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Lithospheric topography, tilted plumes, and the track of the Snake River–Yellowstone hot spot

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[1] The trace of the Snake River–Yellowstone hot spot is the world's best example of a mantle plume that has been overridden by continental lithosphere. The "standard model" calls for the plume head to rise under northern Nevada and be forced northward to form basalts of the Columbia Plateau; subsequent movement of North America to the southwest over the plume tail created a hot spot trace on the surface. We present a new conceptual model for the origin of this feature that resolves inconsistencies in the current standard model and explains the recent documentation of a thermal anomaly in the mantle below Yellowstone today that plunges ~65° WNW. Our model implies that the plume tail was forced beneath thinned cratonic lithosphere to the SE along with part of the plume head and has remained in this orientation for the last 12 Ma. We infer that almost all of the volcanism in SE Oregon and SW Idaho prior to 12 Ma results from overriding the southern extension of the plume head, not the plume tail, and that a distinct plume tail hot spot track was not established until formation of the Bruneau-Jarbidge eruptive center around 12 Ma. The plume tail track may also be controlled by a preexisting structural boundary in lithosphere that is thinner than adjacent lithosphere. This model demonstrates the potential importance of lithospheric topography on controlling the surface manifestation of plume volcanism and the complexity that may arise when lithospheric thickness is nonuniform. **Citation:** Shervais, J. W., and B. B. Hanan (2008), Lithospheric topography, tilted plumes, and the track of the Snake River–Yellowstone hot spot, *Tectonics*, *27*, TC5004, doi:10.1029/2007TC002181.

1. Introduction

[2] The trace of the Snake River–Yellowstone hot spot is the world's best example of a plume "tail" that has been overridden by continental lithosphere (Figure 1). The plume track is attested by the time-transgressive onset of volcanism [*Armstrong et al.*, 1975], the chemistry of plume-derived basalts, and by its spectacular topographic expression, which cuts through Basin and Range structures up to its culmination at the Yellowstone plateau. A number of competing, nonplume models have been proposed, including edge effects of cratonic lithosphere [e.g., *King and Anderson*, 1995] and "hotlines" [e.g., *Christiansen et al.*, 2002; *Humphreys et al.*, 2000]. However, the recent discovery of a cylindrical thermal anomaly ~100 km across in upper mantle below Yellowstone implies a plume origin [*Yuan and Dueker*, 2005; *Waite et al.*, 2006].

[3] The "standard model" that is now accepted by most workers has the Snake River–Yellowstone plume impacting the lithosphere around 17 Ma under northern Nevada (McDermitt caldera) and melting to form a plume head flood basalt province (Columbia River Basalt Group, CRBG), followed by the initiation of time-transgressive volcanism trending NE from its initial position under McDermitt caldera to its final position under the Yellowstone caldera [e.g., *Smith and Braile*, 1994; *Pierce and Morgan*, 1992; *Pierce et al.*, 2002] (Figure 1). The plume head–plume tail paradigm [*Richards et al.*, 1989; *Hill et al.*, 1992] is generally accepted as the model that best explains the large volume of the initial flood basalts and the subsequent time-transgressive nature of the younger volcanics. The track of the plume tail is assumed to be marked by the position of rhyolite caldera complexes that become younger to the northeast (age of initial volcanism). These include (from oldest to youngest) the McDermitt, Owyhee-Humboldt, Bruneau-Jarbidge, Twin Falls, Picabo, Heise, and Yellowstone caldera complexes (Figure 2).

[4] The standard model poses two significant conundrums: (1) the plume head flood basalt province is located far to the north of the purported plume track and (2) the purported plume track deviates significantly from North American plate motion in both trend and velocity prior to ~12 Ma (NUVEL [*Gripp and Gordon*, 1990, 2002; *Anders et al.*, 1989]). There are two plausible models for the first conundrum: tilting of the plume northwestward by the subducting Farallon slab [*Geist and Richards*, 1993; *Pierce et al.*, 2002], and flattening of the plume head against the margin of the continental lithosphere [*Camp*, 1995; *Camp and Ross*, 2004], represented by the $^{87}Sr/^{86}Sr = 0.706$ line in central Idaho [*Armstrong et al.*, 1977; *Fleck and Criss*, 1985, 2004] (Figures 1 and 2). The tilted plume model suggests that curvature in the plume track results as the plume returns to a vertical orientation after sinking of the Farallon slab [*Geist and Richards*, 1993]; the flattened plume head model implies that the plume was always located farther south, but does not address the misfit between observed plume track and North American plate motion.

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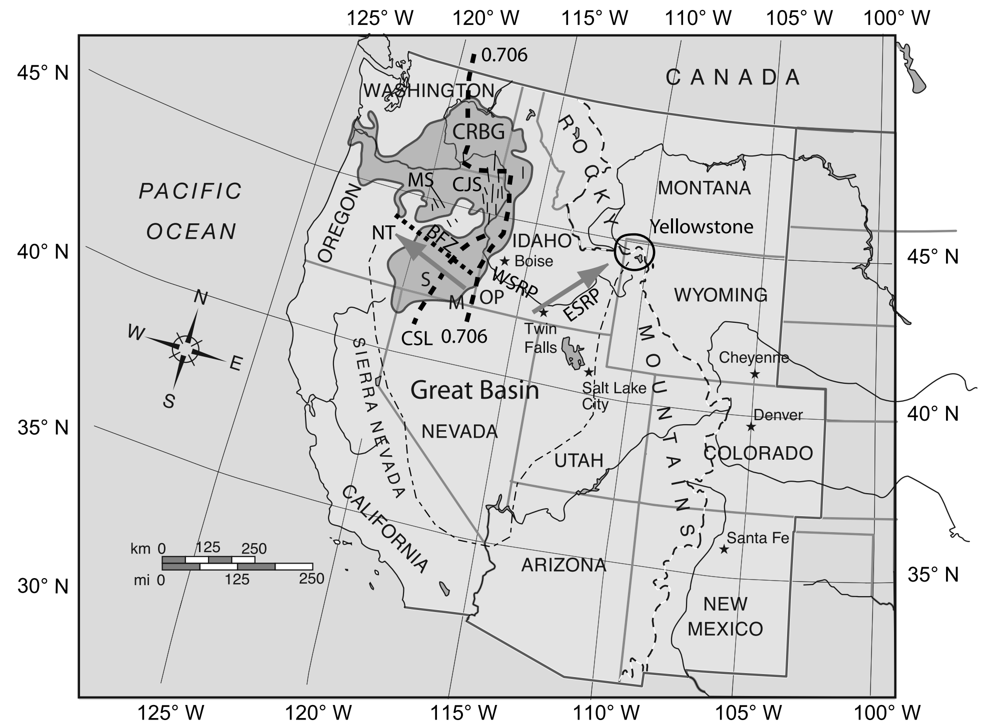

Figure 1. Outline map of the western United States showing the eastern Snake River Plain (gray arrow, ESRP), the western Snake River Plain (WSRP), the Newberry trend (gray arrow, NT), the Columbia River Basalt Group (CRBG, dark gray shading), the $^{87}Sr/^{86}Sr$ = 0.706 line and "Cenozoic Sr line" (CSL) of *Leeman et al.* [1992] (heavy dashed lines), the Brothers Fault zone (BFZ, heavy dotted line), the outline of the Great Basin (light dash-dotted line), the continental divide (light dashed line), and the Yellowstone plateau volcanic complex (heavy circle, Yellowstone). The ESRP trend of time-transgressive volcanism has been interpreted to reflect movement of the North American Plate over the Yellowstone hot spot; movement of the plate is to the SW, in the opposite sense of the time-transgressive volcanism. CJS, Chief Joseph Dike Swarm; M, McDermitt volcanic complex; MS, Monument Dike Swarm; OP, Owyhee Plateau; S, Steens Mountain. Distribution of CRBG is modified after *Camp and Ross* [2004].

[5] *Camp and Ross* [2004] use the age progression of feeder dikes and volcanic centers in eastern Oregon and Washington to document the impact and subsequent deformation of the plume head against the cratonic margin. This deformation ultimately led to decapitation of the plume head from its tail as the plume system was overridden by North American lithosphere [*Camp and Ross*, 2004]. They infer a focus of initial plume head volcanism at Steens Mountain in SE Oregon on the basis of the age progression of basalt volcanism. This model has been elaborated by *Hooper et al.* [2007], who present extensive geochemical evidence for a plume origin. *Glen and Ponce* [2002] have proposed that the plume head impacted the lithosphere ~44°N (near the NW end of the western Snake River Plain) on the basis of dikes, faults, and magnetic lineaments that radiate from a focus in this area.

[6] We present here a new model that examines the influence of lithospheric topography on deflecting thermochemical mantle plumes, and interactions between the plume and its overlying lithosphere. We build on the models of *Camp and Ross* [2004] and *Hooper et al.* [2007] and extend these to explain the subsequent hot spot track as defined by the topographic Snake River plain and by potential field data (gravity, magnetics). We also incorporate the results of new seismic tomography that images the position of the Yellowstone plume today [*Yuan and Dueker*, 2005; *Waite et al.*, 2006] and the seismic structure of the mantle beneath the eastern SRP [*Saltzer and Humphreys*, 1997; *Schutt and Humphreys*, 2004].

2. Regional Geology and Plume Head Volcanism

[7] The western margin of the North American cratonic lithosphere is marked by the $^{87}Sr/^{86}Sr$ = 0.706 line in Mesozoic plutons [*Armstrong et al.*, 1977; *Fleck and Criss*, 2004]. The $^{87}Sr/^{86}Sr$ = 0.704 line is coincident with the $^{87}Sr/^{86}Sr$ = 0.706 line in central Idaho, which implies a near-vertical cratonic boundary with accreted oceanic terranes to the west [*Armstrong et al.*, 1977; *Fleck and Criss*, 2004] (Figures 1 and 2). The coincidence of these isopleths in central Idaho despite Sevier and Laramide thrusting implies that this thrusting was deep seated [*Leeman et al.*, 1992]. *Leeman et al.* [1992] also showed that basalts south of

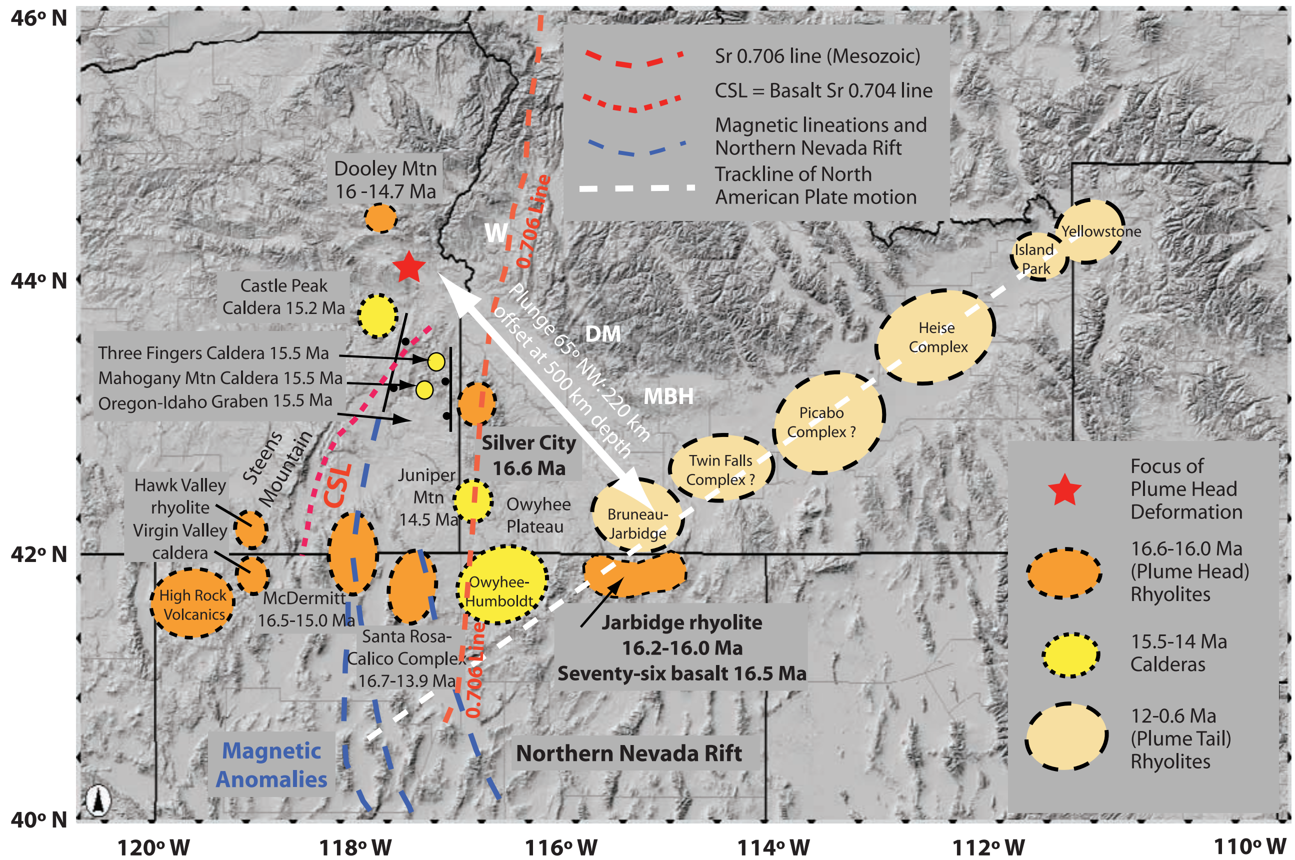

Figure 2. Digital topography of southern Idaho, eastern Oregon, and northern Nevada, showing location of ~16.6 to 16.0 Ma rhyolite eruptive complexes, 15.5 to 14 Ma (largely calc-alkaline) eruptive complexes, and caldera complexes associated with the Snake River–Yellowstone hot spot track. Linear features include the vector trend of North American plate motion (white dashed line), the $^{87}Sr/^{86}Sr$ = 0.706 line (red dashed line), the $^{87}Sr/^{86}Sr$ = 0.704 line (CSL, red dotted line), and magnetic anomalies that represent mafic intrusions into crust (blue dashed lines). Also shown are the Oregon-Idaho graben, the expected offset of the Snake River–Yellowstone plume at 12 Ma based on a plunge of 65° NW and depth of 500 km (solid white arrow), and the inferred center of the plume head at circa 12 Ma (red star). Not shown are extensive 10–12 Ma rhyolites associated with the rhyolite flare-up of the western SRP and Owyhee Plateau. DM, Danskin Mountains; MBH, Mount Bennett Hills; W, Weiser embayment.

latitude 43.5°N define a zone of transitional lithosphere between longitude 116.5°W (the $^{87}Sr/^{86}Sr$ = 0.706 line) and 118.5°W (their "Cenozoic Sr Line" or CSL; Figure 1). Basalts erupted within this transition zone have $^{87}Sr/^{86}Sr$ ratios that range from ~0.704 to 0.707 and $^{143}Nd/^{144}Nd$ ratios that range from 0.5124 to 0.5128 (Figure 3) [*Leeman et al.*, 1992; *Shoemaker and Hart*, 2002]; that is, the CSL is essentially the basalt $^{87}Sr/^{86}Sr$ = 0.704 line.

[8] It is generally accepted that the first manifestation of plume-derived volcanism in the Columbia River Basalt Group (CRBG) was eruption of "lower" Steens-type basalt from vents in southeastern Oregon over a very short time interval (circa 100 ka) around 16.6 Ma [*Camp et al.*, 2003; *Camp and Ross*, 2004; *Jarboe et al.*, 2006; *Hooper et al.*, 2007]. This was followed almost immediately by widespread eruptions of rhyolite ash flows and ignimbrites at around 16.5 Ma from many caldera complexes in northern Nevada, SE Oregon, and SW Idaho [*Rytuba and McKee*, 1984; *Ekren et al.*, 1984; *Brueseke et al.*, 2008; *Henry et al.*, 2006] (Table 1). These rhyolites erupted for the most part east of the CSL, as defined by *Leeman et al.* [1992], with some exceptions (Virgin Valley eruptive center [*Henry et al.*, 2006] and Dooley Mountain rhyolites [*Evans*, 1992]).

[9] West of the CSL volcanism was dominantly basaltic, with lower Steens basalts followed in quick succession by the Imnaha basalt (coeval to upper Steens at Steens Mountain), and the Grande Ronde–Picture Gorge basalts (Table 1). Although it is not possible to distinguish this succession using radiometric dates, it is clearly established by stratigraphic successions in deep canyon exposures along the Owyhee, Malhuer, Imnaha, and Snake rivers [*Reidel et al.*, 1989; *Tolan et al.*, 1989; *Hooper*, 2000; *Hooper et al.*, 2002; *Camp et al.*, 2003; *Camp and Ross*, 2004]. It is now recognized that the bulk of these basalts and rhyolites

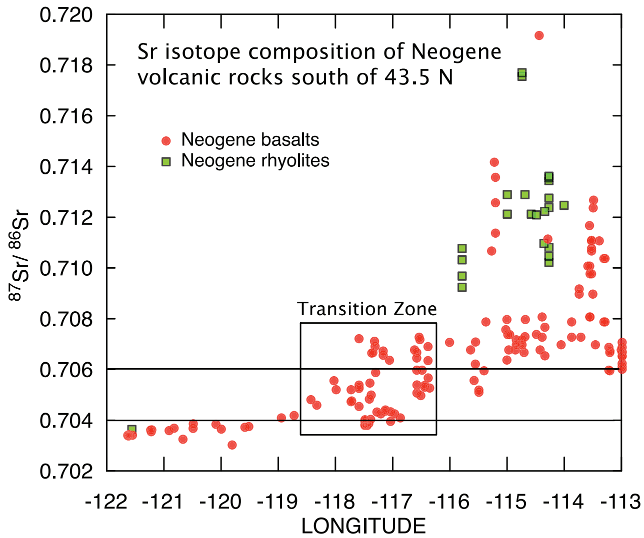

Figure 3. Isotopic composition of Sr in Neogene volcanic rocks south of 43.5°N in SE Oregon, SW Idaho, northern Nevada, and NE California, as a function of longitude [after *Leeman et al.*, 1992]. Basalts and rhyolites east of ~116.5°W have $^{87}Sr/^{86}Sr > 0.706$ (except for a few high-K basalts of Boise River Group [*Vetter and Shervais*, 1992]); this corresponds approximately with the $^{87}Sr/^{86}Sr = 0.706$ line in Mesozoic plutons [*Fleck and Criss*, 2004]. Basalts and rhyolites west of ~118.5°W have $^{87}Sr/^{86}Sr < 0.704$, whereas basalts in the transition zone between ~116.5°W and 118.5°W have $^{87}Sr/^{86}Sr$ between 0.704 and 0.706. The boundary between basalts with $^{87}Sr/^{86}Sr < 0.704$ and those with $^{87}Sr/^{86}Sr > 0.704$ represents the "Cenozoic Sr Line" (CSL) of *Leeman et al.* [1992]. Data are from the North American Volcanic and Intrusive Data Base (NAVDAT) geochemical database.

erupted within a period of <1 Ma on the basis of magnetostratigraphy and high-precision Ar-Ar dates of the lavas [*Jarboe et al.*, 2006], although many continued until around 15 Ma [e.g., *Hooper et al.*, 2002, 2007].

[10] It is less commonly recognized that Steens-type basalts erupted over a much broader area than previously recognized, including the Seventy-six basalt in the Jarbidge Mountains, NE Nevada (16.5 Ma [*Rahl et al.*, 2002]), olivine basalt of the Silver City Range in SW Idaho (16.6 Ma [*Pansze*, 1975]), and possibly the Dooley Mountain basalt in NE Oregon (>16 Ma [*Evans*, 1992]). All of these basalts are overlain by slightly younger rhyolite complexes (Table 1) and together they define a broad area of plume head volcanic activity that occurred over a short time span across northern Nevada, eastern Oregon, and western Idaho (Figure 1). This activity is coeval with a larger area of dominantly basaltic volcanism that is concentrated west and north of the CSL (Figures 1 and 2).

[11] North of 44° latitude, younger volcanism was dominantly basaltic and included the Wanupum and the less voluminous Saddle Mountain basalts, which continued to erupt sporadically until ~6.5 Ma. Younger volcanism in the southeast (east of the CSL and largely west of the $^{87}Sr/^{86}Sr = 0.706$ line) was dominantly rhyolitic, and much of it was associated with formation of the Oregon-Idaho graben and Northern Nevada Rift, starting around 15.5 Ma and continuing until at least 13.7 Ma [*Zoback et al.*, 1994; *Cummings et al.*, 2000; *Hooper et al.*, 2007]. Volcanism in the Oregon-Idaho graben was calc-alkaline in nature and is not clearly related to plume head volcanism; it has been linked to continued back-arc extension subsequent to plume head volcanism [*Hooper et al.*, 2002; *Camp and Ross*, 2004]. Other 15.5–14 Ma rhyolite volcanism in the southeast included centers such as the Juniper Mountain complex and the Swisher Tuff of the Owyhee Plateau [*Ekren et al.*, 1981, 1982, 1984] and the Humboldt eruptive center in northern Nevada (Figure 2).

[12] Many of the ~16.5 Ma rhyolite complexes form an east–west trending array along the Nevada-Oregon border (i.e., High Rock, Virgin Valley, Hawk Valley, McDermitt, Santa Rosa–Calico, Jarbidge), but this trend is broken by complexes located to the north (Silver City, Dooley Mountains) and south (volcanics of the Northern Nevada Rift, Figure 2). This apparent E–W trend may reflect lithospheric delamination at the southern edge of the spreading plume head or may be an artifact of later volcanism and limited exposures beneath the cover of younger (15.5 to 13.7 Ma) volcanism of the Owyhee Plateau (e.g., Juniper Mountain center) and Oregon-Idaho graben (Figure 2 and Table 1). The E–W trend has been accentuated by younger E–W extension associated with opening of the Oregon-Idaho graben [*Cummings et al.*, 2000] and with NW trending right-lateral transtension along the southern margin of the western Snake River Plain [*Wood and Clemens*, 2002; *Hooper et al.*, 2002, 2007]. This right-lateral movement cannot exceed 60 km, however, on the basis of the apparent offset of the $^{87}Sr/^{86}Sr = 0.706$ line north and south of the plain (Figure 2), and total extension associated with the Oregon-Idaho graben is modest, on the basis of its high-angle fault boundaries [*Cummings et al.*, 2000].

[13] Plume track volcanism became firmly established with the formation of the Bruneau-Jarbidge eruptive center starting around 12 Ma [*Bonnichsen*, 1982]. This eruptive center lies at the SW end of the eastern Snake River Plain trend and is the first rhyolite eruptive center to form on the trend that parallels North American plate motion back from Yellowstone (Figure 2).

3. Model Constraints

[14] In order to understand the model which follows, we need to elaborate the constraints that have led us to this model, and which must be addressed by any model that seeks to explain the track of the Snake River–Yellowstone hot spot. The five topics discussed in sections 3.1–3.5 must be addressed by any successful model for this system.

3.1. Orientation of the Plume Today

[15] New seismic tomographs document a low-velocity (thermal) anomaly in the upper mantle below Yellowstone that plunges ~60°–70° to the WNW and extends to a depth

Table 1. Comparison of Ages[a]

	Age (Ma)	Comments	References
		Owyhee Plateau Rhyolite Centers	
McDermitt complex	16.54 ± 0.11 to 16.15 ± 0.13	overlies Steens basalt	*Henry et al.* [2006]
Jarbidge rhyolite	16.2 ± 0.1	overlies Seventy-six basalt	C.D. Henry (unpublished data, 2008)
Silver City complex	16.6–16.3 ± 0.1	ash flow tuffs and flow-banded rhyolites	*Bonnichsen et al.* [2008]
Hawks Valley rhyolites	16.5–16.0	SE Oregon	*Henry et al.* [2006]
High Rocks rhyolites	16.6–16.1	NW Nevada	*Noble et al.* [1973] and *Swisher et al.* [1990]
Santa Rosa-Calico complex	16.6–15.4 ± 0.1	north end Northern Nevada Rift	*Brueseke et al.* [2008]
Virgin Valley Caldera	16.40–16.22 ± 0.06	Idaho Canyon tuff, west of Steens, overlies	*Henry et al.* [2006]
Oregon Valley tuff	16.47 ± 0.03	overlies lower Steens	*Henry et al.* [2006]
Trout Creek Mountains	16.42 ± 0.03	overlies lower Steens	*Henry et al.* [2006]
Dooley Mountain complex	>16–14.7 ± 0.4	date on youngest rhyolite dome; underlain by Steens-type basalt	*Evans* [1992]
Juniper Mountain complex	14.5–13.7	Swisher Mountain tuff plus four other tuffs	*Manley and McIntosh* [2002] and *Bonnichsen et al.* [2008]
Three Fingers, Mahogany Mountain, and Saddle Butte calderas	15.5–15.2	LOVC–OIG (tuff of Leslie Gulch, tuff of Spring Creek)	*Rytuba et al.* [1991] and *Rytuba and Vander Meulen* [1991]
Castle Peak caldera	15.3–14.7	LOVC–Dinner Creek tuff	*Rytuba et al.* [1991]
Rhyolite of Cottonwood Mountain	15.24 ± 0.3	OIG	*Lees* [1994]
Tuff of Leslie Gulch	15.8–15.5 ± 0.5	OIG (Sucker Creek formation)	*Ekren et al.* [1984] and *Vander Meulen et al.* [1987]
		Owyhee Plateau Basalt	
Birch Creek (Malhuer Gorge)	15.7 ± 0.1	Grande Ronde equivalent	*Cummings et al.* [2000] and *Camp et al.* [2003]
Hunter Creek basalt (Grande Ronde)	15.3	interbedded with Dinner Creek tuff	*Cummings et al.* [2000] and *Hooper et al.* [2002]
		CRBG Main Phase Basalts	
Steens basalt at Pueblo Mountain	17.03 ± 0.28	lower Steens	*Hart et al.* [1989]
Steens basalt at Steens Mountain	16.6 ± 0.2	lower Steens (C5Cr Chron)	*Hooper et al.* [2002] and *Jarboe et al.* [2006]
Imnaha basalt	16.1–15.0	equivalent upper Steens	*Hooper et al.* [2002]
Grande Ronde–Picture Gorge basalt	16.1–15.0	overlies Imnaha	*Hooper et al.* [2002]
Seventy-six basalt (Jarbidge Mountains)	16.5 ± 0.2	Steens equilvalent, under Jarbidge rhyolite	*Rahl et al.* [2002]
Silver City basalt (Silver City Range)	∼16.6	Steens equivalent, under Silver City rhyolite	*Pansze* [1975]
Santa Rosa–Calico basalt	16.7 ± 0.2	basalt below rhyolite	*Brueseke et al.* [2008]
Lower Pole Creek (Malhuer Gorge)	16.9 ± 0.8	lower Steens equivalent	*Cummings et al.* [2000] and *Camp et al.* [2003]
Upper Pole Creek (Malhuer Gorge)	16.5 ± 0.3	upper Steens, Imnaha equivalent	*Cummings et al.* [2000] and *Camp et al.* [2003]

[a]Rhyolite eruptive centers of Owyhee Plateau area (SE Oregon, north Nevada, SW Idaho) with main phase basalts of the Columbia River Group and related basalts of SE Oregon. LOVC, Lake Owyhee volcanic field; OIG, Oregon-Idaho graben. Note that all ages may be adjusted upward by 1% to account for error in ^{40}K decay constant [*Min et al.*, 2000].

of at least 600 km (Figure 4) [*Yuan and Dueker*, 2005; *Waite et al.*, 2006]. This anomaly is ∼100 km in diameter, i.e., in the same size range as the Hawaiian and Iceland plumes at depth [*Li et al.*, 2000; *Wolfe et al.*, 1997; *Shen et al.*, 1998; *Allen and Tromp*, 2002], and extends upward to the base of the lithosphere at ∼50 km depth [*Yuan and Dueker*, 2005; *Waite et al.*, 2006]. The velocity anomaly appears to be offset to the SE at ∼250–300 km depth, with an appendage to the SE that may be a roll of return flow from the plume (Figure 4a). Another interesting aspect of this tomography is the distinct high-velocity anomaly (blue) SE of the plume that dips to the NW, parallel to the low-velocity plume anomaly (Figure 4a). *Yuan and Dueker* [2005] suggest that this anomaly may represent a slab of delaminated lithosphere.

3.2. Topography of the Lithosphere-Asthenosphere Boundary

[16] The lithosphere-asthenosphere boundary under the eastern SRP has been imaged along a detailed NW–SE

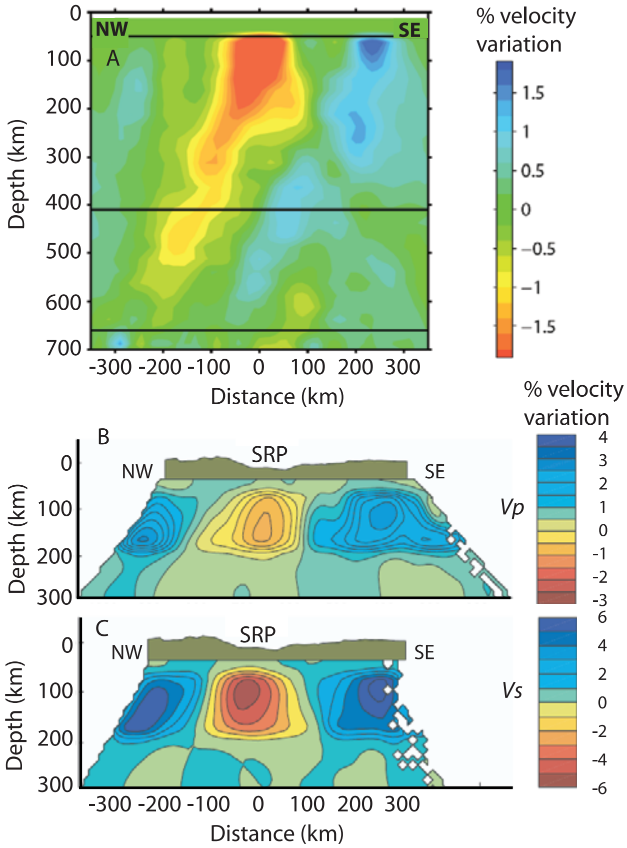

Figure 4. Seismic tomographs of the Yellowstone–SRP system. (a) Seismic tomograph along NW–SE transect across active Yellowstone plume [from *Yuan and Dueker*, 2005]. The low-velocity anomaly (red, yellow) forms a plume of buoyant material ∼100 km in diameter that plunges ∼65 NW, with a distinct offset to the SE at ∼300 km depth. *Yuan and Dueker* [2005] have proposed that the high-velocity anomaly (blue) SE of the plume represents a slab of delaminated lithosphere. (b, c) seismic tomographs of *P* wave velocity (Figure 4b) and *S* wave velocity (Figure 4c) beneath the eastern Snake River Plain, showing a NW dipping low-velocity channel that underlies the volcanic plain and high-velocity shoulders that underlie the margins of the plain [from *Schutt and Humphreys*, 2004]. The low-velocity anomalies are interpreted to represent hot partially molten mantle with up to 1% partial melt, whereas the high-velocity anomalies are interpreted to represent colder, depleted mantle.

transect by several studies (Figures 4b and 4c). *Saltzer and Humphreys* [1997] document a low-velocity anomaly that is confined to a channel <200 km wide, which has been carved into the cratonic lithosphere of North America. This sublithospheric channel dips steeply to the NW and extends to a depth of ∼300 km [*Schutt and Humphreys*, 2004], consistent with the observed plunge of the plume tail today [*Yuan and Dueker*, 2005; *Waite et al.*, 2006] and confirmed by recent Rayleigh wave modeling [*Schutt et al.*, 2008; *Stachnik et al.*, 2008]. The lithosphere is typically around 150–200 km thick beneath the Wyoming craton, making this sublithospheric channel at least 100–150 km deep. *Schutt and Humphreys* [2004] estimate that mantle in the low-velocity zone is ∼200°C hotter than the surrounding mantle and contains up to 1% partial melt on the basis of the observed V_p/V_s structure. The presence of this sublithospheric conduit implies removal of previously existing lithosphere by a combination of thermal and mechanical erosion. *Saltzer and Humphreys* [1997] observe high-velocity shoulders on this channel that extend to depths of around 300 km, which they interpret as ''rolls'' of cooler, depleted plume material that sinks in response to localized convection and the negative buoyancy of refractory plume mantle after melt extraction. *Schutt and Humphreys* [2004] calculate that this residual mantle is ∼80°C cooler then the surrounding mantle and depleted by ∼5% melt removal.

3.3. Temporal Distribution of Volcanism

[17] The oldest rhyolite eruptive center that lies near the Snake River–Yellowstone plume track (as defined by the trend of the eastern SRP) and parallels North American plate motion vectors calculated from magnetic anomaly lineaments in the central Atlantic (e.g., NUVEL [*Gripp and Gordon*, 1990, 2002]) is the Bruneau-Jarbidge eruptive center, SW of Twin Falls, Idaho [*Bonnichsen*, 1982] (Figure 2). Previous workers have suggested that volcanic centers which lie west of the Bruneau-Jarbidge complex mark the inception of plume-related eruptive activity along the plume track (McDermitt, Owyhee-Humboldt). However, these centers are not consistent with the absolute motion of North America in either azimuth or velocity, as based on magnetic anomalies in the central Atlantic, and lie far to the north of the projected plume track based on North American plate vectors (Figure 2 and Table 2). In addition, the purported plume track prior to 12 Ma ignores coeval caldera complexes that do not support a consistent time-transgressive trend based on their location (e.g., Silver City, Jarbidge), and concatenates two coeval but separate eruptive centers (Juniper Mountain, Humboldt) that lie, respectively, north and south of the purported plume track (Figure 2).

[18] In fact, there is a broad area north and east of the putative plume track that is underlain by basaltic and rhyolitic lavas which are essentially coeval with eruption of the Steens and Imnaha basalts at around 16.5 to 16.1 Ma [e.g., *Christiansen and Yeats*, 1992; *Christiansen et al.*, 2002]. Steens/Imnaha-type basalts include the Pole Creek basalt in Malheur Gorge [*Hooper et al.*, 2002], the Silver City basalt [*Ekren et al.*, 1984], the Seventy-six basalt in the Jarbidge Mountains [*Rahl et al.*, 2002], subrhyolite basalt of the Santa Rosa–Calico complex (Northern Nevada Rift [*Brueseke et al.*, 2008]), and the Dooley Mountain basalt [*Evans*, 1992], all of which are now recognized as part of the Columbia River Basalt Group. Rhyolite ash flows and eruptive centers that just postdate Steens-type basalts include the McDermitt, High Rock, Virgin Valley, Silver City, Jarbidge, Santa Rosa–Calico, and Dooley Mountain eruptive centers (Table 1). The oldest rhyolite complexes on the Owyhee plateau include the Silver City eruptive center (16.6–16.3 ± 0.1 Ma), which is underlain by Steens-type basalt [*Ekren et al.*, 1984; *Bonnichsen et al.*, 2008], the

Table 2. Comparison of North America Plate Motion Trend and Velocity[a]

	Trend	Velocity (mm/a)	Reference
		Model	
NUVEL-1A	N56°E ± 17°	22 ± 0.8	*Gripp and Gordon* [1990]
HS3 NUVEL-1A at 112°W	N56°E	18	*Gripp and Gordon* [2002]
HS3 NUVEL-1A at 116°W	N52°E	18.5	*Gripp and Gordon* [2002]
		Calculated	
SRP trend 0–10 Ma	N56°E ± 5°	29 ± 0.5	*Pierce and Morgan* [1992]
SRP trend 10–16 Ma	N75°E	70	*Pierce and Morgan* [1992]
SRP trend 0–9 Ma	N50°E	34.3	*Pollitz* [1988]
SRP trend 0–16 Ma	N56°E	45	*Rodgers et al.* [1990]

[a]Using hot spot reference frame (NUVEL-1A) and results calculated from apparent age progressions in Snake River Plain plume track. Calculated SRP trends for last 10 Ma are close to values for hot spot reference frame, whereas values for 10–16 Ma differ in both trend and velocity.

Santa Rosa–Calico complex (16.6–13.9 ± 0.1 Ma), underlain by 16.7 Ma basalt [*Brueseke et al.*, 2008], and the Jarbidge rhyolite (16.2 Ma), which is underlain by the Seventy-six basalt (16.5 ± 0.2 Ma [*Rahl et al.*, 2002]). The Silver City complex is adjacent to the western SRP (north of McDermitt caldera), whereas the Santa Rosa–Calico and Jarbidge rhyolites erupted up to 200 km east of McDermitt. Only the youngest unit of the Dooley Mountain rhyolite has been dated, but it is estimated that the oldest rhyolites and the underlying basalt are circa 16.5–16.0 Ma [*Evans*, 1992]. Taken together with coeval rhyolites in SE Oregon and northern Nevada, these units define an immense area of dominantly rhyolitic eruptive volcanism that is coeval with the McDermitt caldera and overlies Steens-type basalt.

[19] The remarkable coincidence of these ages over such a widespread area is consistent with plume head-induced volcanism that is dominantly basaltic to the north and west (eastern Oregon and SE Washington), and rhyolitic to the south and southeast, across a broad swath of northern Nevada, southeastern Oregon, and the Owyhee Plateau of SW Idaho (Figures 2 and 5a). There is no evidence that links the volcanism that occurred prior to 12 Ma to the absolute plate motion of North America and thus no support for previous assertions of time-transgressive progression toward the east before 12 Ma.

3.4. Impact of the Plume Head on Lithospheric Mantle

[20] Four areas have been proposed for the location beneath which the central axis of the plume head impacted the lithosphere: McDermitt caldera [*Pierce and Morgan*, 1992; *Pierce et al.*, 2002]; Steens Mountain [*Camp*, 1995; *Camp and Ross*, 2004; *Camp et al.*, 2003]; the central Snake River Plain [*Jordan et al.*, 2004] and an area in east central Oregon approximately 44°N latitude [*Glen and Ponce*, 2002]. *Pierce and Morgan* [1992] infer a plume head locus at the McDermitt eruptive center on the basis of its location at the western end of postulated plume track. This ignores the widespread occurrence of coeval rhyolites from vents located to the east, west, and northeast, which show that the track of the plume tail does not exist prior to ~12 Ma, as discussed in section 3.3. Camp and coworkers favor a location under Steens Mountain on the basis of the age progression of basaltic volcanism, which appears to initiate at progressively younger ages from Steens Mountain in south to the Chief Joseph dike swarm in the north [*Camp*, 1995; *Camp and Ross*, 2004; *Camp et al.*, 2003] (Figure 1). However, Steens basalt crops out as far north as the Wallowa Mountains in NE Oregon [*Hooper et al.*, 2007] and as far east as the Jarbidge Mountains in northern Nevada [*Rahl et al.*, 2002], suggesting a more complex progression. *Jordan et al.* [2004] propose a site beneath the central Snake River Plain (near the Bruneau-Jarbidge eruptive center), but this site lies far to the east of most ~16.5 Ma volcanism.

[21] A wide range in geological and geophysical data point to an area ~44°N latitude near the northwestern tip of the western Snake River Plain as the central focus of the Snake River–Yellowstone plume head, as proposed by *Glen and Ponce* [2002]. These include: annular distribution of flood basalts (Steens, Imnaha, Picture Gorge, Grande Ronde) in Oregon, Washington, and Idaho [*Reidel et al.*, 1989; *Camp et al.*, 2003; *Camp and Ross*, 2004] (Figure 1); orientations and age progressions in the Chief Joseph and Monument dike swarms [*Camp*, 1995; *Camp and Ross*, 2004] (Figure 1); orientation of magnetic lineaments [*Glen and Ponce*, 2002]; and major fractures (Northern Nevada Rift, Weiser embayment, Oregon-Idaho graben, protowestern SRP [*Zoback et al.*, 1994; *Cummings et al.*, 2000; *Wood and Clemens*, 2002; *Glen and Ponce*, 2002; *Camp and Ross*, 2004]) (Figure 2); and a major long-wavelength gravity anomaly that underlies the western Snake River Plain, but not adjacent crust to the south or north, and which culminates at the focus of the dikes, fractures, and magnetic lineaments (Figure 5). Models of continental lithosphere uplift in response to plume impact that use a multilayered lithosphere (elastic-brittle-ductile) show that topographic response to this impact may be complex [e.g., *Burov and Guillou-Frottier*, 2005; *Burov et al.*, 2007]. These models suggest that the central uplift over the plume head may be surrounded by an annulus or moat of lower elevation [e.g., *Burov and Guillou-Frottier*, 2005; *Burov et al.*, 2007] that may become a sink for subsequent basalt eruptions. This is consistent with the distribution of CRBG lavas, which surround a central uplift that was never completely covered

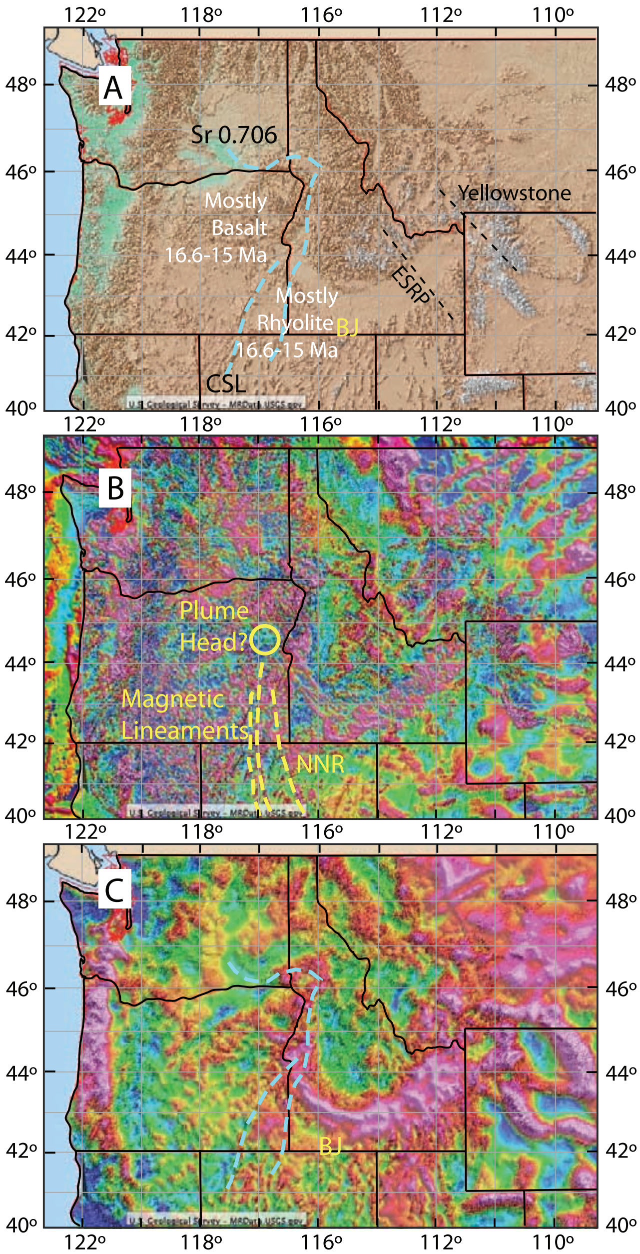

Figure 5. Topographic and geophysical character of the NW United States: (a) digital topography showing outlines of dominantly basaltic plume head volcanism at 16.5–15.0 Ma (red polygon) and dominantly rhyolite plume head volcanism at same time period (white polygon); ESRP trend ends at the Bruneau-Jarbidge eruptive center (BJ); (b) aeromagnetic map showing magnetic lineaments associated with impingement of the plume head (white lines), including the Northern Nevada Rift, and inferred location of plume head center at time of impact with lithosphere (dashed white circle); and (c) isostatic gravity map of western United States, showing strong positive gravity anomaly associated with western Snake River Plain; dashed lines are CSL and $^{87}Sr/^{86}Sr = 0.706$ lines. All maps from U.S. Geological Survey Mineral Resources Web site (http://tin.er.usgs.gov/).

with basalt (Figure 1). All of these observations point to an initiation of plume impact at ~44°N, as proposed by *Glen and Ponce* [2002], in east central Oregon (Figure 5b). This is over 220 km north of the McDermitt caldera complex, and even farther north of the projection of the plume track to the southwest from Yellowstone (Figure 2).

3.5. Distribution of Transitional Lithosphere

[22] It has long been realized that basalts of the CRBG erupted almost exclusively through new continental crust formed by the accretion of Paleozoic and Mesozoic oceanic terranes (largely island arc complexes) to the cratonic margin of North America [e.g., *Camp and Ross*, 2004]. These terranes lie west of the $^{87}Sr/^{86}Sr$ 0.706 line, which is interpreted to represent the western edge of cratonic North America [*Armstrong et al.*, 1977; *Fleck and Criss*, 1985, 2004] (Figure 2), and for the most part thick sections of flood basalt are confined to areas west of this boundary. This boundary coincides with the western margin of the Idaho batholith in central Idaho, and continues southward across the Owyhee Plateau and into Nevada (Figure 2).

[23] In central Idaho, north of 44°N latitude, this boundary is also coincident with the $^{87}Sr/^{86}Sr$ 0.704 line, defining a steep gradient from thin lithosphere of the Phanerozoic accreted terranes to thick lithosphere of cratonal margin (Archean to the south, Proterozoic to the north). Near the southern end of the Idaho batholith (around 44°N), the $^{87}Sr/^{86}Sr$ 0.704 line based on basalt (CSL) diverges from the $^{87}Sr/^{86}Sr$ 0.706 line based on Mesozoic batholiths (Figure 2). The area between these two isopleths lines must be underlain by lithosphere that is transitional between the thin, relatively young oceanic lithosphere to the west and thick Archean lithosphere to the east. This lithosphere may represent Archean lithosphere thinned by later extension, or the stacking of older lithosphere onto younger lithosphere during convergent orogenesis. Confirmation of continental lithosphere west of the $^{87}Sr/^{86}Sr$ 0.706 line and east of the CSL ($^{87}Sr/^{86}Sr$ 0.704 basalt line) comes from xenoliths of granitic and metamorphic rock found in volcanic rocks of the Oregon-Idaho graben; farther west xenoliths of metabasalt and chert document basement of oceanic affinity [*Evans et al.*, 2002].

[24] There are two scenarios that may account for the transitional lithosphere inferred to underlie the region between the CSL and $^{87}Sr/^{86}Sr$ 0.706 lines, as noted above. *Leeman et al.* [1992] propose tectonic erosion and stacking of younger lithosphere onto older lithosphere during convergent orogenesis in the late Mesozoic. This is required to maintain the close spacing of the $^{87}Sr/^{86}Sr$ 0.704–0.706 lines in central Idaho, but does not easily explain the transitional zone noted farther to the south. We suggest that south of ~44°N latitude, the Archean lithosphere was thinned by later extension to form the transition zone. *Wells and Simpson* [2001] have documented circa 45° of clockwise rotation in the Oregon Coast Ranges, and northern Basin and Range, which predates the mid-Miocene eruption of the CRBG. The pole of rotation is located near the Oregon-Washington border in easternmost Oregon, which resulted in small amounts of extension along the western

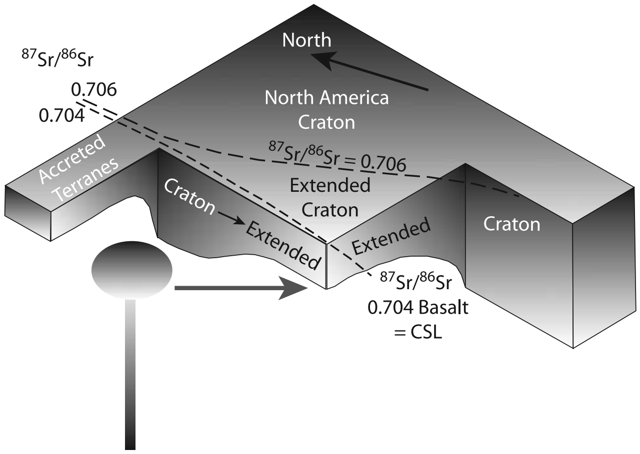

Figure 6. Schematic isometric model of lithosphere topography in NW United States, with thin accreted oceanic lithosphere to the west, thick Archean and Paleoproterozoic lithosphere to the east, and an region of transitional lithosphere in between that is confined to SE Oregon, SW Idaho, and northern Nevada. As plume head is compressed against the cratonic lithosphere, it is extruded to the north (along the steep cratonal boundary) and to the south, where the plume head and tail are overridden by the transitional lithosphere. Once the plume tail has etched a channel into the overlying lithosphere (circa 12 Ma), it remains trapped in that orientation through the buoyant flux of material into the channel sink. See text for details.

margin of the Idaho batholith and progressively greater extension farther south and west.

4. The Model

[25] The model constraints discussed in section 3 show that the Snake River–Yellowstone plume impacted the base of the lithosphere circa 17 Ma at ~44°N in east central Oregon, somewhere near the northwestern tip of the western SRP, and west of the boundary between cratonic North America and Phanerozoic oceanic terranes accreted to North America during the Mesozoic. The plume head was flattened against the thick lithospheric roots of the North American craton in response to SW directed plate motion, forcing it largely to the north as basalts erupted from central vents and dikes that become younger to the north and west [*Camp and Ross*, 2004]. The question remains: how did the plume tail become positioned under the accepted plume track, as defined by the eastern Snake River plain, and constrained by plate motion studies?

[26] We believe that there are two salient observations that address this issue: the broad area of 16.5 to 15.5 Ma bimodal volcanism underlain in part by extended cratonic lithosphere between the CSL (^{87}Sr/^{86}Sr = 0.704 basalt) and ^{87}Sr/^{86}Sr = 0.706 isopleths (Figure 2) and the 60° to 70° NW plunge of the plume tail under Yellowstone today (Figure 4) [*Yuan and Dueker*, 2005; *Waite et al.*, 2006].

[27] We propose that the broad zone of bimodal (but largely silicic) volcanism that characterizes northern Nevada, southeastern Oregon, and the Owyhee plateau of SW Idaho represents the southern half of the plume head as it was overridden by the extensionally thinned lithosphere that lies between the CSL and ^{87}Sr/^{86}Sr = 0.706 lines. As noted in section 3.5, this extension is largely pre-Miocene in age and may have formed in response to clockwise rotation of the Oregon Coast Ranges [*Wells and Simpson*, 2001]. The lithosphere and crust were thin enough and cold enough that some basalts were able to erupt relatively early in the cycle, but for the most part this volcanism is dominated by rhyolite caldera complexes that overlap in age with the main phase CRBG eruptions (16.5–15.5) and which continued to erupt until around 14 Ma (Figure 2 and Table 1).

[28] The geometry of this extended margin and adjacent areas is shown schematically in Figure 6. At the northern end of the diagram, thin lithosphere of the accreted terranes butts directly against thick lithosphere of the North American craton. This is documented in part by the close spacing of the ^{87}Sr/^{86}Sr 0.704–0.706 lines in Mesozoic plutons, which parallel one another along this part of the craton margin (Figure 2). Farther south, the CSL (^{87}Sr/^{86}Sr 0.704 basalt) and ^{87}Sr/^{86}Sr = 0.706 lines diverge from one another as the cratonic lithosphere thins. The extended lithosphere here is not as thin as the lithosphere that underlies the accreted terranes, but it is much thinner than the normal cratonic lithosphere. The southern half of the plume head was extruded beneath the thinned lithosphere (with the plume tail still attached) until it was finally compressed against the thick cratonic lithosphere (which was moving continuously to the SW). Eventually the plume head under the transitional lithosphere was detached from the plume tail, which at this time plunged 60°–70° to the WNW.

[29] This sequence of events is depicted schematically in Figure 7, which shows the plume head rising under central Oregon (Figure 7a), impacting the continental lithosphere and being flattened in a N–S direction (Figure 7b), and extruding to both the north (beneath Phanerozoic accreted terranes) and the south (beneath the thinned cratonic lithosphere: Figure 7c). The sequence ends with the plume head under eastern Oregon, northern Nevada, and the Owyhee Plateau of SW Idaho, and with the plume tail plunging some 65° WNW (Figure 7d). Figure 8 depicts this same sequence in cross section, looking east toward the cratonic lithosphere.

[30] As the plume tail encountered the thick cratonic lithosphere (some 200 km southeast of its initial focal point) it began to erode a channel into the lithosphere similar to that observed beneath the eastern Snake River Plain today. The plume tail continued to be fed by new material flowing in from below; this material was chemically and thermally buoyant so it continued to rise to the shallowest depth possible, that is, below the extended terrane and the new channel eroded into the cratonic roots. Once this feed-back loop was established, the plume tail would no longer rise vertically along a track defined by its initial point of impact and North American plate motion: the plume followed the path of least resistance to the shallowest levels possible

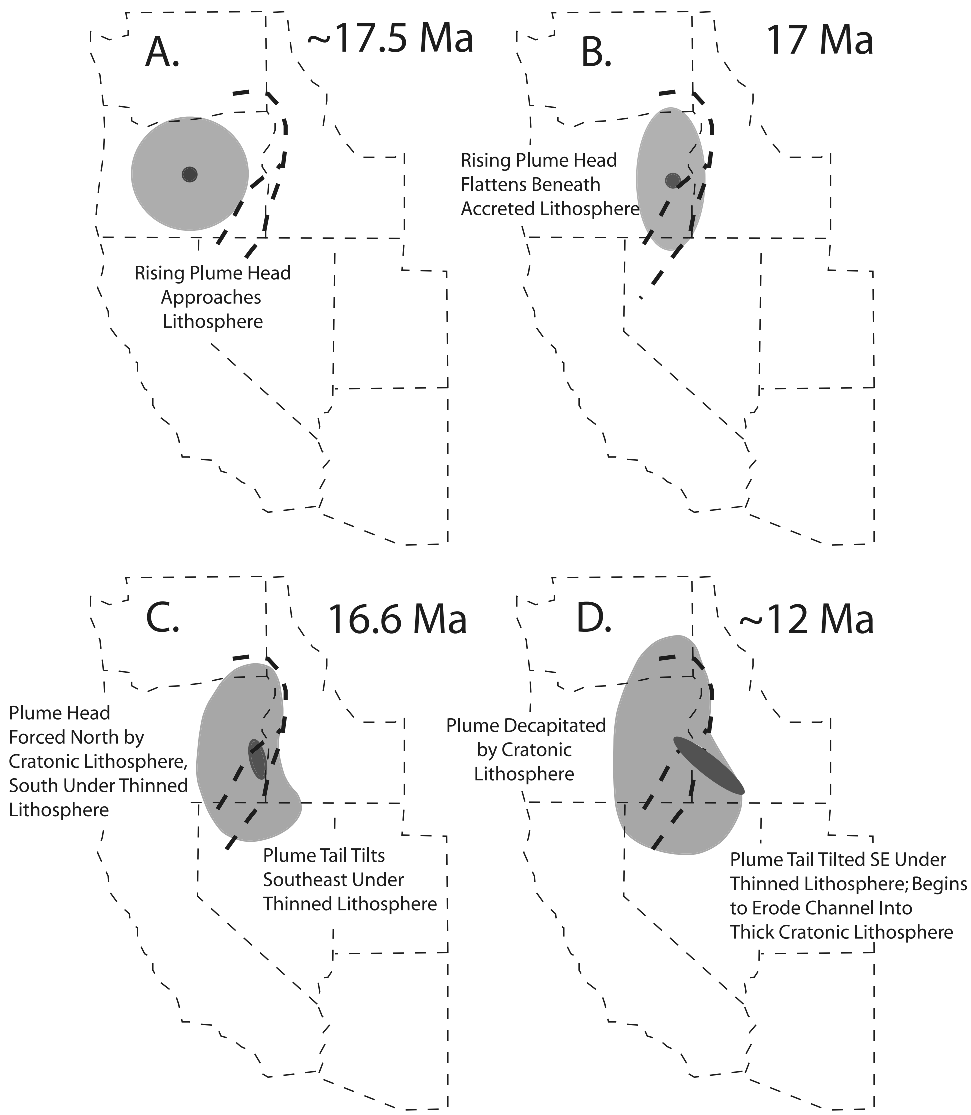

Figure 7. Sequential model showing the rise and deformation of the plume head (light gray) and plume tail (dark gray) as it interacts with lithosphere of western North America. (a) Plume head and tail are concentric as they rise through the asthenosphere beneath east central Oregon. (b) Plume head impacts against base of lithosphere to form extensive decompression melts; at the same time, it is squeezed by the thick lithospheric roots of the craton. (c) Plume head continues to deform against the western cratonic margin; plume head material escapes to north (forming Chief Joseph dikes, CRBG) and to south, beneath thinned lithosphere of the cratonic margin; plume tail follows the head south; main phase CRBG erupts to north, Owyhee–northern Nevada rhyolites erupt to south. (d) Plume head continues to expand both north and south, as well as west; plume head and plume tail are forced under thinned cratonic lithosphere. Tail becomes established in thermally eroded sublithospheric channel sink, guided by preexisting zone of thin lithosphere.

beneath the cratonic lithosphere, maintaining its plunge of 60–70° NW.

[31] An additional factor is the formation of the sublithospheric channel at this location, which acts as a sink for the buoyant plume material, may have been preexisting crustal structure. *Mankinen et al.* [2003, 2004], using maps of long-wavelength isostatic residual gravity and aeromagnetic data transformed into magnetic potential (“pseudogravity”), have shown that the deeper crust north and south of the Snake River Plain are distinct. They propose that the discontinuity in potential field character marked by the present Snake River Plain represents a preexisting crustal boundary exploited by the plume tail [*Mankinen et al.*, 2003]. This implies that the boundary was underlain by thinner lithosphere which acted as a sink for the plume tail, and focused thermal and mechanical erosion of the lithosphere.

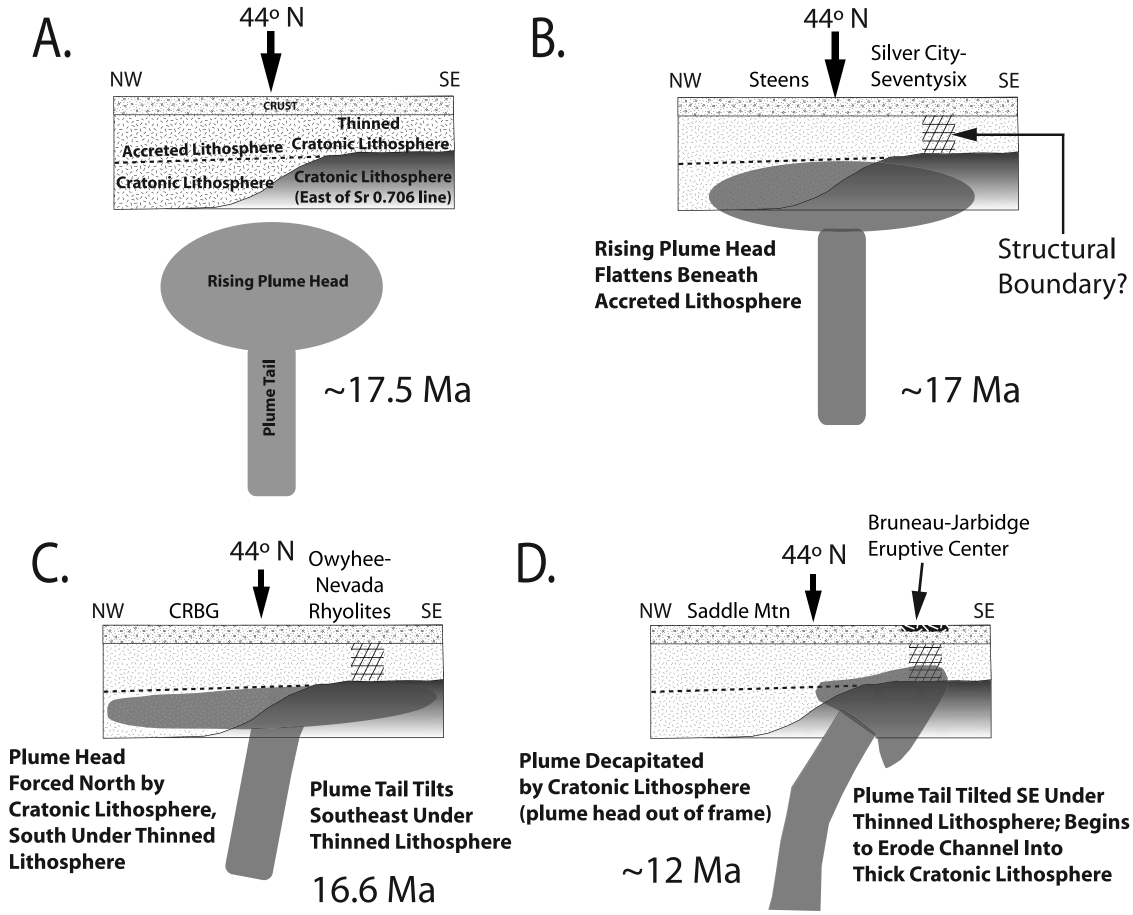

Figure 8. Sequential cross sections of the model showing the rise and deformation of the plume head (light gray) and plume tail (dark gray) as it interacts with lithosphere of western North America; view is looking east toward the cratonic lithosphere. The cross sections are "compressed" views showing lithosphere thickness under accreted terranes (dotted line), cratonal lithosphere (pattern), and thinned cratonal lithosphere (gradient). (a) Plume head and tail are concentric as they rise through the asthenosphere beneath east central Oregon prior to impact with lithosphere. (b) Plume head impacts and flattens against base of lithosphere to form extensive decompression melts; at the same time, the plume head is compressed to north and south as it flattens against accreted lithosphere and beneath thinned lithosphere. (c) Plume head continues to deform against the western cratonic margin; plume tail follows the head south. (d) Plume tail becomes established in thermally eroded sublithospheric channel sink, guided by preexisting zone of thin lithosphere.

Christiansen et al. [2002] proposed a similar boundary on the basis of an inferred magnetic discontinuity that extends beyond the trend of the Snake River Plain, but this inferred structure is not observed in the surface geology. In contrast, the Great Falls Tectonic Zone, a Proterozoic structure in SW Montana, lies well north of the proposed magnetic discontinuity [*Foster et al.*, 2006].

[32] We believe that this sublithospheric channel was fully established by ~12 Ma, when the Bruneau-Jarbidge eruptive center formed as the first large rhyolite caldera complex which lies directly on the plume track defined by the eastern Snake River plain and North American plate motion. This corresponds with initiation of fissure-fed rhyolite volcanism in the western Snake (circa 11.6–11.2 Ma at its NW end, 11.0–10.0 in the center, and 9.8–9.2 Ma at its SE end) and formation of the western SRP graben [*Wood and Clemens*, 2002; *Shervais et al.*, 2002]. These high-temperature rhyolites (900–1000°C) formed by relatively dry melting of mid to lower crustal granitoids in response to the intrusion of large volumes of mafic magma into the crust [*Boroughs et al.*, 2005]. Much of this mafic magma now forms a mafic sill complex beneath the western SRP, as documented by the large, long-wavelength gravity anomaly [*Mankinen et al.*, 2004; *Saltus and Jachens*, 1995] and seismic reflection profiles [*Hill and Pakiser*, 1967; *Prodehl*, 1979]. In effect, the western SRP graben formed by fracture of the crust between two focal points: the area of initial plume impact near 44°N latitude in eastern Oregon and the Bruneau-Jarbidge eruptive center (Figure 2). This fracturing appears

to have initiated in the NW and propagated to the SE over a period of about 2 Ma [*Shervais et al.*, 2002, 2005].

5. Discussion

5.1. Implications for Plume Tail Volcanism

[33] If we place the top of the plume tail beneath the Bruneau-Jarbidge eruptive center circa 12 Ma and assume a plunge of ~65° to the WNW, the base of the observed plume tail at 500 km depth would lie ~230 km to the WNW, that is, at the NW terminus of the western SRP and near the initial point of impact of the plume head circa 17 Ma (Figure 2). Thus, the line of fracture defined by the western SRP graben would have lain above the sloping plume tail as it connected back to the region of initial plume head impact. Extension of the upper crust to form the western SRP graben would be facilitated by thermal softening and erosion of the underlying lithosphere, allowing the rift to propagate from NW to SE over some 2 Ma [*Shervais et al.*, 2002].

[34] We further propose that the plume tail has maintained its current orientation for at least the last 10 to 12 Ma, and that the plume track defined by the eastern SRP has always lain some 200+ km southeast of the plume tail at a depth of 500–600 km. This asymmetry is supported in part by the distribution of deformation zones in the "tectonic parabola" which has its focus under the Yellowstone plateau [*Anders et al.*, 1989; *Pierce and Morgan*, 1992]. Deformation zone IV of *Pierce and Morgan* [1992], defined by large inactive Tertiary faults, is absent north of the plain but wide and well-defined south of the plain. Other deformation zones are broader north of the plain and roughly parallel the plume track, whereas the equivalent zones south of the plain are narrow and oriented at a high angle to the plume track (approximately N15°E versus N56°E for plume track).

[35] Why has the plume tail remained in this orientation for some 10–12 Ma? We suggest that once a topographic "sink" has been created within the subcontinental lithosphere, the buoyant plume material will continue to rise into this sink as the lithosphere moves above it, gradually eroding a channel into the lithosphere. This phenomenon has been thoroughly documented in the East African Rift zones, where plume-related volcanism is concentrated in rift zones that follow the thinner lithosphere of Proterozoic mobile belts between the Archean cratons [*Sleep*, 1996, 2002, 2005; *Sleep et al.*, 2002]. Formation of this sink beneath the Snake River Plain may be enhanced by lithospheric delamination, as suggested by the seismic tomographs of Yellowstone (Figure 4a) [*Yuan and Dueker*, 2005].

5.2. Implications for Plume Head Volcanism

[36] The model presented here suggests that most of the volcanic activity prior to 10–12 Ma across a broad swath of northern Nevada, SE Oregon, and SW Idaho was related to decompression melting of the plume head as it was overridden by thinned continental lithosphere, not the passage of North America over the plume tail. Simultaneously, extrusion of the plume head northward and westward fed main phase eruptions of the Columbia River basalt group [*Camp and Ross*, 2004] and plume-related flood basalt volcanism as far away as California [*Wagner et al.*, 2000]. The transition to plume tail dominated volcanism seems to have occurred circa 10–12 Ma, after the plume tail had already been deflected some 200 km to the SE and eroded a deep channel in the subcontinental lithosphere beneath the Bruneau-Jarbidge eruptive center. This eruptive center lies slightly north of the trend defined by Yellowstone and North American plate motion, suggesting that even at this time deflection of the plume tail southward was not yet complete. The timing of this transition also corresponds approximately to a change in rhyolite eruptions, from more voluminous, Fe-rich, high-T rhyolites prior to 8–10 Ma, to less voluminous, higher-silica, lower-T rhyolites after 8 Ma [*Hughes and McCurry*, 2002; *Nash et al.*, 2006].

[37] The approximate E–W alignment of rhyolite eruptive centers in northern Nevada and SE Oregon with near simultaneous eruptive ages ~16.5 Ma bears comment. While we have shown that there are other eruptive centers of similar age located to the north and west, the concentration of these centers along this trend suggests that another process is involved in their formation. *Burov et al.* [2007] have modeled the interaction of non-Newtonian plumes on multilayered continental lithosphere and show that the spreading plume head is likely to delaminate the mantle lithosphere beneath the continental crust, juxtaposing hot melt-rich plume against the base of the crust [*Burov and Guillou-Frottier*, 2005; *Burov et al.*, 2007, Figure 1]. The juxtaposition of a plume head directly against the base of the crust (or thin remnants of lithosphere) will result in extensive anatexis and the near simultaneous formation of rhyolite melts. Since the plume head is constrained to expand largely to the south by unextended cratonic lithosphere, delamination is likely to occur along the leading edge of the expanding plume head, forming the observed E–W trending array of eruptive centers. This model is supported by the observation of a slab of high-velocity mantle in the seismic tomography of the active plume tail at Yellowstone, interpreted by *Yuan and Dueker* [2005] to represent delaminated cratonic lithosphere (Figure 4).

5.3. Implications for Lithosphere Rheology

[38] The model presented here assumes that the upward path of a thermochemical plume in the shallow mantle is controlled in large part by preexisting lithospheric topography [e.g., *Sleep*, 1996, 2002; *Sleep et al.*, 2002]. That is, thinner than normal cratonic lithosphere acts as a sink for buoyant plume material, whether head or tail, which will flow laterally in order to achieve the greatest rise. In this regard, the effective thickness of cratonic lithosphere appears to exceed the thickness of the conductive thermal boundary layer (approximately seismically defined lithosphere) to include the rheological boundary layer of *Sleep* [2005]. The rheological boundary layer is transitional between the conductive thermal boundary layer and the fully convective asthenosphere. It projects the influence of the

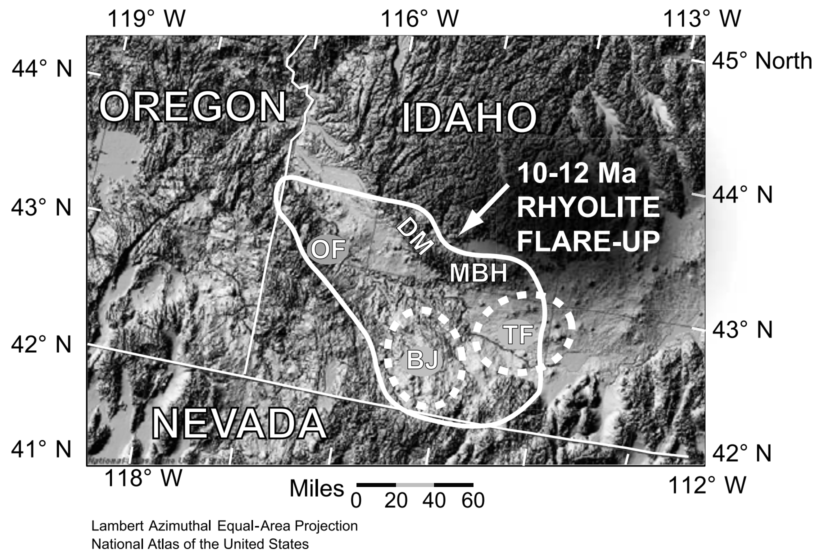

Figure 9. Map of southwestern Idaho showing the location of rhyolites associated with the ~10–12 Ma rhyolite flare-up of *Bonnichsen et al.* [2008]. Heavy white line outlines the main exposures of 10–12 Ma rhyolites; OF, Owyhee Front; DM, Danskin Mountains; MBH, Mount Bennett Hills; BJ, Bruneau-Jarbidge eruptive center; TF, Twin Falls eruptive center.

lithosphere deeper into the mantle and appears to affect the flow of low-velocity material in the underlying asthenosphere. This is implicit in seismic tomographs of the Snake River–Yellowstone plume, which exhibits a distinct shift to the SE at ~300 km depth (Figure 4a). *Schutt and Humphreys* [2004] note that the low-velocity channel beneath the Snake River Plain also extends to ~300 km depth (Figures 4b and 4c), consistent observations farther east beneath Yellowstone [*Yuan and Dueker*, 2005; *Waite et al.*, 2006; *Schutt et al.*, 2008; *Stachnik et al.*, 2008].

5.4. The 10–12 Ma Rhyolite Flare-Up

[39] *Bonnichsen et al.* [2008] have documented an extensive rhyolite flare-up between about 11.7 Ma and 10.2 Ma in south central Idaho. This includes rhyolite ignimbrites and lava flows of the Bruneau-Jarbidge and Twin Falls eruptive centers, as well as rhyolite vitrophyres of the western Snake River plain (Owyhee Front, Danskin Mountains, Mount Bennett Hills). *Bonnichsen et al.* [2008] attribute this flare-up to the intrusion of massive volumes of basaltic magma into the middle and lower crust; however, they conclude that the widespread distribution of this event precludes any relationship to plume-derived magmatism (Figure 9).

[40] We propose that this rhyolite flare-up can be understood best in terms of our model for (1) extensive plume head volcanism circa 16.6 to 16.0 Ma that effected much of southwestern Idaho and (2) a tilted plume tail that underlay western Snake River Plain and the Bruneau-Jarbidge eruptive center circa 12–10 Ma (Figure 7). The distribution of ~16.6 Ma Steens-type basalts in northern Nevada (Santa Rosa–Calico complex, Jarbidge Mountains) and SW Idaho (Silver City Range) show that most of SW Idaho was underlain by the plume head at this time. The occurrence of extensive ~16.6 to 14.0 Ma rhyolites overlying these basalts in all three areas documents prolonged heating and partial melting of the crust during this time interval.

[41] If we accept that the location and orientation of the western SRP graben reflects tilting of the plume tail from its initial point of impact under eastern Oregon to its 10–12 Ma focus beneath the Bruneau-Jarbidge eruptive center (see section 5.1), then the distribution and intensity of 10–12 Ma rhyolites can be understood as the confluence of (1) extensive preheating of the crust by plume head basalts and (2) the concentrated input of heat into the crust by basaltic magmas derived from the tilted plume tail. This explains both the fissure-fed rhyolite vitrophyres adjacent to the western Snake River Plain graben (Owyhee Plateau, Danskin Mountains, Mount Bennett Hills), formed over the tilted plume tail, as well as the central vent eruptives of the Bruneau-Jarbidge and early Twin Falls eruptives, formed over the top of the tilted plume tail.

5.5. Newberry Trend, Oregon High Lava Plains

[42] One of the most commonly cited arguments against a plume origin for the Yellowstone hot spot is the so-called Newberry trend of the Oregon High Lava Plains (Figure 1), in which silicic volcanic centers in eastern Oregon become younger to the NW along a trend antithetical to the eastern SRP trend and North American Plate motion [*Jordan et al.*, 2004; *Jordan*, 2005]. *Humphreys et al.* [2000] and *Christiansen et al.* [2002] have argued that the existence of this trend and the lack of a coherent plume track prior to 12 Ma are not consistent with a plume origin for the Yellowstone–Snake River Plain system. They prefer models in which convective rolls or edge effects explain the contrasting volcanic trends, one younging to the NE (Yellowstone) and the other younging to the NW (Newberry). In contrast,

Jordan et al. [2004] have shown that magmatism associated with the Newberry trend is much less voluminous than the Yellowstone–Snake River Plain system, and that its trend is controlled in part by the WNW trending Brothers Fault Zone. The decrease in age of volcanism to the NW has been attributed to mantle corner flow above the Farallon subducting slab [*Draper*, 1991; *Humphreys et al.*, 2000] or to upflow along the base of the lithosphere at the cratonic margin [*Jordan et al.*, 2004]. Neither explanation is at odds with a plume origin for the Yellowstone–Snake River Plain system, and *Jordan et al.* [2004] have specifically embraced this model for the Columbia River Basalt Group and the adjacent Newberry trend.

5.6. Persistent Volcanic Activity

[43] Another issue that is commonly cited as an argument against the plume model is the persistence of volcanic activity along the plume track long after passage of the lithosphere away from the hot spot. This includes Pleistocene basaltic volcanism in the eastern Snake River Plain, which postdates the rhyolite eruptive complexes [e.g., *Hughes et al.*, 2002; *Kuntz et al.*, 1992; *Shervais et al.*, 2005, 2006], and Pleistocene volcanism that postdates lacustrine sediments in the western Snake River Plain [*Vetter and Shervais*, 1992; *White et al.*, 2002; *Shervais et al.*, 2002]. In the western plain this volcanism is as young as 200 ka [*Howard et al.*, 1982], while in the eastern plain, basalts as young as 2 ka occur [*Kuntz et al.*, 1992].

[44] The model presented here does not address this issue, but the observations summarized here suggest possible explanations. Geophysical modeling has shown that plume tail mantle remains under its point of impact with the lithosphere even as the lithosphere continues to move relative to the plume [e.g., *Ribe and Christensen*, 1994, 1999]. The attached plume mantle spreads laterally and thins, potentially inducing additional melting. This has been modeled for the western United States by *Lowry et al.* [2000]. Alternatively, plume tail material may continue to flow downstream (in the direction of plate motion) in response to an edge effect that channels plume material through the sublithospheric conduit, as suggested by *Hanan et al.* [2008]. This model assumes that the top of the sublithospheric conduit is shallower in the west (where the lithosphere is thinner), providing a sink for the buoyant, partially molten plume tail mantle. This model is similar to that proposed by *Jordan et al.* [2004] for the Newberry trend.

5.7. Nonplume Models

[45] A range of nonplume models have been proposed for the Yellowstone–Snake River Plain volcanic system that must be considered before any plume-centered model can be accepted. These models include back-arc basin extension behind the Cascades arc [*Hart and Carlson*, 1987], edge-driven convection [*King and Anderson*, 1995; *King*, 2007], and a convective roll or hotline driven by self-sustaining convection [*Humphreys et al.*, 2000]. All of these models assume that the source of basaltic magmatism is shallow asthenosphere or subduction-modified asthenosphere that underlies the lithosphere. Since shallow asthenosphere is the source of mid-ocean ridge basalts (MORB), which are typically depleted in incompatible trace elements, this source is inconsistent with the geochemistry of the observed basalts, which resemble ocean island basalts in their major and trace element geochemistry [e.g., *Vetter and Shervais*, 1992; *Hughes et al.*, 2002; *Shervais et al.*, 2006]. Further, none of the nonplume models predicts the sudden outpouring of flood basalt in <1 Ma or the time-transgressive progression of silicic eruptive complexes, which represent the influx of huge volumes of mafic magma into the crust, now represented by subcrustal and midcrustal sill complexes [e.g., *Peng and Humphreys*, 1998]. The back arc basin and cratonic edge effect models attempt to address the CRBG flood basalt province but do not explain the time-transgressive Snake River Plain. In contrast, the convective roll model ("hotline") specifically addresses the eastern Snake River Plain, but does not explain the CRBG flood basalts. As pointed out by *Pierce and Morgan* [1992], *Pierce et al.* [2002], *Camp and Ross* [2004], *Jordan et al.* [2004], and *Hooper et al.* [2007], only the plume model presents a coherent explanation for all aspects of this system, without special pleading.

6. Conclusions

[46] The model presented here combines data from a wide range of geological and geophysical investigations to solve long-standing problems in understanding the Neogene history of the Snake River–Yellowstone plume system. Most significant of these problems is the misfit between established North American plate motion and the purported track of the hot spot tail prior to 12 Ma (Table 2). We show that there was no well-established track prior to $\sim$12 Ma. Rather, there was a broad area of plume head volcanism that extended from eastern Oregon across the Owyhee Plateau and into northern Nevada, as well as north into Washington and western Idaho along the craton boundary. Volcanic activity north and west of the $^{87}Sr/^{86}Sr$ 0.704 basalt isopleth (CSL) was dominantly basaltic, reflecting the underlying substrate of young, Phanerozoic age lithosphere. Volcanic activity south and east of the $^{87}Sr/^{86}Sr$ 0.704 basalt isopleth was dominantly rhyolitic, reflecting the underlying substrate of extensionally thinned cratonic lithosphere.

[47] This model implies that the currently observed $\sim$65 NW plunge of the Yellowstone mantle low-velocity (thermal) anomaly (plume) was established by $\sim$12 Ma, and that the eastern SRP plume track formed over this steeply plunging plume conduit since that time. This inferred forcing of plume geometry in response to lithospheric topography underscores the potential importance of buoyancy in allowing plumes to rise to the shallowest levels attainable, and thus to extract the maximum proportion of decompression melt possible [e.g., *Thompson and Gibson*, 1991; *Sleep*, 1996, 2002, 2005; *Sleep et al.*, 2002].

[48] Finally, semiconfined expansion of the plume head south from its initial region of impact and beneath thinned cratonic lithosphere may have delaminated part of that lithosphere at the leading edge of the plume head. This

delamination may have resulted in the formation of the E–W trending linear array of rhyolite eruptive centers that formed essentially simultaneously. It is this array of caldera complexes that created the impression of a plume track prior to 12 Ma, aided by the spread in ages of early formed and later caldera complexes along the same trend.

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