

Genesis of post-hotspot, A-type rhyolite of the Eastern Snake River Plain volcanic field by extreme fractional crystallization of olivine tholeiite

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Received: 1 August 2005 / Accepted: 8 March 2007 / Published online: 20 June 2007
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Abstract Rhyolites occur as a subordinate component of the basalt-dominated Eastern Snake River Plain volcanic field. The basalt-dominated volcanic field spatially overlaps and post-dates voluminous late Miocene to Pliocene rhyolites of the Yellowstone–Snake River Plain hotspot track. In some areas the basalt lavas are intruded, interlayered or overlain by $\sim 15 \text{ km}^3$ of cryptodomes, domes and flows of high-silica rhyolite. These post-hotspot rhyolites have distinctive A-type geochemical signatures including high whole-rock $\text{FeO}_{\text{tot}}/(\text{FeO}_{\text{tot}}+\text{MgO})$, high Rb/Sr, low Sr (0.5–10 ppm) and are either aphyric, or contain an anhydrous phenocryst assemblage of sodic sanidine $\pm$ plagioclase + quartz $>$ fayalite + ferroaugite $>$ magnetite $>$ ilmenite + accessory zircon + apatite + chevkinite. Nd- and Sr-isotopic compositions overlap with coeval olivine

tholeiites ($\epsilon_{\text{Nd}}=-4$ to -6 ; $^{87}\text{Sr}/^{86}\text{Sr}_i=0.7080-0.7102$) and contrast markedly with isotopically evolved Archean country rocks. In at least two cases, the rhyolite lavas occur as cogenetic parts of compositionally zoned ($\sim 55-75\% \text{ SiO}_2$) shield volcanoes. Both consist dominantly of intermediate composition lavas and have cumulative volumes of several 10^3 of km^3 each. They exhibit two distinct, systematic and continuous types of compositional trends: (1) At Cedar Butte (0.4 Ma) the volcanic rocks are characterized by prominent curvilinear patterns of whole-rock chemical covariation. Whole-rock compositions correlate systematically with changes in phenocryst compositions and assemblages. (2) At Unnamed Butte (1.4 Ma) the lavas are dominated by linear patterns of whole-rock chemical covariation, disequilibrium phenocryst assemblages, and magmatic enclaves. Intermediate compositions in this group resulted from variable amounts of mixing and hybridization of olivine tholeiite and rhyolite parent magmas. Interestingly, models of rhyolite genesis that involve large degrees of melting of Archean crust or previously consolidated mafic or silicic Tertiary intrusions do not produce observed ranges of Nd- and Sr-isotopes, extreme depletions in Sr-concentration, and cogenetic spectra of intermediate rock compositions for both groups. Instead, least-squares mass-balance, energy-constrained assimilation and fractional crystallization modeling, and mineral thermobarometry can explain rhyolite production by 77% low-pressure fractional crystallization of a basaltic trachyandesite parent magma ($\sim 55\% \text{ SiO}_2$), accompanied by minor (0.03–7%) assimilation of Archean upper crust. We present a physical model that links the rhyolites and parental intermediate magmas to primitive olivine tholeiite by fractional crystallization. Assimilation, recharge, mixing and fractional melting occur to limited degrees, but are not essential parts of the rhyolite formation process.

This paper constitutes part of a special issue dedicated to Bill Bonnichsen on the petrogenesis and volcanology of anorogenic rhyolites.

Editorial responsibility: W. Leeman

Electronic supplementary material The online version of this article (doi:10.1007/s00445-007-0143-4) contains supplementary material, which is available to authorized users.

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Keywords Yellowstone–Snake River Plain · Hotspot · Rhyolite genesis · Bimodal volcanism · A-type magma · Basalt differentiation · Fractional crystallization

Introduction

Rhyolite genesis, particularly in basalt-dominated, intra-plate continental volcanic provinces, is a long-standing and fundamental problem of igneous petrogenesis and crustal evolution (Bachmann and Bergantz 2004; Bryan et al. 2002; Hildreth 1981; Smith 1979; Bowen 1928). Relative contributions from crust and mantle sources, and processes and rates of rhyolite formation continue to be strongly debated (e.g. Eichelberger et al. 2000; de Silva 2001); of special interest is the potential for establishing a single liquid line of descent that might connect rhyolites to mantle-derived basaltic magmas (e.g. Nekvasil et al. 2000; Whitaker et al. 2007; Sisson et al. 2005). Continuing volcanism in the wake of the world's largest active

continental hotspot, the Yellowstone–Snake River Plain volcanic track (e.g. Smith and Braile 1994; Pierce and Morgan 1992) (Figs. 1a) has produced a spectrum of basaltic to rhyolitic volcanic rocks in the Eastern Snake River Plain (Figs. 1d and 2). Genesis of these rhyolites has been decoupled from that of coeval mafic and intermediate rocks because of discordances in some of their geochemical characteristics (e.g. Spear 1979). Recent work has questioned the basis for that geochemical argument (McCurry et al. 1999), and new experiments have defined plausible magma evolution pathways for Eastern Snake River Plain-type magmas (Nekvasil et al. 2000; Whitaker et al. 2007). We re-examine the origin of Quaternary Eastern Snake River Plain rhyolites, and document new field, petrological, compositional and isotopic data for the rhyolites and associated intermediate rocks, focusing on two intermediate to silicic volcanic shields that we refer to as Cedar Butte and Unnamed Butte volcanoes. Collectively, Eastern Snake River Plain rhyolites are well suited for examination of their origin owing to their young age, excellent exposures,

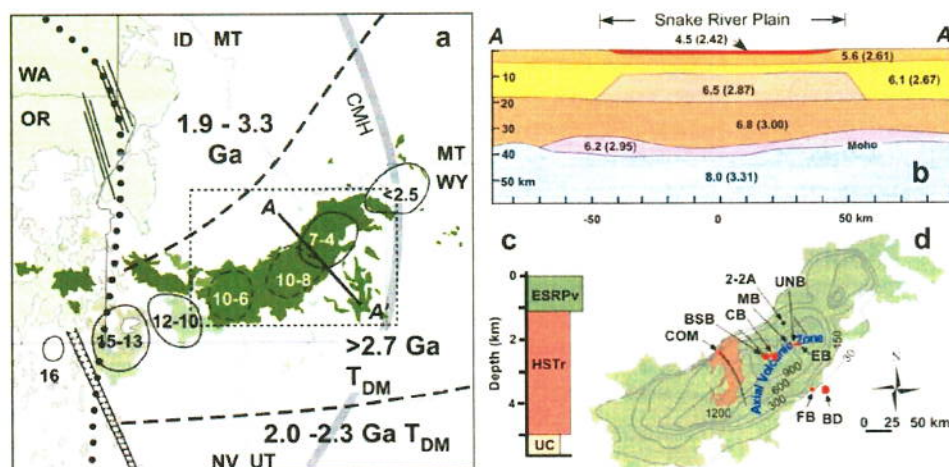


Fig. 1 **a** Geologic setting for the Eastern Snake River Plain and surroundings, illustrating Neogene basalt-dominated terrains < 5 Ma (dark green), 5–10 Ma (medium green) and > 10 Ma (light green). Locations and ages (Ma) for silicic volcanic fields of the Yellowstone–Snake River Plain hotspot track are circled (e.g. Perkins and Nash 2002). Bold dashed lines separate major lithospheric domains yielding Archean and Proterozoic Nd model ages and radiometric dates (Bennett and DePaolo 1987). Bold dotted line separates accreted Paleozoic and younger lithosphere to the west from the Precambrian craton (Leeman et al. 1992). Chief Joseph dike swarms and Northern Nevada rift are illustrated diagrammatically in the northwest and southwest parts of diagram, respectively. State lines are shown for reference (WA Washington, ID Idaho, MT Montana, WY Wyoming, NV Nevada, UT Utah). CMH—approximate location of the Cordilleran miogeosyncline hinge line (e.g. Link et al. 1993). A–A' is a cross-section line for Fig. 1b. Dashed box outlines the area expanded in Fig. 1d. **b** Geophysical profile across the Eastern Snake River Plain

illustrating mean crustal and mantle seismic V_p velocities (km/s), densities—in parentheses (gm/cm³), and layer thicknesses (after Sparlin et al. 1982; Peng and Humphreys 1998). **c** Simplified stratigraphy of the Eastern Snake River Plain. UC pre-Tertiary upper crust; HSRr Yellowstone–Snake River Plain hotspot track rhyolites; ESRPv basalt-dominated rocks of the Eastern Snake River Plain volcanic field. **d** Expanded view of the Eastern Snake River Plain volcanic field (shaded region). Gray lines are isopachs of “Quaternary basalt” thickness (in meters) defined by electrical resistivity surveys by Whitehead (1992). Evolved Quaternary volcanic centers include COM Craters of the Moon; BSB Big Southern Butte; CB Cedar Butte; UNB Unnamed Butte; EB East Butte. Quaternary Cryptodomes include MB Middle Butte; FB Ferry Butte; BD Buckskin Dome. 2-2A—location of INEL borehole 2-2A (Champion et al. 2002). Three prominent northwest trending rift systems are illustrated with short line segments. From southwest to northeast these include, respectively, the Great Rift, Arco Rift and Lava Ridge–Hells Half Acre rifts (Kuntz et al. 1992)

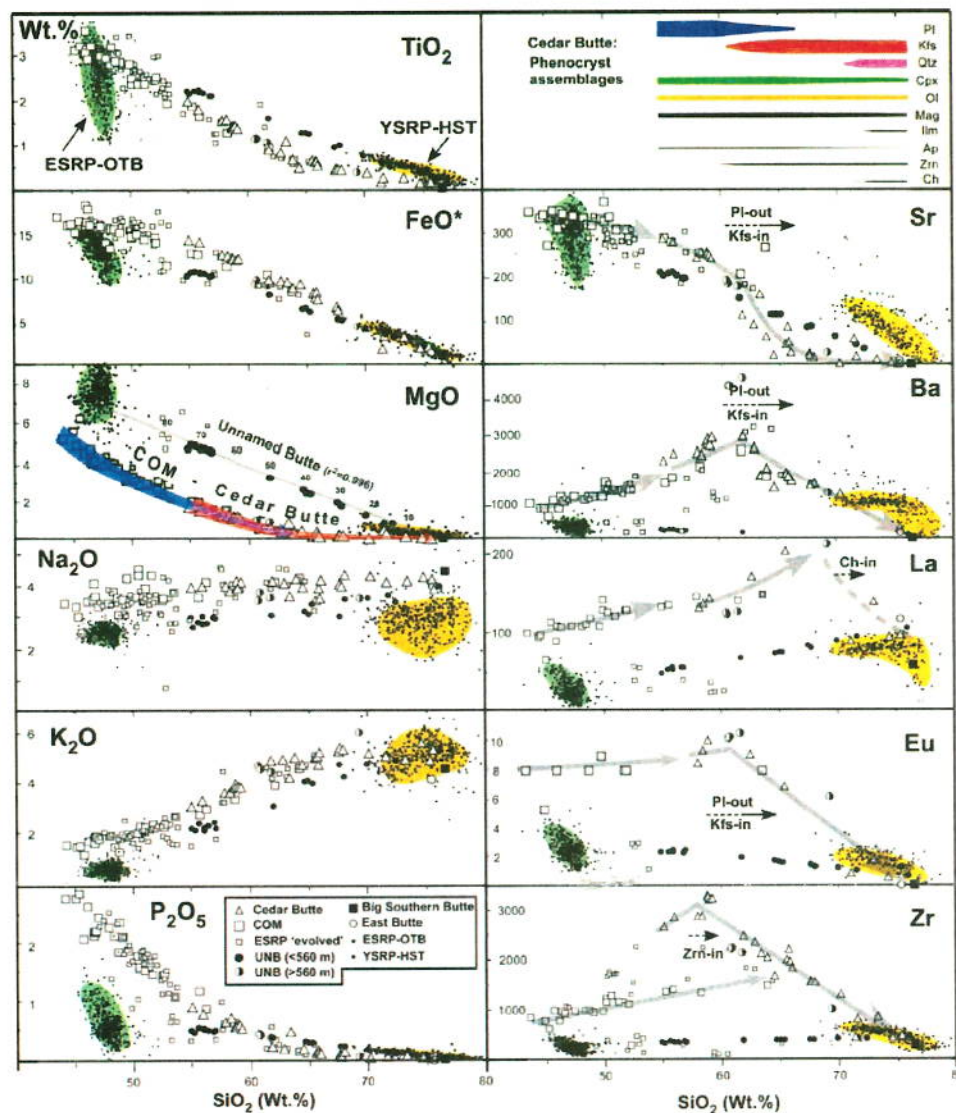


Fig. 2 Harker diagrams illustrating major- and trace-element variations for selected rock suites from the Yellowstone–Snake River Plain volcanic track. Cedar Butte—lavas, dikes and pyroclastic rocks from Cedar Butte volcano; COM—lavas from Craters of the Moon volcanic field; ESRP ‘evolved’—mafic to intermediate composition rocks from the Eastern Snake River Plain volcanic field, excluding Cedar Butte, COM and Unnamed Butte, and having ‘evolved’ geochemical characteristics ($\text{MgO} \leq 4\%$ or $\text{SiO}_2 > 52\%$); ESRP-OTB—olivine tholeiite basalts from the Eastern Snake River Plain volcanic field (Hughes et al. 2002a) (green shaded region highlights the dominant compositional field for ESRP-OTB); UNB Unnamed Butte volcano; EB East Butte volcano; BSB Big Southern Butte volcano; YSRP-HST rhyolites that are coeval with the Yellowstone–Snake River Plain hotspot (tan shaded region highlights the dominant compositional

field for YSRP-HST). Bold colored arrows on the MgO – SiO_2 diagram illustrates overlapping covariation curves for COM (blue) and Cedar Butte (red). Hachured line for MgO is a linear regression fit of data from UNB (r^2 = square of the correlation coefficient); hachures indicate percentages of a hypothetical OTB end-member in a mixture of OTB and rhyolite parent magmas. Bold gray lines indicate inferred trace-element fractionation trends at Cedar Butte (dashed where poorly defined). Horizontal arrowed lines illustrate salient changes in assemblages of phenocrysts with increasing silica content for rocks from Cedar Butte volcano. Top right panel illustrates coexisting phenocryst assemblages in lavas from Cedar Butte; line widths are proportional to the relative abundances of phenocrysts. Abbreviations are from Kretz 1983 (Ch chevkinite)

close spatial and temporal association with more mafic volcanic rocks, and because of the highly distinctive geochemical characteristics of lower- and upper crust and mantle in the region. We evaluate the respective roles of fractional crystallization, assimilation, and fractional melt-

ing in the formation of rhyolites, and specifically test the idea that rhyolite formation is an extension of processes that appear to produce evolved mafic to intermediate lavas in the region, i.e. extreme polybaric fractionation of primitive olivine tholeiite magma (e.g. Thompson 1975; Leeman

1982b; Menzies et al. 1984; Hildreth et al. 1991; Kuntz et al. 1992; Stout et al. 1994; Putirka et al. 2003a; Whitaker et al. 2007).

Eastern Snake River Plain volcanic field

The Eastern Snake River Plain is a well-known region of Pliocene to Quaternary olivine tholeiite volcanism (e.g. Leeman 1982a; Kuntz et al. 1992; Reid 1995; Hanan et al. 1997; Hughes et al. 2002a, b; Geist et al. 2002; Shervais et al. 2006). In some areas these relatively primitive basaltic lavas are also interlayered with, overlain and intruded by compositionally distinctive evolved mafic to intermediate lavas (e.g. Leeman 1982b; Kuntz et al. 1986; Stout and Nicholls 1977; Stout et al. 1994) and rhyolite domes and cryptodomes (Spear and King 1982; McCurry et al. 1999). We refer collectively to these stratigraphically and spatially associated volcanic deposits as the Eastern Snake River Plain volcanic field (Fig. 1c, unit ESRPv; Fig. 1a, d).

Mafic to silicic rocks of the Eastern Snake River Plain volcanic field spatially overlap, and stratigraphically overlie voluminous Miocene to early Pliocene rhyolitic ignimbrites and lava flows of the Yellowstone hotspot track (Fig. 1c, unit HSTr) (Armstrong et al. 1975; Leeman 1982c; Pierce and Morgan 1992; Pierce et al. 2002; Smith and Braile 1994; Bonnicksen et al. 2007). We distinguish the younger rhyolites of the Eastern Snake River Plain volcanic field from the older hotspot-track-related rhyolites on the basis of their stratigraphic association with basaltic and intermediate rocks, lower eruption volumes, and more primitive Nd- and Sr-isotopic composition (Christiansen and McCurry 2007; Branney et al. 2007; McCurry et al. 1999). This study is limited to the younger, post-hotspot group of rhyolites, intermediate and mafic rocks.

The Eastern Snake River Plain volcanic field overlaps an arcuate band of petrologically similar basalt-dominated volcanism that extends over large parts of the Yellowstone–Snake River Plain and Oregon–High Lava Plain volcanic tracks (Camp and Ross 2004; Christiansen et al. 2002), and the northern part of the Neogene–Quaternary Basin and Range Province (Anders et al. 1989; Rodgers et al. 1990, 2002; Pierce and Morgan 1992). Pliocene to Quaternary basaltic lavas increase from 10's of meters thick near the margins of the Eastern Snake River Plain to ~1.5 km near the center of the field (Whitehead 1992; Kuntz et al. 1992) (Fig. 1d). Interpolation of basalt thickness based upon geophysical (Whitehead 1992; Greensfelder and Kovach 1982; Sparlin et al. 1982; Josten and Smith 1997) and borehole data (e.g. Champion et al. 2002) suggests that the field has a volume of basalt of at least ~12,300 km³ (Fig. 1d). Basaltic lavas erupted from numerous, widely scattered small shields and northwest-trending vent zones (Kuntz et al. 1992; Hughes et al. 2002b). They are

interlayered with fluvial, eolian and lacustrine sediments near northern, western and southern margins of the Eastern Snake River Plain (e.g. Geslin et al. 2002).

Formation of the Eastern Snake River Plain volcanic field began within 1 m.y. of cessation of regional hotspot-track-related silicic volcanism. Basaltic lava flows as old as 4.2 Ma (Champion et al. 2002) overlie silicic tuffs of the 6.6–4.5 Ma Heise volcanic field (Morgan and McIntosh 2005). The sequential relationship from rhyolite to basalt-dominated volcanism may be related to waning stages of activity as the hotspot migrated northeast (Leeman 1982c; Anders and Sleep 1992; Smith and Braile 1994), or to some more complicated dynamic mantle process (Humphreys et al. 2000; Ducker et al. 2001; Christiansen et al. 2002; Camp and Ross 2004).

Champion et al. (2002) document a nearly constant basalt accumulation rate for the Eastern Snake River Plain volcanic field of ~20 m/100 k.y. over the last ~4.5 m.y. in one of the few boreholes that penetrate the entire basalt sequence (corehole INEL 2-2A, Fig. 1d). This rate is comparable to the average tectonic subsidence rate of ~25 m/100 k.y. (Rodgers et al. 2002). Accumulation rates increase towards the center of the volcanic field (Champion et al. 2002). Local hiatus' of a few 100,000 years appear common, and apparently result from shifts in volcanic foci. Basalts also exhibit little large-scale spatially, or temporally systematic variations in major, trace or isotopic compositions (Hughes et al. 2002a), excepting variations related to the evolution of specific eruptive centers (Hughes et al. 2002a, b; Geist et al. 2002; Shervais et al. 2006). Subsidence and magma output rates, and the compositions of lavas of the Eastern Snake River Plain volcanic field therefore appear to be nearly constant over the last 4–4.5 m.y.

Previous studies identify two compositionally and petrogenetically distinct rock suites in the Eastern Snake River Plain volcanic field: A volumetrically dominant, olivine tholeiite suite (Kuntz et al. 1992; Leeman 1982a) and an evolved mafic to silicic suite (Leeman 1982b; McCurry et al. 1999). Olivine tholeiites (Figs. 2 and 3) have relatively primitive to moderately evolved elemental compositions, narrow ranges of silica content ($47.3 \pm 2.3\%$ SiO₂ for 620 analyses; Hughes et al. 2002a), and have slightly evolved isotopic compositions (e.g. $\epsilon_{\text{Nd}} \sim -3$ to -6 ; $^{87}\text{Sr}/^{86}\text{Sr}_{(i)} \sim 0.706$ to 0.708 ; Hughes et al. 2002a; Hanan et al. 1997) that are consistent with derivation from an enriched, possibly heterogeneous lithospheric-mantle (Leeman 1982a; Stout and Nicholls 1977; Hughes et al. 2002a; Geist et al. 2002; Hanan et al. 1995; Reid 1995; Menzies et al. 1984). Many, and perhaps all, have undergone minor to moderate amounts of fractional crystallization, and little or no assimilation of Archean crust (Leeman 1982a, Hughes et al. 2002a). Leeman (1982a) and Hughes et al. (2002a) further subdivide the olivine tholeiite suite into subgroups

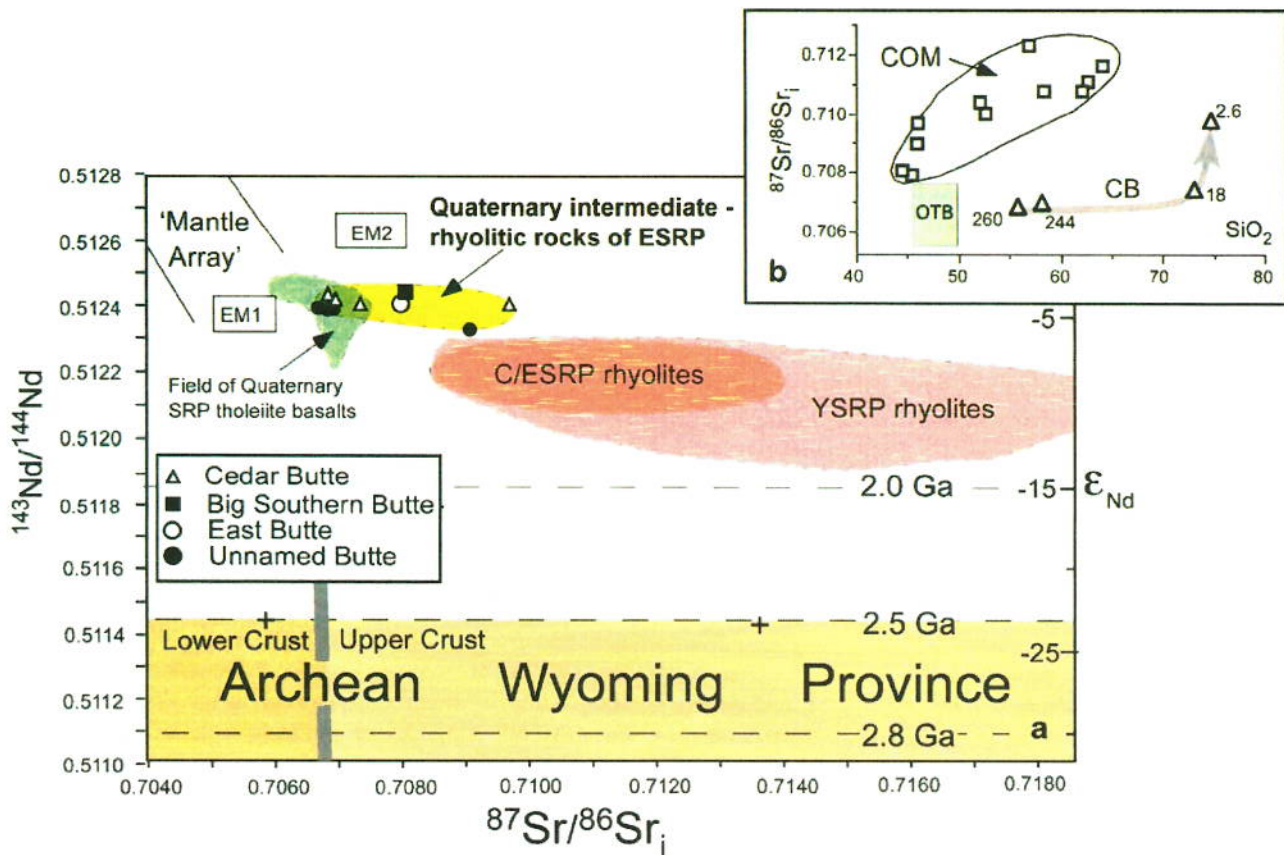


Fig. 3 Nd- and Sr-isotopic characteristics of major Neogene–Quaternary volcanic rock suites (Hughes et al. 2002a; Leeman et al. 1992; Wright et al. 2002; Hildreth et al. 1991), cratonic rocks (Leeman et al. 1985), and plausible mantle reservoirs (Mantle array, EM1 and EM2 from Zindler and Hart (1986) along the Yellowstone–Snake River Plain volcanic track. The isotopic range of samples of Quaternary intermediate to silicic rocks of the Eastern Snake River Plain is shown with yellow shading. Fields representing the range of Nd- and Sr-isotopic composition for most rhyolites from the Central and Eastern Snake River Plain (C/ESRP) and Yellowstone region (YSRP), are shaded dark red and light red, respectively. Triangles—Cedar Butte, open squares—Unnamed Butte, filled square—East Butte, filled circle—Big Southern Butte. Plus symbols represent

compositions of xenoliths in SRP basalts (Leeman et al. 1985). Leeman et al. (1985) document 14 more xenoliths, and surface samples from the Albion Mountains that plot off this diagram, and have ϵ_{Nd} from -23 to -52 and $^{87}\text{Sr}/^{86}\text{Sr}$ of 0.7024 to 0.8926 . Horizontal dashed lines and vertical shaded line indicate model ranges of Nd- and Sr-isotopic compositions of lower and upper crust as a function of age (DePaolo 1988). **b** Inset diagram contrasts the $\text{Sr}_{(i)}$ -isotopic characteristics of rocks from Craters of the Moon (+ symbols), with those from Cedar Butte (triangles), and with olivine tholeiites (OTB). The gray arrow indicates the inferred fractionation path for Cedar Butte. Numbers adjacent to Cedar Butte samples indicate their respective whole rock Sr concentrations

based upon differences in trace element contents and Nd- and Sr-isotopes. Differences are interpreted to result from differing degrees of melting in source rocks and to low-pressure fractional crystallization. Geist et al. (2002) and Shervais et al. (2006) point out patterns in trace elements, K_2O and P_2O_5 contents of some olivine tholeiites that they interpret as resulting from combined fractional crystallization and assimilation of mid-crustal ferrogabbro.

The evolved mafic to silicic suite varies from alkaline, high-K trachybasalt to subalkaline, high-K rhyolite (Figs. 2 and 3). Rocks of this suite also have high Fe/Mg ratios, high concentrations of incompatible elements, anhydrous, pyroxene-bearing phenocryst assemblages, and primitive to mildly evolved isotopic ratios (Menzies et al. 1984; Kuntz

et al. 1986; Futa et al. 1988; McCurry et al. 1999). Mafic to intermediate compositions (45–64% SiO_2) are best represented at the late Pleistocene–Holocene Craters of the Moon volcanic field (Kuntz et al. 1992; Kuntz et al. 1986; Stout et al. 1994; Leeman et al. 1976; Leeman 1982b). These rocks are strongly differentiated, mostly by polybaric fractional crystallization (e.g. Thompson 1975; Leeman 1982b; Putirka et al. 2003a), and exhibit variable amounts of contamination by crustal xenoliths (Leeman 1982b; Menzies et al. 1984; Kuntz et al. 1992). More felsic rocks (55–75% SiO_2) are best represented at Cedar Butte (McCurry et al. 1999; Spear 1979). They overlap compositionally and mineralogically with intermediate lavas at Craters of the Moon (Leeman et al. 1976; Stout et al. 1994),

and with high-silica rhyolite domes and flows of the central Eastern Snake River Plain volcanic field (McCurry et al. 1999; Spear 1979). Collectively, volcanic rocks at Craters of the Moon and Cedar Butte define a nearly continuous and strongly correlated spectrum of rock types from trachybasalt to high-silica rhyolite that we refer to as the Craters of the Moon-Cedar Butte suite (Fig. 2).

Compositional gaps occur between the rocks of olivine tholeiite and evolved mafic suites (e.g. Thompson 1975; Leeman 1982b); the latter having lower MgO and CaO, and higher alkali metals, P_2O_5 and incompatible trace-element contents (Fig. 2). Lavas of the two suites, however, broadly overlap spatially and stratigraphically (Leeman 1982b; McCurry et al. 1999; Hughes et al. 2002a; Fishel 1993). Therefore despite their petrologic and geochemical differences, both suites of rocks are correlated in space and time.

Crustal structure

Eastern Snake River Plain volcanic field is located on the western margin of the Archean Wyoming province (e.g. Chamberlain et al. 2003), west of the Cordilleran miogeoclinal hingeline (Allmendinger et al. 1987) (Fig. 1a). Archean rocks occur in exposures and as xenoliths in basaltic rocks as far west as $115^\circ W$. They consist of diverse granitic, metamorphic and mafic lithologies with highly evolved Nd-isotopic ratios (Fig. 3) (Leeman et al. 1985; Matty 1984). Major Mesozoic and Eocene batholiths and volcanic fields overlap parts of the Yellowstone–Snake River Plain volcanic track in regions to the west (Norman and Leeman 1989), north (Moye et al. 1988) and east (e.g. Feeley 2003), but none of these appear to extend across the Eastern Snake River Plain region.

Major deep crustal and mantle structures are summarized in Fig. 1b. The Moho is approximately flat at a depth of ~42 km. The crust consists of ~20 km thick mafic lower and ~20 km thick felsic upper layers, overlain by ~2–5 km of low-density late Precambrian–Mesozoic miogeoclinal sedimentary rocks (Sparlin et al. 1982; Link et al. 1993). Features that are restricted to parts of the crust and mantle directly beneath the Eastern Snake River Plain are: (1) A structurally downwarped basin that is mostly infilled by voluminous rhyolites of the Yellowstone–Snake River Plain hotspot track, and olivine tholeiite and interbedded intermediate to silicic rocks of the Eastern Snake River Plain volcanic field (Fig. 1c) (e.g. Morgan and McIntosh 2005; Rodgers et al. 2002; Kuntz et al. 1992). (2) A ~10 km thick mid-crustal positive density anomaly, likely produced by intrusions of mafic, mantle derived basaltic magma (e.g. Sparlin et al. 1982; Peng and Humphreys 1998); (3) A layer of partial melt at the crust–

mantle boundary (Peng and Humphreys 1998). (4) A region of mantle distinguished by anomalously low seismic velocities that extends from the base of the crust to at least 270 km depth (Saltzer and Humphreys 1997; Yuan and Dueker 2005).

Quaternary rocks of the Eastern Snake River Plain volcanic field are structurally distinguished from regions marginal to the plain by an absence of major normal faults, and the presence rift zones that are most likely related to northwest-trending dike swarms (Kuntz et al. 2002; Hughes et al. 2002b; Rodgers et al. 2002, 1990).

Rhyolitic volcanic centers

Four Quaternary rhyolite volcanic centers, Cedar Butte, Unnamed Butte, Big Southern Butte and East Butte, occur near the center of the Eastern Snake River Plain volcanic field, along a subtle curvilinear topographic high formed by numerous overlapping basaltic shields (Kuntz et al. 1992) (Fig. 1d “Axial Volcanic Zone”). Additional cryptodomes likely cored by rhyolite also occur in the Axial Volcanic Zone (Middle Butte: Kuntz and Dalrymple 1978), and near the southeast margin of the Eastern Snake River Plain volcanic field (Ferry Butte and Buckskin Dome; Fig. 1d) (Houser 1992; Kellogg and Embry 1986). We describe and interpret the volcanology, petrology and geochemistry of the four exposed rhyolite volcanoes, expanding upon previous work by Spear (1979), Kuntz and Dalrymple (1979) and Kuntz and Kork (1978), emphasizing those that exhibit a wide spectrum of lava compositions.

Big Southern Butte

Big Southern Butte is a prominent 730-m-high rhyolite cumulodome located near the intersection of two major structural lineaments, the Arco Rift Zone and Axial Volcanic Zone (Figs. 1d and 4) (Kuntz and Kork 1978; Spear 1979). The dome evolved from a laccolith that formed near the Tertiary rhyolite-basalt contact at a depth of ~1 km (Spear and King 1982). Continued magma input resulted in inflation of the laccolith, uplift of the overlying basalt, and formation of a cryptodome. Continued inflation of the cryptodome resulted in failure of the basaltic roof rocks and extrusion of rhyolite lava (Spear and King 1982). Remnants of the roof of the cryptodome occur as a ~900-m-thick, north-tilted flap of basalt exposed on the north flank of the dome (Fishel 1992).

Contacts between the dome and uplifted basalt vary from glassy to devitrified, and are weakly hydrothermally altered. Formation of the cumulodome occurred in two stages by extrusion of two overlapping endogenous domes (Spear and

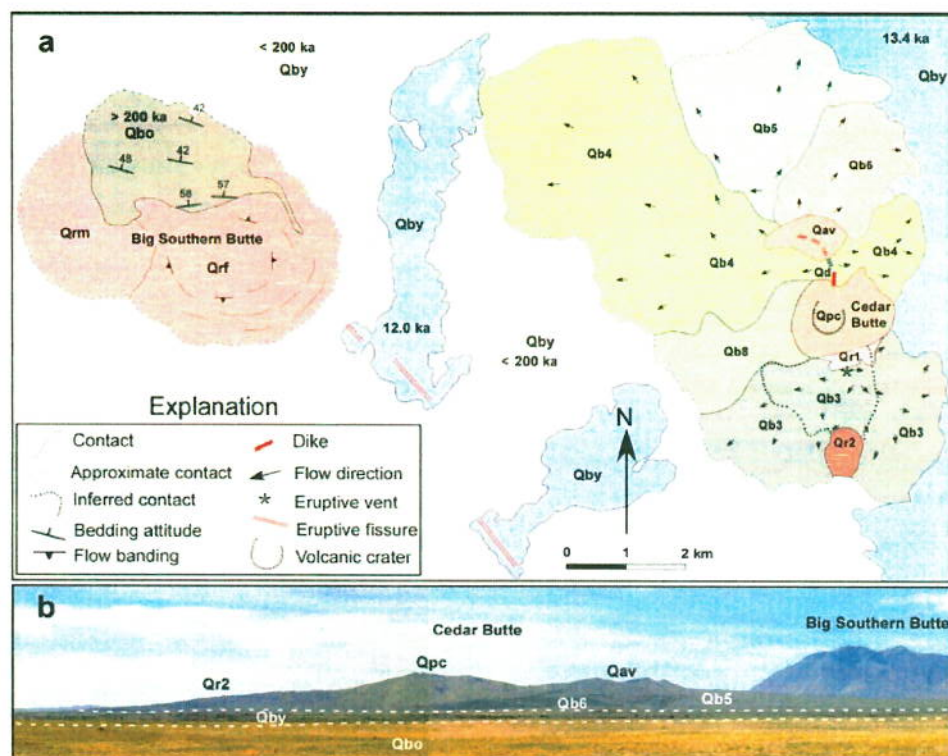


Fig. 4 a Geology of Cedar Butte and Big Southern Butte volcanoes. Qbo—basalt more than 200 ka; Qby—basalt younger than 200 ka (Kuntz et al. 1994); Qrf and Qrm are older and slightly younger lava domes, respectively, of the Big Southern Butte cumuldome (simplified from Spear 1979). Unit labels for Cedar Butte, in sequence from oldest to youngest, are Qr2—high-silica rhyolite lava flow; Qr1—clastogenic siliceous trachyte and low-silica rhyolite lava; Qpc—a tephra cone consisting dominantly of trachyandesite to trachyte scoria

and bombs, and also minor but prominent white pumiceous rhyolite, and obsidian; Qb5—mafic trachyte lava flow; Qb8—trachyandesite lava flow; Qb4—trachyandesite lava flow; Qb6—siliceous trachyte lava flow; Qb3—basaltic trachyandesite lava flow. **b** View of Cedar Butte shield volcano as viewed looking to the southwest. The exposed part of the volcano is about 8 km across, and has a relief of 220 m. Unit labels are from part (a)

King 1982). Whole-rock K/Ar dates from the respective domes overlap at 309 ± 10 and 294 ± 15 ka (Kuntz et al. 1994). The two are virtually identical in major and trace element composition, and consist of nearly aphyric high-silica rhyolite. Spear and King (1982) report rare microphenocrysts of quartz, fayalite, ferroaugite and magnetite that occur in a felsic, felty to devitrified flow-banded to massive matrix. Poorly sorted layers of ash, lapilli and breccia occur near the top of the dome in some places, suggesting that mild explosive activity accompanied dome formation. Minimum volume of the rhyolite is $\sim 6 \text{ km}^3$, and may be as much as $\sim 20 \text{ km}^3$, depending on extrapolations of the dome into the subsurface (Appendix Table 1). The lava dome is surrounded by 12–200 ka lava flows. Contacts between these flows and the dome are covered by colluvium and debris fans shed from the lava dome (Kuntz et al. 1994).

Big Southern Butte is remarkably homogeneous in composition (Spear 1979), and is the most evolved of the Eastern Snake River Plain volcanic field rhyolites, having the highest silica (76.4%), the highest Sm/Eu (51) and Rb/Sr (529) ratios, and the lowest contents of MgO (<0.01%), Zr (309 ppm), Eu (0.26 ppm), Sc (0.1 ppm), LREE (e.g.

La = 59 ppm), Sr (0.52 ppm), Ba (18 ppm), and LREE/REE ratio (Fig. 2; Appendix Table 2). It is also among the least isotopically evolved of the rhyolites, having ϵ_{Nd} of -4.0 and $^{87}\text{Sr}/^{86}\text{Sr}_i$ of 0.70805 (Fig. 3; Table 1).

East Butte

East Butte is a monogenetic rhyolite lava dome located ~30 km east-northeast of Big Southern Butte (Fig. 1d) (Spear 1979; Bretches 1984; Kuntz and Dalrymple 1979; McCurry and Ganske 2005; Ganske 2006). K/Ar dated at 580 ± 90 ka (Armstrong et al. 1975), the lava dome has a relief of 350 m and volume of ~ 0.3 km³. Margins of the dome are partially buried by younger olivine tholeiite lavas. The southeast side of East Butte lava dome has a distinctive concave inward shape that is likely produced by buttressing of that part of the dome by an older basaltic shield volcano (McCurry and Ganske 2005).

East Butte is compositionally similar to Big Southern Butte (Fig. 2). It is distinguished by abundant phenocrysts (10%) of sanidine > quartz > fayalite, ferroaugite, Fe-Ti oxides, and accessory zircon, chevkinite and apatite; rare

Table 1 Whole-rock Sr and Nd isotopic analyses (see footnote a)

Sample	Location	Age ^a	Rb (ppm)	Sr (ppm)	⁸⁷ Sr/ ⁸⁶ Sr ±2s (measured)	⁸⁷ Sr/ ⁸⁶ Sr (initial)	Sm (ppm)	Nd (ppm)	¹⁴³ Nd/ ¹⁴⁴ Nd ±2s (measured)	ε _{Nd}
7	Cedar Butte	400 ka	69.4	260	0.70684±1	0.70684	29.4	146	0.512426±9	−4.1
DI1R	Cedar Butte	400 ka	224.9	18.3	0.70739±1	0.70734	28.8	122	0.512397±20	−4.7
21	Cedar Butte	400 ka	260	2.56	0.71142±2	0.70967	21.5	80.7	0.512401±7	−4.6
DI5	Cedar Butte	400 ka	89.7	244	0.70695±1	0.70694	28.5	139	0.512413±6	−4.4
MC-23	Big Southern Butte	300 Ka	275	0.52	0.714361±13	0.70805	13.2	7.5	0.512434±6	−4.0
MC-24	East Butte	580 Ka	213	9.4	0.708493±11	0.70797	25.2	8.7	0.512403±6	−4.6
1209	CH-1 Unnamed Butte	1.4 Ma	215	10	0.710240±8	0.709043	24.5	104	0.512328±6	−6.1
1493	CH-1 Unnamed Butte	1.4 Ma	180	39	0.707172±8	0.706915	22.5	93	0.512398±6	−4.7
1727	CH-1 Unnamed Butte	1.4 Ma	67	154	0.706854±8	0.706829	13.5	54	0.512392±6	−4.8
1902	CH-1 Unnamed Butte	1.4 Ma	(66) ^a	207	0.706679±6	0.706661	(13) ^a	(54) ^a	0.512399±5	−4.7

One 10-kg sample of aphyric rhyolite is broken down to gravel sized fragments; two ~50 gm separates are split from the gravel and ground to powder using a tungsten-carbide puck-and-ring mill. Three samples from each separate are taken into solution using a standard Li-metaborate fusion technique. ICP-AES analyses of the six samples yield a mean value of 0.52±0.04 ppm (1σ).

^a Isotopic data for samples from Cedar Butte are from the University of Colorado isotope geochemistry laboratory using a Finnigan-MAT mass spectrometer; concentrations of elements are determined by isotope dilution, and are considered accurate to about 1% for Rb and Sr and 0.5% for Sm and Nd. Measurements of the La Jolla Nd isotopic standard yield a mean of 0.511840±5 (2σ; n=10). Analyses of the two MC- and CH-1 samples are from the Rice University isotope geochemistry laboratory using a Finnigan Mass Spectrometer model 262. Sr-standard NBS987 (⁸⁷Sr/⁸⁶Sm=0.71025); Nd-standard AMES, ¹⁴³Nd/¹⁴⁴Nd_{std}=0.51214. Nd analyses are obtained by INAA and Sr by ICP-AES and estimated 1s RSD of ~5%. Numbers in parentheses are concentrations inferred by interpolating respective trace element data between samples have similar major-element chemistry in borehole CH-1. Whole-rock K/Ar dates are from Kuntz et al. (1994). Rb, Sm and Nd concentrations for sample 1902 are estimates, and are shown in parentheses. Sample numbers correspond to those in Appendix Tables 2 and 5.

plagioclase phenocrysts are heavily mantled by plagioclase. Mafic enclaves are common albeit volumetrically minor (Spear 1979; Bretches 1984). Most are magmatic enclaves of basaltic trachyandesite; others are olivine tholeiite xenoliths (McCurry and Ganske 2005). The dome also contains rare felsic granitoid xenoliths (McCurry and Ganske 2005).

Like Big Southern Butte, East Butte rhyolite appears to be chemically homogeneous (Bretches 1984; Spear 1979; McCurry and Ganske 2005) (Fig. 2; Appendix Table 2, MC-24). Slightly higher concentrations of Eu (1.4 ppm) and Sr (9 ppm) may reflect some accumulation of sanidine or less fractionation of the East Butte rhyolite. Sr- and Nd-isotopes are among the least evolved of Eastern Snake River Plain volcanic field rhyolites, having ε_{Nd} and ⁸⁷Sr/⁸⁶Sr_i of −4.0 and 0.70805, respectively (Table 1; Fig. 3).

Cedar Butte volcano

Cedar Butte is a compositionally zoned ~400 ka basaltic trachyandesite–trachyandesite–trachyte–rhyolite volcanic center located ~10 km east of Big Southern Butte (Fig. 4) (Kuntz and Kork 1978; Kuntz et al. 1994; Spear 1979; Hayden 1992). Exposed parts of the volcano define a broad shield, ~4 by 9 km across and 120 m high, capped by a north to northwest trending, ~1-km-long curvilinear spatter rampart, and a 100-m-high tephra cone. Vents and exhumed dikes are north–northeast to northwest trending and are

concentrated near the center of the shield. The volcano has a minimum volume of 5 km³, however younger olivine tholeiite lavas bury much of it. Interpolation between surface exposures and correlative lavas in nearby boreholes indicate that the original volcano has a radius of 10–13 km, height of ~300 m (excluding the tephra cone), and cumulative volume of ~40 km³ (borehole data and volume calculations are summarized in Appendix Fig. 1 and Appendix Table 1, respectively).

Volcanic evolution

Based upon overlapping stratigraphic contact relationships of the exposed part of the shield (Hayden 1992) we identify five sequential stages of eruptive activity:

Stage 1 (oldest)

Effusive eruption of a ~0.25 km³ high-silica rhyolite lava flow from a vent located beneath the tephra cone that now caps Cedar Butte (Fig. 4, unit Qr2). The flow is up to 70 m thick, and is covered in many areas by a thin veneer of younger basaltic trachyandesite lava (Fig. 4, unit Qb3). Top and margins of the rhyolite flow are mantled by glassy and devitrified, dense to pumiceous flow breccia. Phenocrysts are sparse (~1–2% total volume) consisting mostly of euhedral to subhedral sodic sanidine (Or₄₅) > fayalite > quartz > Fe–Ti oxides and accessory zircon and apatite; some sanidine and fayalite crystals are rounded and embayed.

Stage 2

Mild explosive eruptions produce a compositionally zoned, clastogenic, siliceous trachyte to low-silica rhyolite lava flow (Fig. 4, unit Qr1). Orientations of stretched vesicles, glassy ribbon bomb orientations, and rheomorphic fold axes indicate the lava erupted from a vent that is now buried under the Stage 3 tephra cone. The lavas consist of ~1% euhedral to subhedral phenocrysts of anorthoclase (Or_{23}) > fayalite (Fa_{99}), ferroaugite, Fe–Ti oxides, and accessory apatite and zircon, in a very fine-grained hyalopilitic to intersertal, weakly flow-layered matrix. Glomerocrysts of sanidine occur in some samples.

Stage 3

Vigorous strombolian eruptions produce a 0.1 km^3 , 120-m-high tephra cone consisting dominantly of sparsely porphyritic black to reddish trachyandesite to trachyte scoria and bombs, and also minor but prominent white pumiceous rhyolite, and black obsidian (Fig. 4, unit Qpc). Exposures consist of moderately to weakly agglutinated, moderately to poorly sorted, weakly bedded scoria and breccia. Small, rootless flows occur in some areas. A loose carapace of scoria and breccia mantles other parts of the cone. Some pyroclasts are intricately interlayered with bands of contrasting lithologies, most commonly plagioclase-phyric black to brownish mafic, and sanidine-phyric white silicic glass. Age relationships between the cone and surrounding lava flows are evident in the erosionally breached northern part of the cone. In that area excellent exposures occur of lavas that onlap the original tephra cone, and indicate that the cone is an older, rather than a younger part of the Cedar Butte shield (cf. Spear 1979).

Mineral separates from a white, pumiceous, sparsely porphyritic rhyolite pyroclast (Appendix Table 2, sample 6249705A1) consist of sodic sanidine, quartz > fayalite ($\text{Fa}_{0.999}$) > ferroaugite ($\text{En}_{0.002}\text{Fs}_{57}\text{Wo}_{43}$) > magnetite solid solution (magnetite_{ss}) > ilmenite, accessory zircon, apatite and chevkinite. Ferromagnesian phases are remarkable for their extremely high Fe/(Fe+Mg) ratios (Appendix Table 3). A partial microprobe analysis of chevkinite is listed in Appendix Table 3.

Three en echelon, north–northeast-trending dikes are erosionally exhumed in the extreme northern flank of the tephra cone (Fig. 4). The southernmost dikes are wedge shaped, tapering towards the top, indicating that the original magma did not propagate to the surface at this location. Contacts with the host tephra are bulbous on a scale of 10's of centimeters to a meter. Tephra adjacent to the contacts has been fused into dense glass.

Southerly dike segments are strongly compositionally zoned, consisting of a core of rhyolite and a rind of

trachyandesite (Fig. 5). Contacts between the two lithologies vary from sharp to gradational over a few centimeters. Also in some places wispy blebs of one type occur within the other, indicating both were at least partially molten when they originally came into contact with other. Features of the dikes are consistent with processes of viscous segregation and encapsulation described by Carrigan (1994), likely resulting from simultaneous tapping of felsic and more mafic parts of an underlying magma chamber or chambers. Compositions overlap with those of pyroclasts of the tephra cone, which, when combined with the dike trends, suggests that the dikes are an extension of those that fed the tephra-cone-forming eruptions. The northernmost dike merges imperceptibly upward into trachyandesite lavas of Stage 4 (unit Qb4).

Stage 4

Mild strombolian eruptions and lava fountaining produce a prominent 1-km-long curvilinear spatter rampart located ~1 km north of the tephra cone (Qav, Fig. 4). The spatter rampart merges outward into four voluminous trachyandesite and trachyte lavas (Qb4,5,6 and 8 in Fig. 4 and Appendix Table 2). Both trachyandesite and trachyte lavas exhibit fluidal pahoehoe morphology. Similar lava flows are exposed in a north-tilted block of mafic rocks on the north flank of Big Southern Butte volcano (Fishel 1993), and in core from borehole USGS 124 at a depth of 114 m (Appendix Fig. 1 and Appendix Table 2), suggesting that Stage 4 lavas extend in the subsurface to at least 10 km west and southwest of Cedar Butte. The lavas contain 1–6% phenocrysts of plagioclase, sanidine, olivine, ferroaugite and magnetite, and accessory apatite ± zircon. Most phenocrysts are unzoned and euhedral to subhedral, and have compositions that correlate with whole-rock composition (Fig. 2). Plagioclase is boxy cellular in some lava flows. It is the dominant phenocryst in trachyandesite flows whereas sanidine is the dominant phenocrysts in trachyte lavas (Fig. 2). Most lavas contain phenocrysts that exhibit multiple textural varieties (e.g. both euhedral and anhedral, sieve textured or spongy cellular sanidine) and also scattered wispy magmatic enclaves of rhyolite up to several centimeters across.

Stage 5

Mild lava fountaining from a north-trending vent located south of the tephra cone produce a thin basaltic trachyandesite pahoehoe flow, mantling older deposits on the south side of Cedar Butte (Fig. 4). The lava flow is clastogenic near the source vent, and it contains sparse xenoliths of rhyolite resembling unit Qr2. Samples from the trachyandesite flow contain sparse subhedral to euhedral pheno-

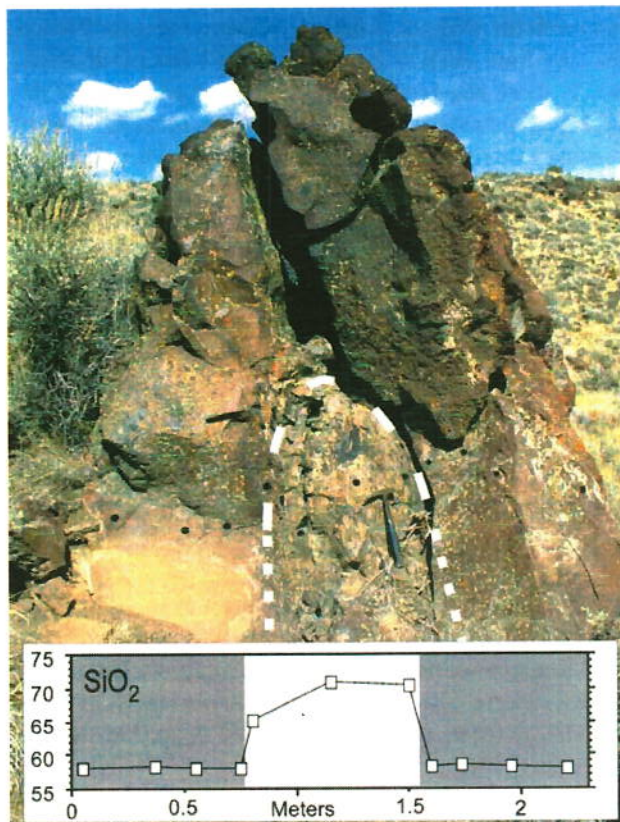


Fig. 5 One of three north–northeast trending, en-echelon dikes erosionally exhumed from beneath the northern flank of the tephra cone that caps Cedar Butte volcano (Qd in Fig. 4). A dashed line is drawn between inner, lighter colored, rhyolitic and outer trachyandesitic parts of the dike. Whole-rock silica contents of samples cored from across the dike are illustrated at the bottom of the figure

crystals of plagioclase (An_{39}), olivine (Fo_{39}), augite, magnetite and accessory apatite in a very fine-grained intersertal to intergranular matrix. Xenocrysts of quartz occur in some samples. They are rounded, strongly resorbed, and armored by fine-grained clinopyroxene.

No erosion or weathering is apparent at contacts between lava flows or pyroclastic units at Cedar Butte suggesting that the volcano is monogenetic. Additionally, samples from lavas of Stages 1, 2, 3 and 5 yield identical paleomagnetic inclinations, limiting formation of exposed parts of the shield to < 100's of years (Champion, USGS, personal communication).

Whole-rock and phenocryst compositions

Compositions of lavas, dikes and pyroclastic rocks from Cedar Butte are illustrated in Fig. 2. The rocks are high-K, dominantly metaluminous, and strongly ferroan, having

$FeO_{tot}/(FeO_{tot}+MgO)$ ratios of 0.89 to 1.00. They plot near the boundary between alkalic and alkali-calcic rock suites using the modified alkali-lime index of Frost et al. (2001). One rhyolite (Appendix Table 2, sample 3R) is slightly peralkaline ($Na_2O+K_2O/Al_2O_3=1.04$). Harker covariation patterns vary from linear or slightly curvilinear (e.g. CaO , FeO_{tot} , Na_2O) to strongly curvilinear (e.g. TiO_2 , MgO , P_2O_5 , many trace elements).

Sr- and Nd-isotopes of basaltic trachyandesite to rhyolite samples from Cedar Butte are illustrated in Fig. 3. $^{87}Sr/^{86}Sr_i$ increases systematically 0.70684–0.70967 with increasing silica (from 55 to 75%) and decreasing Sr (260 to 2.6 ppm). ϵ_{Nd} decreases from –4.1 in the most mafic sample (basaltic trachyandesite, sample 7) to –4.6 to –4.7 in rhyolitic samples DIR1 and 21 (Table 1).

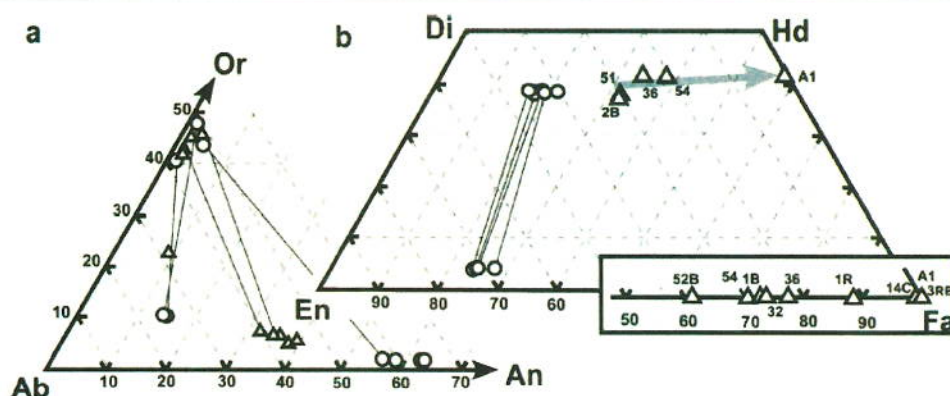
Assemblages of phenocrysts are illustrated in Fig. 2, and compositions in Fig. 6, and Appendix Table 3. Euhedral to subhedral phenocrysts are generally unzoned to weakly zoned, and vary systematically in composition with whole-rock silica content (Fig. 6). Clinopyroxene and olivine compositions exhibit strong increases in $Fe/Fe+Mg$ (Fig. 6b), with the most felsic rocks containing nearly pure Fe-endmembers. Alumina contents of pyroxenes in the most siliceous rocks (0.39% Al_2O_3) are lower than those in more mafic rocks (1.64–2.01% Al_2O_3). Plagioclase An-content decreases slightly from $\sim An_{39}$ to $\sim An_{32}$ between 55 and 66% silica. Pyroxene and olivine outliers in Fig. 6 likely resulted from inmixing of phenocrysts derived from magmatic enclaves.

Intensive variables

Textures of phenocrysts in most of the samples examined from Cedar Butte are indicative of multiple origins including precipitation from the host magma, formation as partially consolidated parts of the chamber wall, and xenocrysts derived from mafic or felsic magmatic enclaves. We infer equilibrium assemblages based upon phenocryst textures (e.g. Vernon 2004), and plausible mineral exchange K_d 's (e.g. Putirka et al. 2003a; Reubi and Nicholls 2005 and references therein). We specifically exclude crystals occurring in glomerocrysts, or that exhibit strong evidence of resorption or skeletal growth.

We apply a variety of mineral geothermometers and barometers to estimate magmatic intensive variables for Cedar Butte lavas (Appendix Table 4). Fe–Ti oxides are weakly to moderately exsolved, do not yield equilibrium Mg/Mn compositions (Bacon and Hirschmann 1988), and are therefore not used in determining temperature or oxygen fugacity. QUILF (Andersen et al. 1993), SolvCalc (Wen and Nekvasil 1994) and clinopyroxene-liquid geothermometers (Putirka et al. 2003a) yield similar temperatures that decline with increasing silica from $\sim 1,000^\circ C$

Fig. 6 Compositions of feldspar (a), pyroxene (b) and olivine (b–inset) in mole percent components (abbreviations from Kretz 1983), from Cedar Butte (triangles) and Unnamed Butte (circles). Coexisting minerals are connected by tie lines. Gray arrow in (b) highlights the pattern of Fe-enrichment of Cpx with increasing whole-rock rock silica content (numbers refer to whole-rock samples in Appendix Table 2)



for basaltic trachyandesite to $\sim 900^{\circ}\text{C}$ for trachyte (Appendix Fig. 2). The augite-olivine geothermometer of Loucks (1996) yields higher temperatures for the most mafic rocks ($\sim 1,100^{\circ}\text{C}$) but overlaps with the other methods for more silicic rocks. The clinopyroxene-liquid thermobarometer of Putirka et al. (2003b) yields temperatures similar to those of the augite-olivine geothermometer, and pressures of <1 Kb. A low equilibration pressure is also consistent with the clinopyroxene geobarometer of Nimis (1999). The Putirka (2005) Pl-L thermobarometer yields the highest temperatures ($\sim 1,115^{\circ}\text{C}$ for basaltic trachyandesite), and also indicates that those phenocrysts may have evolved polybarically (4–8 Kb) and under hydrous conditions (up to 4% H_2O). Zircon saturation temperatures (Watson and Harrison 1983) overlap with temperatures produced by the other methods for trachyte, and decrease to $\sim 870^{\circ}\text{C}$ for the most evolved rhyolites. Discordances among some of these preliminary data likely indicate that some of the selected mineral pairs were not in equilibrium in the original magma.

Discussion

Exposed parts of Cedar Butte volcano were produced by rapid (< 100 's years) sequential eruption of phenocryst-poor rhyolite, followed by trachyte, trachyandesite, and finally basaltic trachyandesite magma. Intermediate compositions dominate. Composite dikes, magmatic enclaves and disequilibrium phenocryst textures, compositions and assemblages indicate that diverse magma compositions were simultaneously present, and that mixing occurred at various times and scales prior to eruption. Xenocrysts and glomerocrysts are common, but many other phenocrysts have morphologies and compositions that are consistent with equilibrium crystallization.

Harker diagrams in Fig. 2 illustrate that the whole-rock compositional variations are strongly correlated, and produce patterns that vary from linear (e.g. FeO^*) to curvilinear (e.g. MgO); others have pronounced inflections (e.g. K_2O , Ba). The latter two patterns are inconsistent with

two-component magma mixing. However both correlate with changes in the assemblages and compositions of enclosed phenocrysts in the rocks (top right panel of Fig. 2) in a manner that is consistent with fractionation of those minerals from the parent magmas. For example, between 60 and 62% silica, plagioclase gives way to sanidine as the dominant phenocryst type. In this same silica range the gradient in K_2O flattens out, the rate of Sr-decline is increased, the positive gradient of Rb increases, and positive gradients in Ba and Eu exhibit pronounced reversals to negative gradients. Leeman and Phelps (1981) and Ren (2004) define respective sanidine/melt partition coefficients of Ba (3–36), Sr (8–21), Rb (0.3–0.6) and Eu (6–9) for rocks similar to those at Cedar Butte. Each are consistent with sanidine-dominated fractionation in rocks containing $>61\%$ SiO_2 . Other distinctive patterns include a steady decline in Sc that is consistent with continuous pyroxene fractionation; a decline in Zr correlates with first appearance of zircon at ~ 58 –61% SiO_2 . A prominent drop in La occurs, and a possible decline in the Th occurs above 70%, correlating with appearance of Th- and LREE-rich chevkinite (Appendix Table 2, sample 6249705A1). Chevkinite fractionation has the effect of reversing LREE enrichment patterns typical of intermediate to felsic rocks (e.g. MacDonald and Belken 2002; Wolff and Storey 1984). Rocks at Cedar Butte therefore record features of both magma mixing and fractional crystallization, however magma mixing appears to have had little general effect on original intermediate magma compositions.

Cedar Butte and Craters of the Moon lava fields

Lavas exhibiting evolved compositions overlapping those at Cedar Butte lavas occur in many parts of the eastern Snake River Plain (e.g. Leeman 1982b). The most voluminous and compositionally diverse occur at Craters of the Moon (e.g. Kuntz et al. 1986, 1992; Stout et al. 1994). Figure 2 illustrates overlapping patterns of whole-

rock composition between Craters of the Moon and Cedar Butte. Collectively these exhibit a remarkably wide and coherent patterns of major and trace element variation (e.g. SiO_2 45–75%; Rb and Sr from 40 to 280 and 1 to 320 ppm, respectively). Zirconium is a rare exception. In lavas from Cedar Butte whole-rock Zr-concentrations increase from 2,700 to 3,300 ppm between 55 and 59% SiO_2 , thereafter decreasing in an approximately linear fashion to ~470 ppm in the most siliceous rhyolites. The decline in Zr-concentration correlates closely with the appearance (with increasing silica) of zircon as a phenocryst phase (Fig. 2) suggesting that the decline in whole-rock Zr concentration with silica increase is a result of zircon fractionation. In contrast, Zr-concentrations in lavas at Craters of the Moon increase in an approximately linear fashion between 45 and 63% SiO_2 , from ~750 to ~1,500 ppm; consistent with observations by Stout et al. (1994) and Kuntz et al. (1986) that zircon is absent or xenocrystic in all but the most evolved rocks. Origin of this particular discordance between Cedar Butte and Craters of the Moon is not clear. Samples from other evolved volcanoes of Eastern Snake River Plain volcanic field (small squares on Fig. 2) overlap, or are intermediate between Craters of the Moon and Cedar Butte types, or plot along a Unnamed Butte-type mixing line.

As at Cedar Butte, lavas at Craters of the Moon have isotopic compositions that are intermediate between those of Archean crust and olivine tholeiite (Leeman 1982b; Futa et al. 1988). However for rocks of similar whole-rock composition, those at Craters of the Moon are more radiogenic (Fig. 3b), indicating that magmas in that area

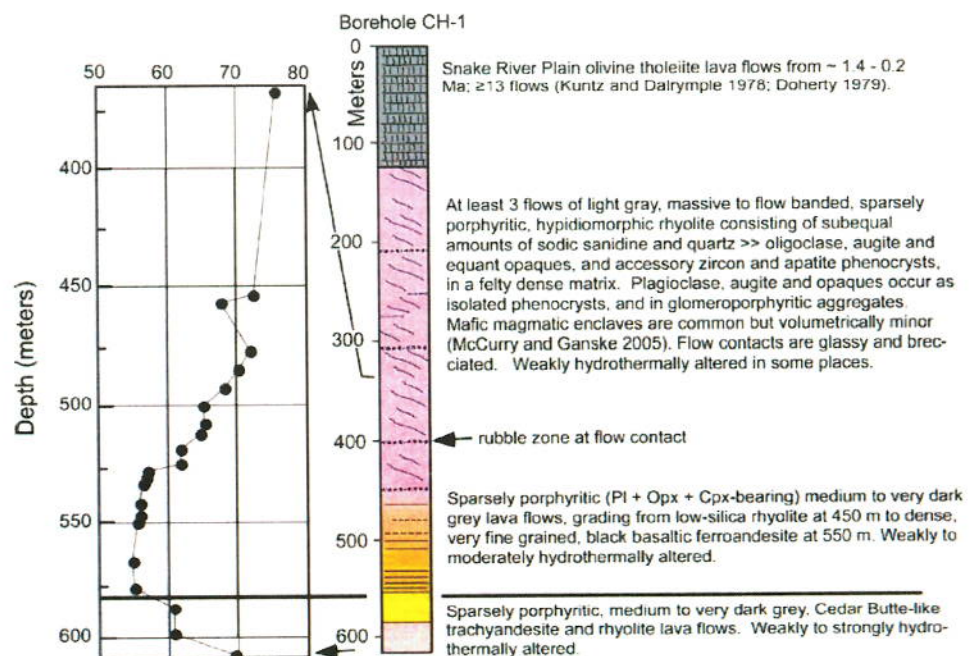
interact more strongly with Archean continental crust than those along the Axial Volcanic Zone (Menzies et al. 1984; Futa et al. 1988).

Unnamed Butte

We refer to an intermediate to silicic volcanic center located ~21 km northeast of Cedar Butte volcano as Unnamed Butte (Fig. 1d). Only the upper 50 m of the volcano is exposed ($43^\circ 29.12'N$, $112^\circ 4.35'W$); lower parts have been buried by younger olivine tholeiite lava flows. Stratigraphically lower parts are cored in Idaho National Laboratory borehole CH-1 (Kuntz and Dalrymple 1979; Doherty et al. 1979; Spear 1979; Morse 2002; Ganske 2006) located ~600 m northeast of surface exposures (Fig. 1d). Kuntz et al. (1994) dated rhyolite from the upper, exposed part of the volcano at 1.400 ± 0.028 Ma.

Examination of core from borehole CH-1 indicates that, beneath ~120 m of olivine tholeiite lavas, there are at least three thick high-silica rhyolite lava flows, cumulatively 330 m thick (Fig. 7). The flows are sparsely porphyritic containing ~2% euhedral to subhedral sanidine > quartz > plagioclase, ferroaugite and minor fayalite (?). The rhyolite lavas in turn overlie a 160-m-thick, ~6 flow sequence of subalkaline low-silica rhyolite, trachydacite, trachyte and ferroandesite, and then ~8 flows of basaltic ferroandesite (Figs. 2 and 7). The flows are systematically more mafic downward in the borehole to a depth of 560 m. The lavas are sparsely porphyritic containing varying proportions of

Fig. 7 A simplified lithologic log of Borehole CH-1, located ~600 m northeast of Unnamed Butte. Whole-rock silica contents are illustrated for samples from a segment of the borehole between ~350 and 610 m depth. Shaded line at 580 m depth separates shallower lavas (120–580 m) that exhibit linear, magma-mixing-like compositional covariations (Fig. 2), and also identical paleomagnetic inclinations, from deeper lavas (580–610 m) that have Cedar Butte-like compositions, and complex, possibly hydrothermally overprinted, paleomagnetic signatures (D. Champion, USGS, personal communication)



plagioclase, sanidine, quartz, hypersthene, augite, and magnetite. Plagioclase-pyroxene-bearing glomerocrysts and aphanitic mafic magmatic enclaves are common but volumetrically minor throughout the lava sequence (McCurry and Ganske 2005). Enclaves are typically aphyric and resemble olivine tholeiite petrographically. Flows in this depth range (120–560 m) have nearly identical secular paleomagnetic field inclinations, indicating that they accumulated within <100's years (D. Champion, USGS, personal communication).

At a depth of ~560 m in borehole CH-1, there is a prominent downward break in the trend in lava composition and phenocryst assemblages (Fig. 7). Sequentially downward in the borehole are trachyandesite (560–590 m) and low-silica rhyolite (590–610 m) lavas. The borehole bottoms out within the low-silica rhyolite flow. Trachyandesite lava consists of ~2% phenocrysts of subhedral to euhedral plagioclase > augite ± olivine (psuedomorphed by smectite clay) > magnetite ± sieve-textured sanidine ± accessory apatite; low silica rhyolite lava is sparsely porphyritic containing <1% phenocrysts of sanidine + fayalite (psuedomorphed by smectite clay) + ferroaugite + magnetite.

Whole-rock composition

High-silica rhyolites of Unnamed Butte are similar in major and trace element composition to those at Cedar Butte (Fig. 2; Appendix Table 5). However compositions of low-silica rhyolite and intermediate lavas between 120 and 560 m depth in borehole CH-1 contrast markedly with Cedar Butte lavas or other lavas of the Eastern Snake River Plain volcanic field. They have strikingly linear major- and trace-element covariation (Fig. 2). Highly correlated ($r^2 \sim 1$) least-square-fit lines extend from Eastern Snake River Plain volcanic field high-silica rhyolite toward the compositions of olivine tholeiite. This spectrum of Unnamed Butte lava compositions could be produced by simple two-component mixture of 0–72% of olivine tholeiite and a typical Eastern Snake River Plain-type high-silica rhyolite (Fig. 2). We refer to this sequence of rock types as the “Unnamed Butte suite.”

Low silica rhyolite and trachyandesite lavas occurring deeper than 560 m in borehole CH-1 plot well off compositional trends defined by the shallower rocks (Fig. 2). Both flows have evolved major- and trace-element compositions, and phenocryst assemblages that overlap those of Cedar Butte lavas, rather than shallower lava flows of borehole CH-1.

Mineral chemistry and paragenesis

Compositions and assemblages of phenocrysts are illustrated in Fig. 6. Rhyolites contain 3–5% euhedral to subhedral

phenocrysts of sodic sanidine (Or_{40}), quartz, plagioclase (An_{15}), ferroaugite, augite ($\text{En}_{41}\text{Fs}_{21}\text{Wo}_{39}$) ± hypersthene, Fe–Ti oxides, zircon and apatite (Appendix Table 6, samples 1209 and 1493). Spear (1979) reports occurrence of minor amounts of fayalite. Low silica rhyolites contain bimodal populations of plagioclase (oligoclase and labradorite) phenocrysts. Trachydacite, trachyte, ferroandesite and basaltic ferroandesite lavas contain 3–5% phenocrysts and glomerocrysts of subhedral to euhedral labradorite (An_{55-63}), hypersthene ($\text{En}_{69-73}\text{Fs}_{24-27}\text{Wo}_{3.9-4.1}$) > augite ($\text{En}_{43-45}\text{Fs}_{17-18}\text{Wo}_{38-39}$), ± magnetite_{ss}. Magnetite phenocrysts increase in abundance from trace to minor amounts with increasing whole-rock silica content. Rare rounded and embayed quartz and anhedral, embayed and sieve-textured sanidine occurs in some samples.

Intensive variables

Estimates of intensive variables at Unnamed Butte are complicated by strong textural and geochemical evidence for phenocryst-liquid disequilibrium. Fe–Ti oxides of the most siliceous rhyolite may be an exception. Their compositions satisfy the Bacon and Hirschmann (1988) Mn–Mg partitioning equilibrium test, and yield reasonable temperatures and near-QFM oxygen fugacity, albeit with large uncertainties ($993 \pm 157^\circ\text{C}$ and $\log f_{\text{O}_2} = -11.04 \pm 1.4$). In contrast, phenocrysts that dominate in intermediate rocks (labradorite, augite and hypersthene) fail similar mineral-melt equilibrium exchange tests (e.g. Reubi and Nicholls 2005; Putirka et al. 2003a). There is also no systematic correlation between the mineral and whole rock compositions.

Although pyroxene and labradorite phenocrysts are not in equilibrium with the surrounding melt, as represented by their host rocks, the observation that they occur frequently, have narrow ranges of compositions, and have subhedral to euhedral textures suggests that they are inherited from a common magmatic source. Coexisting ortho- and clinopyroxene phenocrysts in basaltic ferroandesite, trachydacite and low-silica rhyolite yield similar QUILF temperatures of $\sim 1140^\circ\text{C}$ (Appendix Table 4), consistent with that source having a composition that is intermediate between basalt and basaltic ferroandesite. This hypothetical magma is unlikely to be olivine tholeiite because pyroxene phenocrysts, particularly the assemblage augite and hypersthene, are absent elsewhere on the Snake River Plain (e.g. Leeman 1982a). If true, then this hypothesis indicates that there are at least two stages of magma mixing involved in the production of the Unnamed Butte lavas, one which forms the parent for labradorite and pyroxene phenocrysts, and a second in which that magma mixes with Unnamed-Butte-type high-silica rhyolite.

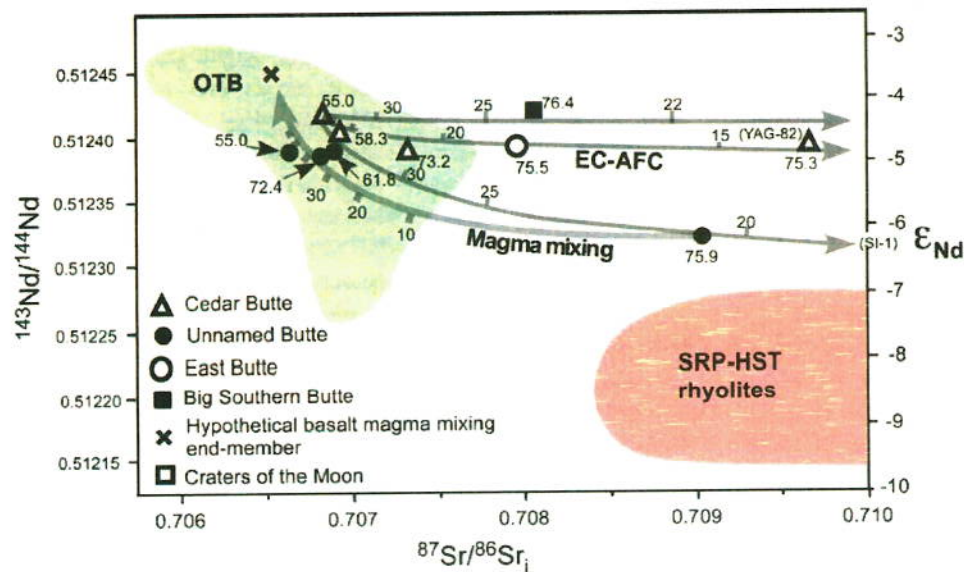


Fig. 8 Plot illustrating Nd- and $\text{Sr}_{(i)}$ -isotopic compositions of intermediate to rhyolitic rocks from the Eastern Snake River Plain. Whole-rock silica contents are shown adjacent to corresponding samples. Colored fields represent the range of isotopic compositions for Quaternary olivine tholeiite basalt (green) and Tertiary Snake River Plain hotspot track rhyolites (red). Curves illustrate EC-AFC (Bohrson and Spera 2001) and magma mixing models. Tick marks on EC-AFC curves (right directed arrowed curves) indicate percent

residual liquid, assuming a trachyandesite parental magma (sample 52B, Appendix Table 2), and assimilant compositions as represented by Archean crustal xenoliths from the region (from Leeman et al. 1985). The left-directed arrow represents a hypothetical magma-mixing curve for Unnamed Butte-type rhyolite (sample 1209, Appendix Table 2) and SRP-olivine tholeiite basalt (sample 2-2A-54, Hughes et al. 2002a). Tick marks indicate the mass percent of olivine tholeiite basalt in the mixture

Nd- and Sr-isotopes

Nd- and Sr-isotopic analyses of Unnamed Butte lavas are illustrated in Fig. 8. Basaltic ferroandesite, ferroandesite and low-silica rhyolite samples overlap isotopically with olivine tholeiite. High-silica rhyolite (sample 1209) has a higher $^{87}\text{Sr}/^{86}\text{Sr}_i$ (~0.709) and has the lowest observed ϵ_{Nd} (–6.1) of Eastern Snake River Plain volcanic field rhyolites.

Basaltic ferroandesite and ferroandesite samples plot near to a mixing hyperbola that connects hypothetical parent basalt (or basaltic ferroandesite) and rhyolite magmas (Fig. 8). End-member basalt Nd- and Sr-isotopic values are chosen to be typical of Eastern Snake River Plain olivine tholeiite (Hughes et al. 2002a). Mixing proportions of 75 and 85% olivine tholeiite, respectively, are similar to those determined from major and trace element mass balance (50 and 72%, respectively). However, low silica rhyolite sample 1493 (72.4% SiO_2 ; Fig. 8) has higher ϵ_{Nd} (–4.7) and lower $^{87}\text{Sr}/^{86}\text{Sr}_i$ (0.7069) ratios than anticipated from simple mixing (–5.5; 0.7072). Simple two-component mixing cannot produce these Nd- and Sr-isotopic data. The isotopic data are consistent with the hypothesis that the mixing evolution of this system is more complex than is apparent from compositional and stratigraphic data alone.

Discussion

Although Unnamed Butte volcano is largely buried, we infer that it is dominantly shield-like in morphology, with stubby silicic flows or domes near the summit region, and more spatially extended intermediate flows, probably much like Cedar Butte. We correlate an 8-m-thick trachyte flow (Appendix Table 5, samples 1931 & 1965) cored at 565–573 m depth with a flow 11 km north of CH-1 identified at 580–604 m depth in borehole ANL-OBS-AQ-014 (Champion and Lanphere 1997). The correlation is based upon similarities in whole-rock composition, paleomagnetic signatures and inferred age (Duane Champion, USGS, personal communication). Borehole data indicate that lavas of Unnamed Butte volcano have a cumulative thickness of ≥ 490 m, at least twice the maximum thickness of the Cedar Butte. Simple extrapolation of those measurements indicates that Unnamed Butte has cumulative volumes of rhyolite and mafic to intermediate lavas of ~1 and 20 km^3 , respectively (Appendix Table 1).

Downward changes in lithologies at 560 m depth in borehole CH-1 coincide with a prominent change in the secular paleomagnetic field inclinations of the rocks (D. Champion, USGS, personal communication). In contrast to shallower rocks, those below 560 m exhibit widely variable

secular paleomagnetic field inclinations, probably owing to low-T hydrothermal alteration (Champion, USGS, personal communication), and possibly to a hiatus between the groups of lavas. We infer that Unnamed Butte evolved in two distinct stages, an older stage resembling Cedar Butte, and a younger stage, perhaps following a gap in time, which was dominated by magma mixing.

Eastern Snake River Plain volcanic field: summary and discussion

We distinguish three coeval petrogenetic suites of volcanic rocks in the Eastern Snake River Plain volcanic field: a volumetrically dominant olivine tholeiite suite, a high-K trachybasalt to high-silica rhyolite suite (Craters of the Moon–Cedar Butte suite), and a subalkaline basaltic ferroandesite to high-silica rhyolite suite (Unnamed Butte suite) (Fig. 2). These rocks stratigraphically overlie, and are petrogenetically and volcanologically distinguished from late Miocene to early Pliocene rhyolites of the Yellowstone–Snake River Plain hotspot track (Christiansen and McCurry 2007). Exposures of Quaternary high-silica rhyolite occur where the rate of olivine tholeiite accumulation is highest, and where the crust is most mafic (Sparlin et al. 1982; Peng and Humphreys 1998). Compositionally and mineralogically similar rhyolites have been erupted at least four times over the last 1.4 m.y. As illustrated in Fig. 2, the rhyolites have low CaO, MgO, Sr, Ba, Eu, Sc and high total alkali-contents, high Ta and Rb contents, and high $K_2O/(Na_2O+K_2O)$ and $FeO_{tot}/(FeO_{tot}+MgO)$ ratios, compared to most older rhyolites of the Yellowstone–Snake River Plain hotspot track (Christiansen and McCurry 2007), but overlap in composition with some of the most evolved extracaldera Yellowstone rhyolites (Hildreth et al. 1991; Halliday et al. 1991). Eastern Snake River Plain rhyolites are even more distinct when compared to Tertiary rhyolites from elsewhere in the Basin and Range (e.g. Christiansen and McCurry 2007). The compositions of these rocks, and their anhydrous phenocryst assemblages (e.g. sanidine + quartz + fayalite + ferroaugite), strongly resemble those of A-type granites (e.g. Frost et al. 2001; Eby 1990, 1992), and the Eastern Snake River Plain volcanic field rhyolites may be extrusive equivalents of those types of rocks (e.g. Frost and Frost 1997; McCurry and Ganske 2005). Nd- and Sr-isotopic compositions (Figs. 3 and 8) of Eastern Snake River Plain volcanic field rhyolites overlap isotopically primitive olivine tholeiite in ϵ_{Nd} , and have only slightly more radiogenic $^{87}Sr/^{86}Sr_i$ (0.706–0.708 vs. 0.708–0.710).

At least two Eastern Snake River Plain volcanic field rhyolites (Cedar Butte and Unnamed Butte) erupted from otherwise intermediate-composition shield volcanoes. A third, East Butte, contains magmatic enclaves of interme-

diat composition. Major and trace element, and isotopic compositions of rhyolites plot along smooth extensions of covariation curves defined by the intermediate rocks (Figs. 2, 3 and 7). We interpret this repeated composition association as indicating that a genetic relationship exists between the intermediate and rhyolitic rocks.

The single exception to the intermediate-silicic rock field association is Big Southern Butte. It is also the most voluminous ($\sim 10 \text{ km}^3$) and has a unique evolution that is distinguished by storage and inflation of a laccolith at shallow depth ($\sim 1 \text{ km}$), followed by extrusion (Spear and King 1982). We speculate that shallow storage of the large volume of rhyolite inhibited ascent of denser, more mafic magma either from within the laccolith, or from a deeper, compositionally zoned parental magma chamber.

Our more detailed study of evolved volcanic centers near the center of the Eastern Snake River Plain supports a preliminary inference by Leeman (1982b) that there is a repeating process of magma generation in the Eastern Snake River Plain volcanic field that has resulted in the production of rocks having similar highly evolved compositions. We extend that idea to include compositions as evolved as high silica rhyolite. We test two petrogenetic models of rhyolite formation, the first involving partial melting of preexisting Archean crust (e.g. Leeman 1982b and c; Spear 1979), and Archean crust strongly modified by intrusion of Tertiary–Quaternary basalt (e.g. Hildreth et al. 1991; Frost and Frost 1997; Christiansen and McCurry 2007), and the second, fractional crystallization with and without assimilation.

Rhyolite petrogenesis

Crustal anatexis

Leeman (1982c) proposes a model of rhyolite formation along the Yellowstone hotspot track in which mantle-derived basaltic magma heats the crust sufficiently to produce felsic fractional melts. Spear (1979) applied a similar idea to rhyolites of the central Eastern Snake River Plain volcanic field. Hildreth et al. (1991) modified this idea for Yellowstone, by adding that the basalts may effectively hybridize with preexisting crust, acting to make the crust more mantle-like in its isotopic and chemical characteristics. Strong isotopic contrasts, particularly with regard to Nd, between mantle-like Eastern Snake River Plain volcanic field rhyolites ($\epsilon_{Nd} = -4$ to -6 ; $^{87}Sr/^{86}Sr_i = 0.708\text{--}0.710$), and radiogenic Archean crust ($\epsilon_{Nd} = -23$ to -47 ; $^{87}Sr/^{86}Sr = 0.702\text{--}0.892$; values from Leeman et al. 1985) precludes the possibility that Eastern Snake River Plain volcanic field rhyolites (or magmas parental to the rhyolite) formed solely by fractional melting of lower or

upper Archean crust (Fig. 3). Simple mixing calculations done with various combinations of olivine tholeiite (Hughes et al. 2002a) and Archean crust (Leeman et al. 1985) limit the amount of crustal mass that can be added to Big Southern-, East- or Cedar Butte, to <1 to 10%. Some of the most compositionally primitive low-K basalts (e.g. 2-2A-717; Hughes et al. 2002a) have an even lower ϵ_{Nd} (−5.6) than these rhyolites.

Geophysical (Peng and Humphreys 1998; Sparlin et al. 1982) and structural analyses of the Eastern Snake River Plain (Rodgers et al. 2002; McQuarrie and Rodgers 1998; Anders and Sleep 1992) suggest that large amounts of basaltic magma have been added to the crust, likely producing domains of gabbroic or metagabbroic composition with mantle-like isotopic signatures. Frost and Frost (1997) proposed that partial melting of mafic rocks produced by solidification of olivine tholeiite magma ponded near the Moho could produce Yellowstone-hot-spot-type rhyolites. Fractional melting of those domains, particularly if hydrated, might be able to produce rhyolitic magma with mantle-like isotopic ratios, or intermediate magmas that could subsequently differentiate to form such rhyolite (e.g. Patiño Duce 1995, 1999; Springer and Seck 1997; Streck and Gruner 2007). However, it is unlikely that such a process could simultaneously produce the observed wide range of crystal poor, evolved mafic to rhyolitic lavas observed at Cedar Butte, Unnamed Butte, East Butte and Craters of the Moon.

Fractional crystallization

Spear (1979) first proposed that trachyte at Cedar Butte is produced by fractional crystallization of basaltic trachyandesite. He based this proposal on similarities of the rocks with those at Craters of the Moon, adopting fractionation models proposed by Leeman et al. (1976) and Thompson (1975). However, Spear specifically excluded a fractional crystallization origin for coeval rhyolites on the basis of their unusually low LREE/HREE ratio. We have documented the occurrence and potential fractionation of LREE-rich chevkinite at Cedar Butte, effectively removing the basis of Spear's argument. McCurry and Ganske (2005) describe chevkinite at East Butte as well, and we expect that it has generally played a role in the REE evolution of Eastern Snake River Plain volcanic field rhyolites. We quantitatively test the plausibility of a fractional crystallization origin for Eastern Snake River Plain volcanic field rhyolites, first by major element least-squares mass-balance (Bryan et al. 1968), and then by combined assimilation and fractional crystallization, using a recent energy-constrained model (EC-AFC) of Spera and Bohrsen (2001) and Bohrsen and Spera (2001).

Mass balance model

Figure 9 illustrates our preferred mass-balance model. Basaltic trachyandesite sample 52B (Appendix Table 2) is used as the parent magma composition. Sample 52B, and other samples used in the modeling sequence, are sparsely porphyritic, therefore their whole-rock compositions approach a liquid. We divide the fractionation path into three steps in order to accommodate changes in phenocryst assemblages (Fig. 2), most importantly the cessation of plagioclase and initiation of alkali feldspar fractionation in Step 2 (Fig. 9). Samples 48 (trachyandesite), G124 (low-silica rhyolite) and 21 (high-silica rhyolite) are taken to represent sequential compositional states of the magmatic system. Compositions of phenocrysts are taken from those occurring in rocks within the respective fractionation step (shown as subscripts in Fig. 9). Cumulative percentages of residual liquid are shown at the end of each fractionation step. The model indicates that fractionation of basaltic trachyandesite would yield 23% residual high-silica rhyolite magma. Comparisons of modeled and observed compositions are indicated by sum of squares of residuals for each fractionation step (e.g. Bryan et al. 1968). All are less than 0.25 indicating a close match for the major elements. The fractionation model is therefore consistent with major-element mass balance limitations.

Figure 9b and c illustrate examples of compatible and incompatible trace element fractionation that are implied by the mass balance model. Bulk partition coefficients are calculated for Ba, Sr and Rb by combining modes and amounts of fractionation determined in the mass balance model with selected published mineral partition coefficients (mineral and calculated bulk partition coefficients are listed in Appendix Table 7). Trace element Rayleigh fractionation paths are shown with dashed lines. Shaded areas encompass variations in fractionation pathways that would occur by making plausible variations in the assumed mineral partition coefficients. Most encompass the observed spectrum of rock compositions. Exceptions include Sr (Fig. 9c, Step 1) and possibly Rb (Fig. 9c, Step 3). The former may result from earlier than expected fractionation of alkali feldspar; the latter may be a result of a poor selection of the low-silica rhyolite model parent composition. Despite these differences we believe the fractionation models match the observed trace element data reasonably well.

Our mass balance model implies that cumulates varying from ferrogabbro to mela-syenite reside in the upper crust. Unrupted evolved magmas could solidify to form a spectrum of rocks varying from quartz monzodiorite–quartz-monzonite–leuco-quartz-monzonite–granite–leucogranite (cf. Anderson et al. 2003; Mushkin et al. 2003; Vander Auwera et al. 2003). This assemblage of rocks would be consistent with on-going development of an A-

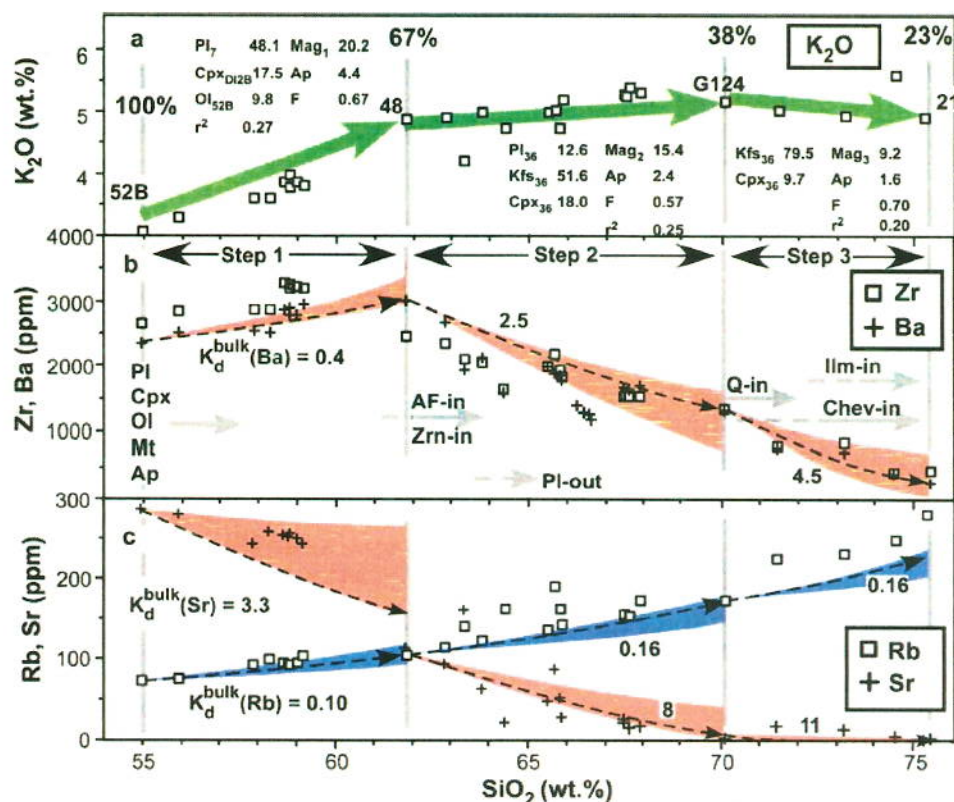


Fig. 9 a A least-squares mixing model of fractional crystallization for Cedar Butte volcano. Four samples are selected from Appendix Table 2 to represent the extremes of the observed compositional range (52B and 21), and whole-rock silica contents at which important changes in mineral assemblage occur (48 and G124). Mineral compositions are from Appendix Table 3. Model results are overlain in Fig. 9a with analyses of rock samples from Cedar Butte (squares). Bold arrows connect model parent–daughter magma compositions for the three-step model. Percentages of fractionated minerals are shown for each model step (abbreviations from Kretz 1983); mineral subscripts refer to specific minerals in Appendix Table 3; F = liquid fraction; r^2 = sum of the squares of the differences between model and observed compositions. Cumulative percent residual liquid is shown above shaded vertical lines at the ends of each fractionation step. Fig. 9b and c

illustrate models of trace element fractionation. Dashed lines illustrate model Rayleigh fractionation curves for Ba, Rb and Sr. They are based upon mineral assemblages and masses calculated from part (a), and estimates of the respective mineral partition coefficients (e.g. Leeman and Phelps (1981); Ren et al. 2003; Ren 2004; White 2003), and references therein) (Appendix Table 7); corresponding bulk K_d s are shown next to each line. Shaded regions bounding the curves encompass model compositions produced by plausible variations of individual mineral K_d s. Important changes in phenocryst assemblages are also illustrated. Zircon data are plotted in Fig. 9b to illustrate a correlation between the appearance of zircon as a phenocryst in the lava series, and a corresponding decline in whole-rock Zr content with increasing silica content

type batholith beneath the Eastern Snake River Plain (e.g. Frost and Frost 1997).

EC-AFC model

To accommodate small, but statistically significant systematic changes in Nd- and Sr-isotopes, we apply the Energy-Constrained Assimilation and Fractional Crystallization (EC-AFC) model of Spera and Bohrsen (2001) and Bohrsen and Spera (2001). Three solutions, constrained to satisfy Sr-concentration and Nd- and Sr-isotopic data, as well as plausible energy constraints (e.g. initial thermal conditions of magma and country rock, etc.), are illustrated in Fig. 8. The models yield fractionation factors (i.e. mass fraction of residual liquid) that decrease from 1 for the parent magma, to 0.24 to 0.14 for rhyolite residual liquids. These are similar to

values determined independently from major-element mass balance (~0.23, Fig. 9). The EC-AFC model also requires 0.03–0.04% (Cedar Butte, East Butte, Big Southern Butte) to 7% (Unnamed Butte) assimilation of plausible types of Archean crust. Assimilating less radiogenic materials such as Tertiary volcanic or plutonic rocks beneath the Eastern Snake River Plain would require more, but still small amounts of assimilation. Assimilation of shallow (<10 km) rocks would also introduce thermal energy problems (e.g. Huppert and Sparks 1988) unless the ratio of Quaternary basalt intrusive to extrusive rates is high (cf. Kuntz et al. 1992). Although a high ratio is possible, we believe it is unlikely given the low rate of flexure-related tectonic subsidence in the region during the late Pliocene and Quaternary (Rodgers et al. 2002; McQuarrie and Rodgers 1998). Although our models are simplistic, given the possible complexities of crustal

structures and rock bodies, we think it is clear that the rhyolites are dominated by a source that ultimately resides in the “mantle”. We conclude that mass- and energy-constrained AFC magma evolution models are consistent with observed compositional and mineralogical features of the Eastern Snake River Plain volcanic field rhyolites and intermediate rocks.

Fractionation mechanism

Bachmann and Bergantz (2004) summarize four principal mechanisms for melt removal from a crystalline residue: crystal settling; fractionation in a solidification front; melt extraction from crystal mushes; and convective fractionation. Extensive crystal settling in near-liquidus magma stored in cool, upper-crustal magma chambers is unlikely owing to viscosity and convection problems (Bachmann and Bergantz 2004), particularly over the short residence times likely for the small magma chambers inferred to have been parental to the Eastern Snake River Plain volcanic field rhyolites (e.g. Hawkesworth et al. 2000; Reid 1995). Alkali feldspar would also have near neutral buoyancy in Fe-rich anhydrous intermediate magmas.

Solidification front and melt extraction ideas produce highly evolved, phenocryst poor rhyolite magma from crystal rich intermediate magma reservoirs. Whitaker et al. (2007) demonstrate experimentally that large amounts of equilibrium crystallization of Snake River olivine tholeiite can produce a wide spectrum of melts resembling those of the Craters of the Moon-Cedar Butte suite. However it is unclear how any of these processes would produce, essentially simultaneously, a wide spectrum of crystal-poor intermediate to silicic melt compositions.

Convective fractionation (i.e. sidewall crystallization) is a process in which parental magma cools and precipitates crystals at contacts with country rocks, producing buoyant residual melt (Sparks et al. 1984; McBirney et al. 1985). Evolved melt layers ascend along chamber walls, incrementally collecting at the top of a magma chamber, thus producing a vertically compositionally stratified reservoir. This model is consistent with key features of Cedar Butte lavas that: (1) Erupted materials are crystal-poor and vary systematically and continuously in their compositional and mineralogical characteristics. (2) Compositions vary in a fashion that is consistent with fractional crystallization. (3) Rhyolites are erupted first, followed in rapid sequence by more mafic compositions.

Bachmann and Bergantz (2004) and Trial and Spera (1990) describe weaknesses of the convective fractionation model. However at least some of these do not apply to Eastern Snake River Plain volcanic field rhyolites. For example phenocrysts in lavas from Cedar Butte tend to be weakly

zoned, and exhibit euhedral to subhedral, or boxy-cellular textures, suggesting that they precipitated rapidly, rather than slowly and repeatedly. Also parental mafic, intermediate and rhyolitic rocks are sparsely porphyritic, suggesting that the source chamber was dominantly liquid. We speculate that other features of some lavas at Cedar Butte, such as crystal-poor mafic enclaves and xenocrysts, may in fact record intra-magma chamber entrainment processes (Bachmann and Bergantz 2004), as well as magma-drawdown-related mixing (e.g. compositionally zoned dikes) from a stratified chamber (e.g. Blake and Ivey 1986).

Magma output rates

Kuntz et al. (1992) calculate an average magma output rate for the Eastern Snake River Plain over the last 15 k.y. of $3.3 \text{ km}^3/1,000 \text{ years}$. Based upon simple interpolations of geophysical and borehole data we infer that $\sim 12,300 \text{ km}^3$ of olivine tholeiite accumulated over $\sim 5 \text{ Ma}$ (Fig. 1d). This leads to a similar calculated average eruption rate of $2.5 \text{ km}^3/1,000 \text{ years}$. The flux is obviously dominated by olivine tholeiite, but some limits can also be placed on evolved (i.e. Craters of the Moon-Cedar Butte series) magma fluxes. We estimate that the volumes of evolved rocks from Craters of the Moon ($\sim 30 \text{ km}^3$; Kuntz et al. 1992), Cedar Butte ($\sim 40 \text{ km}^3$), Unnamed Butte ($\sim 21 \text{ km}^3$), Big Southern Butte ($\sim 10 \text{ km}^3$), East Butte ($\sim 0.3 \text{ km}^3$) and Middle Butte, Ferry Butte and Buckskin Dome cryptodomes ($\sim 3.9 \text{ km}^3$ total) sum to $\sim 105 \text{ km}^3$. Volume estimates are documented in Appendix Table 1. These rocks have been extruded over 1.4 m.y. yielding an average flux of evolved rocks of $\sim 0.075 \text{ km}^3/1,000 \text{ years}$, or $\sim 3.3\%$ of the total magma output rate for Eastern Snake River Plain volcanic field. We estimate that of this total, the rhyolite volume is $\sim 15 \text{ km}^3$; the largest uncertainties being the amounts produced at Unnamed Butte and in cryptodomes (Kuntz and Dalrymple 1979; Houser 1992). Rhyolite output rate is therefore about $\sim 0.011 \text{ km}^3/1,000 \text{ years}$, yielding a ratio of rhyolite to total evolved rock of 0.15. This ratio is comparable to what is expected from our fractionation models (~ 0.14 to 0.24). However the ratios of total evolved rock and rhyolite to olivine tholeiite (3 and 0.4% , respectively) are at least a factor of ~ 10 lower than expected (e.g. Leeman 1982b). These low evolved magma output rates may be an indication the inefficiency of the evolved magma formation process, or that most evolved magmas crystallize within the crust.

Conceptual model

A simple conceptual model for genesis of Eastern Snake River Plain volcanic field rhyolites is illustrated in Fig. 10. The

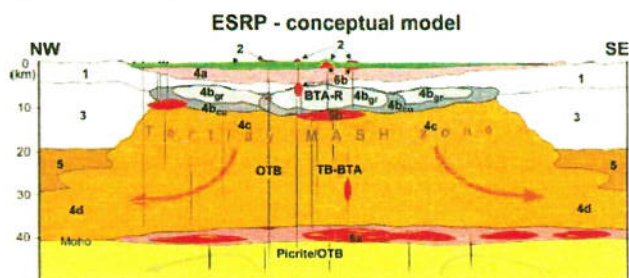


Fig. 10 A conceptual model for Quaternary crustal structures, lithologies and magma genesis of the Eastern Snake River Plain, extending from the geophysical crustal model of Sparlin et al. (1982) and Peng and Humphreys (1998): 1—Paleozoic–Mesozoic sedimentary rocks; 2—interlayered basalt, clastic sediment, and evolved Craters of the Moon–Cedar Butte series lavas, shields and domes; 3—Archean granitic upper crust; 4a—Tertiary hotspot track rhyolite; 4b—cogenetic granitoid plutons (gr-subscript) and mafic cumulates (cu-subscript); 4c—densified upper crust produced by MASH-like interactions between basalt intrusions and Archean upper crust during Tertiary hotspot volcanism; 4d—displaced densified crust (McQuarrie and Rodgers 1998); 5—original Archean lower crust; 6—Quaternary magma reservoirs and pathways. 6a, 6b, and 6c represent the hypothetical polybaric magma reservoir system in which strong magma fractionation is inferred to occur: Picrite-OTB—primitive magma (picrite or high-Mg olivine tholeiite basalt); TB-BTA—trachybasalt-basaltic trachyandesite; BTA-R—basaltic trachyandesite-rhyolite. Bold arrowed lines illustrate possible patterns of mantle convection (Humphreys et al. 2000), and lower crustal flow (McQuarrie and Rodgers 1998)

model conforms to petrological, as well as geophysical and geological constraints on crustal structure (Sparlin et al. 1982; Josten and Smith 1997; Peng and Humphreys 1998; Rodgers et al. 2002; McQuarrie and Rodgers 1998). In this model, primitive olivine tholeiite or picrite magmas ascend buoyantly from convecting underlying mantle (Saltzer and Humphreys 1997). Partial loss of buoyancy at the Moho results in capture of magma bodies into a melt rich layer (Fig. 10, layer 6a) (Peng and Humphreys 1998). Some magma batches ascend rapidly to shallower depths, undergoing “low” degrees of fractionation, solidification (i.e. as dikes and sills), and in some cases eruption as olivine tholeiite (e.g. Kuntz 1992). Remelting of gabbro produced by sequential mafic intrusions appears limited (cf. Streck and Gruner 2007; Frost and Frost 1997). Magmas that are ‘trapped’ in the crust undergo strong polybaric fractionation of olivine+plagioclase+clinopyroxene (e.g. Whitaker et al. 2007; Putirka et al. 2003a; Thompson 1975). Melts likely evolve through a density maximum (e.g. Nekvasil et al. 2000; Christiansen and McCurry 2007) inhibiting eruption of the most Fe-enriched compositions. In our conceptual model these melts continue to evolve via either fractional or equilibrium crystallization (e.g. Whitaker et al. 2007) to produce less dense trachybasalt to basaltic trachyandesite magmas. Those magmas then ascend through dike systems into lower density regions, probably within or near the top of the mid-crustal density high (Fig. 10, unit 6b). In that region the chambers continue to evolve by fractional

crystallization. Ferrogabbro and ferrodiorite cumulates may contribute to contamination of some younger olivine tholeiite magmas (e.g. Geist et al. 2002; Shervais et al. 2006). Juxtaposition of magmas evolved at various depths would lead to high probability of mutual interaction, so we expect that mixing and recharge would be common elements of the evolution of Eastern Snake River Plain volcanic field rocks. Crustal assimilation by evolved magmas is inhibited by the presence of refractory country rock compositions. Minor to moderate assimilation occurs at some locations (e.g. at Craters of the Moon), but it appears to have little effect on magma evolution (e.g. Leeman 1982b). Continued fractionation of magma in the upper crust leads to formation of compositionally zoned, basaltic trachyandesite—high-silica rhyolite magma chambers, parts of which may ascend to shallower depths, or the surface.

Conclusions

Field, petrologic and geochemical data support the hypothesis that Quaternary rhyolites of the Eastern Snake River Plain volcanic field are the product of extensive fractional crystallization of mantle-derived magmas (e.g., Nekvasil et al. 2000). In contrast to older hotspot-track-related rhyolites in the same area (e.g. Christiansen and McCurry 2007) derivation of these Quaternary rhyolites by partial melting of crustal rocks is inconsistent with observed compositional and isotopic trends (cf. Spear 1979; Leeman 1982b).

At Cedar Butte rhyolite lava flows and pyroclastic rocks are cogenetic with more voluminous basaltic trachyandesite-trachyte lavas. Mass, energy and mineral thermobarometry calculations are consistent with rhyolite evolution by ~77%, low-pressure fractional crystallization, most likely by convective fractionation, of a basaltic trachyandesite parent magma, and only a trace (< 1%) of crustal assimilation. Cedar Butte lavas also compositionally overlap with many other compositionally evolved lavas in Eastern Snake River Plain volcanic field, most prominently at Craters of the Moon. This suggests a similar, underlying process of magma evolution during the Quaternary that is dominated by fractional crystallization, and exhibited by various evolved Eastern Snake River Plain volcanic field eruptive centers. Varying degrees of magma recharge, mixing, and crustal assimilation are apparent in some systems (e.g. Unnamed Butte and Craters of the Moon) but apparently played no essential role in the formation of highly evolved magmas including rhyolite.

Thompson (1975) and others (Leeman et al. 1976; Kuntz et al. 1986; Stout et al. 1994; Putirka et al. 2003a; Whitaker et al. 2007) have argued, on petrological, geochemical and experimental grounds, that Craters of the Moon-series lavas are genetically linked to mantle-derived olivine tholeiite suite magmas by fractional or equilibrium

crystallization. It may therefore be possible, in continental environments, to produce rhyolite with little or no involvement of pre-existing continental crust (e.g. Nekvasil et al. 2000; Whitaker et al. 2007). In a planetary context, the implication is that rhyolite and granite can form by fractionation of basalt at low fO_2 (e.g. Whitaker et al. 2005; Bonin 2005).

Eastern Snake River Plain volcanic field rhyolites and cogenetic intermediate rocks have distinctive A-type geochemical characteristics. Cumulates and unerupted magma bodies may be contributing to the formation of an A-type batholith beneath the Eastern Snake River Plain. If this is the case, Eastern Snake River Plain volcanic field magmas would exemplify formation of such batholiths by mantle derived magmas (e.g. Vander Auwera et al. 2003; cf. Frost and Frost 1997). They also support the hypothesis that such magmas contribute to formation of continental crust (e.g. Rudnick 1995; Frost et al. 2003).

Acknowledgements This work is supported by EG&G, Idaho Contract C84-110421 to M. McCurry. We thank Bill Hackett (ISU) and Mel Kuntz (USGS) for introducing us to the many wonders of Eastern Snake River Plain volcanic geology. We gratefully acknowledge and thank Hanna Nekvasil and Don Lindsley (SUNY Stony Brook), Bill Bonnichsen (IGS), Marty Godchaux, Dick Smith (INL), Carol Frost (UW), Bill Leeman (NSF), Anita Grunder (OSU), Dennis Geist (UI), Scott Hughes and Dave Rodgers (ISU) for many helpful and insightful discussions, Duane Champion (USGS) for assistance in drilling rock samples at Cedar Butte and for unpublished paleomagnetic data from Cedar Butte and CH-1, Scott Hughes for assistance in INAA analysis, Linda Davis (USGS) for assistance with core from the USGS core library, and Regan Grandy for preparation of heavy mineral separates from Cedar Butte. Mertzman would like to acknowledge the financial support of the NSF and Franklin and Marshall College in the purchase of the Panalytical 2404 X-ray spectrometer and to thank Karen and Julie Mertzman for their long hours of careful work in the lab. We also thank Keith Putirka, Anita Grunder and Eric Christiansen for reviews that substantially improved the manuscript.

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