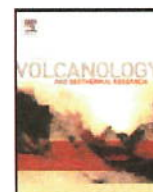




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Mass transfer along the Yellowstone hotspot track I: Petrologic constraints on the volume of mantle-derived magma

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ARTICLE INFO

Article history:

Received 15 May 2008

Accepted 4 April 2009

Available online 17 April 2009

Keywords:

Snake River Plain

Yellowstone

hotspot

rhyolite

continental crust

NCI

ABSTRACT

This paper estimates mantle mass transfer to the crust along the Yellowstone–Snake River Plain (YSRP) hotspot track by applying a simple mass-balance analysis to rhyolite petrogenesis. Mass-balances are based primarily upon distinctive Nd- and Sr-isotopic characteristics of rhyolite, basalt, and pre-existing crust. We present new Nd- and Sr-isotopic data for basalts that are coeval with hotspot volcanism. Many of these basalts exhibit moderate to strong crustal interaction having ϵ_{Nd} as low as -18 and $^{87}\text{Sr}/^{86}\text{Sr}_i$ of 0.715 , while others overlap petrographically, compositionally and isotopically with primitive post-hotspot olivine tholeiites ($\epsilon_{\text{Nd}} = -4.4 \pm 1.3$; $^{87}\text{Sr}/^{86}\text{Sr}_i = 0.7067 \pm 0.001$). We also present new Nd- and Sr-isotopic, petrologic, and U–Pb zircon geochronologic data for YSRP rhyolites associated with Picabo and Heise volcanic fields, some of which occur in 3.2 and 1.6-km-deep boreholes.

Rhyolites of the central and eastern YSRP have clustered and primitive ranges of ϵ_{Nd} (-8 ± 2) and $^{87}\text{Sr}/^{86}\text{Sr}_i$ (0.712 ± 0.003) composition, in contrast to heterogeneous Archean country rocks ($\epsilon_{\text{Nd}} = -23$ to -52 and $^{87}\text{Sr}/^{86}\text{Sr}_m = 0.702$ to 0.893). Using these values for country rock and basalt, we calculate a conservative Neodymium Crustal Index of ~ 0.4 for YSRP rhyolite. We present a simple mass-balance model of rhyolite petrogenesis which budgets sequentially for mantle-derived mass within erupted rhyolites, and cogenetic pluton and restite masses. Our mass-balance estimates indicate that the amount of mantle-derived mass added to the crust between 11 and 4 Ma is equivalent to a layer of gabbro ~ 14 -km-thick or $\sim 340,000 \text{ km}^3$. The corresponding average mass flux of $0.05 \text{ km}^3/\text{yr}$ is comparable to many other global hotspot systems.

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1. Introduction

It is widely recognized that intermediate and silicic magmatic systems may contain substantial fractions of mass derived cogenetically from mantle as well as crustal sources (e.g. DePaolo, 1982; Hildreth et al., 1991; Charlier et al., 2005). In such systems mass transfers from the mantle would constitute a significant mechanism of continental crust formation (e.g. Johnson, 1991, 1993; Kay et al., 1992), particularly in association with some LIPs (e.g. Bryan et al., 2002, 2008). Neodymium isotopes have been particularly useful in distinguishing between the mantle- and crustal sources, and in establishing their relative mass contributions (e.g. DePaolo, 1982; DePaolo et al., 1992; Farmer, 1992). However, unambiguous interpretation of the mass balances has generally been limited by the high degree of geochemical complexity of source rock reservoirs in comparison to the amount of analytical data available, or by a lack of sufficient contrast between the respective mass sources (i.e. 'reservoirs').

In this study we seek to improve on previous estimates of mantle–crust interaction by developing and applying a simple mass-balance model to rocks formed in association with the Yellowstone–Snake River Plain hotspot track (YSRP), focusing on the eastern Snake River Plain (Fig. 1). This system of rocks is uniquely well suited for such an analysis. It is young and well exposed, it has produced tens of thousands of km^3 of rhyolite volcanic rocks and is genetically associated with the well-known Columbia River Basalt large igneous province (Hooper, 1997; Camp and Ross, 2004; Bryan and Ernst, 2007). Additionally, there are abundant analytical data for silicic and mafic magmas produced along the hotspot track, as well as pre-existing continental rocks. These data make clear that there are strong isotopic contrasts between pre-existing Archean crust rocks and likely mantle-source regions (Leeman et al., 1985; Wolf et al., 2005; Nash et al., 2006). Finally, there is also a detailed three-dimensional characterization of crustal and upper mantle architecture from previous geological, borehole and geophysical studies (e.g. Braile, 1982; Morgan et al., 1984; Smith and Braile, 1994; Peng and Humphreys, 1998; Stachnik et al., 2008).

Large contributions of mantle-derived mass to voluminous rhyolites of the YSRP hotspot track have previously been proposed (e.g. Hildreth et al., 1991; McCurry et al., 2002; Wright et al., 2002;

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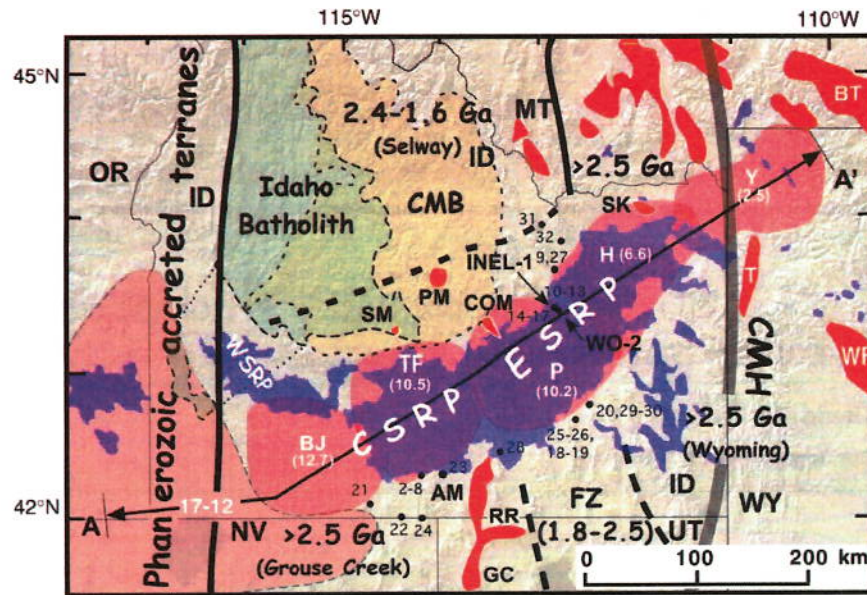


Fig. 1. Digital relief map illustrates the large-scale crustal architecture in the vicinity of the Yellowstone–Snake River Plain hotspot track. Boundaries (dashed where uncertain), names, and ages (in Ga) of inferred Precambrian and Phanerozoic terranes, and between Archean and Proterozoic Precambrian terranes are from Foster et al. (2006). The southern boundary of the Selway terrane has been adjusted to the north of the Pioneer Mountains to accommodate recent identification of Archean rocks in that area (Link et al., 2007). Exposures of Archean rocks are shown in red (AM – Albion Mountains; RR – Raft River Range; GC – Grouse Creek Range; PM – Pioneer Mountains; T – Tetons; WR – Wind River Range; BT – Beartooth Mountains) (Egger et al., 2003; Link and Fanning, 2005; Foster et al., 2006; Mueller and Frost, 2006); Archean xenolith localities are shown in red with white outline (SM – Square Mountain; COM – Craters of the Moon; SK – Spensor–Kilgore area) (Leeman et al., 1985; Wolf et al., 2005; DuFrane et al., 2007). FZ – Farmington Zone of 1.8–2.5 Ga rock separating rocks of the Archean Grouse Creek and Wyoming cratons (>2.5 Ga). CSRP – central Snake River Plain; ESRP – eastern Snake River Plain; WSRP – western Snake River Plain; OR – Oregon; ID – Idaho; MT – Montana; WY – Wyoming; NV – Nevada; UT – Utah. Mid-Miocene to Quaternary volcanic fields of the Yellowstone–Snake River Plain are shown with a pink color; time of inception of silicic volcanism is shown in Ma. From southwest to northeast they include widely dispersed 17–12 Ma silicic volcanics; BJ – Bruneau–Jarbridge; TF – Twin Falls; P – Picabo; H – Heise; Y – Yellowstone Plateau (Pierce and Morgan, 1992; Christiansen et al., 2002; Nash et al., 2006; Brueseke et al., 2008). Blue color depicts Quaternary olivine tholeiite lavas (Kuntz et al., 1992; Christiansen et al., 2002). CMH – Cordilleran Miogeoclinal Hinge Line (Link et al., 1993). Black dots represent sampled localities (Appendix A). A–A' is the location of the profile shown in Fig. 3.

Nash et al., 2006; Leeman et al., 2007; Stachnik et al., 2008; Bonnicksen et al., 2008; Christiansen and McCurry, 2008). We extend that work by (1) better defining a three-dimensional geochemical characterization of hotspot related rhyolite and coeval basalt of the eastern Snake River Plain, (2) refining estimates of the ranges and means of Nd- and Sr-isotopes of likely mantle and crustal mass sources, and (3) developing a simple mass-balance model, based upon a petrologically and geophysically constrained conceptual model, that integrates masses of volcanic rocks with inferred masses occurring in coeval plutons and crustal restite reservoirs. The intent is to provide a reasonable petrologic estimate of mantle-derived mass transferred to the continental crust.

2. Geologic setting and crustal architecture

The YSRP is a ~700-km-long time-transgressive track of voluminous silicic volcanism that crosscuts Basin and Range faulted late Precambrian to Cenozoic age sedimentary rocks (Fig. 1). Origin of the track is not clear (e.g. Pierce and Morgan, 1992; Camp, 1995; Christiansen et al., 2002; Jordan et al., 2004; Hales et al., 2005), but it is likely genetically linked to eruption of the Columbian River Basalt Province (CRB) (Geist and Richards, 1993; Camp and Ross, 2004), and is currently linked to a mantle plume that roots from the transition zone and is currently located beneath the Yellowstone caldera (Smith and Braile, 1994; Yuan and Dueker, 2005; Schutt and Dueker, 2008; Stachnik et al., 2008; Schutt et al., 2008).

Silicic volcanism initiated at ~17 Ma in a diffuse zone (Christiansen et al., 2002; Brueseke et al., 2008) that overlaps in time and space with southern parts of the CRB large igneous province (Camp et al., 2003; Brueseke et al., 2007). Voluminous CRB-related basalts and subordinate rhyolites both issued from widely distributed vents located near or to the west of a major lithospheric suture zone between Archean

craton and Phanerozoic accreted terranes. While basalt volcanism waned rapidly (Brueseke et al., 2007), rhyolite volcanism continued, producing the northeast-trending YSRP track of large, diachronous volcanic fields (Fig. 1). Each field remained active for ~2–4 m.y., and each produced ~5 to 10×10^3 km³ of distinctive Snake River-type (SR-type) rhyolite deposits (Leeman 1982a,b; Morgan et al., 1984; Pierce and Morgan, 1992; Morgan and McIntosh, 2005; Nash et al., 2006; Branney et al., 2008; Bonnicksen et al., 2008). A coeval but volumetrically lesser track of silicic volcanism also extends to the northwest across the high lava plains of eastern Oregon (e.g. Jordan et al., 2004; Streck and Gruner, 2008). Minor olivine tholeiite lavas overlap with the rhyolites in some areas (e.g. Bonnicksen et al., 2008). Early, spatially diffuse eruptions of rhyolites transitioned towards the northeast, along the YSRP volcanic track, into a ~80-km-wide, physiographically well-defined structural downwarp (Kirkham, 1931) referred to as the eastern Snake River Plain (ESRP).

Structural analysis of tilted rocks north of the ESRP indicate that the surficial rocks have been downwarped 5–6 km beneath the current plain surface (McQuarrie and Rodgers, 1998; Rodgers et al., 2002). Most of the downwarp was filled in by coeval rhyolite and possibly by comagmatic granitic intrusions.

Minor olivine tholeiite volcanism postdates and overlaps spatially with rhyolite volcanism. In the eastern Snake River Plain these basalts accumulated at a nearly constant rate over the last 4 m.y. to a maximum thickness of 1–2 km (Kuntz et al., 1992; Champion et al., 2002). Small volumes of young (<2 Ma) olivine tholeiite lavas also erupted across an arc that overlaps large parts of the western and eastern tracks of silicic volcanism (Fig. 1) (Christiansen et al., 2002; Jordan et al., 2004).

Young volcanic rocks and sediment cover and obscure precise location of the Archean–Phanerozoic boundary along the YSRP hotspot track. However, spatially distinct patterns in the isotopic composition of Mesozoic plutonic and Tertiary volcanic rocks allow

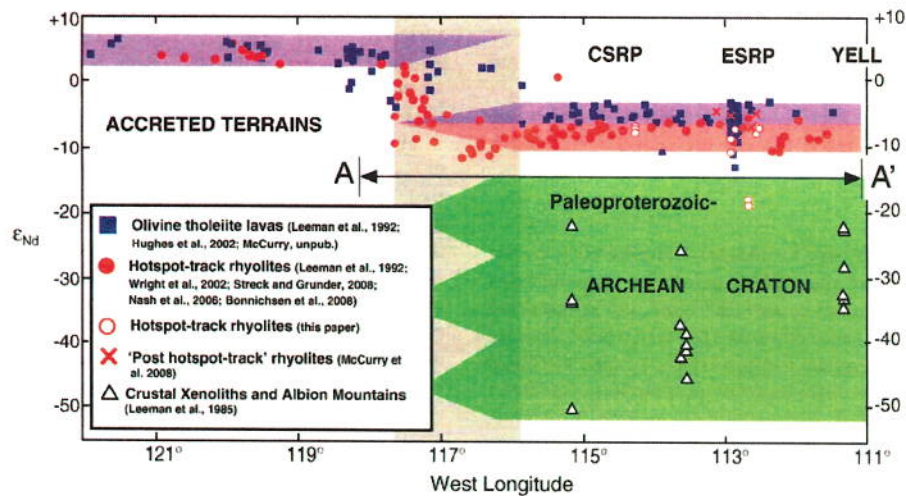


Fig. 2. Plot of Nd-isotopic compositions of mid-Miocene to Quaternary rhyolites and basalts in a transect that parallels the Yellowstone–Snake River Plain hotspot and crosses the Phanerozoic–Archean suture zone (modified after Nash et al., 2006). Also plotted are 18 samples of Archean rocks collected as xenoliths from Quaternary basalts and from surface exposures (Leeman et al., 1982a). The transect A–A' is illustrated in Fig. 1.

the margin to be limited to a north-trending zone between ~116 and 117° west longitude (Fig. 2) (Farmer, 1988; Leeman et al., 1992; Wright and Wooden, 1991; King et al., 2004). Evans et al. (2002) describe

geophysical gradients in that region that are consistent with interpretations of the geochemical data. Presence of buried Archean craton east of this longitude is also indicated by zircon geochronology of crustal xenoliths in late Pliocene to Holocene basalts and inherited phenocrysts in Tertiary rhyolites (Doherty et al., 1979; Wolf et al., 2005; DuFrane et al., 2007) and from surface exposures of gneisses located north and south of the Snake River Plain (Egger et al., 2003; Link and Fanning, 2005; Link et al., 2007). Fig. 1 illustrates a recent model of the craton by Foster et al. (2006).

Three-dimensional crustal and mantle architecture of the Snake River Plain–Yellowstone region has been a subject of detailed geophysical work (e.g. Greensfelder, 1981; Braile, 1982; Sparlin et al., 1982; Greensfelder and Kovach, 1982; Chiang and Braile, 1984; Josten and Smith, 1997; Peng and Humphreys, 1998; Schutt et al., 2008; Stachnik et al., 2008). These studies define a distinctly layered crust that is ~42-km-thick beneath the eastern Snake River Plain (Fig. 3). Regions marginal to the plain are 42 to 44-km-thick and include a ~22-km-thick high-density (3.0 g/cm³) lower crust, an upper crust consisting of a ~17-km-thick felsic layer (2.67 g/cm³), and ~3 km of late Precambrian to Mesozoic sedimentary rock. We infer that this layering structure also originally extended beneath the eastern Snake River Plain, prior to crustal modification by the YSRP hotspot.

Crustal layers beneath the ESRP are distinguished by a prominent, 80-km-wide, ~10-km-thick region of anomalously dense (~2.88 g/cm³) middle crust from ~20 to 10 km depth, and a low-density layer (2.5 g/cm³) from ~5-km-thick consisting of ~1 km of basalt interlayered with sediments, and 2–4 km of dense rhyolite. A deep borehole located within the ESRP, INEL-1, penetrates through 0.65 km of interbedded basalt and sediment, and then into >2.5 km of rhyolite (Doherty et al., 1979; Morgan et al., 1984). Remnants of Precambrian to Mesozoic sedimentary rocks may occur at depth between buried calderas in the ESRP (Josten and Smith, 1997).

Braile (1982) demonstrates that the features defined in the NW–SE cross-section of the ESRP are similar along the axis of the SRP, at least between 112 and 114° longitude (approximately between Yellowstone and Twin Falls volcanic fields).

3. Nd-isotopic characteristics of YSRP rhyolite, crust and subcrustal mantle

Our principal interest is in distinguishing bulk mass transfers between the mantle and crust. DePaolo et al. (1992) propose an Nd-isotope-based index (NCI = Neodymium Crustal Index) as an effective way of discerning

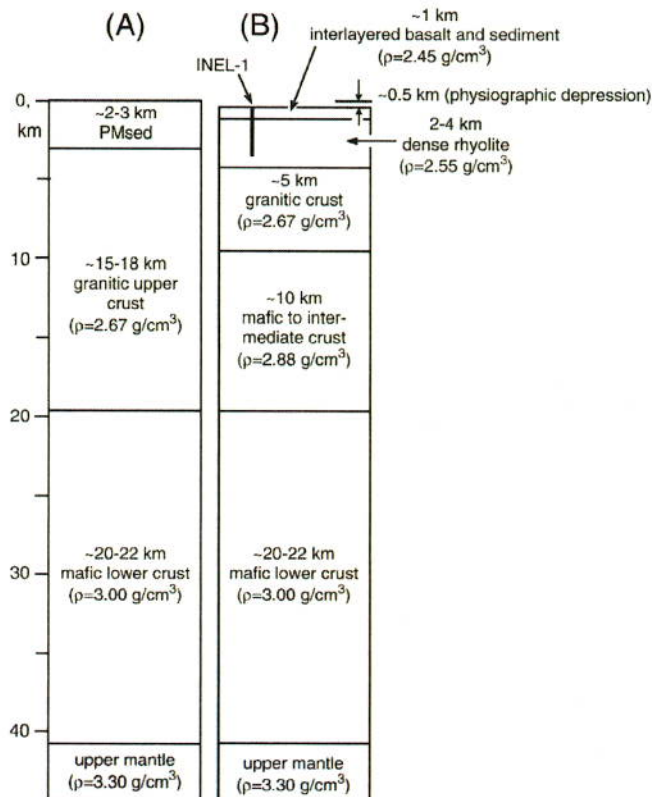


Fig. 3. Diagrammatic representations of crustal layers: A – regions adjacent to the northern and southern margins of the ESRP; B – directly beneath the ESRP. Representative ranges of layer thicknesses; densities and inferred lithologies are selected from Greensfelder (1981), Sparlin et al. (1982), Chiang and Braile (1984), Whitehead (1992), Smith and Braile (1994), Josten and Smith (1997), Peng and Humphreys (1998), and Stachnik et al. (2008). Phanerozoic sedimentary rocks either thin beneath the ESRP (Sparlin et al., 1982) or are broken up by caldera structures (Sparlin et al., 1982; Josten and Smith, 1997). Depth of borehole INEL-1 is illustrated in (B) (Doherty et al., 1979). PMsed – late Precambrian to Mesozoic sedimentary rocks ($\rho = 2.60$ g/cm³; Sparlin et al., 1982).

3.1. Subcrustal mantle

Interestingly, Hanan et al. (2008) argue that the basalts are derived mainly from a CRB-like plume source, but that their heavy isotope ratios have been effectively overprinted east of the craton boundary by radiogenic-element-rich Precambrian mantle lithosphere. They also argue, primarily on the basis of Pb-isotopic variations, for a spatially systematic change in mantle lithosphere consistent with younging from Archean in the east to Proterozoic in the west. However, in our data we see little evidence for spatially or temporally systematic

Of the twelve mafic flows sampled, half are compositionally diverse (olivine tholeiite, ferrobasalt and basaltic andesite) and have low to very low ϵ_{Nd} (-7 to -18) and weakly to strongly elevated $^{87}\text{Sr}/^{86}\text{Sr}_i$ (0.7074 to 0.7123) (Fig. 4). It is beyond the scope of this study to define the petrogenesis of this complex group of rocks. However, their evolved Nd- and Sr-isotopic ratios clearly indicate a comparatively large amount of assimilation of Precambrian crust by plausible mantle-derived primitive magmas, than is the case for other mafic SRP lavas. Leeman (1982c) describes similar isotopic characteristics

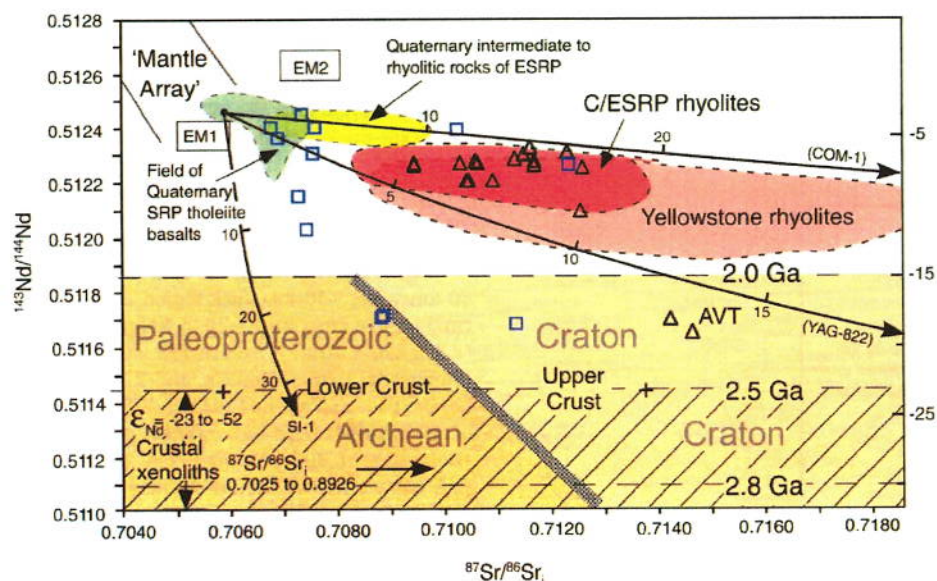


Fig. 4. Nd- and Sr-isotope ratios for volcanic and crustal rocks along central and eastern parts of the Yellowstone–Snake River Plain hotspot track. ‘Mantle Array’ and EM1 and EM2 represent isotopic compositions of possible mantle types (Zindler and Hart, 1986). Colored field represent approximate ranges of isotopic variation for Pliocene to Quaternary basalts (dark green; 128 analyses from Leeman et al., 1985; Hughes et al., 2002b; Hildreth et al., 1991; and Hanan et al., 2008). Quaternary intermediate to silicic rocks of the eastern Snake River Plain (yellow; 10 analyses, from McCurry et al., 2008). Tertiary rhyolites of the central and eastern Snake River Plain (dark red; 126 analyses from Wright et al., 2002; Bonnichsen et al., 2008; Christiansen and McCurry, 2008) and Yellowstone (light red; 20 analyses from Hildreth et al., 1991). New data are illustrated for Tertiary basalts occurring along the northern and southern boundaries of the Snake River Plain (squares) and for Tertiary rhyolites (triangles); AVT = Arbon Valley Tuff (e.g. Kellogg et al., 1994). Horizontal dashed lines represent model isotope ages for ‘upper’ and ‘lower’ Precambrian crust (DePaolo, 1988). Crosses represent analyses of crustal rocks and xenoliths (Leeman et al., 1985); 16 additional $^{143}\text{Nd}/^{144}\text{Nd}$ and $^{87}\text{Sr}/^{86}\text{Sr}$ analyses range from 0.5115 to 0.5100 and 0.703 to 0.893 respectively. Curved lines represent simple mixing lines between selected ESRP olivine tholeiite and Archean rock samples (olivine tholeiite – sample 2-2A-54, from Hughes et al., 2002; Archean samples SJ-1, YAG-822 and COM-1, are from Leeman et al., 1985). Tic marks indicate the mass fractions of Archean mixing components. The curves illustrate possible isotopic ranges for hypothetical mixtures of mantle-derived mafic magma and pre-existing Archean crust.

(for Sr-isotopes) for chemically evolved mafic rocks elsewhere along the margins of the SRP, and suggests that magmas produced from vents near the margin of the SRP could have more readily interacted with Archean rocks. Similar types of chemically evolved rocks also occur along axial parts of the ESRP, but yield little or no evidence of crustal interaction (McCurry et al., 2008; Leeman, 1982c). Paucity of such interaction for post-hotspot basalts extruded within the SRP (Shervais et al., 2006; Hughes et al., 2002a,b) is likely the result of earlier basification of the crust in that area (e.g. Farmer et al., 1991; Peng and Humphreys, 1998).

The other half of the mafic lavas analyzed are olivine tholeiites that have whole-rock major and trace element compositions, and Nd- and Sr-isotopic compositions that overlap with those of post-hotspot-track olivine tholeiites (Fig. 4). This observation suggests that no fundamental change occurred over time in mantle-source regions or melting processes related to basalt production.

A histogram illustrates the ϵ_{Nd} frequency distribution for 136 basalt samples from CSRP, ESRP, and Yellowstone Plateau volcanic field (Fig. 5). Although some samples are derived from different parts of stratigraphically correlated lava flows (e.g. Hughes et al., 2002b), the majority are from different lava flows. Basalts having $\epsilon_{\text{Nd}} < -7$ are inferred to be crustally contaminated and are excluded from our estimate of plausible mantle ϵ_{Nd} . Plausibly isotopically primitive basalts approximate a normal distribution having mean and median values of $\epsilon_{\text{Nd}} -4.4 \pm 1.3$ (1σ) and -4.6 , respectively. We select the median value of -4.6 as the more robust and preferred value. This preferred value is inferred to be representative of large-scale spatially averaged, mantle-derived basalt. Contributions from isotopically distinct mantle sources (e.g. Hatan et al., 2008) are not distinguished.

3.2. Rhyolite

This section evaluates the Nd-isotopic characteristics of YSRP rhyolites. Our objective is to derive a representative ϵ_{Nd} value for rhyolites erupted over those parts of the hotspot track that occur to the east of the Archean–Phanerozoic suture zone (Fig. 1).

Nd- and Sr-isotopic compositions of YSRP rhyolites are illustrated in Fig. 4. The figure demonstrates the marked distinction between the rhyolites and plausible Archean country rocks. Nd- and Sr-isotopic compositions for all but a few rhyolites plot much closer to mantle than to plausible Archean crustal compositions. Also, ranges of ϵ_{Nd} and $^{87}\text{Sr}/^{86}\text{Sr}$ for the majority of rhyolites of -4 to -11 and 0.708 to 0.720 are much smaller than those of the Archean crust (-23 to -52 and 0.7025 to 0.893).

Previous work identifies spatially and temporally systematic variations in YSRP rhyolites (e.g. Nash et al., 2006; Perkins and Nash, 2002; Hughes and McCurry, 2002; Bonnicksen et al., 2008; Christiansen and McCurry, 2008). Fig. 4 illustrates that ranges of ϵ_{Nd} and $^{87}\text{Sr}/^{86}\text{Sr}$ for the rhyolites increase and decrease, respectively, from central to eastern parts of the YSRP. However the least evolved parts of the ranges are nearly constant, having ϵ_{Nd} and $^{87}\text{Sr}/^{86}\text{Sr}$ of ~ -7 and 0.709 , respectively.

Of particular importance to this study, Nash et al. (2006) document temporally and spatially systematic variations in ϵ_{Nd} (Fig. 2). They demonstrate that, for YSRP rhyolites erupted from vents east of the craton margin, ϵ_{Nd} varies in a wave-like fashion from -10 in the west (BJ area, Fig. 1), to ~ -7 in the CSRP, and back to ~ -10 from Heise to Yellowstone Plateau (>1 Ma). These variations appear compelling. However we believe that they are small enough, in comparison to the difference between the rhyolites and Archean country rocks, that they can be neglected for the purposes of this study.

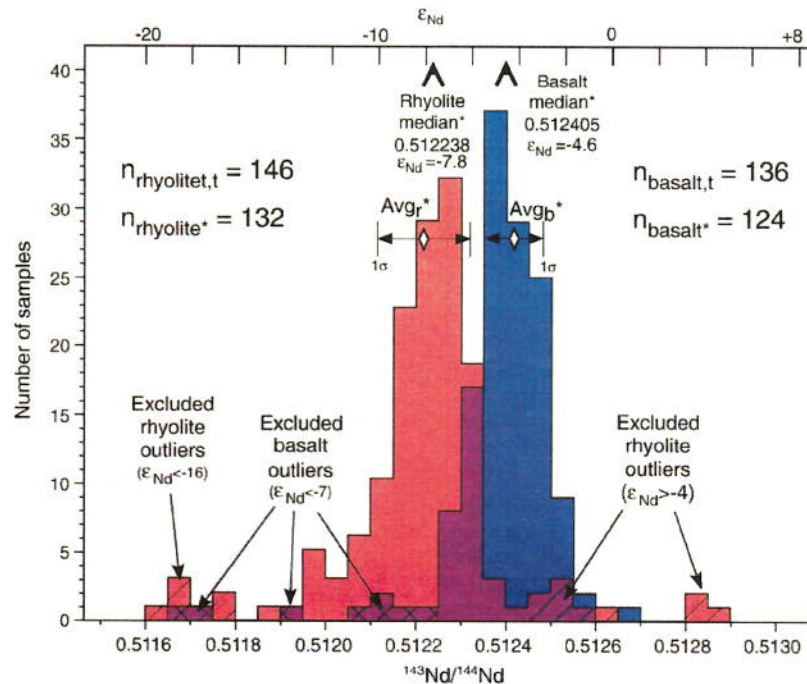


Fig. 5. Histograms representing ranges of Nd-isotopic composition for Yellowstone–Snake River Plain volcanic rocks located east of 116° longitude (from Leeman et al., 1985, 1992; Hildreth et al., 1991; Wright et al., 2002; Hughes et al., 2002b; Bonnicksen et al., 2008; Christiansen and McCurry, 2008; this study). Rhyolites – red (146 analyses); basalts – blue (136 analyses); purple color is overlap between basalt and rhyolite samples. Outliers are identified with arrows; values calculated without outliers are shown with an asterisk. Rhyolite outliers having $\epsilon_{\text{Nd}} > -4$ are fallout tuff deposits (Nash et al., 2006), and are inferred to have been derived from west of 116° longitude. Many rhyolite outliers analyses having $\epsilon_{\text{Nd}} < -16$ are of the same unit (Arbon Valley Tuff; Kellogg et al., 1994). This unit is anomalous in a wide range of its compositional and volcanological characteristics, and it is therefore excluded from the calculated rhyolite average and median. Basalts having $\epsilon_{\text{Nd}} < -7$ are crustally contaminated. Median values for rhyolite and basalt are shown with a hat symbol. Averages and 1σ ranges are shown with diamond symbols and error bars (Avg_r* – rhyolite average; Avg_b* – basalt average). Median values are taken to be representative of the respective sample populations.

A key assumption in our adopting a characteristic ϵ_{Nd} for rhyolite is that we have obtained a representative sample of the silicic hotspot magmas. However, as in the case for the basalts, there is a built-in bias for our existing data set because, in this case, we only have access to ignimbrite outflow sheets along the margins of the SRP, distal fall deposits and post-caldera lavas. Proximal deposits produced during the initial stages of volcanic field formation have the potential to show lower NCI values, since they formed in the early stages of field formation when the crust was cold and difficult to assimilate (e.g. Perry et al., 1993). However, these deposits are inaccessible owing to cover by post-caldera rhyolites and basalt lavas. It is plausible that proximal deposits, particularly intracaldera facies of rhyolites may be systematically biased in composition from those that have been previously documented (e.g. Valentine et al., 1992). We test this hypothesis by describing new geochemical, petrologic and geochronologic data from two deep boreholes, INEL-1 (3.2 km) and WO-2 (1.52 km), which yield unique windows into such rhyolites in the ESRP (Doherty et al., 1979; Morgan et al., 1984; Mazurek, 2004). We also better refine the age and geochemistry of rhyolites associated with the poorly defined 'Picabo' volcanic field.

Borehole INEL-1 penetrates 650 m of olivine tholeiite and then another 2.5 km of rhyolite. Doherty et al. (1979) document 14–15 layers of dense crystalline rhyolite (lavas or welded tuff) interbedded with minor nonwelded tuff or sediment. Borehole lithologies and petrography are documented by Doherty et al. (1979) and McBroome (1981). Some additional petrologic features are documented in Appendix B.1 and B.3. Whole-rock chemical and isotopic compositions and phenocryst modes of core from depths of 0.74, 1.15, 1.50 and 3.15 km are recorded in Appendices A and B.3. Rhyolites from the deepest parts of the borehole (between 1.15 and 3.15 km) have narrow ranges of ϵ_{Nd} (−6.2 to −7.3) and $^{87}\text{Sr}/^{86}\text{Sr}_i$ (0.7117 to 0.7123).

SHRIMP zircon dates from the four cored intervals from INEL-1 yield overlapping ages of ~8.2 Ma (Appendix B.1). Significantly, these data indicate that at least 2 km of rhyolite volcanic rocks, greater than any thickness identified along the margins of the SRP, accumulated in less than a few hundred thousand years, and at a time and location that correlates with the inferred Picabo volcanic field (Pierce and Morgan, 1992). This thick sequence of coeval rocks is likely to be a rare and perhaps unique sample of intracaldera volcanic facies from CSRP and ESRP segments of the YSRP.

Borehole WO-2 is located 5 km southeast of INEL-1. The 1.52 km-deep borehole was cored with almost 100% core recovery (Mazurek, 2004). It penetrates 1.17 km of basalt and interbedded sediment (e.g. Shervais et al., 2006), and another 0.35 km of rhyolite (Appendix B.2). Rhyolite units include two thin welded ignimbrites (WT 1 and 2), a 200-m-thick lava flow, and another welded tuff of unknown thickness (WT3) (Appendix B.2). Anders et al. (1997) record dates of 6.12 and 6.38 Ma for one of the upper ignimbrites and the lava flow, respectively, indicating that those rocks are coeval with the Heise volcanic field (Morgan et al., 1984; Morgan and McIntosh, 2005). However, with one possible exception, they do not appear to correlate with known fall or ignimbrite deposits marginal to the SRP (Morgan and McIntosh, 2005). Morgan and McIntosh (2005) suggest that possible explosive precursor eruptions associated with the lava may be correlative with ash beds exposed near Palisades Reservoir (~150 km east-southeast of WO-2). The thin stratigraphic thickness of welded tuffs WT-1 and 2 (100 and 30 m, respectively) suggests that they are outflow rather than intracaldera ignimbrite facies.

Composition and petrography of rhyolites from WO-2 are included in Appendices A and B.3. Nd- and Sr-isotopic compositions of the rhyolites overlap with, but are slightly more evolved than the majority of SR-type rhyolites, having ranges of ϵ_{Nd} of −8.3 to −10.5 and $^{87}\text{Sr}/^{86}\text{Sr}_i$ of 0.710 to 0.7123.

In summary, rhyolites examined in boreholes INEL-1 and WO-2 overlap in their lithologic, petrographic, geochemical and isotopic compositions with similar age rhyolites documented in regions

marginal to the ESRP (cf. Hughes and McCurry, 2002; Nash et al., 2006; Christiansen and McCurry, 2008). We therefore conclude that there is no strong systematic bias between intracaldera and outflow facies of rhyolites buried at depth along the axis of the SRP (at least to the depths penetrated within the boreholes), and those occurring along the margins of the SRP.

A histogram (Fig. 5) illustrates a frequency distribution of ϵ_{Nd} for 146 rhyolite samples from parts of the YSRP located east of 116° longitude. Approximately 6 samples are duplicates from different parts of stratigraphically correlated units. Some others are also likely duplicates, but the number is unknown owing to lack of a regional correlation of many ignimbrites, particularly in the CSRP. We believe that the majority of analyses are from stratigraphically distinct rhyolite units.

A small number of outliers occur in the rhyolite histogram (Fig. 5). At least three of those with $\epsilon_{\text{Nd}} < -15$ are of the Arbon Valley Tuff (Kellogg et al., 1994). This ignimbrite has many characteristics that distinguish it from other SRP rhyolites (e.g. Kellogg et al., 1994), and it may not be representative of processes associated with other rhyolites of the YSRP hotspot track (e.g. Nash et al., 2006). Rhyolite outliers having $\epsilon_{\text{Nd}} > -4$ are fallout tuff deposits (Nash et al., 2006), and are inferred to have been derived from west of 116° longitude. Excluding those samples from the histogram yields a near-normal distribution having a mean ϵ_{Nd} of -8.3 ± 1.9 (1 σ) and a median value of −7.8 (Fig. 5). The median value is taken as our estimate for the total average mass of rhyolite produced along the hotspot over the last 12 m.y.

3.3. Pre-YSRP hotspot track crustal architecture

A final key assumption of our mass-balance analysis is that we can define a meaningful average Nd-isotopic composition for the region now overprinted by the YSRP hotspot track. Defining such a value in a complex Precambrian terrane may seem an insurmountable task given the possible scales and magnitudes of heterogeneity in such rocks (e.g. Chamberlain et al., 2003). One approach is to assume a limiting condition in which isotopic distinctions between contributing mantle and crustal mass reservoirs are minimized. Nash et al. (2006) applied this approach with the idea of obtaining a minimum limit of mantle contribution to the rhyolites (NCI < 0.6 to 0.8). However, we believe that there is now sufficient additional data to improve on that estimate.

Sources of evidence for the geochemical architecture of the craton across southern Idaho include (1) exposures of Archean rocks (e.g. Chamberlain et al., 2003; Foster et al., 2006), (2) crustal xenoliths in Neogene–Holocene mafic volcanic rocks (e.g. Leeman et al., 1985; Wolf et al., 2005; DuFrane et al., 2007), (3) relic zircon geochronology of grains in Cretaceous and Tertiary intrusive rocks (e.g. Egger et al., 2003; Gaschnig et al., 2007), (4) chemical and mineralogical features of YSRP rhyolites (Leeman et al., 1992; Nash et al., 2006), and (5) Nd-, Sr-, Hf-, Pb- and O-isotopic characteristics of Mesozoic intrusions and Tertiary rhyolite (e.g. Farmer, 1988; Wright and Wooden, 1991; King et al., 2004). Defining an age structure for the craton is important for this study because there is generally a strong correlation between craton age and ϵ_{Nd} (DePaolo, 1988) (Fig. 4). Foster et al. (2006) compiled these data and produced a new model of Archean crustal structure focused in part on the vicinity of the YSRP hotspot track (Fig. 1). They demonstrate that most of the hotspot track is underlain by craton ≥ 2.5 Ga. They also argue that there are likely two Archean terranes, the Wyoming and Grouse Creek terrane, separated by a zone of Paleoproterozoic rocks that they define as the 'Farmington zone' (FZ, in Fig. 1).

Interestingly, the Farmington zone underlies that part of the ESRP having the highest ϵ_{Nd} . We speculate that this would provide an alternate explanation for the high in ϵ_{Nd} of YSRP rhyolites described in that area by Nash et al. (2006). The observed 2–3 epsilon unit increase in ϵ_{Nd} for YSRP rhyolites along that part of the hotspot track could be a

result of a mantle-derived magma interaction with a younger, less radiogenic crust in that area, rather than to changes in magma flux or relative motions of mantle plume and the crust (cf. Nash et al., 2006).

Exposures of Archean age rocks, and in younger intrusive rocks containing inherited Archean zircon grains, occur in metamorphic core complexes to the south and north of the SRP at ~114° longitude (Fig. 1). In contrast to simple post-Archean zircon evolution of the crustal xenoliths, there are multiple populations of zircon grains that include late Archean and also Proterozoic ages, which may record accretion of Archean and Paleoproterozoic terranes (Foster et al., 2006; Mueller and Frost, 2006). Detrital sediment derived from the Albion Mountains includes zircon grains as old as 3.3 Ga (Link and Fanning, 2005). In-situ sampling of orthogneiss yielded an Rb/Sr isochron date of 2.4 Ga (Armstrong and Hills, 1967). Samples of similar orthogneiss from the Grouse Creek Mountains yield Archean zircon dates of 2.55 Ga (Egger et al., 2003).

Archean craton extends at least 50 km north of the SRP. Recent work by Link et al. (2007) indicates that 2.6–2.7 Ga orthogneiss is exposed in the Pioneer metamorphic core complex. The boundary between Paleoproterozoic Selway and Archean (Grouse Creek?) terranes must therefore be located well to the north of the YSRP (Fig. 1).

Archean age rocks also appear as xenoliths within Neogene to Holocene mafic lavas occurring across central and eastern parts of the Snake River Plain (Matty, 1984; Leeman et al., 1985) (Fig. 1). They include a spectrum of granulite facies felsic to mafic compositions, and yield zircon dates ranging from 2.5 to >3.5 Ga (Wolf et al., 2005). Thermobarometry by Matty (1984) indicates that the xenoliths were metamorphosed at 700–800 °C at 4 to 8 kb; consistent with derivation from the mid- to upper crust. Samples yield variable Sr- and very low Nd-ratios respectively varying from 0.7024 to 0.8926 and –23 to –52.

Zircon populations within xenoliths young slightly from the northeast to the southwest, decreasing from ~2.7 to 3.1 Ga at SK and COM, and 2.57 at SM (Wolf et al., 2005) (Fig. 1). Wolf et al. (2005) indicate that the zircons have a complicated Archean history, but more importantly for this study, that there is no evidence of zircon growth occurring in these samples between the Archean and late Tertiary. The latter observation would appear to preclude major pre-Tertiary igneous events that could involve major transfers of mantle-derived mass to the crust in this region.

Hanan et al. (2008) argue that discordances between some chemical and isotopic characteristics of SRP basalts require inputs to those lavas from a mixed Archean–Proterozoic mantle lithosphere. Possible additional influences include emplacement of the late Cretaceous Idaho Batholith, and Eocene rocks of the Challis Volcanic Field (Moye et al., 1988a,b; King and Valley, 2001). Although both produced some transfer of mantle-derived mass to pre-existing Paleoproterozoic and Archean crust in those regions (e.g. Fleck, 1990), recent work indicates that those contributions were probably small (King and Valley, 2001; Gaschnig et al., 2007). Additionally, there is no evidence that either the Idaho Batholith or Challis volcanic field extends across the eastern Snake River Plain, nor do they extend across any but the westernmost part of the central Snake River Plain (e.g. Boroughs et al., 2005).

Establishing a mean composition for the Archean crust is problematic owing to the wide range of values and relatively small number of analyses (13). We therefore take a conservative, limiting case by assuming an ϵ_{Nd} (crust) of –20, corresponding to the least isotopically evolved crustal sample of Leeman et al. (1985). Such a value would allow for possible mixed mass contributions from Proterozoic as well as Archean crust.

3.4. NCI (Neodymium Crust Index)

Estimates of average rhyolite NCI for the YSRP are illustrated in Fig. 6. Histograms for Archean crust, basalts and rhyolites are shown

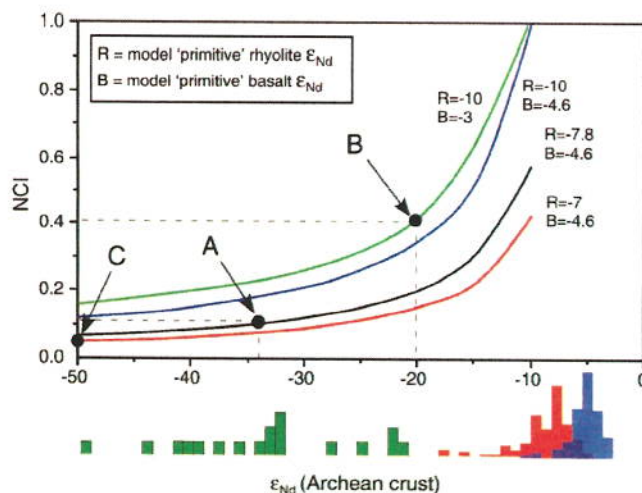


Fig. 6. NCI (Neodymium Crust Index; DePaolo et al., 1992) is the fraction of the rhyolite derived from a binary mixture of crustal and mantle components = $(\epsilon_{\text{Nd,hybrid rock}} - \epsilon_{\text{Nd,mantle component}}) / (\epsilon_{\text{Nd,crustal component}} - \epsilon_{\text{Nd,mantle component}})$. It is plotted against ϵ_{Nd} of a hypothetical crustal component. Histograms below the chart illustrate ranges of ϵ_{Nd} for basalts (representative of the mantle component), rhyolites and Precambrian crust. Curves illustrate ranges of NCI values calculated for plausible ranges of rhyolite and basalt (equals mantle) initial compositions. Point (A) represents a solution where mantle and rhyolite components are given preferred ϵ_{Nd} values from Fig. 5, and the Archean crust is given a median ϵ_{Nd} value of –34, yielding NCI = 0.12. Point (B) represents an estimated upper limit for NCI of 0.4. In this model Archean crust is given the lowest observed value of ϵ_{Nd} = –20; the mantle component has ϵ_{Nd} = –3, and the rhyolite has ϵ_{Nd} = –10 (~1 σ above and below, respectively, the averages shown in Fig. 4). Point (C) represents an estimated lower limit of NCI (0.07) based upon use of the most evolved Archean crust (ϵ_{Nd} = –50); and rhyolite and mantle components having ϵ_{Nd} of –7 and –4.6, respectively.

below the graph. NCI is plotted as a function of ϵ_{Nd} (crust). Separate curves represent plausible ranges of ϵ_{Nd} for rhyolite and basalt components. Our preferred median values of ϵ_{Nd} for basalt and rhyolite, and the median ϵ_{Nd} of Archean crust (–34) yield an NCI of 0.12. Taking the most conservative estimate of Archean crust ϵ_{Nd} of –20, variations in basalt and rhyolite compositions within one-sigma of their statistical means yields NCI estimates of between ~0.07 and 0.4. Subsequent discussion is based on use of the most conservative plausible NCI of 0.4.

4. Conceptual model

Our mass-balance analysis is based upon a conceptual model of YSRP rhyolite petrogenesis adopted from Christiansen and McCurry (2008). The model is illustrated in Fig. 7. In this model magmatism is driven by partial melting of subcrustal mantle in response to the Yellowstone thermal plume (Schutt et al., 2008). Helium and heavy isotopic compositions of the basalts indicate that they were derived at least in part from a sublithospheric mantle source (Graham et al., 2007; Hanan et al., 2008). Primary melts were likely picrite or high-Mg olivine tholeiite having a density of ~2.8 g/cm³. Owing to the high density of the Archean lower crust in this area (ρ = 3.0 g/cm³), the primitive magma is modeled as generally bypassing the Moho and continuing its buoyant ascent until it encounters low-density upper crust (ρ = 2.67 g/cm³), thus intraplating rather than underplating the crust (e.g. Barnes et al., 2002; Charlier et al., 2005; cf. Frost et al., 2001). In effect the upper crust, rather than the lower crust, acted as a primary density filter for the ascending basaltic magmas during this phase of activity.

We prefer intraplating to underplating (i.e. at the Moho) because there is clear geophysical evidence of densification of the middle crust (Sparlin et al., 1982; Peng and Humphreys, 1998), and because Nd- and Sr-isotope AFC and simple bulk mixing require an upper-crust-like reservoir (i.e. crust with high ⁸⁷Sr/⁸⁶Sr) (Fig. 4).

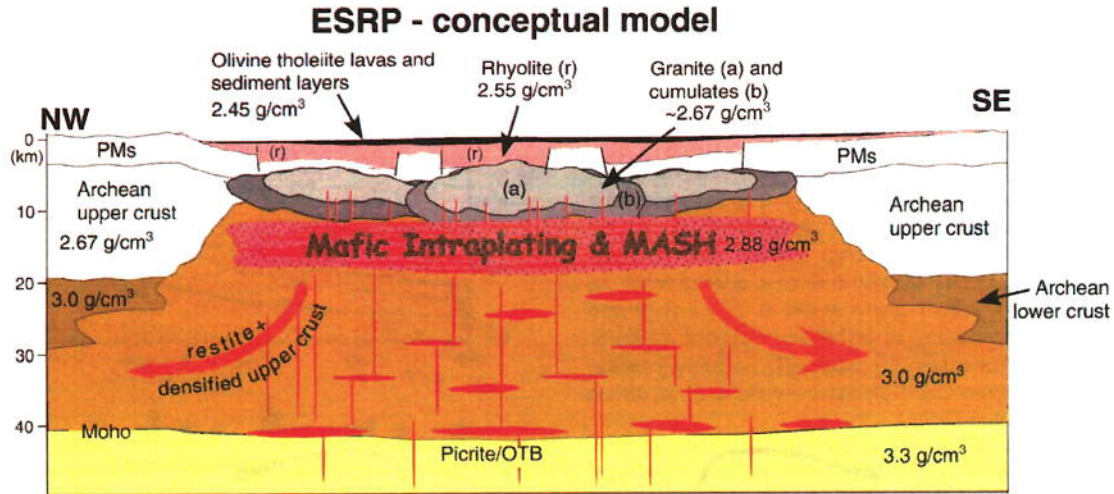


Fig. 7. This diagram illustrates key features of our conceptual model of crustal structures and magmatic processes associated with YSRP magmatism. The NW–SE cross-section is based on geophysical and surficial geologic data for the ESRP. Densities and layer thicknesses are from Fig. 2. PMs – Proterozoic to Mesozoic sedimentary rocks; r – rhyolite. Peak hotspot activity produced a large flux of primitive magma (olivine tholeiite or picrite) from an underlying mantle plume (e.g. Smith and Braile, 1994; Humphreys et al., 2000; Yuan and Dueker, 2005). High original lower-crustal densities allowed for efficient transfer of most of the magma into the lower density upper crust where it is shown producing swarms of dikes and sills. MASH-like processes in that mixing zone result in production of hybrid primitive rhyolite magmas that collect into shallow, high aspect ratio reservoirs. Some vent periodically, commonly producing ~100 to ~1000 km²-scale eruptions and cogenetic calderas. Surficial basalts are produced during waning phases of plume activity (e.g. Leeman, 1982b). MASH-related restite and densified middle crust have negative buoyancy, and are modeled as subsiding ductily into lower-crustal regions marginal to the SRP (e.g. McQuarrie and Rodgers, 1998).

During waxing stages of mantle magma production the mid-crust became a region of MASH-like mixing, assimilation and hybridization (Hildreth and Moorbath, 1988). Fractional melts of the hybridized mid-crust are assumed to have compositions similar to the least evolved SR-type rhyolite.

Conservative estimates of $NCI \leq 0.4$ (Fig. 6) for erupted rhyolite implies that the region in which melting and/or AFC occurred had an effective spatially averaged ratio of $\geq 60\%$ mantle-derived mass and $\leq 40\%$ Archean crust mass. We believe this is an improvement from previous lower limits by Nash et al. (2006) of 20–40% for mantle-

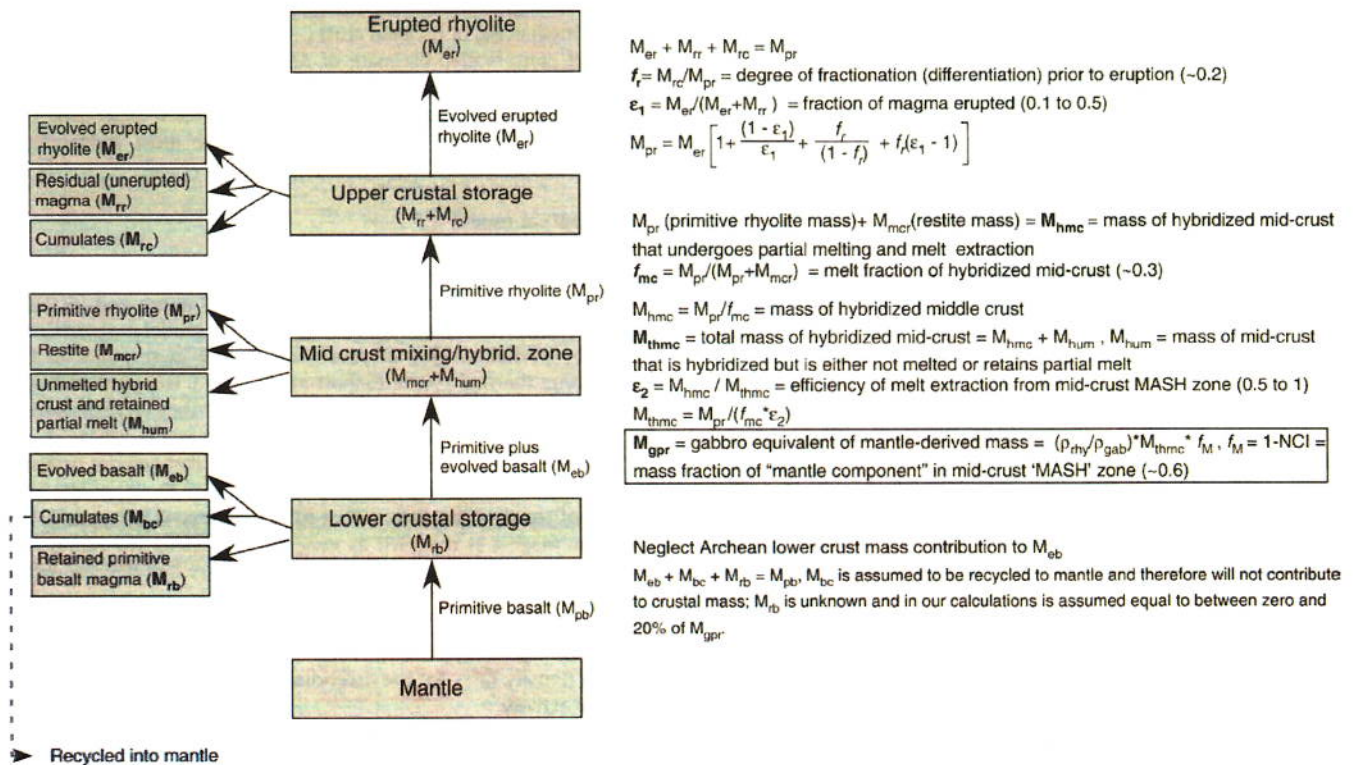


Fig. 8. This flow chart illustrates the design of our mass-balance model and defines our terms and mass-balance relationships. The model is one-dimensional, where masses are expressed in terms of the equivalent thickness of a layer of gabbro with an assumed density of 3 g/cm³, therefore the only length variable is depth (in km). M_{gpr} is the total gabbro equivalent mantle-derived mass (in km). Principle inputs to the model are the thickness of the rhyolite layer, and mass fraction of mantle component in the rhyolite ($f_M = 1 - NCI$).

derived mass contribution. If we assume an initial chemical composition for the Archean crust comparable to 'average' Archean upper crust (Rudnick and Gao, 2003), the hybrid mass (Archean plus gabbro) would approximate the composition of a diorite (Appendix C). 30% melting of such a hybrid source would be needed to produce primitive SR-type rhyolite, leaving a mafic restite with 70% of the original mass. Details of the melting and melt segregation process are almost certainly more complex than this (e.g. Barnes et al., 2002). However, we think it is useful to think of the process in this fashion, since many models involving hybridization followed by partial melting, AFC, or combinations of these and mixing/recharge yield similar mass balances in regards to our Nd-isotope tracer.

During waxing phases of activity, primitive rhyolitic magma is modeled as having assembled into shallow magma chamber systems. Previous existence of large, shallow magma chambers is consistent with caldera formation, which appears to be a characteristic feature of YSRP hotspot track volcanism (e.g. Morgan et al., 1984; Christiansen, 2001; Bonnichsen et al., 2008).

The final stage in this magma transfer system is the eruption of rhyolite. Geophysical data suggest that there may be a layer of rhyolite with an average thickness of 2–5 km (Greensfelder, 1981; Sparlin et al., 1982; Braile, 1982; Chiang and Braile, 1984). Borehole INEL-1 confirms that at least 2.5 km of rhyolite accumulated at that point (Doherty et al., 1979; Morgan et al., 1984), and the McQuarrie and Rodgers (1998) crustal flexure model allows for up to 5 km of rhyolite (plus 1 km of overlying basalt). Volume estimates for individual volcanic fields yield lower limits for mean thicknesses of hundreds of meters, but are severely limited by a lack of access to key proximal parts of the respective fields. We believe a conservative reasonable estimate for mean thickness of the rhyolite layer is ~2 km.

A key unknown in our conceptual model is the characteristic fraction of rhyolite magma that is erupted from a source reservoir. Smith (1979) and Shaw (1985) suggest that ~10% of a chamber is

characteristically erupted. However, Wilson et al. (2006) argue that it is possible to effectively completely evacuate a source chamber. At least some SR-rhyolite supereruptions are followed by eruption of large volumes of cogenetic degassed magma as lava flows (e.g. Hirt, 2002), suggesting that significant volatile-poor or crystal-rich magma is retained during the more explosive eruptions (e.g. Bachmann et al., 2007; Lipman, 2007). From a simple mass-balance view, these details are probably not critical. We think it is highly unlikely that all of the magma that segregates from mid-crustal source regions is delivered by eruptions to the surface. Without being specific about the eruption process or magma reservoir geometry we believe that reasonable estimates of primary rhyolite fraction is 20–30% (cumulates), and also that 20–30% of the differentiated magma is eventually erupted. Thus, the remaining 70–80% remains as unextruded upper crustal granite at depths of 3–10 km.

5. Mass-balance model

Our mass-balance model is illustrated in Figs. 8 and 9. A related spreadsheet is available from the electronic repository Appendix D. Mass of erupted rhyolite (M_{er}), is expressed as a one-dimensional 'thickness'. That mass is then linked in series to cogenetic unerupted parts of the source magma reservoir + cumulates, and mid-crustal restite. We then budget for mantle-derived component of mass in the system based upon the NCI of erupted rhyolites (~0.4). That mass is expressed one-dimensionally as an equivalent layer of gabbro ($\rho = 3 \text{ g/cm}^3$). Curves based on preferred estimates for the pertinent variable are solid; dashed lines illustrate results for ± 0.1 variation of NCI. These best-guess estimates, and reasonable variation of parameters yield a total mantle-derived mass storage in the crust equivalent to a layer of gabbro ~14-km-thick, distributed with erupted rhyolites, comagmatic plutonic and cumulate rocks, and restite mass in the middle crust.

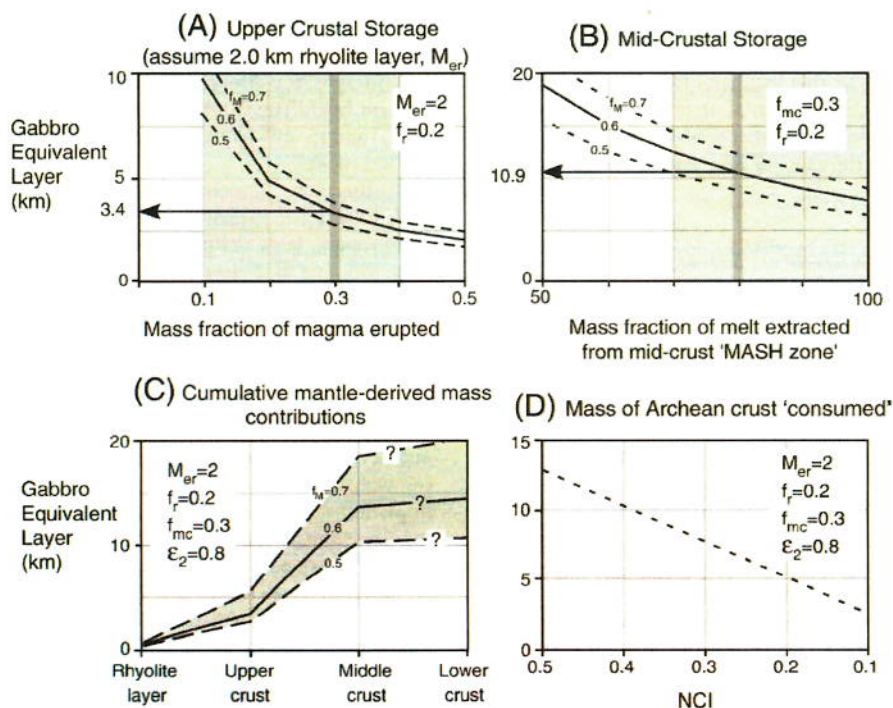


Fig. 9. Plots of solutions to our mass-balance equations given best-guess estimates for the pertinent variables. Mantle-derived mass is expressed as an equivalent layer of gabbro ($\rho_{\text{gabbro}} = 3 \text{ g/cm}^3$). (A) Mantle-derived mass storage in the upper crust in the form of rhyolite and cogenetic plutonic rocks. (B) Mantle-derived mass storage in the middle crust in the form of restite, and hybridized middle crust that either retained partial melt or did not undergo fusion. (C) Cumulative mantle-derived mass in upper, mid-crust and lower-crustal zones. Mass retention in the lower crust is unknown, but is assumed to be 0 to 20% of the total mass (M_{tot}). (D) Effective mass of Archean upper crust that is required to produce calculated hybrid mass. For NCI = 0.3 to 0.4, the amount of Archean crust "consumed" would equate to a thickness of 7.5 to 10 km (assuming $\rho_{\text{Archean}} = 2.67 \text{ g/cm}^3$).

Fig. 9A illustrates step (1) of our mass-balance scenario. The horizontal axis represents the average fraction of magma erupted from upper crustal magma chambers. The chambers are assumed to have undergone ~20% of pre-eruption fractionation (i.e., $f_r = 0.2$). Mass fractions of erupted magma are assumed to be in the shaded range (i.e. 0.1 to 0.4). We use a conservative value of 0.3. Assuming an NCI of 0.4 would yield upper crustal storage of mantle-derived mass equivalent to a layer of gabbro 3.4-km-thick (~1 km of which occurs in the erupted rhyolite layer).

Fig. 9B illustrates mid-crustal mass storage. We assume that mantle-derived magma intraplates within this region in proportions that, when melted, generates an NCI for derivative magmas of 0.4. We estimate melt fractions in this region to be in the range of 0.2 to 0.3 (e.g. Bonnicksen et al., 2008; Christiansen and McCurry, 2008). Modeling of melt fractions greater than 0.3 would require use of an unrealistically felsic Archean crust (e.g. >72% SiO₂; cf. Rudnick and Gao, 2003) in order to satisfy major element mass-balance requirements (Appendix C). Efficiency of melting and melt segregation from the mid-crust are unknown, but we believe it is unlikely to be 100%. That is that some parts of the mid-crust are intruded by basalt and effectively hybridized but either do not melt, or retain partial melt. We conservatively estimate that 80% of the hybridized mid-crust undergoes efficient melt segregation. Lower values yield substantial increases in M_{gppr} . Calculated mass storage in the middle crust, in the form of restite, unmelted hybrid crust and retained partial melt, corresponds to a layer of gabbro 10.9-km-thick.

Fig. 9C illustrates plausible cumulative mantle-derived mass contributions (M_{gppr}) derived from surficial rhyolite and from upper and middle crustal storage zones. In terms of gabbro equivalent layer thicknesses, the rhyolite layer contributes ~1 km, upper crustal storage adds another 2.4 km, and mid-crustal storage another 10.9 km. Contributions of mass to the lower crust are unknown, but would add to the total mass, excepting for dense restite or cumulate components that would likely cycle back into the mantle (Fig. 8). We assume minimal contribution of mass occurs to the lower crust. Variations of NCI from 0.3 to 0.5 yield masses equivalent to layers of gabbro from ~11 to 18-km-thick, and a best-guess estimate of 14 km. Post hotspot-track basalts obviously add to this estimate. However, because the average thickness of the basalts (~100s of meters) is much less than the rhyolite (2–5 km), we believe that that contribution is well within the range of uncertainty for the rhyolite data, and it is not included in this analysis.

6. Discussion

6.1. Why is our NCI different?

Nash et al. (2006) propose conservative upper limits of NCI from 0.8 to 0.6 for most YSRP rhyolites produced east of latitude 116° longitude (the focus of our study). They base their estimate in part on the assumption that ε_{Nd} of the crustal component resembles Proterozoic Selway terrane ~250 km northwest of the SRP ($\varepsilon_{\text{Nd}} = -11$ to -15). However, we argue that it is more likely that the SRP is underlain by Archean-dominated rather than Proterozoic-dominated crust. It is clear from recent zircon work in the Pioneer core complex that Archean crust, or components of it, is located north of the SRP (Link et al., 2007) (Fig. 1). We believe that a more plausible conservative limit of ε_{Nd} is -20 , and that the mean value may be substantially lower than that.

Secondly, Nash et al. (2006) infer a mean mantle isotopic composition of $\varepsilon_{\text{Nd}} = -2$. This value, referenced from Leeman et al. (1992), is based upon the highest ε_{Nd} number they recorded for basalts located east of 116° longitude. However recent work (e.g. Hughes et al., 2002a; Hanan et al., 2008) indicates that, with the exception of obviously contaminated outliers, most of the basalt isotopic variability is a signature of heterogeneity of mantle-source

regions, rather than crustal interaction. We conclude that, excluding the outliers ($\varepsilon_{\text{Nd}} < -6$), a mean of the population of basalt ε_{Nd} values (i.e. $\varepsilon_{\text{Nd}} = -4.6$) is a better choice than the limiting value ($\varepsilon_{\text{Nd}} = -2$) for the mantle component used in calculating NCI.

Nash et al. (2006) recognize the possibility that NCI may be lower than their estimated upper limit of 0.6 to 0.8, and could be as low as 0.4. Given the observed ranges of basalt, rhyolite and crust for the YSRP, we think that an NCI of 0.8 to 0.6 is unlikely, and the CSRP and ESRP rhyolite NCI is more likely ≤ 0.4 .

How do NCIs of YSRP rhyolites compare with rocks elsewhere in the region and the world? DePaolo et al. (1992) and Perry et al. (1993) document NCIs for a wide variety of Cretaceous and Tertiary rhyolites and granites across western North America. NCI values vary from ~0.1 to 1. Interestingly, Cretaceous to Oligocene values are significantly higher (0.6 to 1.0) than late Tertiary to Quaternary values (0.1–0.3). Trumbull et al. (1999) document NCIs of 0.1 to 0.6 for intermediate to silicic, Oligocene to Quaternary volcanic rocks of the central Andean volcanic zone. NCIs of YSRP rhyolites therefore appear well within the range of NCIs for rhyolites being produced elsewhere in North America, particularly at the same time, as well as elsewhere in the world.

Bachmann and Bergantz (2004) and Bachmann et al. (2007) document NCIs of volcanic and plutonic rocks from various igneous provinces. Notably, they show that intrusive and extrusive groups of rocks within each province exhibit strongly overlapping NCIs. Volcanic rocks are therefore likely to be good indicators of the isotopic characteristics of comagmatic, unerupted parts of their reservoir systems.

6.2. $\delta^{18}\text{O}$ constraints

Oxygen isotopes provide an important independent measure of mantle mass contribution to YSRP volcanic rocks. Estimates of $\delta^{18}\text{O}$ for mantle and Archean crust reservoirs may be inferred from YSRP basalts (4.8–5.2‰) and Archean crustal xenoliths (6.7–9.2‰) (Leeman and Whelan, 1983; Bindeman et al., 2007). Bindeman and Valley (2000, 2001) and Bindeman et al. (2007, 2008) document values of 5.5 to 6.1‰ for most rhyolites produced in the Heise and Yellowstone Plateau volcanic fields. The latter indicate mixing of 85–60% basalt (with $\delta^{18}\text{O}$ of 5‰) and 15–40% of Archean crust (with $\delta^{18}\text{O}$ of 8‰), consistent with our NCI model.

Furthermore, Bindeman et al. (2007, 2008) and others (Hildreth et al., 1984, 1991; Bindeman and Valley, 2000, 2001) point out that the first erupted, large volume rhyolites in the Yellowstone Plateau and Heise volcanic fields have the highest $\delta^{18}\text{O}$ (up to 7‰). Lower $\delta^{18}\text{O}$ in many younger rhyolites in those areas (0.7–5‰) is likely a product of remelting and recycling of earlier, strongly hydrothermally altered parts of thick caldera fill and epizonal plutonic rocks (Hildreth et al., 1984, 1991; Bindeman and Valley, 2000, 2001; Bindeman et al., 2007, 2008). The higher $\delta^{18}\text{O}$ for the first erupted rhyolites is therefore the best indicator of original O-isotopic conditions of basalt–Archean crust interaction. In that case, simple mass-balance limits the initial basalt contribution to the rhyolites to <75%, the remainder being contributed from Archean crust (25 to 100%). Combined $\delta^{18}\text{O}$ and NCI data is therefore consistent with rhyolites containing 60 to 75% of mantle-derived mass and 25–40% of Archean crust.

Boroughs et al. (2005, 2007) also document unusually light $\delta^{18}\text{O}$ for rocks in the CSRP, and as far west as 116° longitude. Interestingly, there are few 'normal' ($\delta^{18}\text{O} > 5$) rhyolites in that region. Boroughs et al. (2005) suggest that the low values in that region originate from partial melting of hydrothermally altered parts of the Idaho Batholith. Origin of the low $\delta^{18}\text{O}$ SR-type rhyolites is therefore controversial (Boroughs et al., 2005; Bindeman et al., 2007). It may involve some combination of spatial variation in oxygen isotopic compositions of source rocks (Boroughs et al., 2005, 2007), and remelting of hydrothermally altered

silicic volcanic rocks and cogenetic plutonic rocks (Bindeman et al., 2007, 2008). We believe the latter process is particularly effective. Meteoric water interactions that would strongly alter $\delta^{18}\text{O}$ would have little effect on the isotopic balance of conservative elements such as Nd. That may account for apparent ‘decoupling’ of NCI and O-isotopic data for many of the SR-type rhyolites (e.g. Bonnicksen et al., 2008).

6.3. Error analysis and propagation

Many of the uncertainties in our data and model are admittedly difficult or impossible to quantify. We argue for limits on some key variables such as average ε_{Nd} of rhyolite -8.3 ± 1.9 (median = -7.8), and basalt -4.4 ± 1.1 (median = -4.6). We argue for an upper limit of ε_{Nd} for pre-YSRP hotspot track continental crust of -20 . We also attempt to set limits on some potential sources of systematic error such as might occur in intracaldera versus outflow volcanic facies, and syn- versus post-hotspot track mafic volcanism.

A number of key uncertainties also exist in our conceptual model, and related mass-balance calculations. Among the assumed parameters, greatest sensitivities are likely related to NCI ($1-f_{\text{M}}$), ratio of non-erupted to erupted magma (f_{M}), and the average thickness of rhyolite on the YSRP (M_{er}). In each case we make what we consider to be conservative estimates. Cumulative uncertainties of about a factor of two are generated in mantle-derived-mass calculations by plausible variations of these parameters, as is illustrated in Fig. 9 in shaded fields surrounding best-guess estimates.

7. Conclusions

The main purpose of this paper is to estimate the amount of ‘mantle’ input in the generation of voluminous rhyolites of the Snake River Plain and to develop a model by which rhyolites obtain their characteristic Nd-isotopic values through a multi-stage petrogenetic process. We base our estimate on the distinct Nd-isotopic contrasts between basalts, rhyolites and continental crust and use conservative isotopic values to arrive at the conclusion that the maximum amount of crust involved to generate rhyolites is no more than 40% and may be as low as 12%.

Low crustal input to the YSRP rhyolites implies that voluminous amounts of mass have been transferred to the crust from the mantle. We estimate the mantle mass transfer by adopting a simple model of rhyolite petrogenesis, and then budget sequentially for mass sequestered in the upper-, middle and lower crust. Conservative estimates of pertinent petrologic variables yield a total mantle-derived mass to YSRP equivalent to a layer of gabbro 14-km-thick.

If extended across the CSRP and ESRP, a total mass equivalent to 340,000 km³ of gabbro have been added to the continental crust along the YSRP hotspot track. That is more than the 234,000 km³ extrusive volume of the Columbia River Basalt igneous province (Camp et al., 2003), and ~30% of the combined extrusive and intrusive volume of that province (Coffin and Eldholm, 1994). The corresponding mantle-derived magma flux between 11 and 4 Ma is ~0.05 km³/yr, comparable to many other hotspot systems (e.g. Schubert et al., 2001; West et al., 2003), but 5–10 times less than the Ninetyeast Ridge (Coffin et al., 2002) or late Cenozoic production rate of Hawaii (Vidal and Bonneville, 2004). Structural implications for this mass are discussed in a companion paper (Rodgers and McCurry, 2009-this volume). The mass balance also indicates that a batholith ~5–10 thick has developed beneath the YSRP. The batholith is likely to have a prominent A-type character (Frost et al., 2001; Christiansen and McCurry, 2008), and a volume comparable to some of the large batholiths of the western North America Cordillera.

Acknowledgements

M. McCurry thanks Bill Bonnicksen, Martha Godchaux, Bill Hackett, Bill Leeman, Mel Kuntz, Lisa Morgan, Scott Hughes, Hanna Nekvasil and Mike Branney for help in various stages of field work, and for many stimulating conversations regarding SRP rhyolite and basalt volcanology and petrogenesis. We thank Mark Ford, Kaleb Scarberry and Kathleen Price for constructive participation in a graduate class that served as a key motivation for this work. We thank Linda Davis for the assistance in collecting core from the USGS core library at INL, S. Chen for assistance in geochemical sample preparation and analysis, and Mark Fanning for SHRIMP analyses of zircon grains from borehole INEL-1. This work was supported in part by a funding from EPSCoR Cooperative Agreement OSR-9350539, and U.S. Department of Energy Grant DE-FD0C07-96ID13420 through the Idaho Water Resources Water Research Institute. The manuscript was substantially improved by constructive comments by Martin Streck, Vic Camp, and Ilya Bindeman.

Appendix A. Supplementary data

Supplementary data associated with this article can be found, in the online version, at doi:10.1016/j.jvolgeores.2009.04.001.

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