, through the upper mantle, and into the lower mantle. Contrary to this prediction, the strong, low-wave-speed anomaly observed by teleseismic tomography beneath Yellowstone does not extend deeper than ~200 km and thus appears to be confined to the LVZ and above. Were the anomaly to continue below this depth as a feature too narrow to be resolved by teleseismic tomography, it would have to have a diameter considerably less than 75 km. Furthermore, it necessarily would have maintained this small size for the ~17 m.y. since its postulated inception, despite progressive lateral expansion by heat conduction to its surroundings.

The structure beneath the eastern Snake River Plain and its flanking swell also differs from that predicted by the plume model. A low-wave-speed anomaly ~100 km thick, centered beneath the plain and extending the full width of the swell, is expected for a plume head beneath a hotspot track (Sleep, 1990; Ribe and Christensen, 1994). Instead, a narrow, deep low-wave-speed anomaly is observed that is flanked beneath the swell by high-wave-speed bodies. Saltzer and Humphreys (1997) interpreted these observations as indicating a shallow convection system that forms a core of hot, partially melted material and that is flanked by volumes of high-wave-speed, depleted residuum. Such a model can account for the swell relief and suggests that a thermal origin for the elevated topography, as predicted by the plume hypothesis, is not required (Saltzer and Humphreys, 1997).

A relatively hot transition zone would be expected if the transition zone is traversed by a continuous plume with a temperature anomaly of ~200 K rising from the lower mantle. Such a plume would cause downwarping of the 400 km discontinuity and upwarping of the 650 km discontinuity. No evidence for such systematic discontinuity topography is observed in high-quality receiver-function results. In addition, whole-mantle tomography provides no evidence for a low-wave-speed plume rising from the lower mantle beneath Yellowstone. The resolution of whole-mantle tomography images is ~500 km, though smaller anomalies can be detected if strong. Although a narrow feature could not be detected, the negative Clapeyron slope associated with phase changes at the 650 km discontinuity is expected to hinder upward advection of material from the lower mantle as material would have to rise higher before it could transform to the lower-density phases. A broad, low-wave-speed plume head would thus be expected to spread out below this discontinuity, which is not observed. Furthermore, the broad, high-wave-speed body associated with the cratonic root and underlying Yellowstone in the middle part of the upper mantle is inconsistent with a plume model in that it would provide a barrier to upward flow of material from beneath.

Because the seismic evidence presents several contraindications for the fueling of a Yellowstone hotspot by a plume of relatively hot material rising from the lower mantle, it seems reasonable to question the assumption that the source of the high maximum helium isotope ratios is the lower mantle.

High values of R in terrestrial rocks generally are assumed to stem from excess ^3He , a primordial isotope that was incorporated in Earth when the planet formed. The production of radiogenic ^4He from decay of $\text{U} + \text{Th}$ during the lifetime of Earth has progressively reduced the mean R from an initial value that was in the range of ~100 R_a to 200 R_a (Jeffrey and Anders, 1970; Craig and Lupton, 1981). In individual samples or reservoirs, however, degassing may reduce the absolute abun-

dance of helium ($[\text{He}]$), causing R to decrease more quickly for a given $\text{U} + \text{Th}$ content, or metasomatism may increase $[\text{He}]$ and cause R to decrease more slowly (Zindler and Hart, 1986).

The wide range of R observed in rocks or hydrothermal fluids at hotspots includes values both lower and higher than the mean for MORB. It is typically assumed that the low- R rocks at hotspots initially had higher values of R that lowered as a consequence of eruptive degassing. However, the wide spread is also consistent with greater source heterogeneity or small extraction volumes (Anderson, 2000b).

Models that assign high R to the lower mantle predict that the source of hotspot magmas is a high- $[\text{He}]$ reservoir, but hotspot rocks in general are very low in helium, containing as little as two to three orders of magnitude less than MORB contains. This observation also is generally attributed to eruptive degassing. However, because helium is preferentially retained in the melt relative to the heavier noble gases, degassing is expected to raise the ratio of helium to heavier noble gases. In conflict with this, the values of $^3\text{He}/^{22}\text{Ne}$ and $^4\text{He}/^{40}\text{Ar}$ are commonly depressed in hotspot rocks relative to MORB, indicating that hotspot rocks are less degassed than higher- $[\text{He}]$ MORB rocks, not more degassed (Anderson, 1998a, 1998b, 2000a; Moreira and Sarda, 2000; Ozima and Igarashi, 2000). These data suggest that the sources of high- R hotspot rocks are intrinsically low in $[\text{He}]$.

In the absence of a high- $[\text{He}]$ reservoir, high $^3\text{He}/^4\text{He}$ ratios might reflect either preservation of source domains of past high- R material in the upper mantle or some local process that either enriches ^3He relative to ^4He or depletes ^4He relative to ^3He . High- R values might be preserved in helium stored for long periods in a low- $(\text{U} + \text{Th})$ environment such as a residuum from lithospheric melting (Anderson, 1998a), particularly because helium may be more strongly partitioned into residuum than into melt (B. Wood, 2001, personal commun.). Support for this model comes from diamonds in mantle xenoliths from kimberlite pipes, which have helium isotope ratios up to 285 R_a (e.g., Honda et al., 1987). Upper-mantle models for high $^3\text{He}/^4\text{He}$ ratios require further scrutiny, but the traditional assumption that high R indicates lower-mantle affinity is inconsistent with the seismic data of Yellowstone and elsewhere.

It is interesting to compare Yellowstone with Iceland, which has also been hypothesized to be underlain by a plume rising from the lower mantle (Morgan, 1971; Schilling, 1973). In Iceland too, certain geologic and tectonic features can be attributed only to coincidence if one appeals to a deep-mantle plume model. Iceland lies astride the Mid-Atlantic spreading center at a boundary between regions of contrasting tectonic style to the north and south. This change occurs at a major east-west composite fracture zone dating from opening of the North Atlantic along a reactivated Caledonian structure (Bott, 1985). Like Yellowstone, Iceland is also underlain by a low-wave-speed body and has high helium isotope ratios, the highest ($R_a = 42.9$) found in any hotspot rocks (Breddam and Kurz, 2001). Whole-mantle tomography provides no evidence that the seismic anomaly extends into the lower mantle beneath Iceland (e.g., Ritsema et al., 1999; Megnin and Romanowicz, 2000), and teleseismic tomography contraindicates such a model (Foulger et al., 2000, 2001; Foulger and Pearson, 2001). Thus, despite their different settings, Iceland and Yellowstone both disfavor a deep-mantle plume model.

CONCLUSIONS

We have discussed geologic data that disfavor a deep-mantle plume origin for the Yellowstone hotspot and seismic images that fail to detect any deep-mantle plume beneath Yellowstone or the eastern Snake River

Plain, and we have reasoned that a deep-mantle origin is not required by the helium isotope data. Proponents of a deep-mantle plume propose to explain special tectonic relationships of the Yellowstone hotspot either by coincidence or by special auxiliary hypotheses, according to which a plume may have been either narrow or very broad, either rose vertically or spread laterally in the mantle counterflow, or was deflected either northward or westward by a subducting plate from the west. Rather than appealing to such complex special reasoning, we suggest instead that the features attributed to a Yellowstone hotspot are integral components of regional upper-mantle tectonics of the northwestern United States.

We do not in this paper seek to discuss in detail specific models for generation of the Yellowstone melting anomaly and related magmatic-tectonic features in the upper mantle. However, the requirement that any proposed origin be consistent with the observations discussed herein suggests a general scenario involving positive feedback among enhanced basin-range extension, upwelling and partial melting of upper mantle, emplacement of the resulting mantle melts into the crust, consequent crustal melting, and voluminous surface eruption (p. 129–130 in Christiansen, 2001). These processes, in turn, are controlled by both large-scale and local structure of the lithosphere.

Such a scenario can be outlined as follows. Subduction of the Farallon plate brought about initial contact of the Pacific and North American plates at 30–28 Ma along a zone of right-lateral interaction that then grew at the expense of continental-margin subduction (Atwater, 1970; Atwater and Stock, 1998). By ca. 18–17 Ma, the Cascades arc existed along only a relatively small part of the margin, and stress relaxation—and perhaps right-lateral shear traction (Dickinson, 1997)—caused lithospheric extension and thinning, both inland from the coastal strike-slip system and in the adjacent backarc, which previously had been compressively deformed and magmatically heated. Upwelling of the underlying upper mantle and widespread partial melting increased regional heat flow and decreased rigidity, further enhancing lithospheric extension and thinning. The axis of extension was parallel to the continental margin through the present Great Basin to the Columbia Plateau, and the greatest efficiency in generating partial melts—where the lithosphere was already thinnest—produced flood basalts in the Pacific Northwest. Upward and lateral flow of hotter, relatively undepleted mantle replaced mantle depleted by partial melting. Localized convection was concentrated where the upwelling mantle abutted the cold, thick Archean craton (King and Anderson, 1998), forming the kernel from which both the Yellowstone hotspot and the Newberry melting anomaly would propagate.

The extending region progressively widened from its central axis and by ca. 10 Ma reached nearly to its present west and east margins. As the central zone approached a stable configuration, active extension, seismicity, and basaltic magma generation became increasingly concentrated outward toward the margins (Christiansen and McKee, 1978). Convection and magmatism related to the Yellowstone hotspot kernel migrated outward from its initial center adjacent to the Archean craton. Extension concentrated outward into progressively narrower zones toward the east and west margins of the widening region, and magmatism was focused where these zones intersected the boundary with thicker, cooler, more rigid lithosphere on the north. On the western side of the Yellowstone hotspot kernel, enhanced extension at the north margin of the extending and thinning lithosphere produced crustal melting that, relative to the continental interior, propagated with extensional widening to form the Newberry melting anomaly. On the east, similar enhanced extension, convection, and magmatism became localized and was guided by an ancient structural zone parallel to the direction of

plate motion. Because of this parallelism, the Yellowstone melting anomaly may have been further enhanced by shear melting at the base of the lithosphere and thermal feedback (Shaw and Jackson, 1974) whereby newly generated melts rise buoyantly, reducing pressure in the upper mantle below and promoting further melting. This thermal feedback between lithospheric extension and shear melting may produce a self-sustaining melting anomaly that maintains the most productive magmatic system of the western United States.

The numerous ad hoc variants of the deep-mantle plume model that are required to explain features associated with this and other hotspots and large igneous provinces testify to their considerable variation in structural, tectonic, and petrologic settings and magmatic styles. These variants also underline the poor fit of the plume model to many hotspots and suggest instead that shallow lithospheric architecture strongly influences the character—and is perhaps also responsible for the existence—of other hotspots.

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