

Magma Evolution and Ascent at the Craters of the Moon and Neighboring Volcanic Fields, Southern Idaho, USA: Implications for the Evolution of Polygenetic and Monogenetic Volcanic Fields

KEITH D. PUTIRKA¹*, MEL A. KUNTZ², DANIEL M. UNRUH³ AND NITIN VAID¹

¹CALIFORNIA STATE UNIVERSITY, FRESNO, DEPARTMENT OF EARTH AND ENVIRONMENTAL SCIENCES, 2576 E. SAN RAMON AVE., MS/ST25, FRESNO, CA 93740-8039, USA

²US GEOLOGICAL SURVEY, MS 980, BOX 25046, DENVER, CO 80225, USA

³US GEOLOGICAL SURVEY, MS963, BOX 25046, DENVER, CO 80225, USA

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The evolution of polygenetic and monogenetic volcanic fields must reflect differences in magma processing during ascent. To assess their evolution we use thermobarometry and geochemistry to evaluate ascent paths for neighboring, nearly coeval volcanic fields in the Snake River Plain, in south-central Idaho, derived from (1) dominantly Holocene polygenetic evolved lavas from the Craters of the Moon lava field (COME) and (2) Quaternary non-evolved, olivine tholeiites (NEOT) from nearby monogenetic volcanic fields. These data show that NEOT have high magmatic temperatures ($1205 \pm 27^\circ\text{C}$) and a narrow temperature range ($<25^\circ\text{C}$) at any given depth; NEOT parent magmas partially crystallize within the middle crust (14–17 km), but with little time for cooling or assimilation. In contrast, COME magmas partially crystallize at similar depths, but at any given depth exhibit lower temperatures (by $\sim 40^\circ\text{C}$), and wider temperature ranges ($>50^\circ\text{C}$). Prolonged storage of COME magmas allows them to evolve to higher $^{87}\text{Sr}/^{86}\text{Sr}$ and SiO_2 and lower MgO and $^{143}\text{Nd}/^{144}\text{Nd}$. Most importantly, ascent paths control evolution: NEOT often erupt near the axis of the plain where high-flux (Yellowstone-related), pre-Holocene magmatic activity replaces granitic middle crust with basaltic sills, resulting in a net increase in NEOT magma buoyancy. COME flows erupt off-axis, where felsic crustal lithologies sometimes remain intact, providing a barrier to ascent and a source for crustal contamination.

A three-stage ascent process explains the entire range of erupted compositions. Stage 1 (40–20 km): picrites are transported to the middle crust, undergoing partial crystallization of olivine $\pm$ clinopyroxene. COME magmas pass through unarmored conduits and assimilate 1% or less of ancient gabbroic crust having high Sr and $^{87}\text{Sr}/^{86}\text{Sr}$ and low SiO_2 . Stage 2 (20–10 km): magmas are stored within the middle crust, and evolve to moderate MgO (10%). NEOT magmas, reaching 10% MgO, are positively buoyant and migrate through the middle crust. COME magmas remain negatively buoyant and so crystallize further and assimilate middle crust. Stage 3 (15–0 km): final ascent and eruption occurs when volatile contents, increased by differentiation, are sufficient (1–2 wt % H_2O) to provide magma buoyancy through the middle (and upper) crust.

KEY WORDS: Craters of the Moon; Snake River Plain; geothermometry; geobarometry; geochemistry; assimilation; crustal contamination; feldspar; clinopyroxene; mineral chemistry

INTRODUCTION

The Craters of the Moon lava field, within the Snake River Plain (SRP) in southern Idaho, is an example of a

*Corresponding author. E-mail: kputirka@csufresno.edu

Table 1: Age, stratigraphy and flow volumes

	Age (years BP)*	Stratigraphic order†	Eruptive episode‡	Volume (km³)¶
<i>Craters of the Moon flows (COME)</i>				
Broken Top Flow		44	A	0.1
Blue Dragon	2076	43	A	3.4
Big Craters Flow	2400	40	A	0.05
Serrate		39	A	0.4
Devils Orchard		38	A	0.1
Highway Flow		37	A	0.03
Devil's Cauldron	3660	32	B	0.9
Minidoka	3590	31	B	3
Larkspur		30	B	1.2
Range Fire	4510	29	B	1
Indian Wells North		27	C	0.2
Indian Wells South		26	C	0.1
Sawtooth	6020	25	C	0.6
Sheep Trail Butte		23	C	0.5
Sentinel		21	C	0.3
Silent Cone		20	D	0.1
Carey Kipuka	6600	19	D	0.6
Little Park	6500	18	D	1
Little Laidlaw Park		17	D	0.8
Grassy Cone Flow	7360	16	E	1.2
Laidlaw Lake	7470	15	E	1
Lava Point	7840	14	E	2.6
Pronghorn Flow	10240	13	F	0.8
Heifer	10670	12	F	0.4
Bottleneck Lake Flow	11000	11	F	1.4
Sunset Flow/Cone	12010	10	G	1
Carey Flow	12000	9	G	2.8
Lava Creek	12760	8	G	0.7
Kimama	15100	7	H	
Bear Den Lake		6	H	
Little Prairie		4	H	
<i>Neighboring volcanic vents and fields (NEOT)</i>				
Kings Bowl	2222	36	A	0.005
Wapi	2270	35	A	1.5
South Robbers	11980			0.03
North Robbers	11980			0.05
Cerro Grande	13380			2.3
Rock Coral Butte	59000			
Split Top	114000			
Rock Lake	182000			

*Age dates, volume estimates and stratigraphic order are from Kuntz *et al.* (1986, 2007).

†Stratigraphic order is relative to 44 flows in the Craters of the Moon region listed by Kuntz *et al.* (1986, table 1); stratigraphically lower flows have lower numbers.

‡Kuntz *et al.* (1986) divided Craters of the Moon flows into eight eruptive packages; A is the youngest, H the oldest.

¶Eruptive volumes are from Kuntz *et al.* (1986). It should be noted that Table 2 and the Electronic Appendix tables list the flows segregated in two suites as here, but with flows listed in alphabetical order.

polygenetic volcanic field, having erupted a broad array of intermediate and evolved magma compositions. This lava field is something of an anomaly, as it contrasts with the spatially dominant and compositionally restricted 'monogenetic' basaltic lava fields that carpet the eastern SRP (ESRP). There can be little doubt that contrasts between the polygenetic Craters of the Moon and neighboring monogenetic volcanic fields are greatly influenced by their ascent paths and rates of ascent, and that ascent is in turn affected by crustal structure. However, the ties between these phenomena are poorly understood, in part because we often lack information regarding the depths and temperatures at which magmas are stored prior to eruption. We nevertheless understand that magma transport is controlled by density contrasts or stress states within the lithosphere or crust (e.g. Rubin & Pollard, 1987; ten Brink & Brocher, 1987; Gans *et al.*, 1989; Kuntz, 1992; Parsons *et al.*, 1992; Parsons & Thompson, 1993; Putirka & Condit, 2003; Putirka & Busby, 2007; McCurry *et al.*, 2008)—and such ideas can be tested at any particular location because they imply predictions regarding the depths at which magmas are stored. Mineral–melt thermobarometers can play a crucial role in such tests because by determining crystallization depths we can establish where magmas are stored prior to eruption. Such information is also crucial for evaluating magma buoyancy, in part because by knowing stagnation depths, we have better constraints on ambient rock density, but also because magma densities themselves depend on temperature and pressure. Here, we combine such observations to delimit ascent paths and magma buoyancy in the ESRP.

The Craters of the Moon lava field, and neighboring flow fields in the ESRP, provide an ideal place to compare the origins of polygenetic and monogenetic lava fields. The flows in this area consist of two remarkably distinct magma suites, with nearly coeval eruptions (Table 1): (1) dominantly Holocene evolved lava flows of the polygenetic Craters of the Moon lava field; (2) olivine tholeiitic lava flows erupted in neighboring monogenetic volcanic fields; the latter are typical of most of the ESRP (Fig. 1). Our comparison makes use of a range of data, including new whole-rock major-element and isotope analyses, new mineral compositions, and existing geophysical data. These data lead to a new model for magma storage and ascent for these two contrasting magma suites.

GEOLOGICAL BACKGROUND

Craters of the Moon Lava Field and Snake River Plain lava flows

There are two types of flows in the ESRP. (1) 'Non-evolved' olivine tholeiite (NEOT) lava flows are very similar to one another in composition, with SiO₂ in the range 45.2–48.0%, MgO 4.9–9.5%, FeO 11.5–16%, and total

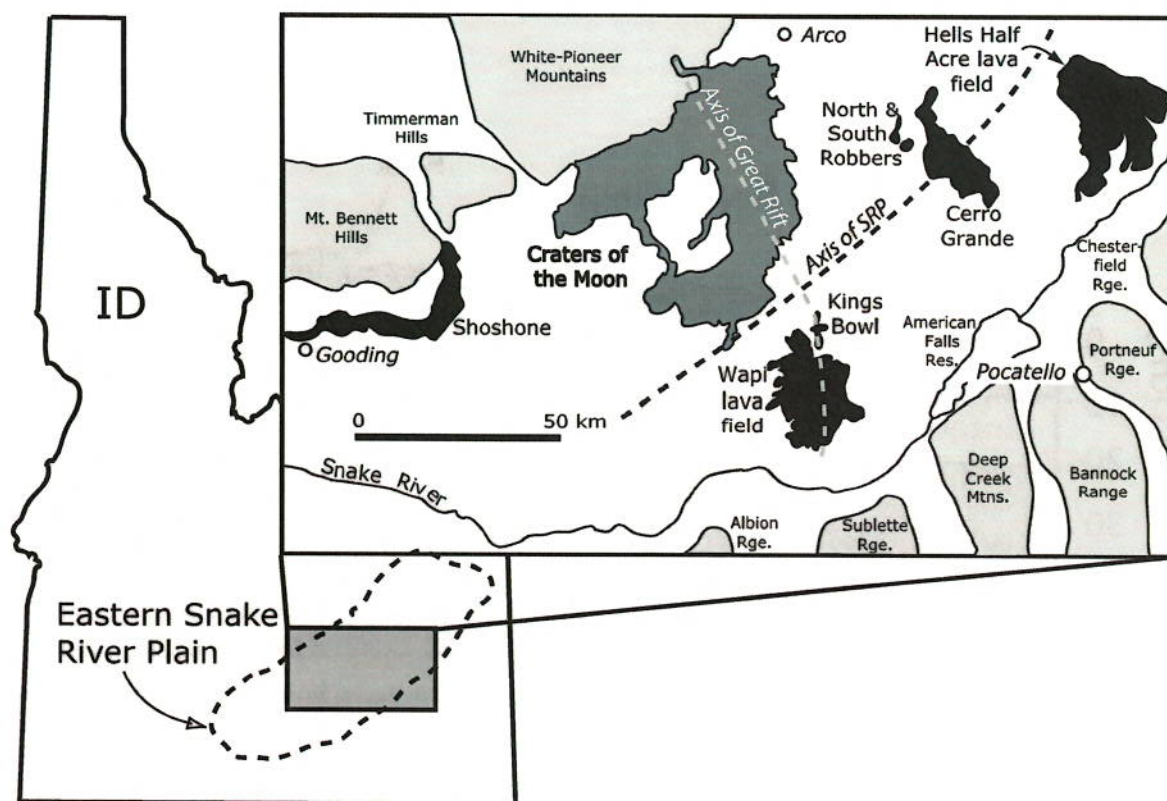


Fig. 1. Location map for the eastern Snake River Plain, Idaho, the Craters of the Moon lava field, and neighboring lava fields, adapted from Kuntz *et al.* (1992). Evolved lavas of the Craters of the Moon lava field (COME) are shown in dark gray; neighboring fields erupt non-evolved olivine tholeiites (NEOT), which are shown in black. The dark gray dashed line shows the approximate axis of the Snake River Plain (SRP) (see Kuntz *et al.*, 1992). The light gray line shows the axis of the Great Rift; nearly all vents for COME flows are concentrated between the northern tip of this line, to the G in Great Rift (Kuntz *et al.*, 1986). (See text for discussion.)

alkalis 1.4–4.7%. (2) ‘Evolved’ lava flows at the Craters of the Moon lava field (COME), in contrast, exhibit much broader ranges of SiO_2 and FeO , at 44.3–62.9% and 7.8–16.7%, respectively, and have lower MgO contents of 0.5–5.6%, and higher total alkali contents of 3.7–9.3% (Kuntz *et al.*, 1985, 1992; this study). In addition, NEOT flows are widely distributed throughout the ESRP. In contrast, although rhyolites are sometimes erupted near the SRP axis (McCurry *et al.*, 2008), volcanic fields of intermediate and evolved (polygenetic) flows are largely restricted to the margins of the ESRP (Christiansen & McCurry, 2008). For example, Leeman & Manton (1971) noted two examples of ‘COM-type lavas’ (Craters of the Moon-type), namely King Hill (~150 km west of the Craters of the Moon) and a field ‘near Blackfoot’, which respectively occur along the northern and southern margins of the ESRP; another example is Spencer High Point (Hughes *et al.*, 2002; Iwahashi & Hughes, 2007), just west of Yellowstone. The dominantly Holocene Craters of the Moon lava field (Fig. 1; Kuntz *et al.*, 1985) is the largest and most diverse of the group, and the most studied.

More than 80% of the ESRP is covered by NEOT lava flows that are magnetically normal and are thus younger than ~780 ka (Kuntz *et al.*, 1986, 2007). The largest volume of the flows is included in coalesced shield and lava-cone volcanoes made up of tube- and surface-fed pahoehoe flows. Deposits of fissure-type, tephra-cone and hydrovolcanic eruptions constitute a minor part of the volume of the ESRP. The eruptions of NEOT produced monogenetic, single-pulse lava fields. There are no examples of composite NEOT lava fields in the SRP. The North Robbers and South Robbers lava fields (11980 ± 300 years BP) and the Kings Bowl lava field (2222 ± 100 years BP) are NEOT lava fields formed in short-duration (a few days), low-volume ($<0.1 \text{ km}^3$), fissure-dominated eruptions. The Hells Half Acre (5200 ± 150 years BP), Cerro Grande (13380 ± 350 years BP), Wapi (2270 ± 50 years BP), and Shoshone (10130 ± 350 years BP) lava fields formed during relatively long-duration (months to decades), high-volume ($1\text{--}6 \text{ km}^3$), lava cone and shield-forming eruptions that were neither preceded nor followed by eruptions at the same or nearby vents (Kuntz *et al.*, 1992).

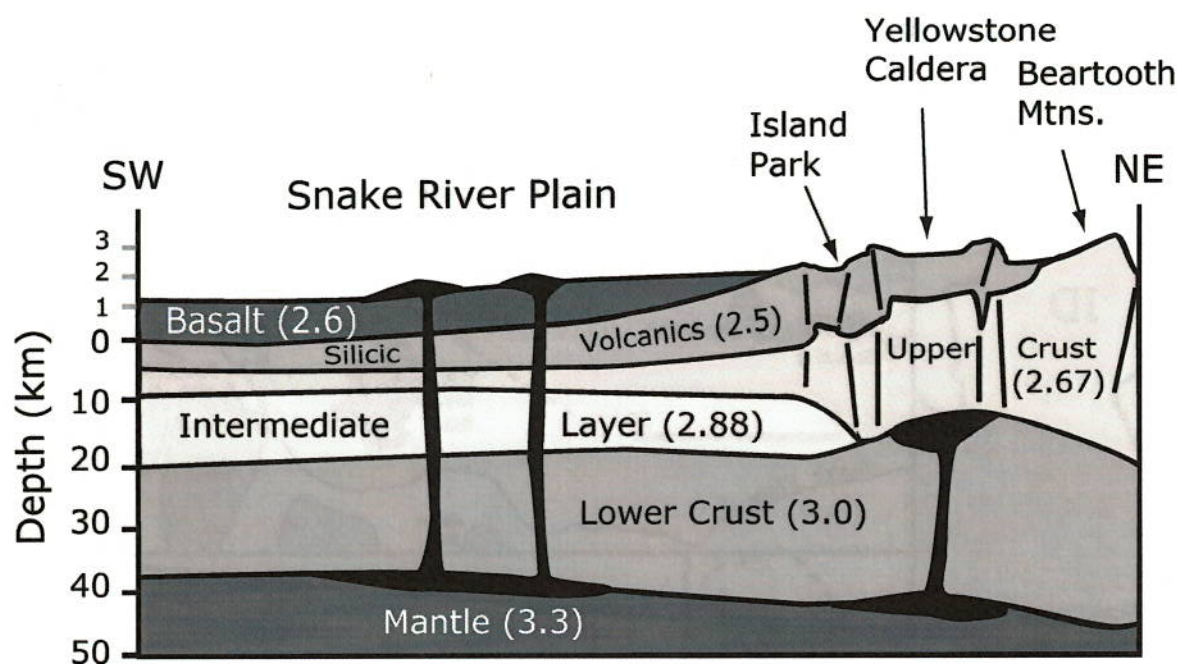


Fig. 2. An SW-NE cross-sectional profile of the crust showing the density stratification of the eastern Snake River Plain crust, adapted from Smith & Braile (1994).

The Craters of the Moon lava field is composite and polygenetic. It consists of more than 60 lava flows that were erupted from ~25 tephra cones and some eight eruptive fissure systems during the last 15 kyr. The closely spaced source vents are aligned along the northern part of the Great Rift volcanic rift zone in a belt about 2–5 km wide and about 50 km long (Fig. 1; Kuntz *et al.*, 1994, 2007). The Craters of the Moon lava field is the largest, dominantly Holocene lava field in the conterminous USA; it covers about 1600 km² and contains about 30 km³ of lava flows and associated vent and pyroclastic deposits. Stratigraphic relationships, paleomagnetic studies, and radiocarbon dating indicate that the Craters of the Moon lava field flows form eight eruptive periods, designated as H, oldest, to A, youngest. The first eruptive period (H) began at about 15 000 years BP and the latest eruptive period (A) at about 2500 years BP and ended ~2100 years BP (Kuntz *et al.*, 1986). Each eruptive period is approximately several hundred to several thousand years in duration and the periods are spaced approximately several hundred to about 3000 years apart. The eruptions appear to be ‘volume-predictable’ and the next eruption, which is expected to happen within the next 1000 years, should erupt between 5 and 6 km³ of material (Kuntz *et al.*, 1986).

For the samples studied here, Table 1 summarizes stratigraphic order and eruptive-period designations for COME flows, and, where measurements have been made, flow volume and age dates for COME and NEOT flows.

Flow locations for all samples and additional volume and age estimates have been given by Kuntz *et al.* (1986, 2007).

The Kings Bowl and Wapi Lava fields, composed of NEOT lava flows along the southern part of the Great Rift volcanic rift zone, are approximately coeval with the Craters of the Moon lava field flows of eruptive period A (Table 1). Details of the basaltic lava fields of the ESRP have been given by Kuntz *et al.* (1985, 1986, 1992). A geological map of the Craters of the Moon lava field, Wapi and Kings Bowl lava fields has been published by Kuntz *et al.* (1994, 2007).

The two eruptive suites—the COME and NEOT—provide a means to assess the effects of magma ascent paths because the ages of eruption and erupted volumes of most flows are well known (Kuntz *et al.*, 1986, 1994). Additionally, although some interpretations of the geophysical data are controversial (e.g. Christiansen *et al.*, 2002), the architecture of the underlying crust and lithosphere (Fig. 2) is reasonably well understood (Smith & Braile, 1994; Peng & Humphreys, 1998). We expand upon earlier studies of the mineralogy (Stout & Nicholls, 1977; Stout *et al.*, 1994) and geochemistry (Leeman & Manton, 1971; Leeman *et al.*, 1976; Menzies *et al.*, 1984; Kuntz *et al.*, 1985, 1986, 1992) of COME and NEOT flows by analyzing samples from each of the eight eruptive COME units, including COME rhyolite and granulite inclusions, and NEOT samples from several neighboring monogenetic basaltic volcanic fields.

METHODS

New data and analytical methods

Mineral textures from COME and NEOT flows were examined in thin section; textures and compositions were examined by electron back-scatter and electron microprobe. Mineral compositions of samples of COME and NEOT flows (Electronic Appendix Table 1, available for downloading at <http://www.petrology.oxfordjournals.org>) were obtained using the Cameca SX-50 electron microprobe at the University of Massachusetts, Amherst. Each mineral analysis represents an average of 2–5 microprobe analyses taken from the entire grain or, where indicated, from multiple spots at the core or rim. Glass compositions (Electronic Appendix Table 2) were obtained with a Cameca SX-100 at UC Davis, using a defocused beam (10 µm) and a 10 nA beam current; each analysis represents an average of 10–20 spot analyses from a single thin section. Although major oxides have been determined for all COME volcanic units, we have reanalyzed samples from which we obtained our mineral compositions, to ensure a close match between mineral and host whole-rock compositions. We also analyzed new NEOT flows and several granulite and rhyolite inclusions contained within COME flows. Whole rocks (Electronic Appendix Table 3) were analyzed by X-ray fluorescence at the California State University, Fresno. Sample preparation and analytical details have been reported by Busby *et al.* (2008). Average deviation between reported compositions and replicate CSU Fresno analyses of USGS standards are: SiO₂, 0.2%; TiO₂, 0.015%; Al₂O₃, 0.045%; Fe₂O₃, 0.03%; MgO, 0.025%; CaO, 0.02%; Na₂O, 0.03%; K₂O, 0.015%; P₂O₅, <0.1%. Loss on ignition (LOI) was determined by heating samples to 850–1000°C. Heating times were 10 min in duration, to minimize oxidation of Fe (see Rhodes & Vollinger, 2004), but even with such short heating times, many LOI values are negative (Electronic Appendix Table 3), most probably because of a combination of high Fe contents and low H₂O.

To test for wall-rock assimilation, we analyzed several COME and NEOT flows and COME inclusions for Sr, Nd and Pb isotope ratios. Analyses (Electronic Appendix Table 4) were conducted at the US Geological Survey in Denver, Colorado. Analytical procedures have been given by Stille *et al.* (1986). Samples were dissolved in PFA–Teflon screw-cap jars using hydrofluoric and nitric acids. Lead was extracted from the sample using anion exchange in HBr medium. Strontium was extracted using cation exchange in HCl medium. Isotopic data were obtained using a single-collector VG54R solid-source mass spectrometer. Lead isotopic data were corrected for mass fractionation during mass spectrometry of $0.13 \pm 0.03\%$ per a.m.u. [95% confidence interval (CI); Ludwig, 1980] based on 12 replicate analyses of NIST standard SRM 981 (Tody *et al.*, 1993). Eight analyses of NIST Sr standard

SRM 987 gave a mean $^{87}\text{Sr}/^{86}\text{Sr} = 0.710258 \pm 0.000010$ (95% CI). Eight analyses of the La Jolla Nd standard yielded a mean $^{143}\text{Nd}/^{144}\text{Nd} = 0.511858 \pm 0.000008$ (95% CI).

Data from published sources

To compare Craters of the Moon lava field flows with other SRP compositions we use SRP Basalts (SRPB) from GEOROC (<http://georoc.mpch-mainz.gwdg.de/georoc/>) and data from Leeman *et al.*, 1976; Leeman, 1982a, 1982b), Kuntz & Dalrymple (1979), Kuntz *et al.* (1985, 1992), Shervais *et al.* (1994), Knobel *et al.* (1995), Reed *et al.* (1997), Stout & Nicholls (1977), Hughes *et al.* (2002) and McCurry *et al.* (2008). We also compare SRPB with Hawaii (Hawaii Scientific Drilling Project; Rhodes & Vollinger, 2004) and test for magma–wall-rock interaction using intrusive rock compositions from Idaho (North American Volcanic Rock Data Base, or NAVDAT, available at: <http://navdat.kgs.ku.edu>) and data from Ratajeski *et al.*'s (2001) study of the Yosemite Intrusive Suite in California.

AFC modeling

As a simple test for magma–wall-rock interaction, we compare our data with curves that describe combined assimilation–fractional crystallization (AFC), using equations from DePaolo (1981). For elemental variations we use

$$C_i^m = C_i^{\text{om}} \left[F^{-z} + \left(\frac{r}{r+1} \right) \frac{C_i^a}{z C_i^{\text{om}}} (1 - F^{-z}) \right]. \quad (1)$$

For a given element or oxide, i , C_i^{om} is the original concentration of i in a magma, C_i^m is the concentration of i in the magma following some interval of AFC, F is the melt fraction following this same interval of AFC, r is the ratio of the mass rate of assimilation to the mass rate of crystallization, and $z = (r + D_i - 1)/(r - 1)$, where D_i is the bulk distribution coefficient describing the partitioning of i between liquid and crystalline phases ($D_i = \sum X_j D_i^{j-\text{liq}}$, where X_j is the fraction of crystalline phase j among an assemblage of minerals growing within a liquid, and $D_i^{j-\text{liq}}$ is the mineral–liquid partition coefficient for a given element between mineral j and liquid, $D_i^{j-\text{liq}} = C_i^j/C_i^{\text{m}}$). For isotopic variations we use

$$\epsilon_m = \frac{\left(\frac{r}{r-1} \right) \frac{C_i^a}{z} (1 - F^{-z}) \epsilon_a + C_i^{\text{om}} F^{-z} \epsilon_m^0}{\left(\frac{r}{r-1} \right) \frac{C_i^a}{z} (1 - F^{-z}) + C_i^{\text{om}} F^{-z}}. \quad (2)$$

Here, ϵ_m^0 is the original isotopic ratio in a magma, and ϵ_m is the isotopic ratio after some interval of AFC. All other quantities are as in equation (1), where terms such as C_i^a and C_i^{om} refer to concentrations for the element i that make up ϵ . In this work, partition coefficients for major elements are derived from averages of observed mineral and coexisting glass or whole-rock compositions (Electronic Appendix); partition coefficients for Nd and Sr are from the GERM database (<http://www.earthref.org/GERM/>);

bulk distributions coefficients are as follows: $D_K = 0.0$; $D_{Mg} = 2.3$; $D_{Sr} = 0.7$; $D_{Si} = 1.3$.

In our AFC calculations, free parameters are (1) r , the ratio of the mass rate of assimilation to the mass rate of crystallization, and (2) F , the amount of residual melt following AFC. We compare curves derived by varying r and F such that the data are most closely described. In the Discussion, we review our choices of potential wall-rock assimilants. In these calculations, the best r values are 0.8–0.9.

Our approach is intended to provide tests of whether particular wall-rock compositions might possibly act as assimilants, and so explain COME and NEOT geochemical variations. Our modeling does not account for magma recharge (e.g. DePaolo, 1985), which is almost certainly required to supply heat sufficient for significant wall-rock assimilation, or for conservation of enthalpy, and other conservation considerations, which Bohrsen & Spera (2001) and Spera & Bohrsen (2001) showed are crucial to describe assimilation-related processes in detail. We acknowledge that these added treatments are important and useful, and could be utilized to refine the work presented here. Our values for F or r thus do not necessarily represent actual melt fractions, or ratios of mass rates of assimilation or crystallization. However, for our present goal of comparing disparate potential wall-rock assimilants, especially given the error involved in defining certain end-members, we judge that the simple expressions used here are useful.

RESULTS

This work expands upon Stout *et al.*'s (1994) study of eruptive period A (2500–2000 years BP) for COME flows by examining whole-rock major, mineral and glass compositions from each of the eight eruptive episodes, A–H, at the Craters of the Moon, which range from 2000 to 15 000 years BP (Kuntz *et al.*, 1986), and from several NEOT flows from neighboring volcanic fields that erupted nearly simultaneously with COME flows (Table 1). We also report new isotopic data from both COME and NEOT flows. All our samples represent volcanic activity that took place within a fairly small area (60 km × 60 km) of the upper mantle, beneath the north-central part of the ESRP (Fig. 1).

Rock and mineral textures, and mineral compositions

COME volcanic rocks are generally fine-grained, consisting of unaltered glassy groundmass (60–95% glass, by visual estimate) with small (typically less than 2 mm) rounded to euhedral plagioclase (and sometimes olivine) microphenocrysts (often pilotaxitic) and microlites (all plagioclase). Several COME flows, such as the Sawtooth, Big Crater, Minidoka, Blue Dragon, Lava Creek, Little

Laidlaw Park and Sunset flows (to name just a few) have very few crystals larger than 2 mm, and are highly vesicular. Otherwise, most COME flows have at least a few plagioclase phenocrysts that range to 2–10 mm. Generally, crystal proportions are 70–100% plagioclase, 0–30% olivine; clinopyroxene occurs as rare microphenocrysts or groundmass crystals. Except for the Indian Wells South and Highway flows, plagioclase phenocrysts and microphenocrysts are largely euhedral and unzoned. In both the Indian Wells South and the Highway flows, some plagioclase phenocrysts and microphenocrysts are rounded and exhibit sieve textures. Also, in the Big Craters flow, one feldspar phenocryst is notably reversely zoned (a core of alkali feldspar is rimmed by plagioclase). Sieve textures can result from decompression (Nelson & Montana, 1992) but are also indicative of magma mixing (e.g. Eichelberger, 1975; Dungan & Rhodes, 1978; Streck, 2008), as is reverse zoning. In addition, the Kimama and Carey flows both show cryptic evidence of 'enclaves' of devitrified glassy material, which suggest mixed magmas. In Electronic Appendix Table 1, crystals described as groundmass have a longest dimension of 0.3–1.0 mm; microphenocrysts range from 1 to 5 mm. Phenocrysts have longest dimensions of 5–10 mm.

NEOT flows are noticeably coarser than COME; they are medium-grained, have glassy groundmasses (50% glass), and contain rounded or euhedral olivine phenocrysts and abundant plagioclase phenocrysts in the size range 2–7 mm. Crystal proportions are approximately 50–75% plagioclase, 50–25% olivine. Clinopyroxene is somewhat more common among NEOT flows, but still forms less than 1% of total crystal content for most flows, and is largely confined to the groundmass. Like COME flows, NEOT crystals are mostly euhedral and unzoned.

In both COME and NEOT suites phenocrysts are homogeneous and inter-grain heterogeneity is more significant than intra-grain heterogeneity (Electronic Appendix Table 1). Mineral compositions are also independent of grain size or texture. In contrast, spot analyses of COME and NEOT glass compositions (Electronic Appendix Table 2) show that groundmass glasses are heterogeneous, with the following standard deviations being typical: SiO_2 , 3%; TiO_2 , 2%; Al_2O_3 , 2.3%; FeO , 4.6%; MgO , 0.8%; CaO , 1.7%; Na_2O , 0.8%; K_2O , 0.3%; P_2O_5 , 1.3%.

H₂O contents and ascent rate calculations (decompression-related crystal growth?)

Minimum water contents for COME and NEOT magmas can be derived from water saturation models (Moore *et al.*, 1995), as all flows are vesicular. At 1 atm (and 900–1200°C) COME and NEOT flows can dissolve between 0.05 and 0.06 wt % H_2O , and so represent a minimum water content. The lack of hornblende in any COME or NEOT rocks allows for a rough estimate of maximum water contents. For example, T estimates for COME

flows are as low as 900°C (see following sections); at temperatures of 900–950°C, experiments by Barclay & Carmichael (2004) showed that amphibole can crystallize at relatively low water contents (*c.* 2 wt % at $P < 50$ to 300 MPa), and that at 1000–1050°C, amphibole is stable down to at least 2.5% H₂O. Given that hornblende is absent even from those flows with the lowest temperatures, we estimate that maximum water contents are *c.* 2 wt %. The plagioclase hygrometer from Putirka (2005) yields a median water content of 0.7 wt %, and generally low water contents (<1%) for the vast majority of samples, as does the hygrometer of Putirka (2008). The root mean square error on these hygrometers, however, is $\pm 1.5\%$ H₂O, so at these low water contents, these models do little more than confirm the broad limits of water concentrations indicated by phase equilibria and water saturation models.

At these relatively low water contents it is unclear that decompression could induce partial crystallization, but we explore the issue. Decompression experiments (Geschwind & Rutherford, 1995; Hammer & Rutherford, 2002) indicate that crystal nucleation and growth can be triggered by magma ascent of water-bearing magmas; the associated loss of pressure results in exsolution of H₂O, causing an increase in the liquidus temperature of the liquid. Such experiments are conducted on water-saturated systems at moderate pressure (150–220 MPa), and even in these systems, crystals that nucleate during the experiments are microlites, with maximum lengths of <100 µm (Geschwind & Rutherford, 1995). Decompression of dry liquids should induce partial melting, not crystal growth, as anhydrous mineral saturation surfaces have positive dP/dT . In any case, to test what crystal sizes might be obtained by decompression, we calculate minimum growth rates and maximum lengths ($l_{\max}$) for plagioclase crystals, assuming they nucleate and grow only during ascent. For these calculations, we use the slowest ascent rates calculated by Kuntz (1992) for ESRP magmas, to allow the longest time for crystal growth. Kuntz (1992) used the following expression from Wilson & Head [1981; their equation (12)] for magma (Newtonian, not volatile saturated) transport within a dike:

$$V = \left(\frac{A\eta}{4K\rho_m v} \right) \left[\left(1 + \frac{64gr^3(\rho_m - \rho_r)K\rho_m}{A^2\eta^2} \right)^{0.5} - 1 \right]. \quad (3)$$

Here, V is ascent velocity, A and K are constants (24 and 0.01, respectively), $(\rho_m - \rho_r)$ is the density contrast between magma and wall-rock (150 kg/m³ or 0.15 g/cm³), g is acceleration due to gravity, v is magma viscosity (300 Pa s), and r is dike half-width; an r value of 0.5 m yields the minimum V of Kuntz (1992) (0.4 m/s). Actual ascent rates for COME magmas by this mechanism [equation (3)] are expected to be greater (so the time for decompression-related crystal growth is shorter), given that (1) fissure

widths are in the range 1–2 m, in which V would range from 0.4 to 1.6 m/s (Kuntz, 1992) and (2) viscosities, calculated from Giordano *et al.* (2008) using 0–1% H₂O, are an order of magnitude lower than used by Kuntz (1992). Paired with the highest growth rates of Hammer & Rutherford (2002) (10^{−6} mm/s), and by assuming that COME magmas rise from 20 km, then $l_{\max} = 0.05$ mm; if the magmas rose from 40 km (Kuntz, 1992), $l_{\max} = 0.1$ mm. Most COME flows, however, have crystals ranging to at least 1–2 mm, greater than the greatest length that can apparently be obtained by ascent or devolatilization alone.

The Hammer & Rutherford (2002) experiments used rhyolitic liquids (albeit water saturated), where growth rates might be less compared to mafic systems. As an alternative test, we use an ascent rate of 0.4 m/s and a transport distance of 20 km (so a transport time of 13.9 h) to calculate a growth rate of 4×10^{-5} mm/s to obtain $l = 2$ mm. This growth rate is faster than the fastest experimentally determined rates of $(1.2\text{--}5.5) \times 10^{-6}$ mm/s reported by Cashman (1990) for the growth of plagioclase from basalt, and these high growth rates apply only at undercoolings of $\geq 100^\circ\text{C/h}$, which in the present case would require a total cooling of $\sim 1390^\circ\text{C}$. Some microlites in COME flows thus undoubtedly formed by decompression, but most crystals, even those <2 mm, formed at greater depths, during an episode of subsurface storage (or very slow transport) prior to their final ascent.

Thermobarometry

Tests of published clinopyroxene and plagioclase–liquid barometers

Before calculating pressure (P) and temperature (T) conditions, new partial-melting experiments from Whitaker *et al.* (2007) provide a means to test clinopyroxene–liquid (Putirka *et al.*, 2003) and plagioclase–liquid (Putirka, 2005) barometers, using liquid compositions similar to NEOT flows. For the tests, we calculate T and P simultaneously for the Whitaker *et al.* (2007) experiments [instead of using the Whitaker *et al.* (2007) experimental values for T as input] to mimic how P and T are calculated for COME and NEOT flows.

For the clinopyroxene–liquid barometers (Putirka *et al.*, 2003), the greatest absolute error is for 1 atm experiments (Whitaker *et al.*, 2007), where the models yield P estimates that are consistently ~ 3 kbar high (Fig. 3). At higher experimental pressures, calculated pressures still exceed experimental values, but experimental and calculated values are highly correlated ($R = 0.95$), capturing 90% of the variation of experimental pressures in the range 3–14 kbar (Fig. 3). This systematic error remains, even if experimental temperatures are used as input, and so is related to the calibration of the barometer.

Because SRP NEOT basalts are high in both FeO and P₂O₅, it is plausible that the compositions used by

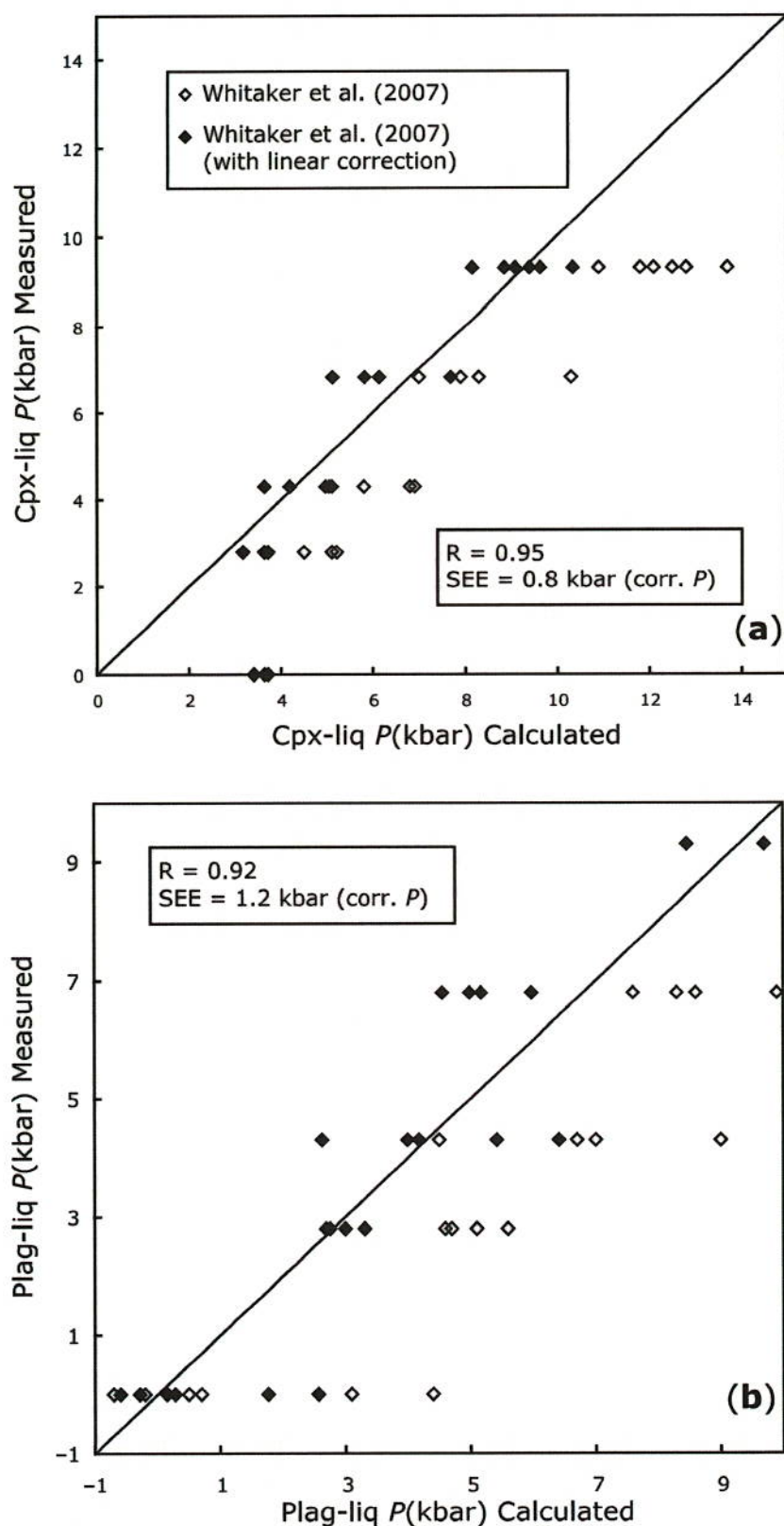


Fig. 3. Tests of geobarometers based on (a) clinopyroxene-liquid and (b) plagioclase-liquid equilibria (Putirka *et al.*, 2003; Putirka, 2005), using the partial melting experiments of Whitaker *et al.* (2007). Pressure estimates are systematically high for both barometers. Open symbols are uncorrected; filled symbols represented 'corrected' values, using equations (1) and (2). Equations (1) and (2) eliminate systematic error (slopes and intercepts for regression lines are 1.0 and 0.0); standard errors of estimate (SEE) are ± 0.8 kbar for clinopyroxene and ± 1.2 kbar for plagioclase.

Whitaker *et al.* (2007) are not adequately explained by the clinopyroxene barometer of Putirka *et al.* (2003). It is also conceivable that the Whitaker *et al.* (2007) compositions are not in equilibrium, or that their experiments have other systematic errors. However, no such problems are obvious from their published methods, so we assume that deviations between predicted and experimental P reflect a model deficiency. Because of the high correlation between experimental and calculated values, a simple linear correction can be applied to the Putirka *et al.* (2003) clinopyroxene barometer:

$$P(\text{SRP} - \text{cpx}) = 0.7787P(\text{P03}) - 0.3319. \quad (4)$$

Here, $P(\text{SRP-cpx})$ is the pressure of clinopyroxene crystallization for SRP compositions, and $P(\text{P03})$ is the pressure derived from Putirka *et al.* (2003); units of P are in kbar. With this correction, the correlation between calculated and experimental pressures is unchanged, but through its use, systematic error for these compositions is eliminated at $P \geq 4$ kbar (the correction is constrained so that the slope and intercept of a regression line through calculated vs experimental pressures is 1.0 and 0.0, respectively). The standard error of estimate, after applying this correction, is ± 0.8 kbar, well within calibration error (Putirka *et al.*, 2003) (Fig. 3). It is not entirely clear why the barometer of Putirka *et al.* (2003) does not accurately recover the pressures of 1 atm experiments of Whitaker *et al.* (2007). Putirka *et al.* (1996, 2003) showed that pressures are more precise for those experiments where precautions against Na_2O volatilization are employed [as in the experiments of Tormey *et al.* (1987)]. However, Whitaker *et al.* (2007) used evacuated glass tubes, which should obviate the problem. However, even with such remedies, calculated pressures often (but not always) exceed 1 atm for such experiments, which prompted Putirka *et al.* (1996, 2003) to exclude 1 atm experiments from calibration, suspecting problems related to diffusion rates in clinopyroxene. We hold out the possibility that calculated pressures ≤ 3 kbar might still be accurate, but if the clinopyroxenes of the 1 atm experiments of Whitaker *et al.* (2007) have indeed equilibrated with their host liquid, all pressure estimates ≤ 3 kbar possibly represent 1 atm crystallization.

As with clinopyroxene liquid equilibria, pressure estimates derived from plagioclase–liquid equilibria (Putirka, 2005) also contain systematic errors (even when experimental temperatures are used as input), over-predicting pressures for plagioclase-saturated experiments from Whitaker *et al.* (2007) (Fig. 3). However, like clinopyroxene, calculated and experimental pressures are highly correlated ($R=0.92$), which again allows for a simple linear correction:

$$P(\text{SRP} - \text{plag}) = 0.6199(P05) - 0.1571. \quad (5)$$

Here, $P(\text{SRP-plag})$ is the pressure of plagioclase crystallization for SRP-like volcanic compositions, and $P(\text{P05})$ is the pressure derived from Putirka (2005) (in units of kbar). As with clinopyroxene, this correction eliminates systematic error; the resulting standard error of estimate is ± 1.2 kbar, which is less than calibration error (Putirka, 2005). Pressures of 1 atm experiments (Whitaker *et al.*, 2007) are not systematically over-predicted when using the Putirka (2005) models: calculated values average 1.3 ± 2.0 kbar.

Mineral compositions and tests for equilibrium

Several lines of evidence indicate that most COME whole-rocks may approximate liquid compositions. Phenocrysts are $<5\%$ by volume, crystallinity is generally low ($<50\%$), and disequilibrium mineral textures are uncommon (the Highway and Indian Wells South flows excepted). NEOT flows are more highly crystalline, but for both NEOT and COME flows (except for the Bottleneck Lake flow), the maximum forsterite (Fo) contents of olivine phenocrysts approach equilibrium with host whole-rocks, based on an equilibrium Fe–Mg exchange coefficient, $K_D(\text{Fe–Mg})^{\text{ol-liq}} = 0.30 \pm 0.03$ (Roeder & Emslie, 1970) (Fig. 4a). Here, $K_D(\text{Fe–Mg})^{\text{ol-liq}} = [X^{\text{ol}}_{\text{FeO}}/X^{\text{ol}}_{\text{MgO}}]/[X^{\text{liq}}_{\text{FeO}}/X^{\text{liq}}_{\text{MgO}}]$, where X^i_j represents the cation fraction of j in phase i (ol is olivine, and liq is liquid).

Tests for equilibrium between olivine phenocrysts and putative liquids are often illustrated using the Rhodes' diagram (Rhodes *et al.*, 1979), which plots the Mg-number of the whole-rock [$\text{Mg-number}^{\text{whole-rock}} = X^{\text{liq}}_{\text{MgO}}/(X^{\text{liq}}_{\text{MgO}} + X^{\text{liq}}_{\text{FeO}})$] against the Fo content of olivine ($\text{Fo} = \text{Mg-number}^{\text{olivine}} = X^{\text{ol}}_{\text{MgO}}/(X^{\text{ol}}_{\text{MgO}} + X^{\text{ol}}_{\text{FeO}})$) (Fig. 4a). Most COME olivines plot below the equilibrium line defined by $K_D(\text{Fe–Mg})^{\text{ol-liq}} = 0.30$ (here, using the whole-rock as a liquid). For $\text{Mg-number}^{\text{whole-rock}}$, we calculate FeO from Fe_2O_3 (Electronic Appendix Table 3), assuming that $f\text{O}_2$ is buffered at quartz–fayalite–magnetite (QFM; Stout & Nicholls, 1977; Christiansen & McCurry, 2008) where, approximately, $\text{FeO} = 0.9[\text{Fe}_2\text{O}_3]$ on a weight % basis. If $f\text{O}_2$ conditions are more oxidizing than QFM, $\text{Mg-number}^{\text{whole-rock}}$ will be higher, resulting in a rightward shift from the equilibrium line (Fig. 4a).

The deviation of COME flows from olivine–whole-rock equilibrium can be explained in at least three ways, as follows. (1) Whole-rocks could represent olivine cumulates (a shift to the right in Fig. 4a), which for the average COME magma, requires addition of 10% olivine; $>20\%$ for those samples furthest from the equilibrium line. That COME flows could represent cumulates seems highly unlikely, as they are glassy and typically have $<2\%$ olivine. Also, few rocks plot above the curve, so the olivine-depleted liquids—a necessary by-product of olivine accumulation—are curiously absent. (2) Whole-rocks are affected by magma mixing recharge (a diagonal

shift in Fig. 4a). Because of the shape of the equilibrium curve (concave down), mineral-whole-rock pairs can plot below the equilibrium line as a result of magma mixing. However, the curvature is not strong, and many COME samples plot outside a possible mixing envelope, or require mixing between extreme compositions that do not exist. In addition, disequilibrium mineral textures are uncommon. (3) Intra-flow variations in mineral compositions reflect cooling and differentiation, without or following wall-rock assimilation (a shift downward; Fig. 4a). Here, olivine phenocrysts with the highest Fo contents are the first olivine crystals to form, and so are in equilibrium with the whole-rock. Olivines with progressively lower Fo contents represent equilibration with later, lower T , lower-MgO residual liquids. Those crystals with the lowest Fo should be in equilibrium with matrix-glass. This last option is most likely, as the lowest Fo-content COME olivines do indeed approach equilibrium with matrix glass

compositions (Fig. 4a). The Bottleneck Lake flow is an exception; it appears that either the Bottleneck Lake magma was contaminated with olivines derived from NEOT-like magmas and/or the Bottleneck Lake flow is related to a complementary olivine cumulate at depth.

To test whether whole-rock or glass compositions represent a plausible equilibrium liquid for plagioclase or clinopyroxene, we use the models of Putirka (2005) and Putirka (1999), respectively. Treating whole-rocks as liquids, and using calculated values for crystallization P (Table 2; Electronic Appendix Table 5 contains additional P - T calculations) as input [into models D-G of Putirka (2005) and models 3.1-3.8 of Putirka (1999)], we calculate the equilibrium plagioclase or clinopyroxene compositions (Fig. 4b and c) that would precipitate from a liquid equivalent to the whole-rock, as well as the temperatures at which these liquids should become saturated with plagioclase (Fig. 4d) or clinopyroxene (Table 2). Nearly every

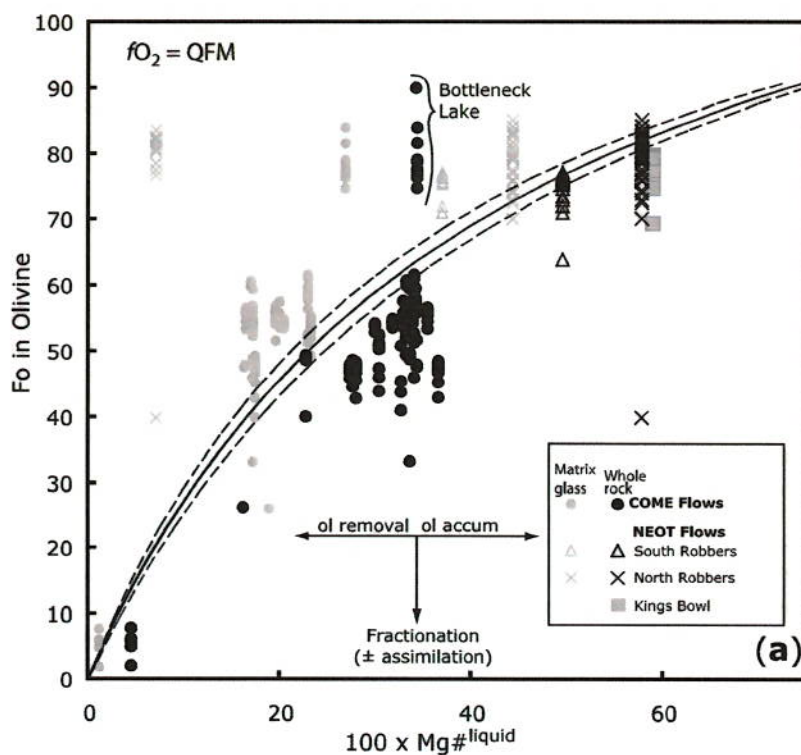


Fig. 4. Tests of mineral-melt equilibrium. (a) Olivine forsterite content ($Fo = 100 \times \text{Mg-number}^{\text{olivine}} = 100[X_{\text{MgO}}^{\text{ol}}/(X_{\text{MgO}}^{\text{ol}} + X_{\text{FeO}}^{\text{ol}})]$) vs $100 \times \text{Mg-number}^{\text{liquid}} = 100[X_{\text{MgO}}^{\text{liq}}/(X_{\text{MgO}}^{\text{liq}} + X_{\text{FeO}}^{\text{liq}})]$, where the 'liquid' is either the host whole-rock or matrix glass. FeO is calculated from Fe_2O_3 (Electronic Appendix Table 3) assuming $f\text{O}_2$ is buffered by quartz-fayalite-magnetite (QFM; Stout & Nicholls, 1977; Christiansen & McCurry, 2008). The two sets of glass data from North Robbers represent glass analyses from two different samples. Continuous and dashed curves represent $K_D(\text{Fe-Mg})^{\text{ol-liq}} = 0.30 \pm 0.03$ (Roeder & Emslie, 1970). (b, c) Predicted vs observed phenocryst compositions for clinopyroxene (cpx) (b) and plagioclase (plag) (c), when treating whole-rocks as liquids, and using calculated P as input (see Putirka, 1999, 2005). (d) Plagioclase saturation temperatures calculated using the whole-rock as a liquid with calculated P as input (Putirka, 2005); saturation temperatures are plotted vs crystallization temperatures calculated using plagioclase and whole-rock compositions as input. (e) Crystallization temperatures for olivine (Beattie, 1993) vs plagioclase (Putirka, 2005). For (b)–(e) 2σ error bars are shown. (f) Pressures calculated from clinopyroxene-liquid vs plagioclase-liquid equilibria.

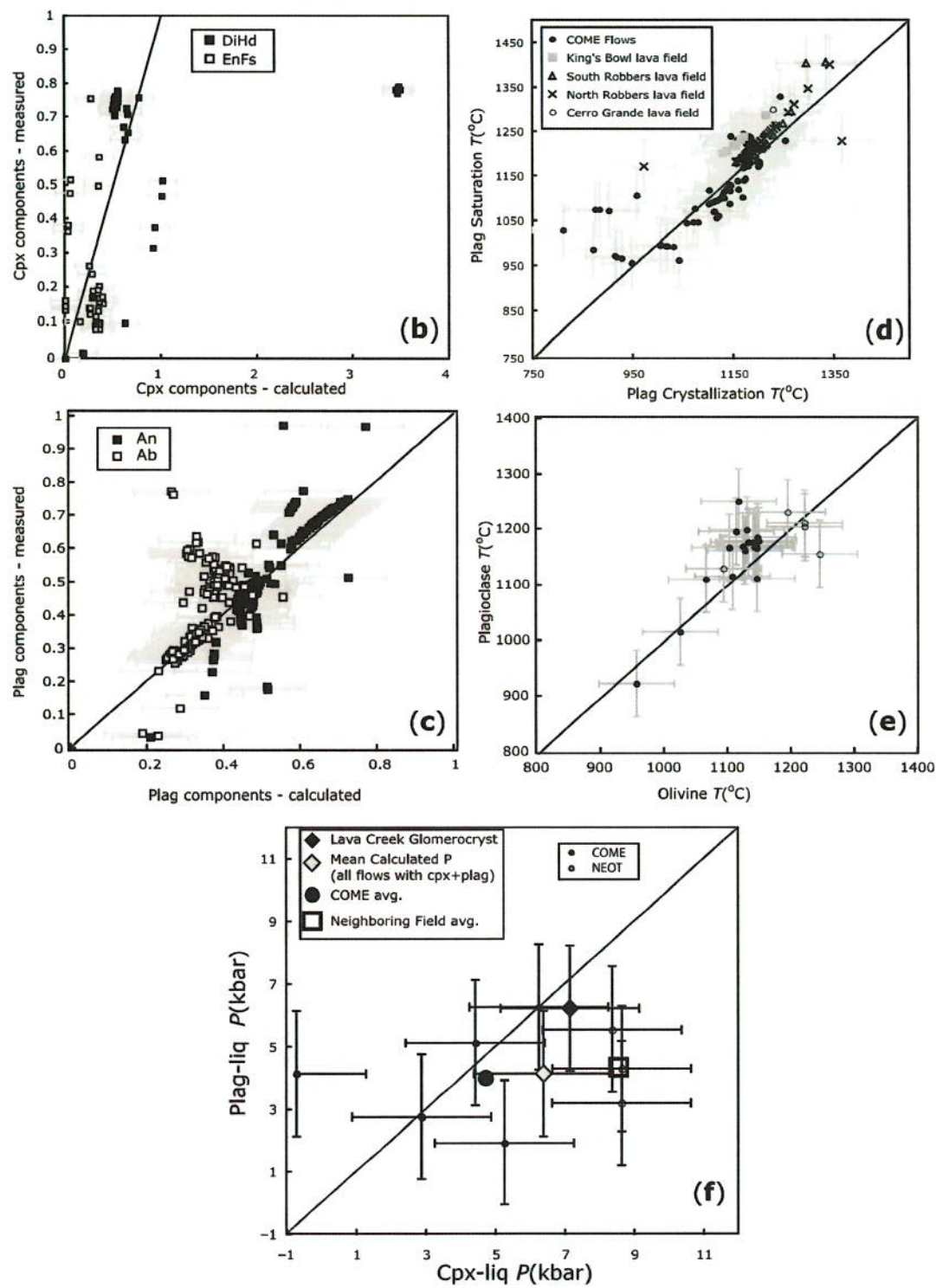


Fig. 4. Continued.

Table 2: P–T estimates

	Olivine <i>T</i>		Clinopyroxene <i>P–T</i>		Plagioclase <i>P–T</i>		
	Whole-rock	Glass	Whole-rock		Whole-rock (anhydrous)		Whole-rock (with 1% H ₂ O)
	<i>T</i> (K)	<i>T</i> (K)	<i>T</i> (K)	<i>P</i> (kbar)	<i>T</i> (K)	<i>P</i> (kbar)	<i>T</i> (K) <i>P</i> (kbar)
<i>Craters of the Moon flows (COME)</i>							
Big Craters	1338	1283			1387	3	1353 2.19
Big Craters	1437	1281			1392	2.8	
Bottleneck Lake	1395		1402	5.3	1388	1.9	1385 1.8
Broken Top					1437	4.5	1352 2.57
Carey Kipuka	1407		1357	4.4	1443	5.1	1382 4
Carey, mid-distal	1463				1452	6.1	1382 5
Carey, proximal	1428	1297			1438	5.1	1372 3.16
Devil's Cauldron	1482	1250			1442	7	
Grassy Cone Flow, Laidlaw Lake	1424						
Grassy Cone, distal	1426	1340			1526	7.2	
Grassy, mid-distal	1448	1245			1453	5.9	1374 3.84
Grassy, proximal	1420	1245			1452	5.2	1374 3.37
Highway Flow	1291	1155	1074	−0.7	1199	4.2	
Heifer	1448						
Indian Wells South	1335		1293	2.9	1292	2.8	
Kimama	1452				1444	6.1	1385 4.112
Lava Creek	1435	1304	1418	6.2	1447	6.3	1384 4.15
Little Prairie	1431				1445	5	1378 3.14
Little Laidlaw Park					1396	5.1	1368 4.49
Minidoka	1467	1302			1455	6.4	1388 4.8
Pronghorn Flow	1450	1256			1453	6.1	1383 4.27
Range Fire	1454	1254			1462	6.5	1387 4.49
Serrate	1393				1406	4.5	1353 3.23
Sunset flow, distal	1429	1313			1475	6.8	1383 4.89
Sunset, medial	1402				1472	6.9	1384 5
<i>Neighboring Snake River Plain (SRP) tholeiites (NEOT)</i>							
Cerro Grande					1465	4	1414 2.85
Kings Bowl					1465	4	1414 2.85
Cerro Grande					1465	4	1414 2.85
Cerro Grande					1465	4	1414 2.85
Cerro Grande					1465	4	1414 2.85
Cerro Grande					1465	4	1414 2.85
Kings Bowl	1542	1512	8.6	1432	3.2	1393	2.8
North Robbers 1	1503	1336	1510	8.6	1480	4.3	1441 3.3
North Robbers 2	1505	1487	4.5	1441	3.3		
South Robbers 2	1480	1481	8.4	1506	5.6	1432	3.4
South Robbers 1	1478	1379			1498	5.2	1432 3.6

Olivine *T*(K): (1) Putirka *et al.* (2007), Eqn. 2; (2) Putirka *et al.* (2007), Eqn. 4; (3) Beattie (1993). Clinopyroxene *P*(kbar)-*T*(K): Putirka *et al.* (2003), using corrections noted in text; Plagioclase *P*(kbar)-*T*(K): Putirka (2005), using corrections noted in text; calculations are shown when no water is used for the calculations, and when 1 wt % H₂O is assumed; see text for details. "Whole Rock" indicates *P–T* conditions calculated using whole rock compositions as a liquid. Glass indicates *P–T* calculations using microprobe analyses of interstitial glass to represent liquid.

flow yields two or more plagioclase phenocrysts or microphenocrysts whose compositions are consistent with an approach to equilibrium with respect to their whole-rock hosts (Fig. 4c) and whose plagioclase saturation temperatures closely match calculated crystallization temperatures (Fig. 4d).

Although some groundmass compositions meet these conditions, we approach their P - T estimates with caution. Rapid crystallization, or incidental fluorescence of adjacent phases during microprobe analysis can yield unreliable mineral compositions, and hence misleading P - T estimates. Clinopyroxene provides an example. Seventy per cent (35 of 50) of clinopyroxene analyses fail the equilibrium test. Of those that pass, all but one yield pressure estimates of <10 kbar. The exception is a phenocryst from the North Robbers lava field, which yields $P = 33.4$ kbar, but this phenocryst also contains 1.32 wt % K_2O . In contrast, all remaining clinopyroxene crystals have $K_2O < 0.8$ wt %. The 33.4 kbar pressure estimate from the North Robbers field is almost assuredly too high, and attributable to an inflated Na_2O -in-clinopyroxene content as a result of fluorescence of adjacent, alkali-rich glass. We thus use K_2O as an additional filter for clinopyroxene P - T estimates. For some clinopyroxene phenocrysts, a slightly better fit between calculated and measured clinopyroxene compositions is observed when matrix glass is used as the putative liquid (Electronic Appendix Table 2).

Plagioclase is no less susceptible to such errors; K_2O contents, in the form of orthoclase (X_{Or}), can be used as a filter, as plagioclase generally contains <10% X_{Or} (Fuhrman & Lindsley, 1988). In the Craters of the Moon lava field, several plagioclase phenocrysts yielding pressures in excess of 11 kbar have (X_{Or}) >0.2, which we use as a liberal upper limit for X_{Or} contents. To filter against contamination with adjacent glass, we also consider TiO_2 contents, as plagioclase should contain little TiO_2 compared with coexisting glass. Most COME plagioclase shows little evidence of glass contamination, and calculated pressures are uncorrelated with plagioclase TiO_2 contents ($R = 0.04$). Interestingly, replacing glass for whole-rock compositions in the P - T calculations results in poorer matches for calculated vs measured plagioclase compositions, and slightly higher P - T estimates overall. For plagioclase, we thus accept P - T estimates (using whole-rocks as liquids) for those crystals whose predicted and measured mineral compositions, and saturation and crystallization temperatures, lie within 2σ of model error (Fig. 4a), and whose X_{Or} contents are <0.2.

Pressures and temperatures of crystallization for COME and NEOT magmas

P - T conditions are calculated using models from Beattie (1993) Putirka (2005) and Putirka *et al.* (2003), and the pressure corrections expressed in equations (4) and (5). Because we infer that the NEOT and COME magmas are

relatively dry, plagioclase T estimates are based on thermometer B from Putirka (2005); this model uses hydrous data for calibration, but does not include a separate term for water (precisely so that T can be estimated for samples containing some water, without knowing actual water contents, as in this case). Equation B is less precise than equation A from Putirka (2005), but contains no systematic error for hydrous samples. Although our judgment is that water contents are low, we also illustrate the effects of water by using equation A from Putirka (2005), with the assumption that COME and NEOT magmas equilibrated with plagioclase with 1% H_2O . P - T estimates are based on core compositions of phenocrysts or microphenocrysts that, when paired with whole-rock compositions, pass our tests of equilibrium.

After applying various mineral-liquid equilibrium filters, clinopyroxene pressure estimates range from -0.7 (effectively 1 atm) to 8.6 kbar, whereas pressure estimates from plagioclase (dry conditions) range from 1.9 to 7.2 kbar (or from 1.8 to 5 kbar if $H_2O = 1\%$). Age and volume of eruption appear to be largely uncorrelated with P or T , although the four highest volume flows (>2 km³) all yield plagioclase-liquid temperatures >1400°C. P estimates are converted to depth using crustal densities from Fig. 2. The broad correspondence of clinopyroxene- and plagioclase-derived pressures (depths) appears to validate the use of whole-rocks as liquids (Fig. 5). As a further test, we compare P - T estimates on a flow-by-flow basis: if P (or T) values derived from different mineral-liquid equilibria are correlated, then inter-flow variations in P or T are more likely to reflect real differences in crystallization conditions. For olivine, temperatures for a given flow are calculated from the mean of all olivine crystals whose compositions yield $K_D(Fe-Mg)^{ol-liq} = 0.30 \pm 0.09$ (a 3σ error) for any given flow; for plagioclase, we use all crystals that fall within 2σ of the equilibrium tests illustrated in Fig. 4c. Except for the Kings Bowl lava field, all olivine and plagioclase crystallization temperatures overlap within 1σ of thermometer errors (Fig. 4e), and Kings Bowl T estimates overlap within 2σ . Of course, these values need not overlap. Olivine crystallization, for example, may well precede precipitation of other phases, and hence yield higher estimates for T (e.g. Putirka, 1997; Putirka & Condit, 2003). The overlap of olivine and plagioclase T estimates for COME and NEOT flows suggest that olivine and plagioclase co-precipitated, and that inter-sample temperature differences reflect real differences in crystallization temperatures.

As a test of P estimates, we also compare P for the few flows that contain both clinopyroxene and plagioclase, and for a clinopyroxene-plagioclase pair from a cognate glomerocryst (Leeman, 1974). For the glomerocryst, clinopyroxene and plagioclase pressure estimates are just 0.9 kbar apart, at 7.1 and 6.2 kbar, respectively. For the

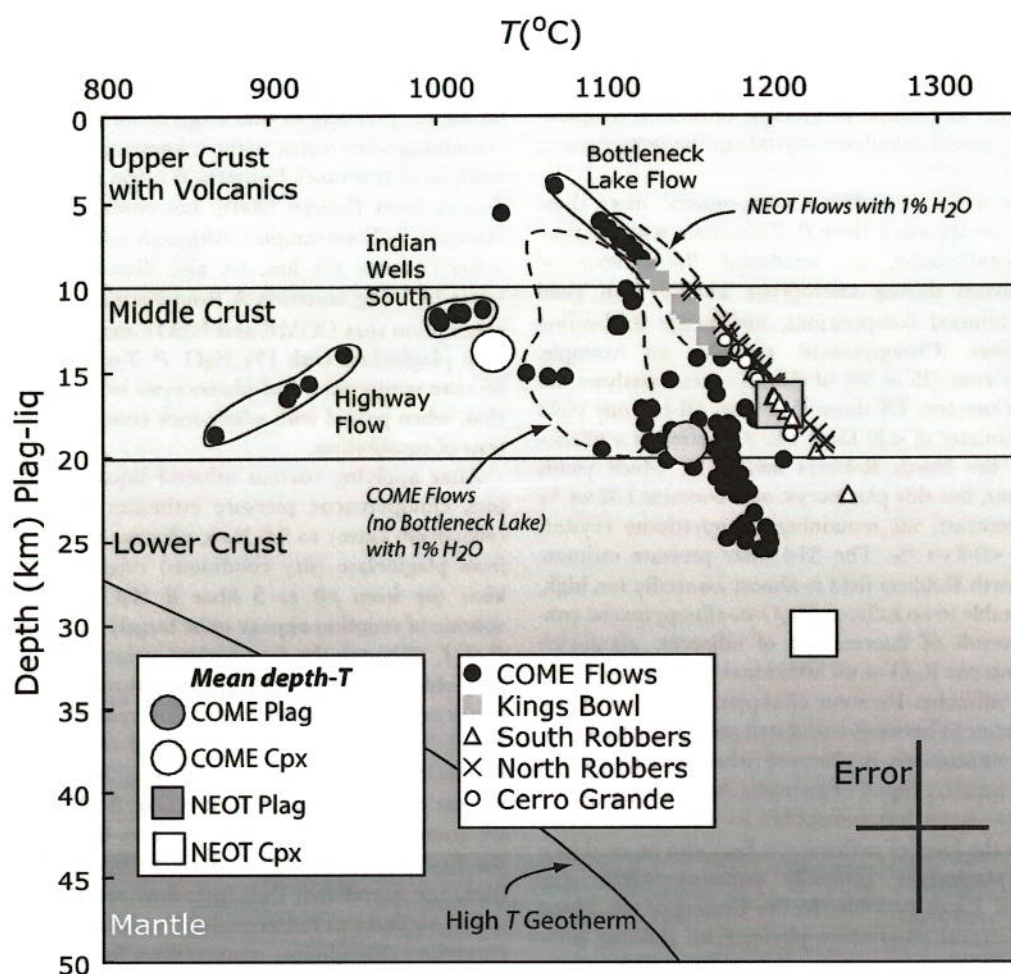


Fig. 5. Depth– T estimates from plagioclase–liquid pairs are compared with inferred crustal stratigraphy from Smith & Braile (1994) and Peng & Humphreys (1998). Depth, d (km), can be calculated from P (Table 2) for the ESRP, with trivial error compared with error on input parameters using: d (km) = $0.4626074 + 3.6592182P$ (kbar) – $0.0211427P$ (kbar)² + $0.000024P$ (kbar)³. 1σ error bars for depth and T are shown. The continuous curve to the bottom left is the high- T geotherm from Bott et al. (1978), which is approximated as T (°C) = $37.0345d$ (km) – $0.2815d$ (km)²; the geotherm at the SRP margin (Bott et al., 1978) can be approximated as T (°C) = $24.05d$ (km) – $0.0802d$ (km)². Each COME and NEOT flow is represented by several points, representing d – T estimates from crystals that pass our equilibrium tests. The highly coherent trends for NEOT-derived crystals and for Bottleneck Lake reflect the interdependence of the P – T calculations, and represent a ‘saturation surface’ for plagioclase for these liquid compositions. The effect of adding 1% H_2O when calculating plagioclase–liquid P and T is also illustrated, as dashed lines that delimit depth– T estimates for the COME (minus Bottleneck Lake) and NEOT.

Highway flow, however, clinopyroxene (–0.7 kbar) and plagioclase (4.2 kbar) pressures are very far apart (Table 2; Fig. 4f). Although these estimates may indicate contrasting saturation conditions, the presence of disequilibrium mineral textures places Highway flow estimates in doubt. For the remaining nine flows that contain both plagioclase and clinopyroxene (using whole-rock compositions as liquids), pressures overlap within 2σ (Table 2) for five flows (Fig. 4f). Pressure estimates for the COME flows are moderately correlated ($R=0.50$), but are few in number (four). The mean of the clinopyroxene and plagioclase pressure estimates for the COME flows are within error, at 4.7 ± 1.4 and 4.0 ± 2 kbar, respectively;

crystallization temperatures are close, with clinopyroxenes yielding slightly lower crystallization temperatures ($1094 \pm 56^\circ\text{C}$) than plagioclase ($1119 \pm 72^\circ\text{C}$), consistent with the findings of Thompson (1975) that plagioclase is the liquidus phase at these pressures. Most COME flows do not contain clinopyroxene, and the mean P for all COME plagioclase-based estimates is $P=4.8 \pm 2.1$ kbar (average $T=1152 \pm 68^\circ\text{C}$). We tentatively conclude that most COME magmas partially crystallized at 4.8 ± 2 kbar (or 2.8 ± 1 kbar if $H_2O=1\%$), which is equivalent to about 17 km depth (14 km if $H_2O=1\%$) (Fig. 5), and where clinopyroxene is present, it co-precipitated or closely followed plagioclase crystallization (Fig. 5).

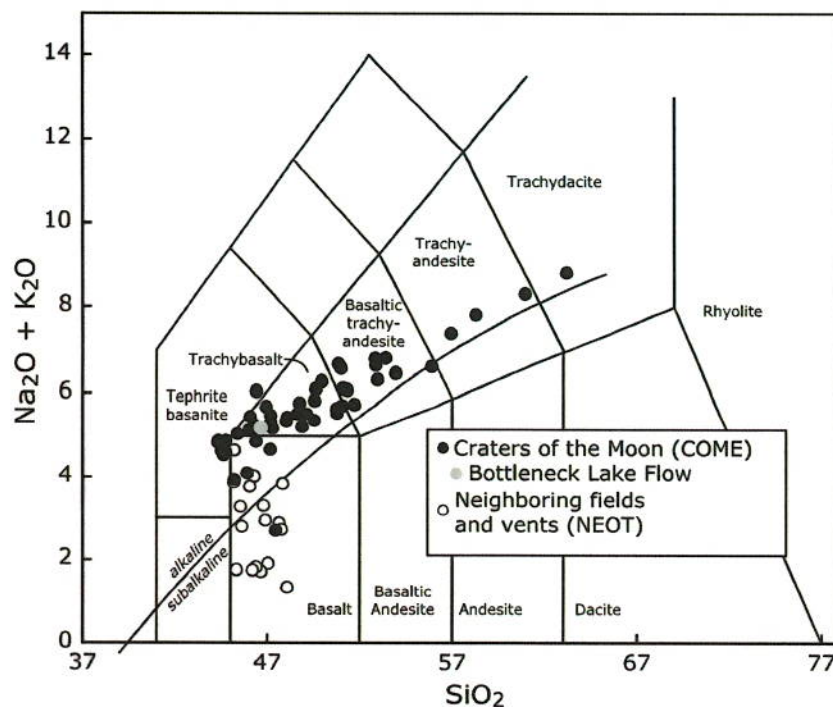


Fig. 6. Alkali-silica diagram, with the classification scheme of Le Bas *et al.* (1986) and the alkalic-subalkalic boundary curve of Irvine & Baragar (1971). COME are more evolved, having higher alkalis and (not shown) lower MgO and Mg-number. The Bottleneck Lake flow (gray symbol) yields NEOT-like mineral-melt P - T estimates (Fig. 5), but plots as part of the COME magma suite with respect to major oxides.

In contrast, for the three NEOT flows that contain clinopyroxene phenocrysts, clinopyroxene pressures (8.5 ± 0.2 kbar; ~ 30 km) clearly exceed those for plagioclase (4.3 ± 1.1 kbar; ~ 15.8 km; 3.2 kbar and 12 km if $H_2O = 1\%$). Clinopyroxenes also yield consistently higher temperatures compared with plagioclase at $1228 \pm 17^\circ\text{C}$ and $1194 \pm 40^\circ\text{C}$, respectively. These results suggest that, where present, clinopyroxene crystallization preceded plagioclase for the NEOT flows, and occurred at much greater depths.

Depths (d) and temperatures (T) of crystallization—differences between COME and NEOT

COME and NEOT flows yield a broad array of temperatures, which at first glance appear to overlap considerably, but there are distinct thermal differences. Because mineral saturation temperatures are strongly affected by P , it is crucial to compare T not over a range of pressures, but rather at a given P (i.e. depth). In effect, our P (depth) estimates provide this necessary correction to T for the effect of P on plagioclase saturation (Fig. 5). At any given depth, COME flows (the sole exception being Bottleneck Lake) yield lower T and a broader T range compared with NEOT flows—differences that must reflect their respective ascent histories.

Although we suspect that water contents for NEOT and COME magmas are low, we test for the effect of adding

1% H_2O (the midpoint between our estimates of water contents for basaltic samples) when calculating P and T (dashed lines in Fig. 5). Some of the lowest T estimates are absent because of a lack of numerical convergence for the simultaneous solution of the barometer and thermometer. Otherwise, equilibration depths and temperatures are shifted to lower values, as expected. However, the general inter-suite relationships hold: stagnation depths are largely within the middle crust, and COME flows are displaced to lower T , and exhibit a broader T range at any given depth compared with NEOT flows.

In summary, three features of the d - T estimates (Fig. 5) are critical. (1) At any given depth, plagioclase phenocrysts from NEOT flows encompass a very narrow T range, whereas COME flows yield a broad T range. (2) d - T estimates from the COME flows span a similar depth range compared with NEOT flows, but yield a nearly non-overlapping array of d - T estimates, with lower absolute T at any given depth. (3) Clinopyroxene crystallization depths are greater for NEOT flows. The Bottleneck Lake flow of the Craters of the Moon lava field is an exception. Although situated within the Craters of the Moon lava field, plagioclase phenocrysts from this flow lie on an extension of the NEOT trend, but displaced to shallower depths—in spite of Bottleneck Lake having a whole-rock composition very similar to other COME flows (Fig. 6). The Highway and Indian Wells South flows of the Craters

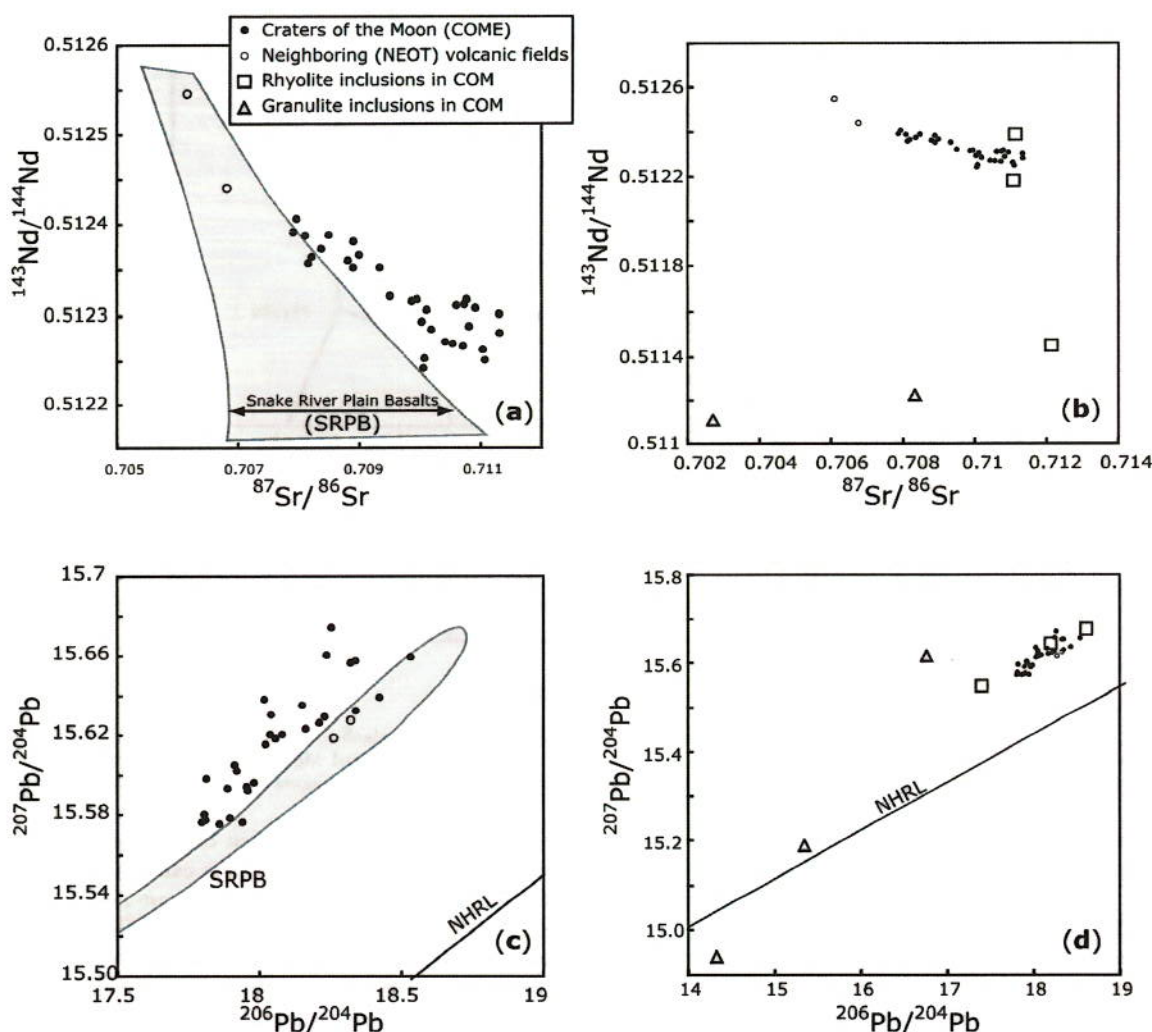


Fig. 7. Isotopic compositions for COME and NEOT, and Snake River Plain Basalts (SRPB; gray fields) that are similar to NEOT. (a) and (b) show the COME trend to higher $^{87}\text{Sr}/^{86}\text{Sr}$ and lower $^{143}\text{Nd}/^{144}\text{Nd}$. (c) and (d) show that NEOT flows are offset to somewhat lower $^{207}\text{Pb}/^{204}\text{Pb}$ at a given $^{206}\text{Pb}/^{204}\text{Pb}$ compared with other Snake River Plain basalts. NHRL is the Northern Hemisphere Reference Line of Hart (1984); the field for SRPB is derived from data from GEOROC (<http://georoc.mpg-mainz.gwdg.de/georoc/>).

of the Moon lava field are also distinct; minerals that pass our equilibrium tests (i.e. some non-sieve textured plagioclase) yield much lower crystallization temperatures compared with other COME magmas (Table 2), so low P - T storage conditions for these flows appear likely.

Whole-rock and inclusion geochemistry

With some exceptions, our new data confirm many of the findings of prior studies regarding the composition and evolution of COME and NEOT flows (Leeman *et al.*, 1976; Kuntz *et al.*, 1985, 1992; Stout *et al.*, 1994). Leeman & Manton (1971) observed that NEOT flows exhibit a very narrow range of $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (0.707 ± 0.00015) and that much higher $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (0.7080–0.7180) characterize

flows from the Craters of the Moon lava field. Our new isotopic data (Electronic Appendix Table 4; Fig. 7) confirm their findings, and those of Menzies *et al.* (1984), in that this time-integrated enrichment for COME flows extends to other isotopic systems. $^{143}\text{Nd}/^{144}\text{Nd}$ ratios are much lower for COME flows compared with NEOT flows, and $^{143}\text{Nd}/^{144}\text{Nd}$ is collinear with $^{87}\text{Sr}/^{86}\text{Sr}$ (Fig. 7), indicating contamination of COME magmas by materials with long-term Nd and Rb enrichments. Lead-isotope data show that the COME flows are displaced further from the Northern Hemisphere Reference Line (Hart, 1984) compared with NEOT flows (Fig. 7), but the NEOT and COME suites exhibit comparable ranges in $^{208}\text{Pb}/^{204}\text{Pb}$, $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{206}\text{Pb}/^{204}\text{Pb}$ (Electronic Appendix Table 4).

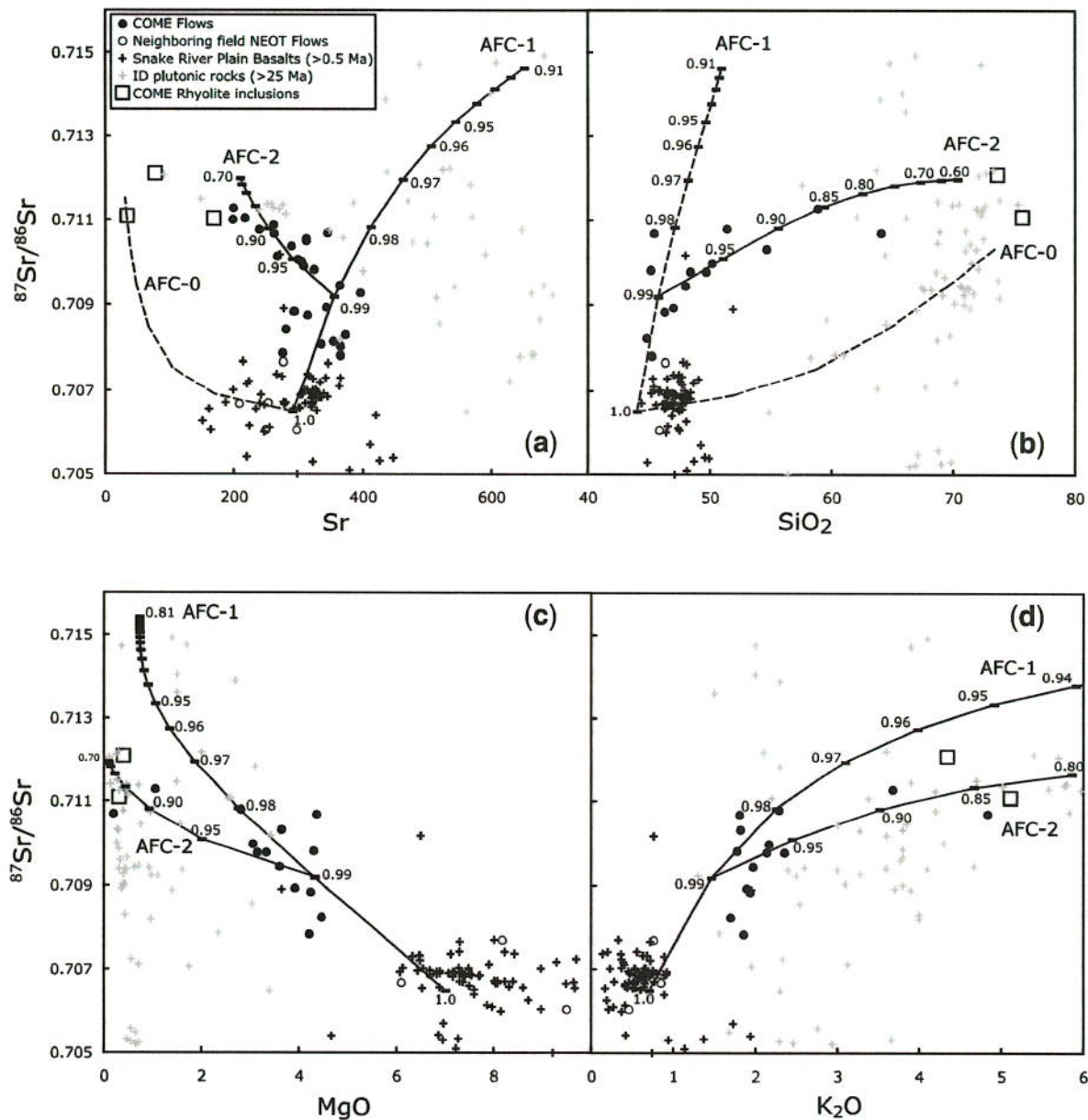


Fig. 8. $^{87}\text{Sr}/^{86}\text{Sr}$ ratios vs (a) Sr, (b) SiO_2 , (c) MgO and (d) K_2O for COME and NEOT flows. Also shown for reference are (a) older (>0.5 Ma) Snake River Plain tholeiitic basalts and (b) plutonic rocks from Idaho; both datasets are from NAVDAT (<http://navdat.kgs.ku.edu/>). Dark curves are calculated assimilation–fractional crystallization (AFC) trends, with tick marks indicating melt fraction, F . The light dashed lines in (a) and (b) (AFC-0) show a combined AFC model (rate of assimilation/rate of fractional crystallization, $r=0.8$; $D_{\text{Sr}}=2.5$; $D_{\text{SiO}_2}=1.02$), using Snake River Plain olivine tholeiite and rhyolite inclusions as end-members. Assimilation curves AFC-1 (of gabbroic crust) and AFC-2 (of felsic crust) illustrate a possible two-stage assimilation path that can explain COME magmas.

Leeman & Manton (1971) rejected a role for assimilation of buried rhyolite to explain the elevated $^{87}\text{Sr}/^{86}\text{Sr}$ ratios for COME flows. Instead, they called on ‘Precambrian metamorphic’ materials as a compositional end-member. Our new isotopic data show that rhyolite and granulite inclusions have different isotopic characteristics: rhyolite inclusions have $^{87}\text{Sr}/^{86}\text{Sr} > 0.712$ and $^{206}\text{Pb}/^{204}\text{Pb} > 17$,

whereas granulite inclusions have $^{87}\text{Sr}/^{86}\text{Sr} < 0.710$ and $^{206}\text{Pb}/^{204}\text{Pb} < 17$ (Fig. 7). The rhyolite inclusions provide a better description of COME isotope variations; however, as we discuss below, our conclusions are similar to those of Leeman & Manton (1971), as assimilation of other local intrusive rocks can also explain COME isotope variations.

DISCUSSION

Geochemistry and liquid evolution

New geochemical data and P – T estimates derived from mineral–melt thermobarometry for COME and NEOT flows indicate differences in ascent paths and total residence times. These differences lead to an explanation of the stark contrasts in their respective magma compositions, perhaps explaining the controls on the eruption of polygenetic evolved (COME-type), and monogenetic basaltic lava fields (NEOT-type).

Wall-rock assimilation (AFC-2)

Strontium and Nd isotopic ratios of COME clearly show that these magmas interacted with enriched crustal components (Figs 7b and 8), similar in composition to the rhyolite inclusions. As we show in the next section, such crustal interaction probably represents a later stage of assimilation, which we refer to as AFC-2. In this late-stage process, Pb isotopes further support assimilation of a component very similar to COME rhyolite inclusions, as the ranges of $^{207}\text{Pb}/^{204}\text{Pb}$, $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$ in COME flows are matched by similar ranges found for these inclusions, but not those for granulite inclusions (Fig. 7b and d; Electronic Appendix Table 4). COME major- and trace-element compositions also trend toward rhyolite-like compositions (e.g. Sr, SiO_2 , MgO, and K_2O ; Fig. 8).

However, the rhyolite inclusions are not necessarily the actual wall-rock with which COME magmas interacted. First, older crustal components cannot be excluded as potential assimilants; a survey of rocks >25 Ma from NAVDAT (<http://navdat.kgs.ku.edu/>) indicated that some plutonic rocks in Idaho have appropriate isotopic and major-element compositions, and so provide wall-rock materials that are suitable to explain COME geochemical variations (Fig. 8). Second, P estimates indicate partial crystallization of COME magmas in the middle crust (Fig. 5), but Tertiary rhyolite flows exist in the upper crust (Smith & Braile, 1994). We thus suspect that COME flows derive their isotopic character from a recycling of a middle crust (curve AFC-2, Fig. 8) source that has a composition very similar to the rhyolite inclusions (Leeman & Manton, 1971; Menzies *et al.*, 1984). Yellowstone may provide an analog of the process. Bindeman *et al.* (2007) and Vasquez *et al.* (in press) showed that rhyolites in the Yellowstone region may be the products of partial melting of older rhyolites, which may themselves result from assimilation of yet older continental crust. Pre-Holocene rhyolites in the SRP also appear to carry a continental-crust geochemical signature (e.g. Hildreth *et al.*, 1991; Bindeman *et al.*, 2007; Bonnicksen *et al.*, 2007; Christiansen & McCurry, 2008). And indeed, thermometry by Vasquez *et al.* (in press) showed that the highest T for Yellowstone rhyolites (900°C) approaches the lowest T for COME

magmas, indicating thermal continuity between these suites. We surmise that the isotopic evidence for assimilation, although consistent with direct assimilation of rhyolite, may instead reflect assimilation of deeper-seated wall-rocks that may have also been the source for pre-Holocene SRP rhyolites (Fig. 2).

Although assimilation is not unique to the Craters of the Moon lava field (e.g. Honjo & Leeman, 1987), COME flows provide an interesting contrast to the highly evolved ESRP flows at Cedar Butte, and Unnamed Butte, which appear to have evolved largely by fractional crystallization (McCurry *et al.*, 2008) of basaltic precursors. For the COME however, it is also clear that no one-stage assimilation process (e.g. Menzies *et al.*, 1984) can successfully link COME magmas to an NEOT parent.

Relationships between NEOT and COME magmas (AFC-1)

Kuntz *et al.* (1992) showed that COME magmas cannot be explained by differentiation of primitive NEOT magmas, and our new data re-emphasize that result (Fig. 8). Fractional crystallization alone, simple mixing, and one-stage AFC models (DePaolo, 1981) using rhyolite as an assimilant and primitive NEOT as a parental liquid, all fail to reproduce COME compositional trends (Fig. 8a and b). In addition, as noted by Leeman & Manton (1971), even the most primitive of COME flows have $^{87}\text{Sr}/^{86}\text{Sr}$ ratios too high to reflect a mantle source. With whole-rock Mg-number [$=X_{\text{Mg}}/(X_{\text{Mg}}+X_{\text{Fe}})] < 0.60$ (Fig. 4a), none of the COME liquids are likely to have been in equilibrium with a mantle mineral assemblage (Asimow & Longhi, 2004).

Some other process is thus needed to generate the most primitive COME compositions. One possibility is that the sub-SRP mantle yields different parental magmas for the COME and NEOT flows. We are inclined to reject this hypothesis as it is unclear why the dominant and highly repeatable magma parental to NEOT would be delivered everywhere except in the area of the Craters of the Moon lava field. Before accepting the added complexity of a distinct mantle source, we test whether COME magmas can be derived from NEOT magmas by a deeper level of assimilation (that precedes AFC-2, as discussed above) which we designate as AFC-1. AFC-1 (Fig. 8) involves a mafic NEOT composition as a parental liquid, and a hypothetical middle (or lower) crust assimilant. Because the most primitive of COME flows have SiO_2 contents nearly identical to NEOT flows, with similar Sr contents, but much lower MgO, it is necessary that NEOT parent magmas assimilate material with high Sr and $^{87}\text{Sr}/^{86}\text{Sr}$, and low to moderate SiO_2 . Materials that successfully connect NEOT magmas to the least evolved COME magmas (path AFC-1, Fig. 8) are 'mafic pods' that occur in granitoids of Yosemite National Park (Ratajeski *et al.*, 2001), which have the requisite high Sr, high $^{87}\text{Sr}/^{86}\text{Sr}$, and low

SiO₂. If such gabbroic rocks form part of the SRP lower crust, then only small amounts of assimilation (1%) are needed to increase ⁸⁷Sr/⁸⁶Sr to COME-like values, and yet still retain basalt-like SiO₂ contents for the most mafic of COME flows. Other major-element components of Yosemite mafic pods are similarly appropriate to explain the primitive end of the COME trends through an AFC process (Fig. 8). After generating this parental COME magma, other COME magmas can then be generated by the second period of assimilation (AFC-2), involving rhyolite, or similar compositions as already noted; in Fig. 8, the AFC-2 curve makes use of the end product of AFC-1 (i.e. contaminated NEOT magmas) as a 'parental COME magma'.

If the gabbros of AFC-1 exist, they are probably part of the older (Archean?) continental crust or upper mantle. Shervais *et al.* (2006) showed evidence of gabbro assimilation in SRP drill core samples, representing 'previously intruded basalt' (see their fig. 3). However, such gabbros will have ⁸⁷Sr/⁸⁶Sr ratios effectively identical to recently erupted NEOT flows; assimilation of relatively young 'previously intruded basalt' thus cannot explain primitive COME compositions with elevated ⁸⁷Sr/⁸⁶Sr (0.7145). To illustrate the problem, if SRP basalts are derived from a parent with (⁸⁷Sr/⁸⁶Sr)_i = 0.707 and Rb/Sr = 2, then 263 Myr are required to yield a rock with ⁸⁷Sr/⁸⁶Sr = 0.7145; this is much too old to be explained by SRP-related magmatism, and even then would require 100% assimilation to yield COME isotope ratios. In addition, NEOT flow compositions do not vary so as to indicate assimilation of high ⁸⁷Sr/⁸⁶Sr materials (Menzies *et al.*, 1984; this study). We thus surmise that older gabbros probably form part of the Archean middle or lower crust (Smith & Braile, 1994) and are better preserved near the SRP margins. A drawback to our model is that no such older gabbroic rocks are reported in the NAVDAT database (<http://navdat.kgs.ku.edu>) for Idaho. Such gabbros, however, are not commonly reported in the Sierra Nevada, California either, perhaps because they represent deep and un-exhumed portions of the lower crust. It seems plausible that, like granites, gabbroic rocks everywhere carry similar geochemical characteristics, and that such rocks are still a likely component in the sub-SRP crust.

Crystallization depths and wall-rock assimilation

The contrasting geochemistry of COME and NEOT flows is matched by contrasts in their conditions of storage and ascent paths. The narrow *T* range for the NEOT flows at any given depth indicates that such magmas are transported quickly enough to disallow significant cooling at any depth. This is not an unexpected result; the non-evolved nature of these flows suggests that storage times are too brief to allow crystals and magma to evolve to low *T*. Interestingly, however, the NEOT depth interval of

partial crystallization is broad, spanning more than half the thickness of the crust, which supports the magma-mush column views of Ryan (1988), Kuntz (1992) and Marsh (1995). In their models, magmas reside within a plexus of dikes and sills that are not restricted to any particular narrow depth interval, so there is no single isolated magma chamber acting as the locus for differentiation, or the source for eruptions.

In contrast, isotopes and *P-T* conditions for COME magmas (Table 2; Fig. 5) indicate longer-term storage and assimilation in middle crust reservoirs, at 3.8–4.8 kbar, or at about 14–17 km (Fig. 5). Because COME minerals appear to be in equilibrium with their host whole-rocks (Fig. 4), it appears that assimilation was either concomitant with or preceded the crystallization of most of the minerals that are contained at present in COME flows—assimilation thus occurred at depths equal to or greater than 14–17 km. Because assimilation is aided by the latent heat of crystallization (and magma recharge; see Bohron & Spera, 2001), magmas parental to the COME crystallized minerals that were largely un-entrained during COME eruptions. Two lines of evidence indicate the presence of a significant fraction of such buried magmatic material: (1) all COME flows have Mg-number <40 (Fig. 4a), and so require significant crystallization of olivine from mafic parent liquids; (2) the Bottleneck Lake flow contains multiple olivine grains with Fo70–Fo90 (Fig. 4a), providing a direct sampling of this high Mg-number crystalline residue.

Mineral melt barometry also implies that wall-rocks at 14–17 km or deeper (Fig. 5), have an isotopic and major-element composition similar to the rhyolite inclusions. The work of Smith & Braile (1994) indicated that silicic volcanic rocks at Yellowstone extend to depths not greater than 5–6 km. Rhyolites near the Craters of the Moon lava field volcanic field should be little different. However, Hildreth *et al.* (1991) showed that Yellowstone rhyolites contain a crustal contribution as high as 30–50%, and a survey of NAVDAT plutonic rocks from Idaho shows that some plutons are remarkably similar to the Craters of the Moon lava field rhyolite inclusions. We surmise that the isotopic compositions reflected by COME lava flows and their rhyolite inclusions probably reflect the isotopic composition of the middle crust.

Magma buoyancy and ascent paths

Ascent of COME and NEOT magmas

To test how density contrasts might control magma transport, liquid densities are calculated from the models of Lange & Carmichael (1990) and Ochs & Lange (1999), using whole-rock compositions and calculated *P-T* conditions as input (Figs. 9, 10). These calculations show that throughout the Craters of the Moon lava field region, magmas are buoyant at the Moho and within the lower

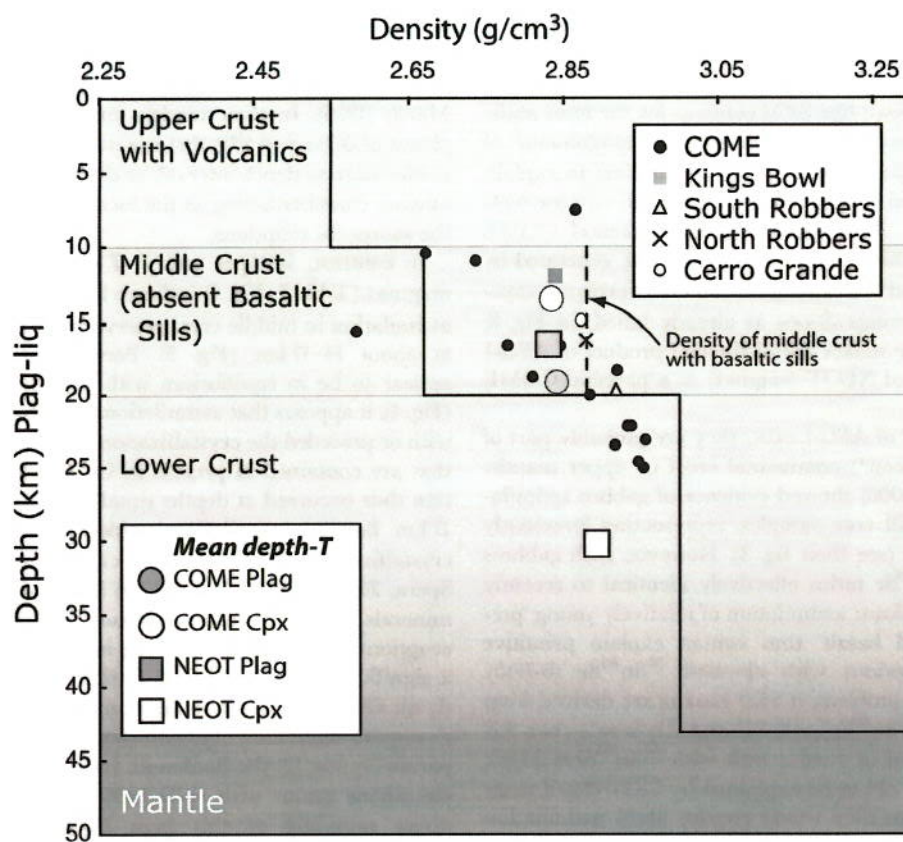


Fig. 9. Crust and liquid densities vs crystallization depths. Liquid densities are calculated from P - T conditions and whole-rock compositions [Lange & Carmichael (1990), using their equation (9) and coefficients from their table 3, and the coefficient for H_2O from Ochs & Lange (1999)]. Continuous line shows crustal densities where basaltic sills in the middle crust are absent; dashed vertical line in the middle crust shows the effect of adding basalt sills (see Fig. 2). In contrast to Fig. 5, each flow is represented as a single point, where the depth estimate derives from the mean of all depths calculated from individual phenocrysts (Electronic Appendix Table 1).

crust—the Moho is thus not the staging area for COME or NEOT eruptions. However, both COME and NEOT magmas (dry) are neutrally to negatively buoyant within the middle and upper crust (Fig. 9), provided such crust consists of felsic materials. For the COME, this result is consistent with the density trap model of Kuntz (1992) and Christiansen & McCurry (2008). Kuntz (1992) suggested that COME magmas might stall just beneath, and partially melt a plutonic body that underlies the Craters of the Moon lava field to a depth of 12 km, or beneath the ‘magnetized crust’, whose base is at 16–18 km—highly consistent with mineral–melt barometry.

Although NEOT magmas would also be neutrally buoyant within a felsic middle crust, isotopic and thermobarometry evidence shows that NEOT magmas do not stall within or interact with middle crust. This result may be explained by the particular ascent path traced by NEOT magmas. Geophysical studies (Brott *et al.*, 1978; Smith & Braile, 1994; Peng & Humphreys, 1998) show that along the axis of the ESRP, where most of the young NEOT magmas are erupted, felsic crust has been replaced by

basaltic sills (NEOT undoubtedly represent a later phase in a continuum of basaltic intrusions and eruptions that have reshaped the surface and subsurface of the ESRP). Earlier basaltic magmas partially melted pre-existing crust to form the pre-Holocene rhyolites (now exposed at the margins of the SRP; e.g. Leeman, 1982b; Perkins & Nash, 2002; Nash *et al.*, 2006; Bonnicksen *et al.*, 2007; Christiansen & McCurry, 2008). Later magmas that follow this same (compositionally armored) path would not show signs of felsic crust assimilation and would be positively buoyant in the middle crust (Fig. 9). Off the axis of the SRP, however, geophysical data indicate that the upper crust is partially preserved, at least in some places such as at the Craters of the Moon, (Kuntz, 1992; Kuntz *et al.*, 1992) (Fig. 11), and so is available to inhibit magma transport and contaminate upwelling magmas. Because NEOT flows show no evidence of assimilation of either rhyolite (path AFC-2; Fig. 8) or the ‘older’ gabbroic (path AFC-1) materials [to be distinguished from the younger gabbros, whose role Shervais *et al.* (2006) have documented], we surmise that both lithologies exist at the margins

of the SRP, but are now rare or absent along near-axis conduits of the SRP.

The Bottleneck Lake flow of the Craters of the Moon lava field is an exception to this model. Thermobarometry on this flow shows that it lies on a low P - T extension of the trend defined by NEOT flows. The Bottleneck Lake flow is among the least evolved of COME flows, and it may well have escaped much of the middle crust assimilation that is otherwise evident in other COME flows. The Bottleneck Lake flow is also one of the largest COME flows, so if its feeder dike system is wider, its transport could be more rapid [equation (3)] than for other COME flows. The Bottleneck Lake vent is also situated closer to the axis of the ESRP than most other COME flows, so it may transit crust of a more mafic composition, perhaps with greater buoyancy compared with other COME flows.

Implications of water contents, crystal growth rates, crystallinity

Our P - T estimates appear to show relatively rapid throughput of magmas parental to NEOT flows, and much slower throughput for COME parental and evolved magmas, especially within the middle crust. However, we would like to caution that instantaneous ascent rates at any given time might not be very different between COME and NEOT magmas, and that ascent velocities calculated from equation (3) do not necessarily apply at all stages of transport. For example, COME crystal growth, which occurred during or after wall-rock assimilation for COME magmas (not before), provides some time constraints. Because fracture transport ascent rates of 0.4–1 m/s from Kuntz (1992) are too rapid to allow COME phenocrysts time to form by ascent-induced devolatilization, a period of cooling and crystal growth following assimilation is implied. However, COME flows have low crystallinity compared with NEOT flows. The low crystallinity of COME flows may reflect efficient segregation of COME magmas from their crystalline residue following assimilation, with relatively rapid transport to the surface AFC processing. This is consistent with recent findings regarding magma storage and ascent times. Costa & Dungan (2005), for example, showed that assimilation can occur on a time scale of months to years, whereas Klügel (2001) has shown that magmas can be stored and partially crystallize for tens to hundreds of years prior to eruption. Our density calculations show that dry COME magmas would be positively buoyant with a density contrast ($\rho_m - \rho_r$) of at least 0.12–0.19 g/cm³ in the lower crust (Fig. 9), but would have zero or even negative buoyancy in the middle crust, where the density contrast is 0.1 g/cm³. Ochs & Lange (1999) showed that by concentrating H₂O, magma densities can be greatly reduced, even if H₂O occurs only at 1–2 wt % levels (Putirka & Busby, 2007). For example, if AFC caused magmatic water to increase

to 1.5% H₂O, COME magmas would become buoyant in the middle crust. If CO₂ were present, as seems likely, density would be further reduced, if for no other reason than that CO₂ is much less soluble in silicate melts (Moore, 2008), and should allow nucleation of a vapor phase even at great depths.

In contrast, the narrow T range and lack of evidence for assimilation indicate that NEOT magmas were not stored for extended periods within the crust; yet NEOT flows carry much larger crystals and are highly crystalline compared with COME flows. Their crystalline character almost certainly reflects their more mafic compositions; NEOT magmas crystallize at higher temperatures, and are much less viscous, which should result in higher crystal growth rates (by several orders of magnitude, Cashman, 1990; Hammer, 2008), compared with SiO₂-rich, MgO-poor, lower- T COME magmas. Differences in crystallinity can also mean that NEOT magmas were more efficient at carrying their crystal cargo. If nothing else, such considerations show that mineral textures taken alone can be misleading with regard to inferences of magma ascent, and that our attempts to calculate crystal growth times have narrow applicability.

Origin and ascent of primitive SRP (NEOT) magmas

McCurry *et al.* (2008) suggested that mafic olivine tholeiites might be erupted from the Moho, whereas Christiansen & McCurry (2008) used more detailed density arguments to suggest that a gabbroic middle crust may inhibit the upward transport of such magmas. We test these arguments further by considering density contrasts between SRP magmas generally, and consider the possibility that liquids more mafic than those erupted might lie trapped beneath the SRP.

Some early studies concluded that the SRP tholeiites represent near-primary peridotite partial melts (Stout & Nicholls, 1977; Kuntz *et al.*, 1992). However, few if any of NEOT flows are direct partial melts of peridotite; such partial melts should have Mg-number between 0.75 and 0.80 (Asimow & Longhi, 2004), but NEOT flows generally have Mg-number <0.7. Only the Kings Bowl (Mg-number = 0.76) is potentially viable as a mantle partial melt; however, even here MgO is <10%, which is low compared with oceanic primitive liquids (where continental crust does not interfere with magma transport) at Siqueiros, Iceland, Hawaii, or Samoa, which have MgO >14% (Putirka *et al.*, 2007). The SRP–Yellowstone system is thought to arise from a thermally driven mantle plume (Pierce & Morgan, 1992), although Christiansen *et al.* (2002) offered an alternative explanation for geophysical features of the region. Leeman *et al.* (in press) suggested that excess temperatures beneath Yellowstone are positive, but low (*c.* 50°C above ambient mantle); although they could not exclude a plume completely. Regardless of the cause of hotspot activity, the absence of high-MgO picritic

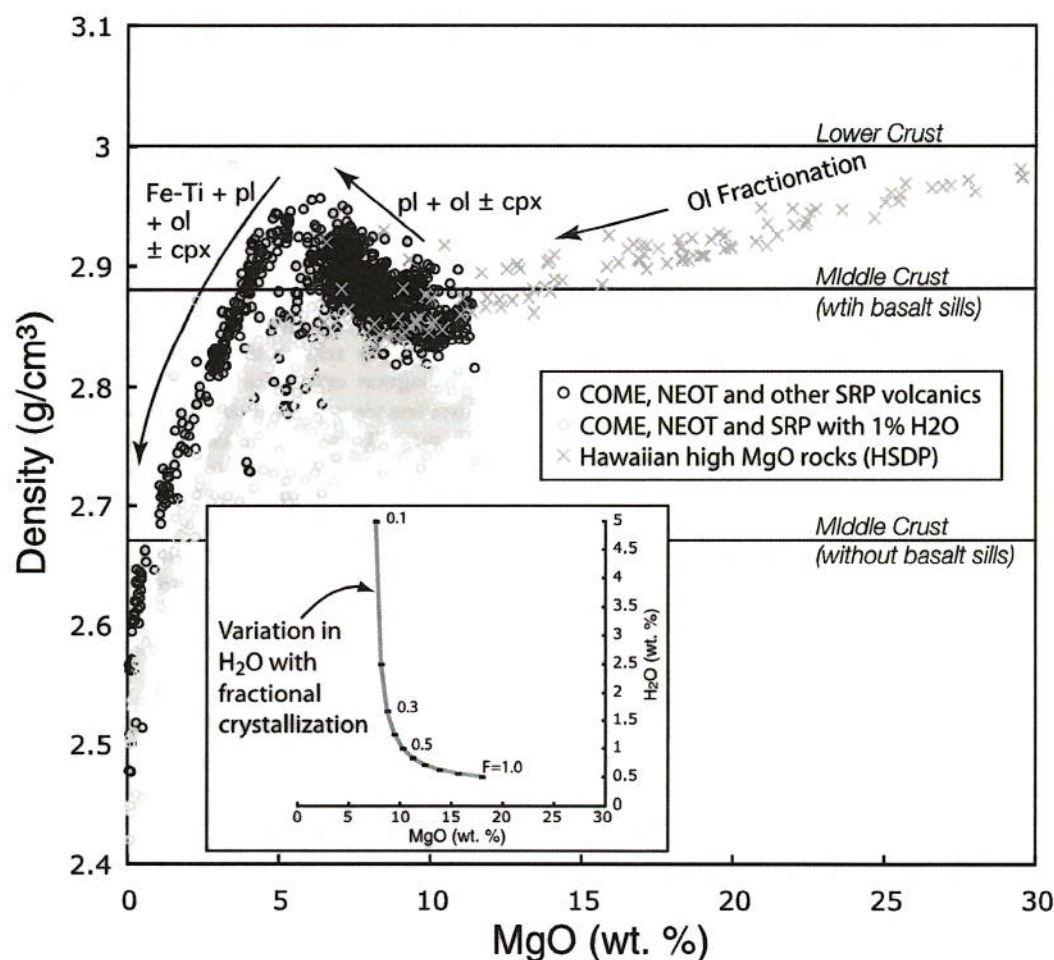


Fig. 10. Magma densities (calculated as in Fig. 9) vs MgO wt % for 1214 volcanic rocks from the Snake River Plain (Leeman *et al.*, 1976; Kuntz & Dalrymple, 1979; Leeman, 1982a, 1982b; Kuntz *et al.*, 1985, 1992; Shervais *et al.*, 1994; Knobel *et al.*, 1995; Reed *et al.*, 1997; Stout & Nicholls, 1977; Hughes *et al.*, 2002; McCurry *et al.*, 2008), and for Hawaiian lava flows from the Hawaii Scientific Drilling Project (HSDP) (Rhodes & Vollinger, 2004). Seismically determined densities for lower and middle crust (Fig. 2) are shown as horizontal lines. Arrows show the effects of fractionation of olivine (ol), plagioclase (pl), clinopyroxene (cpx) and Fe-Ti oxides (Fe-Ti), on liquid density (whole-rocks generally contain the indicated mineral assemblages). The most mafic of SRP flows plot at a density minimum defined by the intersection between the HSDP and SRP trends, at a density that roughly coincides with the middle crust 'Intermediate Layer' of Smith & Braille (1994).

magmas requires that either such high-MgO magmas are not generated or they are generated but not erupted. We favor the latter hypothesis.

The possibility that picrites could be generated in large volumes beneath the ESRP, but not erupted (McCurry *et al.*, 2008), is well illustrated at Hawaii. There, high-MgO picrites are especially abundant in deep-sea dredge samples off volcano flanks (Garcia, 2002), but rare at volcano summits. Continental crust should provide an even greater barrier to the eruption of high-MgO picritic flows compared with the basalts that constitute the subaerial crust at Hawaii. Stolper & Walker (1980) were the first to illustrate a solution to a dilemma initially posed by Carmichael *et al.* (1974), and applicable to the SRP: how is it that mafic continental tholeiites tend to erupt with very

similar compositions? Stolper & Walker (1980) showed that tholeiites from mid-ocean ridges, ocean islands, and continental settings tend to have compositions that plot near a density minimum. This minimum occurs because picritic magmas have olivine on the liquidus, and as olivine is removed, liquid density decreases. Later, with plagioclase $\pm$ clinopyroxene saturation, liquid densities increase, as a result of Fe enrichment. Later still, liquid densities decrease dramatically as SiO_2 continues to increase and as Fe-oxide phases precipitate and deplete liquids of Fe.

To test the model of Stolper & Walker (1980) and, more specifically, the models of McCurry *et al.* (2008) and Christiansen & McCurry (2008), MgO is plotted against anhydrous magma densities [calculated using Lange &

Carmichael (1990)] for various volcanic rocks from the SRP (Fig. 10). Also plotted are picrites and other high-MgO (>10%) rocks from the Hawaii Scientific Drilling Project (HSDP) (Rhodes & Vollinger, 2004). Horizontal lines representing estimates of the densities of lower and middle crust are compared with these densities. This comparison shows that the most mafic of SRP flows fall at a density minimum near 10% MgO, defined by the intersection of SRP and HSDP compositions; this minimum density is slightly lower than that for middle crust when accounting for the intrusion of basalt sills. We suggest that lavas with >10% MgO (Hawaiian-like picrites) are indeed generated beneath the ESRP, as at oceanic hot-spots, but that such compositions are trapped beneath a middle crust density filter. The position of the density minimum near 2.7 g/cm³ is supportive of the Christiansen & McCurry (2008) model, whereby parental NEOT magmas are trapped at the base of or within the middle crust. This plot supports the McCurry *et al.* (2008) hypothesis that picrites underlie the crust throughout the eastern ESRP, although they should exist at middle crust depths, rather than at the Moho.

A natural question is: why do magmas on the low-MgO side of the density minimum erupt if they are no more (and perhaps even less) buoyant than those on the high-MgO side? The answer is that just 1 wt % water is sufficient to decrease densities such that the most Fe-rich magmas (those at the density maximum in Fig. 10) are equal in density to dry magmas falling at the density minimum. And such modest water contents can be attained by differentiation of relatively dry magmas: a magma with 0.5% H₂O, and an initial Mg-number of 0.73 (18% MgO, 12% FeO) can yield a magma with 10% MgO, an Mg-number of 0.6 (very similar to NEOT magmas) and 1% H₂O, with 50% crystallization of olivine (Fig. 10, inset). We suggest that increases in water contents are also critical in explaining the density minima observed by Stolper & Walker (1980).

Residence times?

We caution that our analyses of buoyancy and *P-T* conditions do not lead to estimates of instantaneous ascent rates or absolute residence times. Ascent rates calculated from equation (3), for example, probably apply only to the latest stages of magma transport. Crystal sizes may place some broad constraints on time. For example, the largest crystals in the NEOT and COME suites (10 mm) can be created only if total residence times are of the order of 5 months to 30 years, assuming growth rates of 10⁻⁶ and 10⁻⁸ mm/s, respectively. Such times are minima for residence of parental mafic magmas. Whatever the minima, residence times for magmas parental to COME flows exceed the minima for NEOT flows, to achieve lower *T* and crustal contaminated signatures for COME flows. However, it is important to recognize that COME flows

are 'daughter products', and so leave behind a residue of crystals that are largely un-erupted. This means that crystals entrained by COME flows on the one hand, and NEOT flows on the other, record different aspects of ascent and storage history. It should be noted that NEOT magmas are well above the ambient geotherm and so begin crystallizing as soon as they enter the lower crust, if not before (although such early formed crystals might not be erupted), whereas COME crystals may record only ascent following a period of storage in the middle crust; longer crustal residence times inferred for COME magmas (e.g. from isotopes and thermometry) imply extended residence of parental products, erupted or not. Diffusion profile methods (Costa *et al.*, 2008), and where possible, U-series data (Cooper & Reid, 2008) and melt inclusions (Kent, 2008; Métrich & Wallace, 2008) are clearly needed, and will no doubt add greatly to our understanding of magma storage and ascent rates in the ESRP.

CONCLUSIONS

A key finding is that a main control on whether volcanic fields will be polygenetic (erupting a range of intermediate and evolved compositions) or monogenetic (erupting a narrow range of non-evolved lavas) is the architecture of underlying crust. Polygenetic COME flows occur where the felsic crust remains intact, which, as shown by geophysical data (Brott *et al.*, 1978; Smith & Braile, 1994; Peng & Humphreys, 1998; Shervais *et al.*, 2006), is at present mostly restricted to the margins of the SRP. In contrast to the high-volume, bimodal (basalt + rhyolite) volcanism of Yellowstone, COME-style polygenetic volcanism also appears to occur when the flux of magma from the mantle to the crust is relatively low. In contrast, monogenetic NEOT flows are erupted near the SRP axis, where the crust has been pre-conditioned by the passage of the Yellowstone hotspot, with its concomitant large-scale melting of middle crust (Christiansen, 2001) and emplacement of basalt sills; this history serves to increase middle crust density, and insulate magma conduits from ambient felsic crust. This is not to say that basalts can only erupt at the margins of the SRP—the Shoshone field in Fig. 1 shows otherwise; nor do we suggest that evolved lavas cannot erupt near the SRP axis—the Cedar Butte and Unnamed Butte rhyolites (which are not polygenetic; McCurry *et al.*, 2008) also show otherwise. However, there is no reason that crustal modification need everywhere be uniform.

We propose a three-stage model to illustrate how magma ascent is so influenced (Fig. 11); the model is highly consistent with Kuntz's (1992) model for the Craters of the Moon region, and aspects of other models that describe SRP evolution (Leeman & Manton, 1971; Leeman, 1982a, 1982b; Menzies *et al.*, 1984; Smith & Braile, 1994; Christiansen, 2001; Hughes *et al.*, 2002; Christiansen & McCurry, 2008).

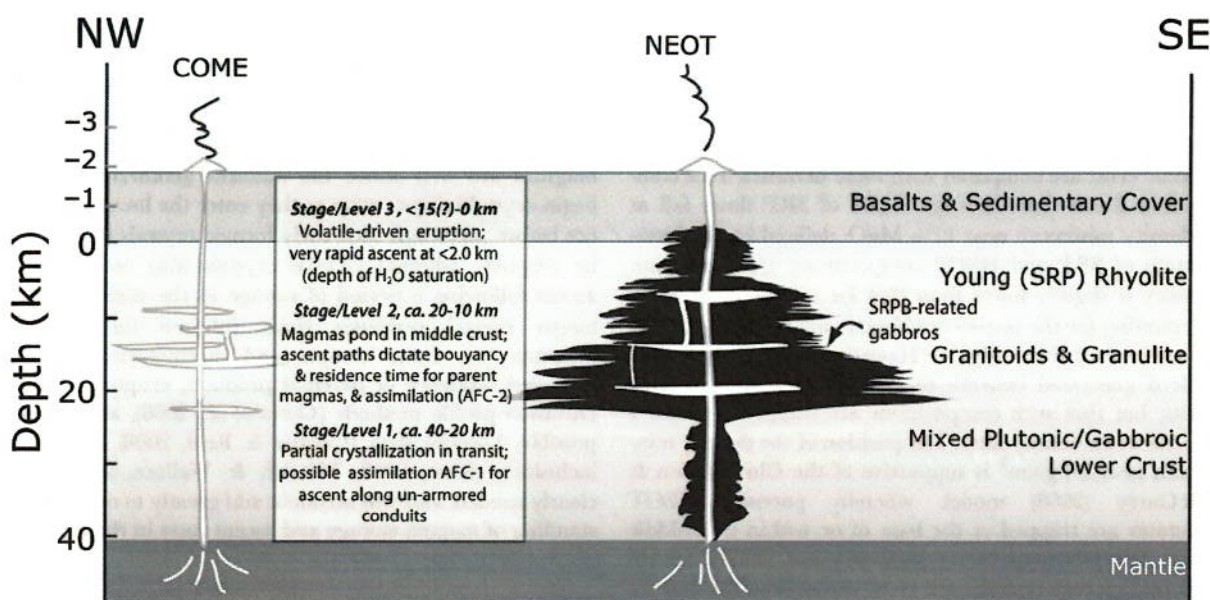


Fig. 11. Model for the development of magma conduits within the Snake River Plain. Cross-section, depth scale and crustal lithologies are adapted from Peng & Humphreys (1998) and Shervais *et al.* (2006). Dikes and sills from older volcanic episodes are shown in black; Holocene-related dikes and sills (NEOT, COME) are shown in white. NEOT exploit conduits that lie nearer the axis of the eastern Snake River Plain (ESRP) compared with COME flows, which erupt further from the ESRP axis. Compositional differences are explained with a three-stage model of magma ascent (see text).

Stage 1 (40–20 km). Mantle-derived high-MgO magmas ($\text{MgO} > 10\%$), parental to all SRP erupted products, are transported through the lower crust. Being much hotter than ambient crust, such magmas experience undercooling and partially crystallize. Rare high-Fo olivine grains (Fig. 4) and lower-crust derived clinopyroxenes (Fig. 5; Putirka *et al.*, 2003) are possible records of this stage. Assimilation of wall-rock is possible, and may be the source of the gabbroic assimilation period AFC-1 evident for COME flows. However, in conduits armored by pre-Holocene magma throughput, interaction may be limited, as is inferred for NEOT magmas. For the COME, the AFC-1 assimilant must have high Sr (400–500 ppm), high $^{87}\text{Sr}/^{86}\text{Sr}$ (> 0.709) and low SiO_2 (45–60%) (Fig. 8), similar to gabbroic material from the Sierra Nevada (Ratajeski *et al.*, 2001). Even small amounts of assimilation ($< 1\%$) of lower crust gabbroic materials can significantly alter $^{87}\text{Sr}/^{86}\text{Sr}$ ratios, which means that it may be difficult to differentiate whether mafic basalts derive from an enriched or depleted mantle source.

Stage 2 (20–10 km). Mafic magmas from Stage 1 reach a level of neutral or negative buoyancy within the middle crust. Here, magmas pond (or migrate slowly) and differentiate (by crystallization of olivine $\pm$ clinopyroxene $\pm$ plagioclase) to $\leq 10\%$ MgO. With the precipitation of plagioclase, Fe increases, but overall liquid densities may continue to decrease as density increases caused by increased Fe are offset by density decreases caused by increases in H_2O (and CO_2). Further evolution depends

upon the ascent path: (a) at the SRP margins, COME magmas remain neutrally or negatively buoyant within felsic middle crust, and follow path AFC-2 (Fig. 8) until H_2O contents reach 1–2 wt % and MgO contents are $< 8\%$; (b) nearer the SRP axis, NEOT magmas with 10% MgO (and up to 1 wt % H_2O ; Fig. 10) are buoyant within, and rise through, a more mafic middle crust, with minimal assimilation.

Stage 3 (15–0 km). This stage represents a final ascent of magma to the surface. Undoubtedly, such ascent involves at least two sub-stages, a shallow stage where magmas reach volatile saturation and are accelerated rapidly to the surface, preceded by a deeper stage where volatiles are undersaturated, and density contrasts and ascent rates are less. The maximum depth at which eruptions initiate is ~ 15 km (for dry magmas); this estimate derives from the highest P estimate at South Robbers, whose crystals yield the greatest depths overall (among NEOT flows that reveal a continuum of depths). As for the shallowest depths of eruption, if upward acceleration is controlled by water solubility (Moore *et al.*, 1995), then with 1% H_2O , NEOT magmas would be saturated at 117 bars, or a depth of about 10 km, and COME magmas with 1–2% H_2O would reach water saturation at 134–440 bars, or depths of 10–20 km. Undoubtedly, CO_2 may also play a crucial role: saturation of CO_2 will allow for rapid ascent, perhaps even from mid-crustal depths for some magmas—so our estimates of 1–2 km for depths of accelerated upward motion are minima.

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SUPPLEMENTARY DATA

Supplementary data for this paper are available at *Journal of Petrology* online.

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