A plume-triggered delamination origin for the Columbia River Basalt Group

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ABSTRACT

The Columbia River Basalt Group reveals a complete and detailed stratigraphic succession to assess the interplay of lithospheric and asthenospheric processes. This record of chemical change through time is used to evaluate genetic models for Columbia River Basalt volcanism. We recognize four primary constraints on source melting: (1) a plume component appears to be the dominant source of Imnaha Basalt; (2) Grande Ronde Basalt is best interpreted as being derived from a mafic pyroxenite or eclogite source; (3) the sequence of source melting must correspond with the stratigraphic record; and (4) working models must explain a step-function chemical change at the Imnaha–Grande Ronde stratigraphic boundary. We can envision only three potential models to satisfy these primary constraints: (1) melting of a mantle plume entrained with eclogite, (2) plume interaction with the Juan de Fuca plate, and (3) delamination triggered by plume emplacement. The first two of these are inconsistent with the time-stratigraphic sequence of melting and cannot satisfy all four primary constraints. In contrast, a model of plume-triggered delamination accurately predicts a progressive sequence of melting that satisfies each of the primary constraints. Such a model is consistent with recent numerical experiments demonstrating that delamination is the expected result of plume emplacement beneath thin Mesozoic lithosphere lying adjacent to a thick cratonic boundary. We test this model by comparing the observed history of uplift and tectonism in eastern Oregon and adjacent Washington to that predicted by the numerical models to reveal consistent stress regimes and strikingly similar topographic and structural profiles.

Keywords: mantle plume, Columbia River Basalt, delamination, Yellowstone, flood basalts.

INTRODUCTION

The geologic record is punctuated by periodic outbursts of voluminous basalt lava, generating Large Igneous Provinces (LIPs) in both oceanic and continental intraplate settings. Such eruptions are characterized by extraordinarily high effusion rates that are difficult to explain by normal plate-boundary processes. The traditional interpretation has been to attribute their intraplate origin to the adiabatic rise and partial melting of plumes from the deep mantle (e.g., Morgan, 1971, 1972; Campbell, 2006). There has been much recent debate, however, on the efficacy of this model, leading some workers to promote alternative, nonplume models for the origin of all oceanic plateaus, flood-basalt provinces, and hotspot tracks (e.g., King and Anderson, 1998; Sears et al., 2004; Lustrino, 2005; Anderson, 2007; Foulger, 2007). The credibility of both plume and nonplume models depends on a detailed assessment of the chemical stratigraphy of these massive volcanic terrains. Most LIPs, however, are poorly studied, with incomplete or poorly exposed stratigraphic successions due to extensive erosion or burial. In contrast, the Columbia River Basalt Group comprises the youngest, best preserved, and most intensely studied flood-basalt province on Earth, with a complete and well-sampled stratigraphy exposed in the canyonlands of northeastern Oregon and southeastern Washington States.

A fundamental goal confronting workers on the Columbia River Basalt Group is the establishment of a realistic petrogenetic model consistent with time-dependent chemical variations in the flood-basalt stratigraphy. The entire volcanic succession cannot be explained by a simple model of decompressional partial melting of a mantle-plume source (e.g., Hooper et al., 2007). Instead, trace-element and isotopic data demonstrate a more complicated scenario requiring a variety of source components, with variations in the degree of mixing between each (Carlson and Hart, 1987; Brandon and Goles, 1988; Brandon et al., 1993; Hooper and Hawkesworth, 1993; Wolff et al., 2008). Workers have focused considerable effort in defining these source components, but less effort in establishing a time line of petrogenetic processes consistent with the stratigraphic record of chemical change through time. Our intent here is to review the existing data on source components in an attempt to establish geochemical and temporal constraints on source melting, which we then use to evaluate models of flood-basalt genesis.

Melt generation has been attributed to one of three potential mechanisms: (1) backarc extension (Hart and Carlson, 1987; Smith, 1992), (2) mantle-plume emplacement (e.g., Brandon and Goles, 1988; Hooper and Hawkesworth, 1993; Camp and Ross, 2004; Hooper et al., 2007), and most recently, (3) lithospheric delamination (Hales et al., 2005; Humphreys, 2007). We demonstrate here that models relying solely on any one of these processes have difficulties in satisfying the temporal constraints on source melting; but instead, the history of melting appears to be more consistent with an evolving sequence of delamination events triggered by the Miocene arrival of the Yellowstone mantle-plume head in an environment of backarc extension.

THE COLUMBIA RIVER BASALTS

The main eruptive phase of the Columbia River Basalt Group, from ca. 16.6 to 15.0 Ma, generated a lava volume of ~220,500 km^3, or 94% of the total volume (Camp et al., 2003). Volcanism began with the eruption of Steens Basalt over a broad region of southeastern Oregon. It then migrated rapidly into northeastern Oregon where Imnaha and Grande Ronde Basalts erupted from the Chief Joseph dike swarm, and where the partly contemporaneous Picture Gorge Basalts erupted from the Monument dike swarm (Camp and Ross, 2004) (Fig. 1). After ca. 15 Ma, these main-phase eruptions were followed by more sporadic eruptions of Wanapum

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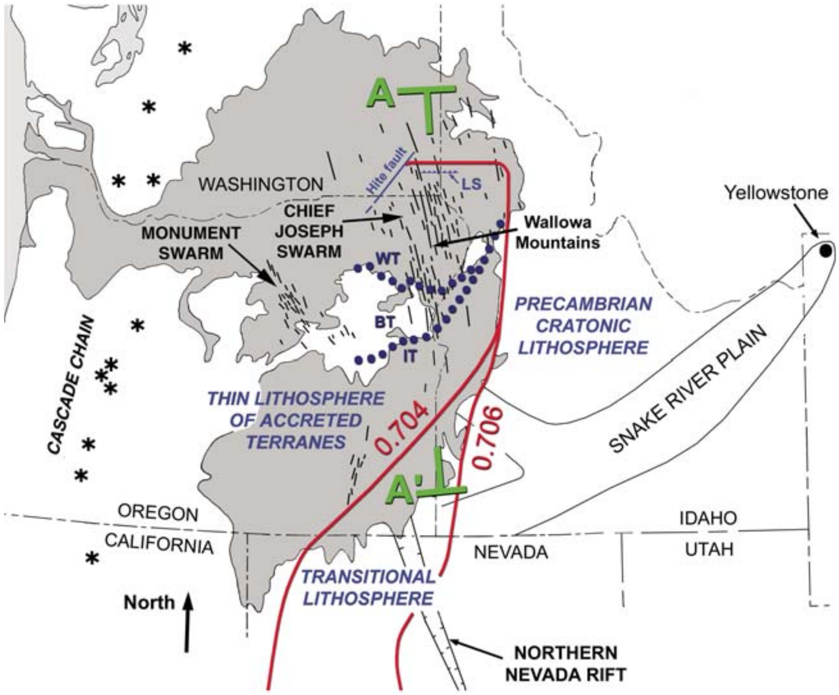

Figure 1. The Columbia River Flood Basalt Province. Red lines correspond with initial $^{87}Sr/^{86}Sr$ isopleths marking the boundaries between accreted oceanic terranes west of the 0.704 line, and the Precambrian craton east of the 0.706 line (Fleck and Criss, 2004; Pierce et al., 2002). Transitional lithosphere exists between the 0.704 and 0.706 lines. Dotted lines separate the Izee (IT), Baker (BT), and Wallowa (WT) terranes. LS is the "Lewiston structure." A–A' corresponds with the cross-section diagrams in Figure 7 and the topographic profile in Figure 8B.

and Saddle Mountains Basalts from the northernmost part of the Chief Joseph dike swarm in southeastern Washington (Fig. 1). Cessation of the main-phase eruptions at ca. 15 Ma was contemporaneous with the initiation of rhyolite eruptions across the cratonic boundary of North America and into the Idaho segment of the Snake River Plain, the eastern terminus of which is marked by a mantle plume extending to a depth of at least 500 km, near the top of the mantle transition zone, beneath the Yellowstone caldera (Yuan and Dueker, 2005; Waite et al., 2006).

The Columbia River Basalt Group appears to have been derived from the melting of a variety of source components identified isotopically using Pb, Nd, Sr, He, Re-Os, and O isotopes (Carlson et al., 1981; Carlson, 1984; Church, 1985; Carlson and Hart, 1987; Brandon and Goles, 1988; Brandon et al., 1993; Hooper and Hawkesworth, 1993; Chamberlain and Lambert, 1994; Dodson et al., 1997; Chesley and Ruiz, 1998; Wolff et al., 2008). Here, we present new isotopic data on the Steens and Imnaha Basalts in Table 1. We combine these data with existing data to delimit the presumed source components for the main-phase lavas in Figure 2.

ROLE OF THE YELLOWSTONE MANTLE PLUME

All workers agree that the Columbia River Basalt Group erupted in an environment of back-arc extension, but they disagree on whether the burst of the main-phase eruptions, from ca. 16.6 to 15 Ma, was triggered by the Miocene arrival of the Yellowstone mantle plume near the western edge of the Snake River Plain hotspot track. During this very brief interval, volcanism from the Chief Joseph dike swarm was sustained with an eruption rate of at least 500–600 km^3/km/Ma (Hooper et al., 2007), and possibly as high as 1200 km^3/km/Ma based on recent $^{40}Ar/^{39}Ar$ and paleomagnetic data supporting a much shorter eruption interval of 0.75 m.y. (Jarboe et al., 2006). The Chief Joseph dike swarm is located in a region of minimal extension, so it is unlikely that the main-phase eruptions were tectonically controlled. The high magma supply rate, over an exceedingly short interval of geologic time, in an environment of little extension, appears to require the abrupt arrival of a thermal and/or fertile anomaly at mantle depths. Plume proponents have long noted that these eruption constraints are consistent with a plume origin but are more difficult to satisfy by any of the alternative nonplume models yet proposed (e.g., Hooper et al., 2007).

The primary rationale for dismissing a plume genesis for both the Columbia River Basalt Group and Snake River Plain was the failure of seismic studies to identify a mantle anomaly beneath the current Yellowstone caldera to a depth >200 km (Christiansen et al., 2002), until such an anomaly was resolved by Yuan and Dueker (2005) extending beneath the caldera to a depth of at least 500 km. A southwestward trend of progressively older rhyolite calderas along the Snake River Plain, mirroring the westward drift of the overriding North American plate, demonstrates that this anomaly has persisted for at least 16.5 m.y. (Pierce and Morgan, 1992). We believe that the Miocene location of this anomaly in southeastern Oregon, at the very site of the flood-basalt initiation at ca. 16.5 Ma, is too compelling to be readily dismissed as coincidence.

Pierce and Morgan (1992) and Camp and Ross (2004) describe the Columbia River Basalt Group and Snake River Plain as integral parts of a single magmatic system, linking flood-basalt volcanism of the Columbia River Basalt Group to the Yellowstone plume head, and hotspot volcanism of the Snake River Plain to the connecting plume tail. Such an interpretation is supported by a variety of interdisciplinary studies, which include geophysical modeling (Parsons et al., 1994; Saltus and Thompson, 1995; Bijwaard et al., 1998; Glen and Ponce, 2002; Montelli et al., 2004; Yuan and Dueker, 2005; Waite et al., 2006), together with geochemical investigations (Brandon and Goles, 1988, 1995; Draper, 1991; Brandon et al., 1993; Hooper and Hawkesworth, 1993; Dodson et al., 1997; Hooper et al., 2007; Wolff et al., 2008; Hanan et al., 2008), and both field and stratigraphic relationships (Pierce and Morgan, 1992; Camp, 1995; Pierce et al., 2002; Glen and Ponce, 2002; Hooper et al., 2002; Camp et al., 2003; Jordan et al., 2004; Camp and Ross, 2004; Hooper et al., 2007).

The geochemical data have been interpreted both for and against a plume signature in the Columbia River Basalt Group lavas. Based largely on $^{208}Pb/^{207}Pb/^{206}Pb/^{204}Pb$, $^{143}Nd/^{144}Nd$, and $^{87}Sr/^{86}Sr$ data, Carlson (1984) defined his C2 mantle-source component as relatively depleted oceanic mantle contaminated by ~2% subducted sediment. Smith (1992) suggested instead that the C2 component was derived from crustal contamination by the accreted terranes beneath the Chief Joseph dike swarm. In contrast, C2 was reinterpreted by Brandon and Goles (1988, 1995), Brandon et al. (1993), Hooper and Hawkesworth (1993), and Wolff et al. (2008) as a plume component (*PL,* in Fig. 2).

TABLE 1. ISOTOPE ANALYSES OF STEENS AND IMNAHA BASALTS*

Sample	Latitude/longitude	$^{87}Sr/^{86}Sr$	$^{143}Nd/^{144}Nd$	$^{206}Pb/^{204}Pb$	$^{207}Pb/^{204}Pb$	$^{208}Pb/^{204}Pb$
Upper Steens						
JS-72†,§	42.62°N/118.58°W	0.703981	0.512841	18.908	15.601	38.512
JS-48†	42.62°N/118.58°W	0.703736	0.512919	18.879	15.609	38.543
JS-36†	42.60°N/118.66°W	0.703691	0.512927	18.934	15.614	38.581
Lower Steens						
MF 94-63†	42.60°N/118.56°W	0.703529	0.513167	18.795	15.566	38.369
MF 94-65†	42.60°N/118.56°W	0.703842	0.512871	18.962	15.621	38.682
JS-08†	42.59°N/118.55°W	0.703669	0.513055	18.894	15.588	38.546
JS-18†	42.60°N/118.53°W	0.703629	0.512958	18.897	15.596	38.560
JS-22†	42.60°N/118.53°W	0.703331	0.512975	18.788	15.595	38.440
JS-25†	42.60°N/118.53°W	0.703406	0.512992	18.817	15.590	38.450
JS-27†	42.60°N/118.53°W	0.703507	0.512971	18.859	15.621	38.575
JS-28†	42.60°N/118.53°W	0.703549	0.512968	18.830	15.598	38.475
JS-30†	42.60°N/118.56°W	0.703548	0.513022	18.813	15.587	38.448
BB-59#	43.69°N/118.56°W	0.703548	0.512925	18.892	15.578	38.476
BB-77#	43.69°N/118.56°W	0.703734	0.512902	18.943	15.593	38.536
Imnaha						
MF 94-34**	43.75°N/118.05°W	0.704094	0.512870	19.028	15.626	38.678
HOR-05**	43.83°N/117.73°W	0.703854	0.512883	18.945	15.607	38.552

*All Pb isotope ratios are normalized on the basis of replicate measurements of National Institute of Standards and Technology (NIST) SRM981, using the values of Todt et al. (1984). Pb was loaded on a loop-type Re filament with silica gel and phosphoric acid. Pb isotopic compositions were determined at San Diego State University (SDSU) using a Micromass SECTOR 54 thermal ionization mass spectrometer (TIMS) (multi-collector static mode). The mass discrimination factors average 1.26‰/amu ± 0.03 2se (standard errors). Total uncertainties in Pb isotope ratios are <0.05% per amu, computed by error propagation of both sample and standard analyses. Pb blanks were <100 pg, and are negligible. Sr isotope analyses were performed on the SDSU Micromass SECTOR 54 TIMS. Sr was loaded on a Re filament with a slurry of Ta_2O_5 in phosphoric acid, and data were collected in multi-dynamic mode. Repeated measurements of $^{87}Sr/^{86}Sr$ for NIST SRM987 at SDSU gave 0.710255 ± 0.000003 2se, n = 2; sample measurements have been normalized to $^{88}Sr/^{86}Sr$ = 0.1194 and referenced to a value of 0.71025 for the standard. Nd isotopes were determined by multi-collector–inductively coupled plasma–mass spectrometer (MC-ICP-MS) on the SDSU Nu Plasma HR. MC-ICP-MS Nd isotope ratios were corrected for instrumental mass fractionation and machine bias by applying a discrimination factor determined by bracketing sample analyses with analyses of the SDSU AMES Nd standard using a value of 0.512130 for the AMES standard. The measured value of La Jolla Nd is 0.511844 ± 4 by MC-ICP-MS on the Nu Plasma HR over the course of this study. Sample data were normalized to $^{146}Nd/^{144}Nd$ = 0.7219 and are reported relative to a value of $^{143}Nd/^{144}Nd$ = 0.511858 for the La Jolla Nd standard. Procedural blanks for Sr and Nd are <50 pg and are negligible for the analyses reported here. The ±2se analytical precision errors in last decimal place, where n is the number of measured isotopic ratios, averaged ±7 for $^{87}Sr/^{86}Sr$ and ±4 for $^{143}Nd/^{144}Nd$ for the MC-ICP-MS data. Chemical procedures are given in Hanan et al. (2004).

†Samples analyzed from the type Steens Mountain section of Johnson et al. (1998).

§Trachyandesite sample from the uppermost part of the Steens Mountain section of Johnson et al. (1998).

#Sample originally mapped as lower Pole Creek basalt of Hooper et al. (2002).

**Sample originally mapped as upper Pole Creek basalt of Hooper et al. (2002).

It is significant that Imnaha Basalt appears to be the only stratigraphic unit derived solely from this mantle source (Fig. 2). These lavas have the trace-element signatures of oceanic island basalt (OIB), similar in composition to Hawaii and Iceland (Hooper and Hawkesworth, 1993; Bryce and DePaolo, 2004), which is a necessary requirement for a plume origin. Perhaps more importantly, the Imnaha Basalt lavas also have high $^3He/^4He$ ratios of 11.4 ± 0.7 Ra, and both $^{20}Ne/^{22}Ne$ and $^{21}Ne/^{22}Ne$ ratios characteristic of a plume source derived from the deep mantle (Dodson et al., 1997). Figure 2 demonstrates that both Steens and Picture Gorge Basalts are more similar to Hawaiian plume basalts than is Imnaha Basalt (Fig. 3), suggesting that both could have been derived from a mantle plume source. Analysis of the Steens and Picture Gorge lavas for $^3He/^4He$ would help better answer the degree to which a mantle plume may have been involved in their origin.

We believe that the preponderance of data from a variety of geochemical, geophysical, and field studies supports the notion that the Yellowstone mantle plume may have played an important role in triggering the Columbia River flood basalt eruptions. However, the data also suggest that this role was complex, requiring the interaction of several source components, only one of which was the mantle-plume source.

THE ENIGMATIC GRANDE RONDE BASALTS

Variations in the silica content and rock composition of the main-phase stratigraphic units are shown in Figure 3. The Steens, Imnaha, and Picture Gorge Basalts are composed of olivine tholeiites, with subordinate alkali olivine basalts exposed in the upper part of the Steens stratigraphy. Such rocks are typical of flood-basalt provinces worldwide, many of which also contain smaller amounts of primitive picritic lavas containing abundant olivine (Fig. 4). In marked contrast, the Grande Ronde Basalts are composed of high-Fe/Mg, high-silica (52%–58%) basaltic andesites, which comprise ~65% of the flood-basalt volume (Camp et al., 2003), but over 90% of the volume that erupted from the Chief Joseph dike swarm (Tolan et al., 1989). These high-silica rocks are unusual, and perhaps unique, when compared to the most dominant rock types comprising the bulk composition of all other flood basalt provinces (Fig. 4).

Grande Ronde Basalts also differ markedly from the younger, small-volume Saddle Mountains Basalts which erupted sporadically during the waning stages of the Columbia River Basalt Group volcanism, from ca. 13 to 6 Ma. These late-stage lavas display wide variations in major-element, trace-element, and isotopic compositions, forming chemically distinct flows or groups of flows (Hooper, 2000). In contrast, the peak of the flood-basalt event generated Grande Ronde lavas that are chemically more coherent, with very limited isotopic variations and consistent trends in all major and minor elements (Wright et al., 1989; Hooper, 2000; Carlson, 1984).

The origin of these evolved lavas has been controversial, with some workers arguing that they are partial melts of peridotite modified by large degrees of crustal assimilation and fractional crystallization (Carlson et al., 1981; Carlson, 1984; Carlson and Hart, 1987, 1988), whereas others contend that they were generated by large-scale melting of a mafic source composed of eclogite or pyroxenite (Wright et al., 1976; Helz, 1978; Swanson and Wright, 1981; Reidel, 1983; Wright et al., 1989; Takahahshi et al., 1998; Hooper et al., 2007).

Genesis from the Modification of a Peridotite Partial Melt

The quantitative modeling of Carlson et al. (1981) and Carlson (1984) suggests that the most differentiated Grande Ronde lavas could be generated from a melt derived from a depleted peridotite source (the C1 component of Carlson, 1984), modified by a combination of ~20%–30% assimilation of granitic crust combined with ~30%–60% fractional crystallization of plagioclase, clinopyroxene, and olivine, so that the mass of the crystallizing phases removed from the liquid was at least twice the mass of the assimilated crust added to it. However, Carlson (1984) notes that a smaller amount of crustal assimilation (10%–20%) would be necessary if the mafic end-member for Grande Ronde Basalt derives from a more enriched upper mantle contaminated by subduction-derived sediment (his C2 mantle component).

There is disagreement on the amount of fractional crystallization displayed by the Grande Ronde Basalts. Hooper and Hawkesworth (1993) use Sr and Rb distribution coefficients to suggest that fractional crystallization in this

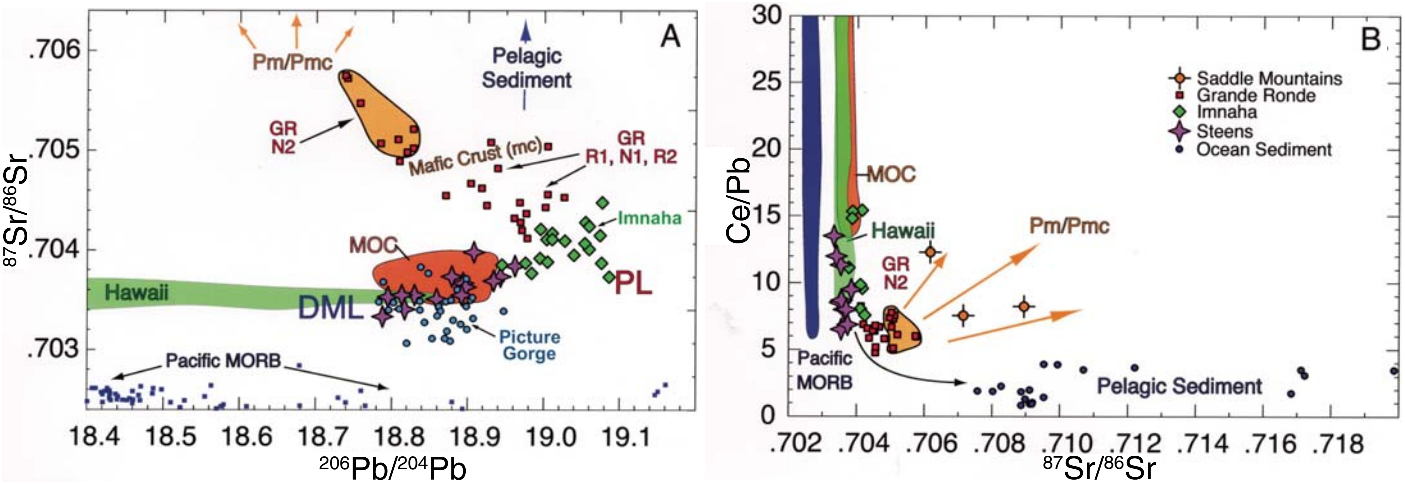

Figure 2. (A) $^{87}Sr/^{86}Sr$ versus $^{206}Pb/^{204}Pb$ diagram of analyses from the main eruptive phase of the Columbia River Basalt Group. Data are from a variety of sources (Carlson, 1984; Carlson et al., 1981; Hooper and Hawkesworth, 1993; Brandon et al., 1993), with new analyses presented here on Steens Basalt (Table 1). Depleted upper mantle *(DML,* the C1 component of Carlson, 1984), plume mantle *(PL,* the C2 component of Carlson, 1984), mafic crust of Paleozoic-to-Mesozoic age *(mc)*, mafic crust of Precambrian age *(Pmc)*, and Precambrian mantle lithosphere *(Pm)* are presumed source components for the Columbia River Basalt Group. Note that the Grande Ronde lavas deviate from the mantle array toward older lithospheric components *(mc, Pmc, Pm)* and/or oceanic sediment components. The Grande Ronde N2 lavas have high $^{87}Sr/^{86}Sr$ and distinctly lower $^{206}Pb/^{204}Pb$ relative to the stratigraphically lower R1, N1, and R2 lavas, consistent with the input of a cratonic component to the younger N2 lava source. Younger lavas from the Wanapum and Saddle Mountains Formations are not shown, but are generally more isotopically evolved (Saddle Mountains lavas all have $^{87}Sr/^{86}Sr$ >0.706), consistent with components from the enriched Archean lithosphere (Carlson, 1984; Hooper and Hawkesworth, 1993). Pelagic ocean sediment has $^{87}Sr/^{86}Sr$ >0.706 and plots off the figure at this scale. Pacific mid-ocean ridge basalt (MORB) represents the East Pacific Rise north of 23°S (Hanan and Graham, 1996; Pet DB). Also shown is the field for Pacific Mesozoic Oceanic crust (MOC; Shervais et al., 2005) and Hawaiian Island shield tholeiites (GeoRoc). Note that the Hawaii intra-plate plume basalts overlap with the Steens and Picture Gorge Basalts, but not with the Imnaha or Grande Ronde Basalts. (B) $^{87}Sr/^{86}Sr$ versus Ce/Pb with the same data set as (A), but with Ce/Pb data for Steens Basalt from J. Wolff (2008, personal commun.). Ce/Pb is a sensitive indicator of crustal components in the mantle. The Pacific MORB, Hawaii shield basalts, and Pacific Mesozoic Oceanic crust (MOC) all show a wide range in Ce/Pb that suggests pollution of their mantle source by pelagic sediment. Note that Hawaii and the accreted MOC basalts are very similar in $^{87}Sr/^{86}Sr$, Ce/Pb, and radiogenic Pb isotopes. This illustrates the difficulty in differentiating between recycled oceanic lithosphere that may be a plume source (e.g., Hawaii) and accreted oceanic lithosphere. The Steens and Picture Gorge Basalts are more similar to the Hawaii plume basalts and MOC than is Imnaha Basalt. We could therefore just as readily tie the Steens and Picture Gorge Basalts to the plume composition. Imnaha Basalt would then represent the plume polluted by the Mesozoic accreted terranes, with some Paleoproterozoic cratonic component. The Grande Ronde is mainly accreted terrane lithosphere and/or crust variably polluted by more of the cratonic component.

volcanic succession could not have exceeded 10%. Indeed, there is some petrographic evidence for small degrees of fractional crystallization between the less evolved and more evolved lavas of Grande Ronde Basalt (Reidel, 1983; Durand and Sen, 2004; Ramos et al., 2005; Caprarelli and Reidel, 2004), but there is little geologic evidence that fractional crystallization was a significant factor in generating Grande Ronde Basalt from the modification of peridotite partial melts (Wright et al., 1989; Takahashi et al., 1998; Hooper et al., 2007). Unlike the Steens, Imnaha, and Picture Gorge lavas, which have abundant large plagioclase phenocrysts commonly >15 mm across, the vast majority of Grande Ronde lavas are totally aphyric, with only a few flows containing sparse plagioclase phenocrysts, typically <4 mm across. Major-element trends led Carlson et al. (1981) to suggest that a substantial fraction (~50%) of the crystallizing assemblage was plagioclase. Significant plagioclase removal should result in large negative Eu anomalies, as expressed by the ratio of analyzed Eu to Eu measured by linear extrapolation on rare-earth element (REE) plots between Sm and Tb. However, such plots reveal small negative Eu anomalies in the Grande Ronde lavas that do not correlate with either MgO or La/Yb ratio, and are therefore unlikely to result from plagioclase fractionation (Wright et al., 1989). Chamberlain and Lambert (1994) suggest instead that these small Eu anomalies could have been inherited from a pyroxenitic source. Although clinopyroxene appears to be a necessary component in the fractional crystallization models of Carlson et al. (1981), there is a total absence of clinopyroxene phenocrysts, and instead only sparse, exceedingly small microphenocrysts or groundmass crystals, typically <0.1 mm across. The near total absence of phenocrysts does not in itself eliminate crystal fractionation as a viable process, but it does require a special mechanism in crustal reservoirs to separate crystallizing phenocrysts. The lack of a hiatus or chemical transition across the widespread Imnaha–Grande Ronde stratigraphic boundary requires this mechanism to have been abrupt and near simultaneous across the entire length of the Grande Ronde dike exposures, a north-south distance of ~400 km.

Whereas the quantitative models of Carlson et al. (1981) and Carlson (1984) attribute the major- and trace-element variations of Grande Ronde Basalt largely to fractional crystallization, the observed isotopic variations are instead attributed largely to crustal assimilation. Although the mixing component of assimilation appears to require evolved, "granitic" rock types, it seems unlikely that such assimilation could have taken place in the granitic upper crust. The broad area of eruption above the Chief Joseph

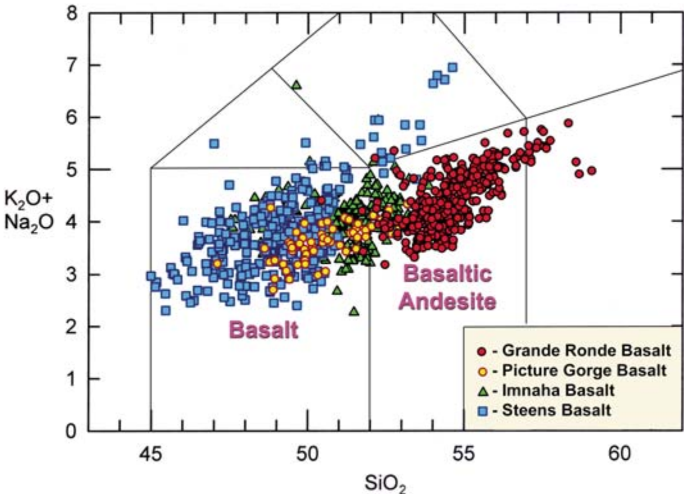

Figure 3. Total alkali versus silica diagram (Le Bas et al., 1986) displaying 1253 representative samples of the Columbia River Basalt Group main phase stratigraphic units. Chemical data from Johnson et al. (1998), Hooper (2000), Hooper et al. (2002), Camp et al. (2003), and the GeoRoc chemical database (http://georoc.mpch-mainz.gwdg.de/georoc/).

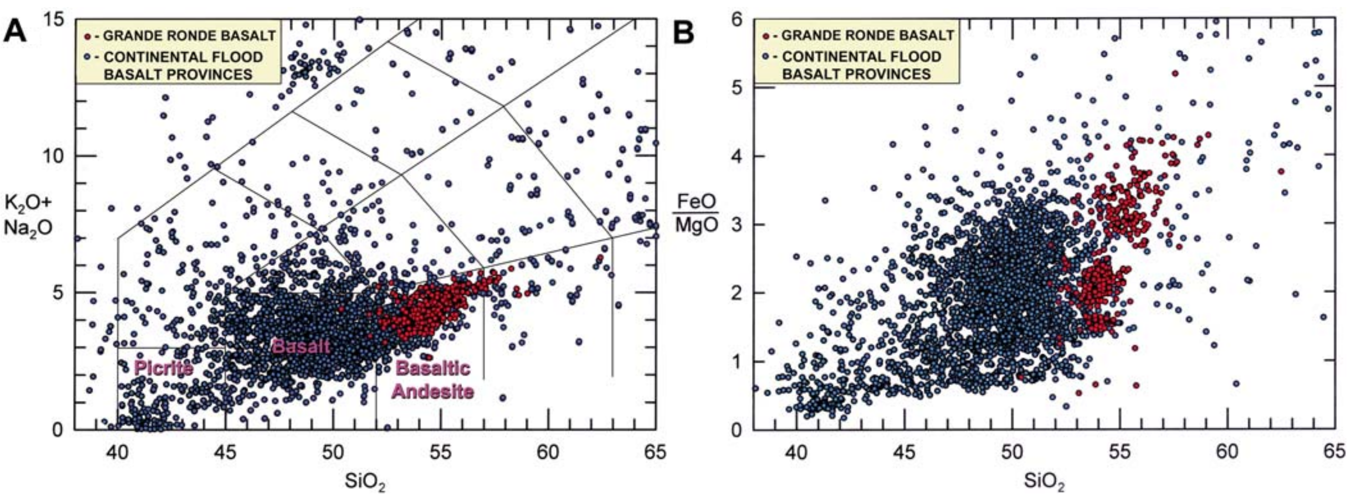

Figure 4. Comparison of 470 analyses of Grande Ronde Basalt with 4466 analyses from well-known continental flood-basalt provinces (Deccan, Ethiopia, Yemen, Parana, Siberia, and Emeishan). (A) Total alkali versus silica diagram (Le Bas et al., 1986). (B) FeO/MgO versus silica diagram. Analyses are from Johnson et al. (1998), Hooper et al. (2002), Camp et al. (2003), and the GeoRoc chemical database (http://georoc.mpch-mainz.gwdg.de/georoc/).

dike swarm is maintained at a high elevation, with only a few local basins along its perimeter. Swanson et al. (1979) note that this lack of regional subsidence at the center of the eruption is inconsistent with the massive evacuation of the Grande Ronde Basalt (~148,000 km^3) from magma chambers in the upper crust. Wright et al. (1979, 1989) and DePaolo (1983) also note that the dike swarm lacks discrete volcanic centers, calderas, and extensive rhyolites, all of which should be expected above large magma chambers in the upper crust. Although the few phenocrysts found in the Grande Ronde lavas appear to have crystallized at crustal depths, where some were also modified by magma mixing processes (Durand and Sen, 2004; Ramos et al., 2005), thermobarometric and mineral stability considerations demonstrate that these magmas could not have spent the substantial amounts of time in shallow-level reservoirs necessary for large-scale assimilation (Ramos et al., 2005; Caprarelli and Reidel, 2004, 2005).

Carlson (1984) acknowledges many of these concerns on the assimilation of upper crust, and suggests instead that substantial assimilation could have taken place at deep crustal levels, or at the crust-mantle boundary. To satisfy the isotopic variation of Grande Ronde Basalt, Carlson (1984) states that a melt derived from his relatively enriched C2 mantle source (the main component of Imnaha Basalt) would still require the assimilation of a crustal component composed of isotopically evolved "granitic" materials, presumably metamorphosed to the granulite facies in the lower crust, but with Sr, Nd, and Pb concentrations equivalent to the average upper crust of Taylor and McClennan (1981). However, Glazner (2007) has shown that assimilation at any scale is severely limited by the energy required to dissolve xenoliths, so that adding only 10% of granitic crust to basalt will result in rapid cooling, with an abrupt increase in viscosity and a crystal content of 30%–40%, all of which are inconsistent with the rapid eruption and large volume of aphyric Grande Ronde Basalt.

We agree that the evolved chemistry of these lavas can be reasonably modeled by combined processes of fractional crystallization and crustal contamination (Carlson et al., 1981; Carlson, 1984). However, as noted above, there appears to be little direct geologic evidence for significant crystallization of plagioclase and pyroxene as required by the model. Although these model calculations also require substantial assimilation of evolved "granitic" crust, the Re-Os data of Chesley and Ruiz (1998) rule out an evolved crustal component in the Grande Ronde source; but instead, these data identify a mafic component, consistent in composition to the lower crust thought to reside beneath the primitive oceanic arc terranes that host the vast majority of Grande Ronde feeder dikes (Vallier, 1995; O'Driscoll, 2007).

Genesis from the Melting of Mafic Crust

Wright et al. (1989) believe that the aphyric nature of the Grande Ronde lavas, their chemical coherency, and their high silica compositions, are characteristics of true melt compositions generated at the climax of the Columbia River Basalt Group eruptions. It is clear that such high-Si, high-Fe/Mg rocks cannot be generated from the partial melting of mantle peridotite; however, several studies suggest that they can be generated from large-scale melting of mafic crust. This is demonstrated by the chemical and experimental results of Helz (1978), Helz and Wright (1987), and Wright et al. (1989), and by the high-pressure melting experiments of Takahahshi et al. (1998). These studies show that the Grande Ronde major- and trace-element trends can be produced by partial melting along an aluminous-clinopyroxene control line, at pressures above ~2.0 GPa (~70 km), consistent with the melting of mafic crust composed of pyroxenite or eclogite at mantle depths (Helz and Wright, 1987; Takahahshi et al., 1998).

Such a process would seem to require a mechanism capable of placing mafic crust at a depth below the presumed crust-mantle boundary underlying the Chief Joseph dike swarm. This could be accomplished by a "bottom-up" process involving the ascent of an eclogite-bearing plume (i.e., Takahahshi et al., 1998), or by a "top-down" process involving the descent of mafic lower crust by detachment and delamination (i.e., Hales et al., 2005).

In summary, the available evidence suggests that there are two equally viable alternatives to generate the Grande Ronde chemical trends: (1) through the modification of peridotite partial melts by large-scale fractional crystallization and assimilation, or (2) through direct melting of a mafic crustal source. We do not rule out a limited role for fractional crystallization and/or crustal assimilation in the genesis of Grande Ronde Basalt. However, we believe that the substantial storage time necessary for these processes to produce the evolved Grande Ronde compositions is inconsistent with the enormous volume and rapid eruption rate (~600 km^3/km/Ma) of these aphyric lavas at the climax of flood basalt volcanism. Geologic evidence discussed here and elsewhere (e.g., Wright et al. 1989; Takahahshi et al., 1998) casts at least some doubt on the genesis of these lavas by combined processes of large-scale fractional crystallization and assimilation. The overall evidence leads us to concur with others who maintain that the Grande Ronde lavas were derived instead from the melting of a mafic pyroxenitic or eclogitic source rock at mantle depths (Wright et al., 1976; Helz, 1978; Swanson and Wright, 1981; Reidel, 1983; Wright et al., 1989; Cordery et al., 1997; Takahahshi et al., 1998; Hooper et al., 2007).

CONSTRAINTS ON SOURCE MELTING

The main stratigraphic units comprising the Columbia River Basalt Group are shown in Figure 5A. After arriving in southeastern Oregon at ca. 16.6 Ma with the eruption of Steens Basalt, the flood-basalt succession migrated rapidly into northeastern Oregon and adjacent Washington, as demonstrated by (1) the northward migration of age-progressive dikes, and (2) the northward thickening and progressive offlap of age-progressive Columbia River Basalt Group stratigraphic and paleomagnetic units (Fig. 5B) (Camp, 1995; Hooper et al., 2002; Camp et al., 2003; Camp and Ross, 2004; Hooper et al., 2007). The lack of evidence for a tectonically controlled propagating rift zone led Camp and Ross (2004) to suggest that this migration was ultimately controlled by the northward advance of upwelling mantle.

The Columbia River Basalt Group stratigraphic succession (Fig. 5A) is consistent with the sequential melting of several source components that mix with a northward-advancing mantle plume (*PL*). These source components are the depleted Paleozoic-to-Mesozoic mantle lithosphere (*DML*), mafic crust of Phanerozoic (*mc*) and/or Archean (*Pmc*) age, and Archean mantle lithosphere (*Pm*) (Fig. 2). The mantle-plume component is least diluted by these other components in the Imnaha Basalt source.

We assess here the feasibility of petrogenetic models in meeting the following four primary constraints on source melting: (1) Imnaha Basalt appears to require a mantle-plume source as its dominant component, consistent with its trace-element content, its OIB character, and its high $^3He/^4He$ ratios (Brandon and Goles, 1988; Draper, 1991; Brandon et al., 1993; Hooper and Hawkesworth, 1993; Dodson et al., 1997; Hooper et al., 2007; Wolff et al., 2008). (2) Grande Ronde Basalt is best interpreted as being derived largely from the melting of a mafic source of pyroxenite or eclogite (Wright et al., 1976; Helz, 1978; Swanson and Wright, 1981; Reidel, 1983; Wright et al., 1989; Takahahshi et al., 1998; Hooper et al., 2007). (3) The time-line of source melting must correspond with the stratigraphic record, thus generating a progression of melting events from the following dominant sources, each variably mixed with a plume component, from oldest to youngest (and, with northward progression): depleted mantle lithosphere (*DML*, Steens

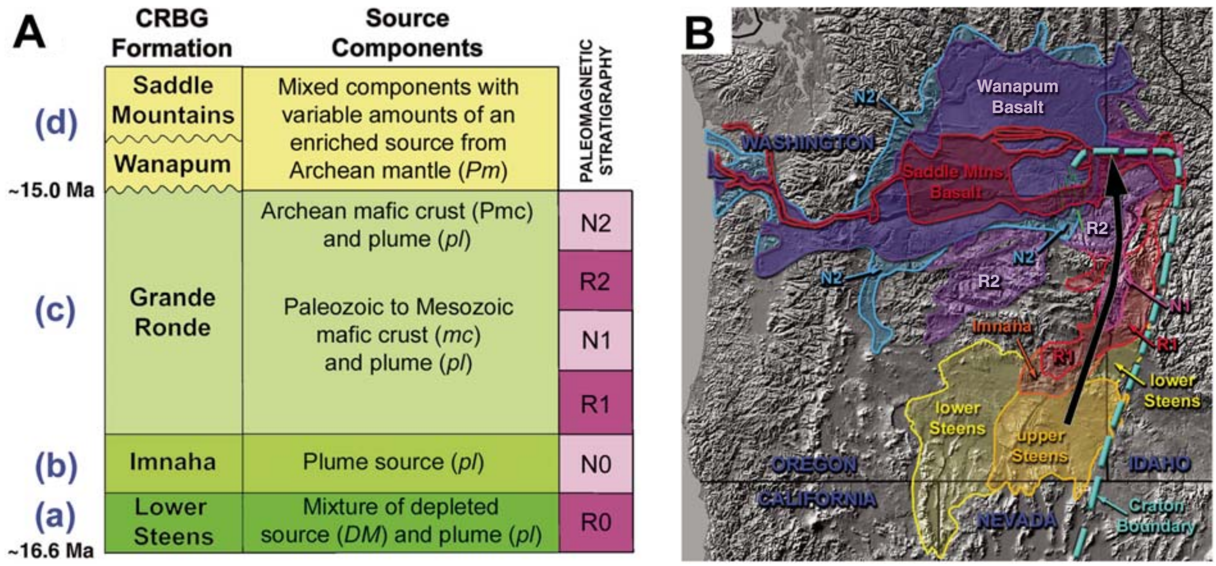

Figure 5. Stratigraphy and map distribution of main Columbia River Basalt Group (CRBG) units. (A) Stratigraphy and source components of the Columbia River Basalt Group units that erupted along cross-section A–A' in Figure 1. Letters (a)–(d) correspond with the evolution of each formation as depicted in the cross-sectional diagrams of Figure 7. Paleomagnetic units R0–N2 correspond with sequential reverse and normal paleomagnetic intervals during the main-phase eruptions. The terms lower Steens and upper Steens Basalts are defined in Hooper et al. (2002) and Camp et al. (2003). Imnaha Basalt clearly overlies lower Steens Basalt in the Malheur Gorge of eastern Oregon (Hooper et al., 2002). The stratigraphic relationship between Imnaha Basalt and upper Steens Basalt is poorly constrained, although they may be interbedded with one another south of the Malheur Gorge region (Camp et al., 2003). (B) Map distribution of main Columbia River Basalt Group units. Northward migration of volcanism is evident in the northward offlap of progressively younger units from southeastern Oregon into northeastern Oregon and adjacent Washington State.

Basalt), largely undiluted plume (*PL*, Imnaha Basalt), mafic crust (*mc* and *Pmc*, Grande Ronde Basalt), and Archean mantle lithosphere (*Pm*, a progressively important component in Wanapum and particularly the Saddle Mountains Basalts) (Fig. 5). Picture Gorge Basalt erupted farther to the west, along the Monument dike swarm during Grande Ronde time (Fig. 1), from a depleted mantle source similar to Steens Basalt (*DML*). (4) There is a clearly defined, step-function change in chemistry at the Imnaha–Grande Ronde stratigraphic boundary, with a complete lack of transitional lavas, interbedded relationships, or an unconformity. Imnaha Basalt was the first to erupt from the Chief Joseph dike swarm, reflecting the abrupt appearance of melts having a plume signature. This was followed by the equally abrupt appearance of Grande Ronde Basalt derived from a mafic source. This step-function change in source melting can only be explained by a mechanism that abruptly terminates Imnaha Basalt volcanism, while simultaneously initiating Grande Ronde Basalt volcanism, without a hiatus in the eruption sequence. A similar step-function chemical change is not as obvious at the Steens-Imnaha stratigraphic boundary, which is only exposed in a few isolated places and not as well defined (Hooper et al., 2002). Some Imnaha Basalts (the Rock Creek chemical type of Hooper et al., 1984) have major- and trace-element compositions similar to the lower Steens Basalts of Johnson et al. (1998) and Camp et al. (2003); however, these same flows have isotopic compositions indicative of a mantle plume source (*PL*) (Hooper and Hawkesworth, 1993).

Realistic petrogenetic models must not only conform to the proper stratigraphic sequence of source melting, but must also provide a mechanism of plume emplacement capable of generating the voluminous Grande Ronde succession from the large-scale melting of a mafic source at mantle depths. We can envision only three potential models capable of meeting these constraints: (1) melting of an eclogite-bearing mantle plume (Takahahshi et al., 1998), (2) plume interaction with the Juan de Fuca plate, and (3) lithosphere delamination triggered by plume emplacement.

MELTING OF AN ECLOGITE-BEARING MANTLE PLUME

Takahahshi et al. (1998) concluded from high-temperature melting experiments that Grande Ronde Basalt could be derived from a heterogeneous plume containing large, lithologically distinct slabs of old oceanic crust. Since these slabs of oceanic crust come from the deep mantle, they should be composed of eclogite, and therefore characterized by high-P mineral assemblages, including garnet. The REE data for Grande Ronde Basalts, however, rule out the involvement of garnet in melting of the mantle source (Wright et al., 1989).

Under close scrutiny, an eclogite-bearing plume has difficulty in satisfying three of our four primary constraints on source melting. Although Takahahshi et al. (1998) envision a plume bearing distinct slabs of oceanic crust, numerical experiments on rising plumes predict the deformation and mixing of such chemical heterogeneities into thin streaks and filaments (Farnetani and Samuel, 2003, 2005; Lin and van Keken, 2006). It is unlikely that partial melting of such well-mixed plumes could generate the abrupt step-function change in source melting seen at the Imnaha–Grande Ronde stratigraphic boundary (constraint 4).

Herzberg et al. (2007) demonstrate that thermal anomalies associated with mantle plumes are typically 200–300 °C above ambient mantle potential temperatures, which vary between

1280 and 1400 °C. At these elevated temperatures, the sequence of source melting appears to be inconsistent with the Columbia River Basalt Group stratigraphy (constraint 3). The melting experiments of Yaxley (2000) show that at a potential temperature of 200 °C above ambient mantle, eclogite will begin to melt at ~270 km depth, and fertile peridotite (i.e., pyrolite) at ~90 km depth (Fig. 6). This is verified by the numerical calculations of Leitch and Davies (2001) showing that pyrolite in a rising plume starts to melt near the base of the lithosphere at about the same time that eclogite has melted completely. Even at the lower plume temperatures and pressures envisioned by Takahahshi et al. (1998), eclogite melts (i.e., Grande Ronde) should arrive at the surface first, followed by peridotite melts (i.e., Imnaha). This sequence is the reverse of the Columbia River Basalt Group stratigraphic succession (Fig. 5).

Inherent in the model of Takahahshi et al. (1998) is the assumption that eclogite magmas, once generated, would rise unimpeded to the surface. Several workers, however, have suggested that eclogite melts generated in a rising plume will instead react with olivine from the mantle peridotite to refertilize the plume (Kelemen et al., 1998; Herzberg and O'Hara, 2002; Herzberg, 2006). Sobolev et al. (2007) have shown that this reaction produces an olivine-free rock enriched in pyroxene, a so-called hybrid pyroxenite. They demonstrate through melting experiments that eclogite-entrained mantle plumes should typically produce picrites and olivine tholeiites from a mixture of peridotite and pyroxenite sources, noting that high-silica lavas derived from 100% melting of the pyroxenite component are rare. These experiments are inconsistent with the generation of voluminous Grande Ronde-like lavas from eclogite-bearing plumes. Although we do not discount the involvement of a heterogeneous plume in the genesis of the Columbia River Basalt Group lavas, the existing data suggest that melting of this source in isolation is an unlikely mechanism for the derivation of Grande Ronde Basalt (constraint 2), the sequence of Columbia River Basalt Group eruption (constraint 3), and the step-function change in source melting evident at the Imnaha–Grande Ronde stratigraphic boundary (constraint 4).

PLUME INTERACTION WITH THE JUAN DE FUCA PLATE

Ihinger et al. (2004, 2005) suggested that the Yellowstone plume may have impinged on the Farallon slab as early as 80 m.y. ago. Other workers, however, have suggested that the plume is much younger and was shielded by the Miocene remnant of the Farallon slab (Juan de Fuca plate) only a few million years before breaking through to generate the surface eruptions at ca. 16.5 Ma (Duncan, 1982; Geist and Richards, 1993; Pierce et al., 2002; Humphreys, 2007). When corrected for 15%–20% extension of the northern Basin and Range province since ca. 20 Ma (Colgan et al., 2004; Lerch et al., 2007), plume-plate interaction would require a subduction angle <45°. This is transitional between the much steeper 60° subduction angle today (Rasmussen and Humphreys, 1988) and the presumed flat subduction that existed beneath the region at ca. 50 Ma (Atwater, 1989; Pierce et al., 2002). If the subduction angle was >45°, however, the slab may have never encountered the plume, instead encountering the 660-km mantle discontinuity before reaching the area of plume ascent in southeastern Oregon and adjacent Nevada.

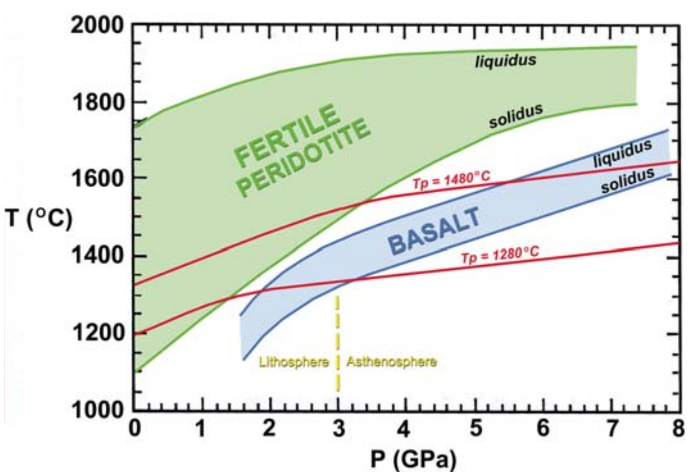

Figure 6. Solidi and liquidi for average mid-ocean ridge basalt (MORB) (Yasuda et al., 1994) and fertile peridotite (McKenzie and Bickle, 1988). Mantle adiabats for potential temperatures of 1280 °C and 1480 °C from Miller et al. (1991a, 1991b). Modified from Yaxley (2000). A pressure interval of 1.0 GPa is equivalent to ~35 km depth.

The immediate response of plume impingement against the Juan de Fuca plate should have been buoyant uplift of the slab leading to a flatter subduction angle. This is consistent with a hiatus in Cascade volcanism at ca. 18 Ma (Priest, 1990), and with a poorly understood, broad region of Oligo-Miocene calc-alkaline to mildly alkalic volcanism that extended across the breadth of eastern Oregon and into western Idaho from ca. 26 to 18 Ma, immediately before the initial Steens eruptions (Ekren, et al., 1981; Langer, 1991; Mathis, 1993).

During impingement, it is conceivable that the initial Steens eruptions were generated by rapid heating of the Juan de Fuca mantle lithosphere, to produce melts derived from a depleted source (*DML*) mixed with melts from the mantle plume (*PL*) (Fig. 2). This was followed by the Imnaha eruptions as the plume began to break through the subducting slab while decompressing, thus producing more enriched melts with OIB (i.e., plume) signatures (*PL*). Before completely breaking through the subducting plate, the plume would have encountered the oceanic crust. With the solidus temperature of basalt lying well below the potential temperatures of a rising plume at asthenospheric depths (Fig. 6) (Yasuda et al., 1994), the oceanic crust would have undergone near-wholesale melting, partially mixing with plume melts to generate Grande Ronde Basalt. Although this scenario appears capable of generating the main compositional types of the Columbia River Basalt Group, it is difficult to envision how such a model could achieve the abrupt compositional change and observed chemo-stratigraphic order of the Imnaha and Grande Ronde successions.

Isotopic ratios for Grande Ronde Basalt appear to fall along broad mixing lines between a plume source and an unknown source showing relatively low $^{143}Nd/^{144}Nd$, and relatively high $^{87}Sr/^{86}Sr$, $^{207}Pb/^{204}Pb$, and $^{208}Pb/^{204}Pb$ relative to $^{206}Pb/^{204}Pb$. Although this trend has been attributed to crustal contamination (Carlson, 1984; Carlson and Hart, 1987), it could also be derived from pelagic sediment that was melted by the thermal flux of the plume and then mixed with basaltic melts derived from the Juan de Fuca plate (Fig. 2).

Plume rupture of an overriding plate has been suggested from seismic data beneath the south-central Sea of Japan (Lebedev and Nolet, 2003), and in the kinematic models of Ranalli et al. (2004) applied to volcanism in the European Alps. The feasibility of slab rupture by the buoyant forces of a rising plume, however, remains poorly constrained by substantive data. Plume ascent could also be attributed to a structural gap in the descending plate (Humphreys, 2007), although such a mechanism is less likely to result in the melting of oceanic crust. The lack of substantive data on the mechanics of slab erosion makes it difficult to assess accurately a model of plume-slab interaction. Based solely on chemical arguments, however, we conclude that this is a plausible model for the generation of the Columbia River Basalt Group main-stage compositional types (constraints 1 and 2), but we are skeptical that it can provide a mechanism for the proper sequence of source melting (constraint 3) and the step-function petrochemical change evident at the Imnaha–Grande Ronde stratigraphic boundary (constraint 4).

PLUME-TRIGGERED DELAMINATION

Delamination is based on the premise that magmatic underplating can thicken lower crust, converting basaltic rocks into garnet pyroxenites, so that the increased density generates gravitational instabilities that drive delamination (Anderson, 2005, 2007; Elkins-Tanton, 2005; Lustrino, 2005). Hales et al. (2005) have applied such a model to uplift and volcanism in the Columbia River Basalt Group source region. They have identified a large, seismically fast volume of material beneath the Chief Joseph dike swarm at a depth of ~75–175 km. Hales et al. (2005) conclude that this volume is the depleted residue of basalt melting, leaving a residual body with increased mantle buoyancy that accounts for the amount of regional uplift during basalt eruption. They note, however, that this basalt-depleted mantle cannot account for more focused uplift of ~2 km in the Wallowa Mountains lying above the center of the high-velocity volume. They attribute this instead to a gravitational instability associated with the dense roots of the Wallowa batholith, resulting in delamination of the mantle lithosphere and lower crust.

Delamination alone, however, has difficulty in explaining the regional extent of flood-basalt volcanism, from as far west as the Monument dike swarm and as far south as the Northern Nevada Rift (Fig. 1). Also, the numerical experiments of Elkins-Tanton (2007) demonstrate that at ambient mantle temperatures, even the most extreme conditions of delamination cannot generate the large melt volumes and the rapid accumulation rates necessary for flood-basalt genesis. These experiments show that anomalously hot mantle must be present to generate these large melt volumes, even where mantle melting temperatures are lowered by the descent and devolatilization of hydrated mantle.

Elkins-Tanton (2007) note that emplacement of a hot mantle plume decreases asthenosphere density and viscous strength, which would enhance the probability of delamination without requiring a thickened crust. This is consistent with numerical experiments showing that the mechanical detachment of mantle and crustal layers is a natural consequence of plume impingement beneath thin continental lithosphere (Burov and Guillou-Frottier, 2005). Hales et al. (2005) speculated on whether a plume head residing beneath the thin lithosphere of arc terranes adjacent to the thick cratonic boundary of North America may have provided a trigger for delamination associated with the Columbia River Basalt Group eruptions. Such a scenario is supported by thermo-mechanical experiments of plume impingement near cratonic boundaries (Burov et al., 2007). To test this idea, we first discuss the susceptibility of the Mesozoic accreted lithosphere to plume-induced delamination; we then examine whether this model can satisfy our four primary constraints on source melting, and its consistency in predicting the Tertiary history of uplift and deformation in eastern Oregon and adjacent Washington.

Instability of the Mesozoic Lithosphere

The calculated density profiles of Poudjom Djomani et al. (2001) suggest that young, thin lithosphere is gravitationally unstable when compared to older, thicker lithosphere, and therefore particularly susceptible to sinking into hot mantle upwellings (Elkins-Tanton, 2007). Numerical models have demonstrated that mechanical decoupling between mantle and crustal layers is the expected result of plume impingement beneath thin, Mesozoic continental lithosphere, similar to that underlying the Chief Joseph dike swarm (Burov and Guillou-Frottier, 2005). These experiments show that Rayleigh-Taylor instabilities naturally develop at the top and sides of mantle-plume heads during impingement and lateral spreading beneath the lithospheric lid.

The peculiar lithospheric architecture beneath northeastern Oregon may have provided an environment particularly conducive to both the focusing of mantle-plume flow and lithospheric detachment. The Yellowstone plume first arrived near the Oregon-Nevada-Idaho tri-state area at ca. 16.6 Ma, centered beneath transitional lithosphere located between the 0.704 and 0.706 Sr-isopleths of Figure 1 (Pierce and Morgan,1992; Camp and Ross, 2004), or perhaps beneath the craton itself (Pierce et al., 2002; Jordan et al., 2004). Upon its arrival, the spreading plume head rapidly flowed into northeastern Oregon, as it advanced "uphill" (Thompson and Gibson, 1991) beneath the thinner lithosphere of oceanic terranes (Fig. 5B) (Camp, 1995; Camp and Ross, 2004). Here, the plume head encountered an abrupt westward bend in the thick cratonic boundary at the northern end of the Chief Joseph dike swarm (Figs. 1 and 5B). The boundary here is well defined by the 0.706 Sr isopleth, the juxtaposition of disparate lithologic assemblages, U-Pb age data, ^{40}Ar-^{39}Ar cooling ages, and structural relationships (Fleck and Criss, 2004; Lund et al., 2005; McClelland and Oldow, 2007). This buttress to mantle-plume flow would have enhanced downward convective flow at the front of the plume head, thus providing a catalyst for delamination. This is consistent with the thermo-mechanical experiments of Burov et al. (2007) demonstrating that slab-like instabilities result from mantle-plume flow against thick cratonic boundaries, thus generating subvertical downthrusting and delamination.

Delamination at the cratonic boundary is likely to incorporate lower crust of the accreted terranes. Mafic lower crust underlying primitive arc terranes has been shown to be gravitationally unstable and therefore highly susceptible to being incorporated into foundering slabs above mantle plumes (Jull and Kelemen, 2001). The gravitational instability and detachment of lower crust is not only demonstrated by numerical models (Burov and Guillou-Frottier, 2005), but also by direct field observations and fractionation modeling of primitive island arcs (Xu et al., 2002; Kelemen et al., 2003; Behn and Kelemen, 2006), together with evidence derived from lower crustal xenoliths, seismic imaging, and observed uplift patterns (Ducea and Saleeby, 1996, 1998; Lee et al., 2000, 2006; Saleeby et al., 2003; Zandt et al., 2004; Behn et al., 2007).

The Chief Joseph swarm is confined to the primitive Wallowa and Baker arc terranes, dominated by mafic magmatic rocks with dense crustal roots (Vallier, 1995) and intrusive bodies that are likely to be underlain by dense residuum of garnet-amphibolite (O'Driscoll, 2007). Delamination may well have taken place along mechanically weak suture zones separating these Mesozoic terranes. Two lithospheric weak zones bound the majority of the dikes associated with the Chief Joseph swarm: the cratonic boundary, juxtaposed against the Wallowa terrane in the north, and a well-defined suture zone separating the Baker and Izee terranes to the south (Fig. 1). A circular feature with a central uplift and a diameter of ~230 km lies between these two zones of crustal weakness. Hales et al.

(2005) attribute this bull's-eye pattern of uplift to a drip-like downwelling near the center portion of the Chief Joseph dike swarm. We suggest that this circular region of massive dike injection and delamination is the result of a predictable sequence of events associated with the Miocene arrival of the Yellowstone mantle plume head.

Spatial and Temporal Evolution of Source Melting

In most plume models, the lithosphere is treated as a single viscous layer with Newtonian rheology. Burov and Guillou-Frottier (2005), however, present a more realistic model that considers the consequences of plume impingement on continental lithosphere having variable rheologic properties, as expected from the variable compositions of lithospheric mantle, ductile lower crust, and brittle upper crust. These experiments have been recently expanded to include the consequences of plume impingement beneath thin Phanerozoic lithosphere lying adjacent to a thicker, older cratonic boundary (Burov et al., 2007). The model conditions for these experiments closely resemble the tectonomagmatic environment of flood-basalt volcanism in eastern Oregon. It therefore seems reasonable to utilize the model results in an attempt to establish a realistic sequence of events during the period of Columbia River Basalt Group eruption. An important observation of these experiments is the identification of two types of Rayleigh-Taylor gravitational instabilities: (1) "drip-like" downwellings located closer to the plume center where heating is most efficient, and (2) "slab-like" downwellings that form at the edges of the advancing plume head. Burov et al. (2007) note that the importance of both types is that they may trigger and enhance melting processes by bringing lithospheric material into contact with hot mantle, particularly near a cratonic margin where the hot plume head stalls.

Here, we test the idea of plume-induced delamination in its ability to predict an evolving sequence of melting events consistent with the record of chemical change through time (i.e., constraints 1–4) (Fig. 7). We base the following scenario on direct field observations, stratigraphic data, isotopic data on source melting, and the precise delamination events predicted in the model experiments. Near the center of the plume in southeastern Oregon, impingement first resulted in drip-like instabilities of lithospheric mantle which could not be supported by the low viscous strength of the plume head. Descent and partial melting of this depleted lithosphere by the elevated temperatures of the plume generated, mid-ocean ridge basalt (MORB)–like melts (*DML,* Fig. 2), which quickly mixed with melts from the mantle plume (*PL*, Fig. 2), thus generating Steens Basalts (Fig. 7A). These magmas were the first to erupt in southeastern Oregon, well south of the Chief Joseph dike swarm, where their ascent was facilitated by lithospheric extension in a backarc environment.

As the northward-advancing plume crossed the suture zone separating the Izee and Wallowa terranes (Fig. 1), a slab-like instability was generated at the edge of the advancing plume head beneath the Wallowa region, as predicted by the model. Slab descent of lithospheric mantle and dense lower crust into the plume head allowed hot peridotite from the plume to rise rapidly into the lithospheric void, where decompressional partial melting of this enriched plume source generated Imnaha Basalt (*PL*, Fig. 2). The Imnaha magmas ponded at the crust-mantle boundary, but then rose quickly to the surface, erupting through fissures at the site of the incipient Chief Joseph dike swarm (Fig. 7B).

Simultaneously, mafic lower crust of the delaminated slab(s) was being emplaced into a thermal environment ~200 °C hotter than the solidus temperature of basalt at asthenospheric depths, and well above its liquidus temperature (Fig. 6) (Yasuda et al., 1994; Herzberg et al., 2007). These pyroxenite-rich rock types melted spontaneously, generating a massive volume of silica-rich, high Fe/Mg melts (*mc*, Fig. 2), perhaps mixing to a very small degree with plume-derived magmas. While largely maintaining near-liquidus temperatures, these high-volume melts would rise rapidly, breaking through the crust-mantle boundary with little residence time in the crust, to erupt at the surface as the voluminous aphyric lavas that comprise the Grande Ronde succession (Fig. 7C). We can envision no other mechanism that so adequately explains the step-function chemical change that takes place at the Imnaha–Grande Ronde stratigraphic boundary (constraint 4).

Grande Ronde volcanism ceased at ca. 15 Ma, shortly after the northward-advancing plume head encountered the thick cratonic boundary of North America. This hiatus was followed by more sporadic eruptions of Wanapum and Saddle Mountains Basalts across the cratonic boundary from ca. 15.0 to 6.0 Ma. These later-stage lavas appear to be composed of variable mixtures of a plume component (*PL*, Fig. 2) with melts having an increasingly greater quantity of an enriched Archean lithosphere component (*Pm*, Fig. 2) (Hooper and Hawkesworth, 1993). Some of the Wanapum and Saddle Mountains magmas may well have ponded at the crust-mantle boundary, or in the crust, where they assimilated Archean granitic rocks having a more enriched isotopic signature (Carlson, 1984).

This model of delamination, triggered by a northward-propagating plume head, appears to satisfy all of our four primary constraints on source melting, generating a series of melting events identical to that observed in the stratigraphic record. This progressive sequence of melting is dictated solely by the delamination events predicted by the model experiments, without requiring additional assumptions on mantle processes. The model also provides a mechanism for generating a high volume of superheated lavas above liquidus temperatures (Fig. 6), consistent with the chemical coherency of Grande Ronde Basalt, and rapid eruption rate of aphyric lavas at the climax of Columbia River Basalt Group volcanism.

Further Constraints on the Grande Ronde Mafic Source

Grande Ronde Basalts have many of the chemical signatures associated with continental crust, such as high K_2O/P_2O_5 and Th/Ta, elevated $^{87}Sr/^{86}Sr$ and $\delta^{18}O$, and low $^{143}Nd/^{144}Nd$ (Carlson, 1984; Carlson and Hart, 1987, 1988). Given the evidence noted above against significant assimilation of granitic upper crust, we believe that these chemical signatures are instead produced by partial melting of mafic lower crust (e.g., Glazner et al., 1991; Dungan and Davidson, 2004). Following Takahahshi et al. (1998), we believe that melting of this mafic crust occurred below ~70 km depth. The lack of heavy rare-earth element (HREE) depletion suggests that the melted source did not contain residual garnet, as one would expect from near-wholesale melting, which would quickly exhaust any garnet in the crustal source. Isotopic considerations, described below, suggest that the age of this crustal source increases toward the top of the Grande Ronde stratigraphy, with the older magnetostratigraphic units (R1, N1, and R2) being derived largely from mafic crust of the Paleozoic-to-Mesozoic arc terranes, and the younger N2 magnetostratigraphic unit being derived largely from mafic crust of the Archean craton.

The idea that mafic lower crust played an important role in the genesis of the Columbia River Basalt Group lavas is demonstrated by Re-Os isotope systematics. In a two-component mixing model, Chesley and Ruiz (1998) showed that the calculated initial $^{187}Os/^{188}Os$ values for Imnaha Basalt are consistent with a mantle-plume source mixed with a very small amount (<10%) of mafic lower crust. More importantly, they demonstrated in the same model that the calculated initial $^{187}Os/^{188}Os$ values for Grande Ronde Basalt (R1) are consistent with a primary source of mafic lower crust, having an age of 250 Ma, mixed with a small component (<20%) of plume material. This 250 Ma time period appears to mark the age of the primary high-silica, high-Fe component of Grande Ronde Basalt.

An older source component is also present in the Grande Ronde lavas, as displayed by samples lying along mixing lines between a plume source and a component that appears to be too enriched in $^{87}Sr/^{86}Sr$ and too depleted in $^{206}Pb/^{204}Pb$ and $^{143}Nd/^{144}Nd$ to be of Phanerozoic age (Fig. 2). Although the highest $^{87}Sr/^{86}Sr$ ratios displayed by these lavas are consistent with Archean crust of either mafic or felsic composition, the Re-Os data of Chesley and Ruiz (1998) are more conducive to an origin in the mafic lower crust. Melting of this older source is consistent with the continued northward advance of the plume head, associated with continued delamination along its frontal edge as the plume impinged upon the cratonic margin in N2 time.

Seismic, Topographic, and Structural Profiles of Plume Emplacement

Burov et al. (2007) note that delaminated slabs of lithospheric mantle should descend into the plume to depths up to 300–400 km. Such depths cannot be resolved by the sparse seismic data currently available for northeastern Oregon. However, the tomographic data of Hales et al. (2005) have generated reasonable imaging to depths of 250 km. They have interpreted the seismically fast volume between ~175 and ~75 km depth to be the mantle residuum of source melting. This high-velocity body has an estimated volume of 2×10^6 km^3. If this body underwent 5% partial melting, it would generate 100,000 km^3 of melt, which is ~70% of the volume of Grande Ronde Basalt. However, since the volume of magma to erupt at the surface is likely to be only a fraction of the melt generated at depth, an alternative fraction of 10% would provide the volume necessary for the eruption of Imnaha Basalt. Without more precise seismic data it is difficult to further evaluate the true nature of this high-velocity body and its role in magma genesis.

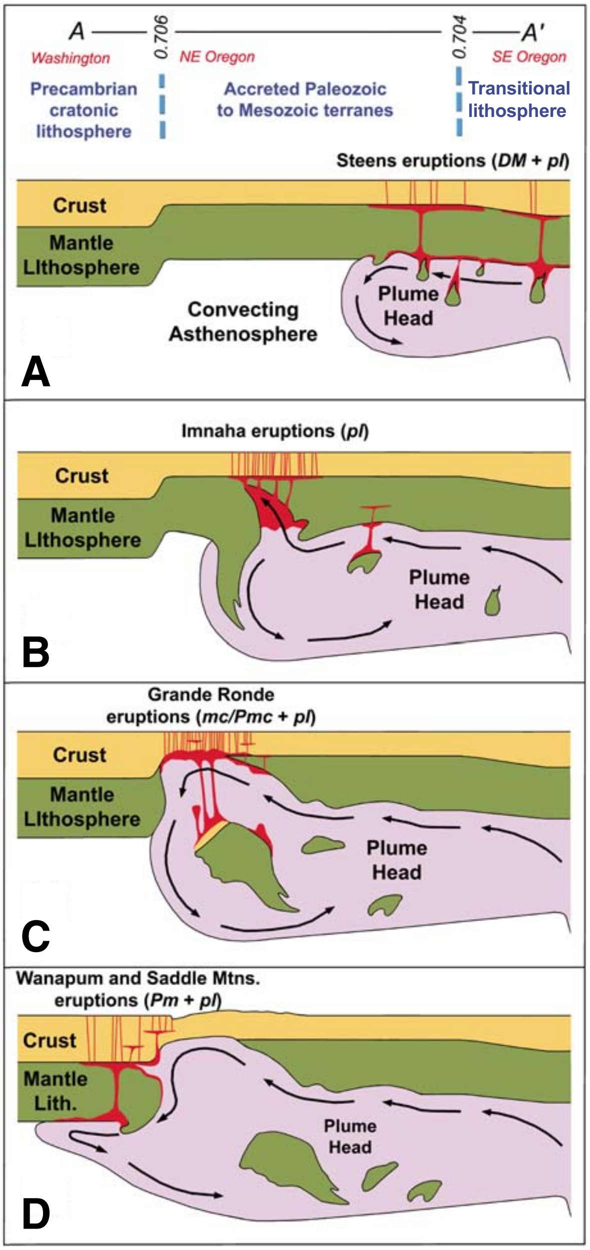

Figure 7. Plume-induced delamination model for the Columbia River Basalt Group, based partly on the thermo-mechanical experiments of Burov et al. (2007). Cross-sections (A)–(D) correspond with the age-progressive evolution of the Columbia River Basalt Group stratigraphy (Fig. 5) as the plume head advanced northward along the cross-section A–A' in Figure 1. (A) Plume impingement in southeast Oregon generates drip-like delamination of depleted lithospheric mantle *(DML)* into the hot plume head, as predicted by the model of Burov et al. (2007), thus generating Steens basalt. (B) As the plume spreads to the north, slab-like delamination predicted by the model allows the mobile plume head *(PL)* to rise into the lithospheric void, thus generating more enriched melts of Imnaha Basalt that erupt from incipient fissures in the Chief Joseph dike swarm. (C) The delaminated slab simultaneously descends into the hot plume head. With the plume temperature lying well above the solidus temperature of basalt, mafic lower crust *(mc)* of the delaminated slab undergoes near-wholesale melting to produce the voluminous Grande Ronde succession. (D) As the plume impinges against the cratonic boundary, more isotopically evolved lavas of the Grande Ronde N2 paleomagnetic unit are generated from the melting of Archean lower crust *(Pmc)*, followed by sporadic eruptions of Wanapum and Saddle Mountains Basalts, generating melts with an increasingly greater component of Archean mantle lithosphere *(Pm)*. After the main-phase Columbia River Basalt Group eruptions, mildly alkaline to calc-alkaline lavas and high-alumina olivine tholeiites erupted discontinuously above the plume head in southeastern Oregon, during a time of crustal extension at the northern margin of the Basin and Range province (Hart et al., 1984; Cummings et al., 2000; Brueseke et al., 2007; Hooper et al., 2002, 2007).

A potential test of the plume-delamination model lies in comparing the uplift topography and surface structure lying above the presumed plume in eastern Oregon to that predicted in the model experiments of Burov et al. (2007). Such a comparison reveals strikingly similar topographic profiles (Fig. 8). The calculated model profiles (Fig. 8B) can be subdivided into four distinct topographic and structural domains, each of which is also present in the observed north-south profile A–A' across eastern Oregon and adjacent Washington (Figs. 1 and 8C). The model experiments generate an unusual juxtaposition of stress regimes, with extension and shortening expressed simultaneously in different domains. This is identical to the observed domains in profile A–A', where Miocene extension is dominant in southeastern Oregon, coeval with shortening in northeastern Oregon and adjacent Washington (Fig. 8) (Walker, 1977; Hooper et al., 1995a; Cummings et al., 2000).

Domain 1

Following an initial period of doming, the thermo-mechanical calculations predict subsidence above the center of the plume, associated initially with drip-like foundering of lithospheric mantle, followed after 3–5 m.y. by surface rifting during plume flattening and relaxation of lithospheric stresses. In southeastern Oregon, initial eruption of Steens Basalt was followed 3–5 m.y. later by rifting above the presumed plume center to generate the Oregon-Idaho graben (Cummings et al., 2000), as predicted by the model.

Domain 2

Uplift and north-south shortening in northeast Oregon is difficult to explain in a backarc setting. However, a similar region of uplift and compression is predicted by the model calculations to form as a peripheral zone at the front of the advancing plume. After 10–16 m.y., the highest elevations are not above the plume center (domain 1), where extension is maintained, but instead near the edge of the plume where it stalls against the cratonic boundary beneath normal lithosphere. This zone of shortening and uplift compares well with the Miocene development of the Blue Mountains anticlinorium, where broad east-west folding and uplift began during the period of flood-basalt eruption and continued after eruption of main-phase lavas (Ross, 1989; Hooper et al., 1995b; Reidel, 2007).

Domain 3

The model experiments produce an abrupt and pronounced subsidence at the cratonic margin caused by slab-like delamination and underthrusting of the cratonic mantle. The experiments predict the development of the Lewiston Basin straddling the Washington-Idaho border, adjacent and parallel to the cratonic boundary. Such striking subsidence and downthrusting in a contractional environment may help to explain the origin of the enigmatic "Lewiston structure" (LS, Fig. 1), an east-west–trending thrust fault with folds in the hanging wall forming a spectacular 500-m–high escarpment on the northern margin of the Lewiston Basin (Figs. 8 and 9) (Camp and Hooper, 1981; Garwood and Bush, 2005). A subduction-like configuration with a narrow thrust-and-fold belt at the edge of basin subsidence accords remarkably well with similar small contractional belts generated in the model experiments. Although other east-west folds in the Pacific Northwest have been attributed to shortening associated with clockwise Tertiary rotation (Wells et al., 1998; McCaffrey et al., 2000), there has been a lack of such rotation in eastern Washington and eastern Oregon over the past 17 m.y. (Wells et al., 1998), so that the Lewiston structure cannot be attributed to that mechanism. However, it is consistent with the surface manifestation of downthrusting mantle against the cratonic margin, as predicted in the model experiments.

Domain 4

Above the craton, the model experiments predict a flat plateau with little variation in surface topography, at an elevation lower than domains 1 and 2, consistent with the Columbia Plateau lying north of the cratonic boundary (Fig. 8).

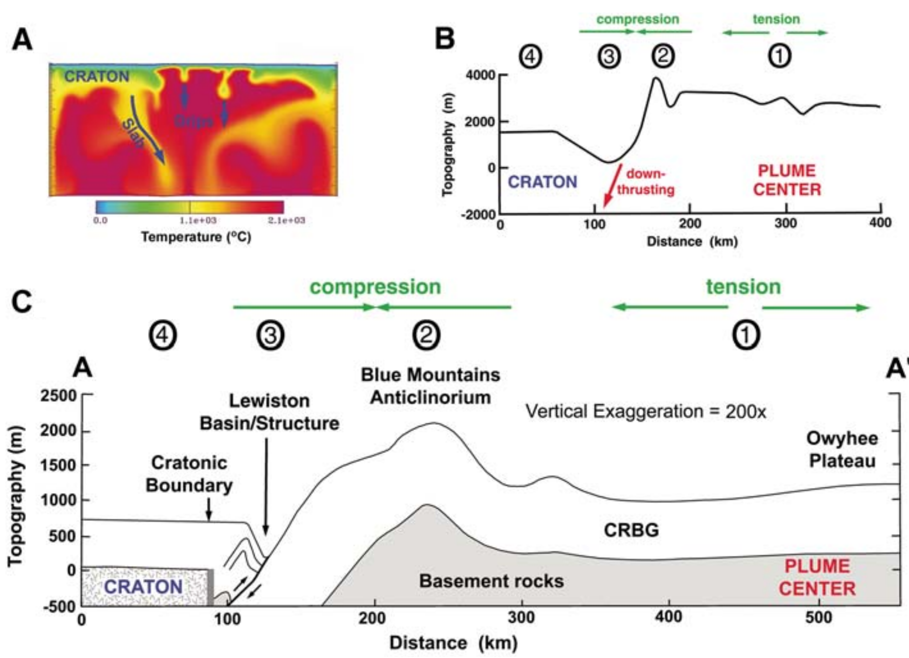

Figure 8. Temperature, uplift, and structural profiles of plume impingement against the cratonic boundary. (A) Temperature profile of Burov et al. (2007) showing predicted drip-like and slab-like instabilities as the mantle plume impinges on the cratonic boundary. (B) Predicted topographic profile generated by the model experiments of Burov et al. (2007), 16.5 m.y. after plume emplacement against a cratonic margin. In the model, the center of the plume occurs beneath normal 150 m.y. lithosphere 300 km away from a thick cratonic boundary. Circled numbers refer to the following topographic and structural domains predicted by the model: Domain 1 is a region of crustal extension and rifting 3–5 m.y after plume emplacement. Domain 2 is a peripheral zone of crustal shortening and uplift where the plume stalls against the cratonic boundary. Domain 3 lies above a zone of slab-like delamination at the cratonic boundary, resulting in crustal shortening, underthrusting, and basin subsidence. Domain 4 is a flat plateau lying above the craton with little variation in surface topography. (C) South-to-north topographic profile and generalized cross-section along A–A' in Figures 1 and 7. The observed domains equivalent to the calculated domains correspond with (1) the Owyhee Plateau and Oregon-Idaho graben, (2) the Blue Mountains uplift and anticlinorium, (3) the Lewiston Basin and Lewiston Structure, and (4) the Columbia Plateau north of Lewiston, Idaho, respectively. See text for more thorough explanations.

CONCLUSIONS

The genesis of continental flood-basalt provinces continues to be the subject of considerable debate. Plume proponents maintain that the characteristic OIB signatures, the high $^3He/^4He$ ratios, and the high magma supply rates of flood-basalt provinces are better explained by a plume origin. However, there is sufficient variability in the composition, structural development, and uplift history of flood-basalt provinces to suggest equally variable tectonic and/or magmatic conditions of plume emplacement. Genesis of the Columbia River Basalt Group, for example, was profoundly influenced by mantle-plume flow against the cratonic boundary of North America in a tectonic environment of backarc extension (Camp and Ross, 2004; Hooper et al., 2007). The thick cratonic boundary provided an effective barrier controlling the propagation direction and underthrusting of mantle-plume flow, and backarc extension provided a stress regime controlling the overall orientation of the Columbia River Basalt Group feeder dikes. The most obvious difference between the Columbia River Basalt Group and other flood-basalt provinces is the unusual high-Si composition of the voluminous Grande Ronde Basalts. We believe that this important distinction requires a genesis that is atypical of the lavas generated during the climax stage of all other continental flood-basalt provinces.

Many genetic models have been proposed for the Columbia River Basalt Group. Our intent here has been to present a review of existing data, not to categorically reject any genetic model, but rather to scrutinize the capability of each in meeting our primary constraints on source melting. We rely on the evidence of others who recognize a plume component in the Columbia River Basalt Group source, variably mixed with other components in the main-phase lava types (e.g., Brandon and Goles, 1988; Hooper and Hawkesworth, 1993; Dodson et al., 1997; Wolff et al., 2008). The most important constraint we use to limit the consideration of genetic models is the evidence for melting of a mafic source component to generate the basaltic andesites of the Grande Ronde Formation (e.g., Wright et al., 1989; Takahahshi et al., 1998; Hooper et al., 2007).

Of the three models we consider, plume-induced delamination is the only one that appears to satisfy all four primary constraints. It provides a realistic mechanism for generating chemically homogeneous, high-Si, aphyric basaltic andesites of the Grande Ronde Formation, while simultaneously generating a time line of melting events consistent with the age-progressive chemical variations seen in the stratigraphic record (Fig. 7). Such a model is not only supported by complementary evidence for both delamination (Hales et al., 2005) and a plume source (Hooper et al., 2007), but also by Re-Os data for a source component of Mesozoic age for most of the Grande Ronde lavas, derived largely from mafic crust of the accreted terranes (Chesley and Ruiz, 1998). The Pb, Sr, and Nd isotopic signature of the Grande Ronde Basalt is also consistent with an accreted terrane source composed of Mesozoic oceanic island-arc crust and associated lithosphere, oceanic sediment, and cratonic detritus welded together by Mesozoic granitoids (Fig. 2). The gravitational instabilities inherent in the model are compatible with field observations in mafic arc terranes (e.g., Jull and Kelemen, 2001), with numerical calculations demonstrating that plume impingement on thin Mesozoic lithosphere results in the decoupling of crustal and mantle layers (Burov and Guillou-Frottier, 2005), and by calculations showing that plume flow against a thick cratonic boundary results in rapid downthrusting of mantle lithosphere (Burov et al., 2007). The resulting surface topography, stress regimes, and structure predicted by these latter calculations are identical to those observed above the presumed plume head in eastern Oregon and adjacent Washington (Fig. 8). The combined evidence leads us to conclude that the genesis of the Columbia River Basalt Group is best ascribed to a model of delamination triggered by the northward-propagating Yellowstone plume head.

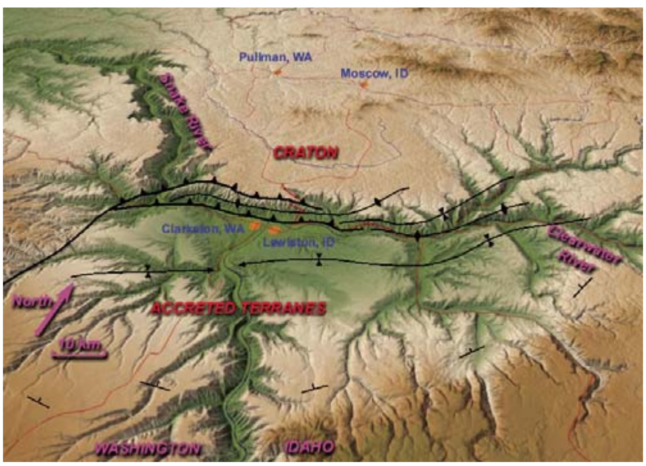

Figure 9. Oblique panoramic view of the Lewiston Basin, with thrust faults and folds of the Lewiston structure marking its northern edge adjacent to the cratonic boundary of North America. Topography courtesy of William Bowen, California Geographical Survey. Geologic structures include thrust and reverse faults denoted as barbed lines, modified from Camp and Hooper (1981) and Garwood and Bush (2005). ID—Idaho; WA—Washington.

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