Origin of Temporal Compositional Trends in Monogenetic Vent Eruptions: Insights from the Crystal Cargo in the Papoose Canyon Sequence, Big Pine Volcanic Field, CA

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Abstract

Many monogenetic vents display systematic temporal-compositional variations over the course of eruption. Previous studies have proposed that these trends may reflect variable degrees of crustal assimilation, or melting and mixing of heterogeneous mantle source(s). Discrimination between these two endmember hypotheses is critical for understanding the plumbing systems of monogenetic volcanoes, which pose a significant volcanic hazard in many areas. In this study, we examine the Papoose Canyon (PC) monogenetic vent in the Big Pine Volcanic Field (BPVF), which had been well characterized for temporal-compositional variations in erupted basalts. We present new major and trace element and Sr-Nd-Pb-O isotopic data from the PC “crystal cargo” (phenocrysts and xenoliths). Comparison of “crystal cargo” and host basalt provides new constraints on the history of magma storage, fractionation, and crustal contamination that are obscured in the bulk basalts due to pre- and syn-eruptive magma mixing processes.

The abundances of phenocrysts and ultramafic xenoliths in the PC sequence decrease up-section. Olivine and clinopyroxene phenocrysts span a wide range of Mg# (77-89). The majority of phenocrysts are more evolved than olivine or clinopyroxene in equilibrium with their host basalts (Mg# = 68-71, equilibrium Fo $\approx$ 85-89). In addition, the ultramafic xenoliths display cumulate textures. Olivine and clinopyroxene from ultramafic xenoliths have Mg# (73-87) similar to the phenocrysts, and lower than typical mantle peridotites. Sr-Nd-Pb isotope compositions of the xenoliths are similar to early PC basalts. Finally, many clinopyroxene phenocrysts and clinopyroxene in xenoliths have trace element abundances in equilibrium with melts that are more enriched than the erupted basalts. These features suggest that the

phenocrysts and xenoliths derive from melt that is more fractionated and enriched than erupted PC basalts. Pressure constraints suggest phenocrysts and ultramafic xenoliths crystallized at ~5-7 kbar, corresponding to mid-crust depths. Correlations between HFSE depletion and Sr-Nd-Pb isotopic compositions, high $\delta^{18}\text{O}$ values in olivines, and radiogenic Os isotopic compositions in whole rocks also suggest incorporation of a crustally contaminated component.

We propose that the phenocrysts and ultramafic xenoliths derive from melts that ponded and fractionated and assimilated continental crust, possibly in mid-crustal sills. These melts were drained and mixed with more primitive melts as the eruption began, and the temporal-compositional trends and decreasing crystal phase abundances reflect gradual deflation and exhaustion of these sills as the eruption progressed. The isotopic variations in the PC sequence span much of the compositional range observed in the BPVF. Evidence for variable crustal contamination of PC basalts suggests that much of the isotopic variation observed in the BPVF may also reflect crustal contamination rather than mantle source heterogeneity as previously proposed. In addition, evidence of pre-eruptive magma ponding and fractionation, if applicable to other monogenetic vents, may have significant implications for monitoring and hazard assessment of monogenetic volcano fields.

1. Introduction

Monogenetic vents are small volcanoes that often occur in clusters and are active for a single eruption or short-lived eruptive cycle (months to decades). Monogenetic vent fields represent a significant volcanic hazard in many regions of the world (e.g., Auckland, Mexico City; Bebbington & Cronin, 2011; Siebe & Macias, 2006). An improved understanding of the mechanisms of melt generation and in particular melt storage within monogenetic vent fields will allow better assessment of volcanic hazards associated with monogenetic vents. If magmas erupted from particular vent fields are typically stored within crustal sills prior to eruption (e.g., Erlund et al., 2010), future eruptions may be preceded by sill emplacement or inflation, providing a possible early warning for impending eruption. However, if melts directly ascend from the mantle source to surface with little or no crustal storage prior to eruption (e.g., Blondes et al., 2008), prediction of such eruptions may be difficult.

Many monogenetic vents display significant chemical and isotopic variations during a single eruptive cycle, which can mimic compositional variations observed in entire volcanic fields or regions over longer timescales (e.g., Garcia et al., 2000; Blondes et al., 2008; Erlund et al., 2010). These geochemical variations may provide important constraints on the processes of melt generation and transport. In some cases these short-term variations are proposed to reflect melt generation and extraction from heterogeneous mantle sources, which would suggest rapid magma ascent from source to surface with limited shallow storage (e.g., Blondes et al., 2008; Rasoazanamparany et al., 2015). In other cases, these temporal variations appear related to temporally varying degrees of crustal contamination, more consistent with protracted pre-eruptive magma storage (e.g., McBirney et al., 1987; Erlund et al., 2010).

In addition, distinguishing between mantle and crustal signatures in continental basalts has been a long-standing and vexing problem. Geochemical variations in basalts from the Basin and Range are widely used to constrain the composition and evolution of their mantle sources (e.g., Beard & Glazner, 1995; Gazel et al., 2012). However, basalts represent only indirect probes of mantle composition. Processes such as melt generation, migration, and melt-crust interaction all can affect the compositions of erupted lavas. Distinguishing between mantle versus crustal signatures is particularly difficult in continental settings, because the signatures of crustal contamination may qualitatively resemble the chemical and isotopic signatures of melt generation from metasomatized continental lithospheric mantle (e.g., Hildreth et al., 1991; Glazner, 1991). Therefore, understanding the history of magma ponding and evolution of continental basalts may allow better assessment of potential crustal contamination and more accurate interpretation of mantle source evolution.

In this study, we examine the crystal cargo of the Papoose Canyon (PC) monogenetic vent sequence from the Big Pine Volcanic Field (BPVF), California, a well-documented monogenetic vent with clear temporal-compositional trends (Blondes et al., 2008). Major and trace element and isotope variations in olivine and clinopyroxene phenocrysts and ultramafic xenoliths provide new constraints on the pre-eruptive magma storage and mixing history at PC vent, and allow us to test crust versus mantle origins for the observed temporal-compositional trends.

2. Background

The BPVF is a monogenetic volcano field situated in Owens Valley, bounded to the east and west by the Inyo-White Mountains and the Sierra Nevada. With the exception of one silicic

vent, lavas erupted at the BPVF are primarily mafic in composition, ranging from basanites to alkalic and sub-alkalic basalts (Bierman et al., 1991). Most vents erupted between 0.1 and 0.8 Ma, with the oldest documented eruption at ~1.2 Ma (Gillespie, 1984; Bierman, 1991; Blondes et al., 2008). The origin of BPVF volcanism is commonly attributed to Basin-and-Range regional extension and associated lithosphere thinning and/or delamination (e.g., Beard & Glazner, 1995; Lee et al., 2001; Wang et al., 2002).

Basalts erupted at the BPVF span a range of major and trace element and isotope compositions. For example, whole rock SiO₂ abundances vary from 44.1 to 53.0 wt.%, olivine-fractionation-corrected Sr concentrations vary from ~900 to 2300 ppm and ⁸⁷Sr/⁸⁶Sr ranges from 0.70534 to 0.70648. (Ormerod et al., 1988 & 1991; Rogers et al., 1995; Beard & Glazner, 1995; Mordick & Glazner, 2006; Blondes et al., 2008). Ormerod et al. (1991) suggested that the variations in trace element abundances reflect different degrees of partial melting of a relatively homogeneous mantle source. However, this cannot account for observed isotopic variations. Beard and Glazner (1995) proposed that the observed compositional variations reflect melt generation from heterogeneous metasomatized continental lithospheric mantle. More recently, Gazel et al. (2012) suggested that melt generation depths of BPVF basalts have shifted from asthenospheric mantle to the lithosphere-asthenosphere boundary over time, resulting in increased incorporation of lithospheric mantle components in younger BPVF basalts.

Several individual eruption sequences in the BPVF display compositional variations (Blondes et al., 2008). This study focuses on the PC sequence (760.8 ± 22.8 ka) because this sequence is well exposed and displays significant temporal-compositional variations, and previous geochemical characterization of PC whole rocks provides important context for the

new crystal cargo data presented here (Blondes et al., 2008). The PC sequence is cut and exposed by the Papoose Canyon valley, which allows stratigraphically-controlled sampling of the entire sequence. Over the course of the PC eruption, the concentrations of highly incompatible trace elements decreased by up to a factor of ~2 (e.g., La concentrations decreased from 66 to 33 ppm). This decrease in trace element concentrations was accompanied by decreasing $^{87}\text{Sr}/^{86}\text{Sr}$ (0.7063 – 0.7055) and increasing ϵ_{Nd} (-3.5 – -1.1) (Blondes et al., 2008). The PC sequence is overlain by two younger vents (Quarry and Jalopy) at the western end of the valley. Basalts from the younger vents are chemically distinct from the older PC basalts (e.g., PC basalts have systematically higher Nb concentrations; Blondes et al., 2008; Gazel et al., 2012).

3. Samples and methods

3.1 Field observations and samples

The PC vent is located at the easternmost part of the BPVF (Blondes et al., 2008). The vent sits on top of dolomitic country rock. The PC sequence contains both massive units and loose scoria. The massive units have sparsely vesicular interiors and more vesicular margins. The interiors of these units appear mostly fresh, with scattered secondary zeolite/calcite minerals filling vesicles. The scoria and massive unit margins are generally more altered and oxidized. No soil horizons are observed between layers within the sequence, consistent with a monogenetic origin for the sequence. Overall, PC basalts have low phenocryst abundances, with phenocryst mineralogy dominated by olivine and clinopyroxene (cpx). Abundant pyroxene-rich ultramafic xenoliths (70-90 % modal cpx and 10-30 % modal olivine) and some carbonate crustal xenoliths

are present in the early eruption sequence. Both ultramafic and carbonate xenoliths are sub-rounded to rounded.

During two field trips in 2012 and 2014, we collected stratigraphically-controlled samples from the interiors of massive flows. More oxidized flow margins and scoria were avoided due to potential post-eruptive oxidization effects on phenocryst compositions (Blondes et al., 2012). Pyroxene-rich ultramafic xenoliths and carbonate crustal xenoliths contained in the PC flows were also collected with host basalts when possible. Sample location and approximate stratigraphic position are reported in Appendix Table A1. Relative stratigraphic positions were calculated assuming an overall bedding dip similar to the angle of repose for most cinder cones ($\sim 30^\circ$), consistent with measured bedding angles. See Appendix Fig. A1 for details.

3.2 Methods

3.2.1 Point counting

Olivine phenocryst abundance was determined by point counting of 6 stratigraphically controlled samples spanning the eruption sequence. Each sample was cut into flat slabs of $\sim 5 \times 8$ cm and polished. Point counting was done under an optical microscope using a transparent 2×2 mm grid sheet overlay. More than 500 points were counted for each sample. Olivine phenocrysts are classified into 3 populations based on size: small phenocrysts (0.5-1 mm), large phenocrysts (1-3.5 mm) and megacrysts (> 3.5 mm).

3.2.2 Major element, trace element, and isotope composition

Whole rock major and trace elements were analyzed using standard XRF and ICP-MS techniques at Washington State University. Olivine and cpx major element abundances were measured by EMPA at the University of Texas at Austin (UT Austin). Clinopyroxene trace

element compositions were measured by LA-ICP-MS (UT Austin). Strontium, Nd and Pb isotopes were measured by TIMS. Osmium isotopes were measured by N-TIMS, and Os concentrations were measured by isotope dilution. Oxygen isotope ratios were measured by laser fluorination. All isotopic analyses were performed at UT Austin. Further discussion of analytical methods is provided in the Appendix.

4. Results

4.1 Point counting

Phenocryst modal abundances (proportion of the whole rock) are presented in Appendix Table A2. PC basalts contain 2-7% olivine and < 0.5% cpx phenocrysts, as well as rare olivine and cpx megacrysts (<0.1%). The modal abundance of large olivine phenocrysts (1-3.5 mm) decreases gradually from 4.1% in the early-sequence sample to 0.6% in the late-sequence sample (Fig. 1). In contrast, the modal abundance of small olivine phenocrysts (0.5-1 mm, 0.7-3.9% modal abundance) does not change systematically over time. Megacryst abundances in most samples are too low to be determined by point counting in this study.

4.2 Major and trace elements

4.2.1 Whole-rock major and trace elements

The major and trace element compositions of PC basalts are presented in Appendix Table A3. PC basalts have relatively high and uniform Mg# (69-70). SiO₂ (45.7-47.4 wt.%) and Al₂O₃ (15.8-16.3 wt.%) increase over the course of eruption, whereas CaO (10.2-9.6 wt.%), TiO₂ (2.1-1.9 wt.%), K₂O (2.0-1.4 wt.%) and P₂O₅ (1.1-0.7 wt.%) decrease. FeO_{total} (8.8-9.1 wt.%), Na₂O

(3.6-3.9 wt.%), and MgO (9.3-9.8 wt.%) do not show systematic temporal trends. Our data are generally consistent with previous studies (Blondes et al., 2008; Gazel et al., 2012; Fig. A2).

PC basalts show Ti and HFSE depletions, slight positive Sr anomalies and no Eu anomalies. Incompatible trace element concentrations (e.g., Sr, Ba, and LREE) decrease significantly over the course of eruption. LREE and large ion lithosphere elements (LILE, e.g., Sr, Ba) show greater variation than high field strength elements (HFSE, e.g., Nb, Ta) and HREE, with the former varying by about a factor of 2 while HFSE vary by ~20-30% (e.g., Fig. 2b). The concentrations of moderately incompatible trace elements (e.g., HREE) remain nearly constant, whereas those of compatible elements (Cr = 185-384 ppm; Ni = 151-226 ppm) increase over the course of eruption. The concentrations of highly incompatible trace elements show linear correlations with those of moderately incompatible trace elements (e.g., La vs. Sm; Fig. 3a).

4.2.2 Mineral major and trace elements

Major element compositions of 215 olivine phenocrysts from 8 samples spanning the entire stratigraphy are reported in Appendix Table A4. Olivine phenocrysts have Mg#s ranging from 77 to 89, with an average Mg# $\approx$ 84 (Fig. A3a). Several individual samples contain olivines spanning a wide range of composition, which is nearly as great as the range defined by PC olivine phenocrysts as a whole (Fig. 4a). 37 cpx phenocrysts from 4 samples were also analyzed (Table A5). These cpx phenocrysts have Mg#s ranging from 81 to 87, with an average of 84 (Fig. A3a). Again, individual samples contain cpx spanning a range in composition (Fig. 4b).

Olivine and cpx from PC ultramafic xenoliths have major element compositions similar to PC phenocrysts (Table A4 and A5). Olivines from PC ultramafic xenoliths have Mg#s ranging from 73-87, whereas cpx from these xenoliths have Mg#s ranging from 73-88 (Fig. A3b). Olivine

and cpx from individual xenoliths have relatively uniform Mg#s, and the Mg# of olivine and cpx are also similar to each other in a given xenolith. The range of Mg#s observed in the ultramafic xenoliths extends to lower values than observed in PC phenocrysts. However, the Mg# distribution and average are similar (Fig. A3).

Clinopyroxene trace element abundances in PC phenocrysts and ultramafic xenoliths are presented in Table A6. Phenocryst cpx and xenolith cpx have similar trace element patterns, (Fig. 2a). Cpx also have greater variations in highly incompatible trace element concentrations (e.g., [La] = 1.1-17.3 ppm) than in moderately incompatible trace element concentrations (e.g., [Lu] = 0.1-0.4 ppm).

4.3 Radiogenic and oxygen isotopes

Strontium, Nd, Pb, Os and O isotope compositions are reported in Table 1. $^{87}\text{Sr}/^{86}\text{Sr}$ decreases from 0.7061 to 0.7058, and ϵ_{Nd} increases from -2.7 to -1.4 over the course of the eruption. Pb isotopic compositions also evolve toward less enriched compositions over time (e.g., $^{206}\text{Pb}/^{204}\text{Pb}$ varies from 19.11 to 18.99). The Sr, Nd, and Pb isotopic ratios are linearly correlated with the inverse of trace element concentrations (e.g., $^{87}\text{Sr}/^{86}\text{Sr}$ vs. $1/[\text{Sr}]$; $^{143}\text{Nd}/^{144}\text{Nd}$ vs. $1/[\text{Nd}]$; Fig. 3b).

Five middle-sequence samples were measured for $^{187}\text{Os}/^{188}\text{Os}$ to fill the gap of the Os-isotope data set previously reported by Blondes et al. (2008). The Os isotopic compositions of PC basalts ($^{187}\text{Os}/^{188}\text{Os} = 0.183\text{-}0.287$) are more radiogenic than depleted MORB mantle or primitive upper mantle ($^{187}\text{Os}/^{188}\text{Os} < 0.130$; Shirey & Walker, 1998). However, Os isotopic ratios do not correlate with Sr-Nd-Pb isotopes or relative stratigraphic position.

Five pyroxene-rich ultramafic xenoliths have Sr, Nd, and Pb isotopic compositions falling on the trends defined by PC basalts (e.g., these xenoliths have $^{87}\text{Sr}/^{86}\text{Sr} = 0.7059 - 0.7063$, $\varepsilon_{\text{Nd}} = -2.5 - -3.3$, $^{206}\text{Pb}/^{204}\text{Pb} = 19.08 - 19.14$). However, all of the xenoliths are isotopically more enriched than their host basalts, with higher $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{206}\text{Pb}/^{204}\text{Pb}$ (Fig. 5).

Olivine phenocryst separates from PC basalts have $\delta^{18}\text{O}$ values ranging from +5.4‰ to +5.8‰, slightly higher than the $\delta^{18}\text{O}$ value of olivine from fresh NMORB and mantle peridotite (+5.2±0.2‰; Eiler, 2001; Matthey et al., 1994). No temporal trend is observed. The $\delta^{18}\text{O}$ values of olivine mineral separates from PC ultramafic xenoliths range from +5.5‰ to +5.7‰.

5. Discussion

Basalts from the PC monogenetic vent display significant temporal variations in major and trace element and isotope composition. Given the short timescales of single eruptive cycles, melt generation conditions and source characteristics are expected to be relatively invariant. The compositional variations observed in single eruption sequences therefore may provide valuable insight into the processes of magma generation and migration that are not possible from studies conducted over broader areas and longer timescales (e.g., Reiners, 2002).

Several recent studies indicate that erupted basalts are often hybrid mixtures of multiple discrete magma batches (e.g., Andrews et al., 2008; Moune et al., 2012). Understanding the history and location (crust vs. mantle) of magma storage and mixing can therefore help constrain the origins of the compositional variations in monogenetic vents. This history is often difficult to constrain from bulk basalt because of the mixing and homogenization processes described above. Instead, crystal cargos often contain more information about the magma

storage and mixing history than bulk basalt (e.g., Rankenburg et al., 2004; Andrews et al., 2008). In the following discussion, the geochemical variations in phenocrysts and xenoliths are combined with compositional variations observed in erupted basalts, to constrain the pre-eruptive storage and mixing history of PC magmas.

5.1 Temporal compositional variations reflect two-component mixing

Over the course of the PC eruption, incompatible trace element concentrations decreased, whereas Sr, Nd and Pb isotopes evolved to less enriched compositions. Several lines of evidence suggest these compositional variations primarily reflect mixing of two distinct melt components. For example, PC basalts show linear correlations between highly and moderately incompatible trace elements (e.g., La vs. Sm, Nb vs. Zr; Fig. 3a). Although variations in degree of partial melting may generate variations in trace element concentrations (e.g., Ormerod et al., 1991), curved correlations are expected for elements with different partition coefficients (e.g., Reiners, 2002). In contrast, two-component mixing will generate linear correlations between different elements regardless of their compatibility (e.g., Reiners, 2002). Partial melting curves may appear linear within a small fragment. However, Sr, Nd and Pb isotopic ratios are also linearly correlated with inverse trace element concentration (e.g., $^{87}\text{Sr}/^{86}\text{Sr}$ vs. $1/[\text{Sr}]$; Fig. 3b), which is again consistent with binary mixing (e.g., Glazner et al., 1991; Reiners, 2002). Overall, the compositional variations suggest mixing of one melt characterized by high incompatible trace element concentrations and enriched isotopic signatures, and another melt that is chemically and isotopically less enriched. PC basalts show greater compositional variation in LILE and LREE (~2x) than in HFSE and HREE (~1-1.2x). LREE/HFSE ratios in PC basalts also correlate with Sr-Nd-Pb isotope compositions. This suggests that the enriched melt is preferentially enriched in LILE

and LREE relative to HFSE and HREE. The observed stratigraphic variations indicate that the proportion of the enriched component contained in the erupted hybrid melt decreased over time.

Several different models have been proposed for these two melt components, which have different implications for melt generation and transportation. For example, partial melting of a lithologically heterogeneous mantle source (e.g., pyroxenite and peridotite) may produce melts with distinct chemical and isotopic compositions (e.g., Reiners, 2002; Blondes et al., 2008). Alternatively, the compositional trends could reflect mixing of lithosphere- and asthenosphere-derived melts, possibly as asthenosphere-derived melts ponded at the asthenosphere/lithosphere boundary (Blondes et al., 2008). Finally, if PC magmas ponded within the continental crust prior to eruption, the compositional trends could reflect variable degrees of crustal contamination. Therefore, when and where melt mixing occurred has important implications for the origin of the observed compositional trends. In the following section, we evaluate the record of magma storage contained within PC phenocrysts and ultramafic xenoliths.

5.2 Constraints on pre-eruption magma storage and fractionation

Previous studies have proposed rapid magma ascent from source to surface with limited shallow storage for many monogenetic volcanoes (e.g., Mattsson, 2012; McGee et al., 2013). PC whole rocks have relatively high and uniform MgO contents (8.7-10.5 wt.%) and Mg# (69-70), indicating they have undergone limited fractional crystallization (Blondes et al., 2008). In addition, early PC sequence basalts contain abundant ultramafic xenoliths that were previously interpreted as mantle xenoliths (Blondes et al., 2008). These features would seem to indicate

PC basalts have experienced limited crustal ponding and fractionation, which would suggest the melt mixing described above occurred within the mantle (Reiner, 2002; Blondes et al., 2008). However, as discussed below, the compositions of PC phenocrysts and xenoliths suggest a more complicated history of magma storage, fractionation and mixing than suggested by whole rock compositional variations.

Although PC basalts span a narrow range of Mg#, olivine phenocrysts in PC basalts span a wide range of composition (Fo = 77-89, with an average of 84). Large variations are observed among olivine grains from individual samples (Fig. 4a). In addition, the majority of olivine phenocrysts appear to be more evolved than olivine in equilibrium with their host basalts. Assuming a $K_D^{\text{ol-melt}}(\text{Fe}^{2+}/\text{Mg}) = 0.3 \pm 0.03$ (Roeder & Emslie, 1970), olivine phenocrysts in equilibrium with PC basalts should have Fo $\approx$ 87-89, but the majority of phenocrysts have significantly lower Fo (Fig. 4a). Clinopyroxene phenocrysts span a slightly narrower range of Mg# than olivines (Mg# = 81-87, with average of 84), but for the most part also have Mg#s lower than in equilibrium with their host basalts ($K_D^{\text{cpx-melt}}(\text{Fe}^{2+}/\text{Mg}) = 0.28 \pm 0.08$; Putirka, 2008; Fig. 4b).

One possible explanation for the apparent disequilibrium between phenocrysts and basalt whole rocks is that the whole rocks contain cumulate olivine and cpx, which may result in overestimation of melt Mg# from whole rock compositions. PC basalts contain up to ~4% large olivine (1-3.5 mm), ~1-4% small olivine (0.5-1 mm), and trace amounts of cpx phenocrysts (<0.5 %). If we assume the majority of large phenocrysts are cumulate and take this effect into account, the melt composition estimated from cumulate subtraction is still not in equilibrium with the phenocryst phases (Fig. 4). Alternatively, the melt $\text{Fe}^{2+}/\text{Fe}_T$ ratio may be higher than

the value we chose (0.85). This $\text{Fe}^{2+}/\text{Fe}_T$ ratio corresponds to melt oxygen fugacity of $\sim\text{FMQ}+1$, which was estimated by Gazel et al. (2012) for BPVF basalts based on V partitioning between olivine and melt inclusions. High oxygen fugacity of BPVF basalts is also suggested by high sulfur contents of the least degassed melt inclusions (Gazel et al., 2012; Jugo et al., 2010). In addition, few melts from the Great Basin have oxygen fugacities lower than FMQ (e.g., Christiansen & McCurry, 2008). A $\text{Fe}^{2+}/\text{Fe}_T$ ratio of 0.9, which would correspond to $\sim\text{FMQ}$ (Botcharnikov et al., 2005), would shift the calculated melt Mg# by only ~ 1 unit, too little to bring the melt and phenocrysts into equilibrium. Therefore, the phenocrysts appear to derive primarily from melts that have lower Mg# and thus experienced greater fractionation than their host basalts.

Ultramafic xenoliths in the early sequence were previously described as likely derived from the mantle (Blondes et al., 2008), which would indicate rapid magma ascent from mantle depths. However, the xenoliths we investigated are primarily wehrlites and clinopyroxenites, and no true 4-phase peridotites (olivine + orthopyroxene + clinopyroxene + spinel/garnet) were identified. Most mantle-derived xenolith suites contain peridotite even if pyroxenite is also present. Therefore, although pyroxenite xenoliths could derive from the mantle, the lack of 4-phase peridotites makes a mantle origin for the pyroxenite xenoliths less likely. In addition, the PC pyroxenite xenoliths have cumulate textural features, such as interstitial microcrystalline material likely derived from interstitial melt and rounded grain boundaries, but lack mantle metamorphic features such as olivine triple junctions (Fig. A4; e.g., Pike & Schwarzman, 1977). Moreover, xenolith olivine and cpx Mg#s range from 73 to 87, extending well below the range of Mg# (88-90) reported from BPVF mantle peridotites from other localities (Beard & Glazner, 1995; Lee et al., 2001 & 2002). Instead, the range of Mg#s are similar to the range defined by

PC phenocrysts. As discussed below, the mineral compositions of these xenoliths are consistent with the expected fractional crystallization products of PC magmas at mid-crust depths. These xenoliths therefore most likely formed as cumulates from the same melts that generated the phenocrysts.

Overall, both the phenocrysts and ultramafic xenoliths appear to derive from melts that on average are more evolved and have lower Mg# than PC basalts. Over the course of the eruption, abundances of both large olivine phenocrysts and ultramafic xenoliths decrease. This trend mirrors the decrease in incompatible trace element concentrations in PC basalts. In contrast, smaller olivines display no systematic variation in modal abundances, and may represent true phenocrysts formed during magma ascent immediately prior to eruption. As discussed in section 5.1, the compositional variations in PC basalts likely reflect mixing of two melt components. Given the discussion above, the mixing between a primitive (phenocryst-poor) melt and a more evolved (phenocryst-bearing) melt may account for these compositional trends. In this case, decreasing abundances of large phenocrysts and ultramafic xenoliths indicate that the proportion of the evolved melt component in the erupted lavas decreased over time. In the following discussion, we examine the likely depths of magma storage and fractional crystallization.

Previous studies have estimated the phenocrysts crystallization depths of younger (< 500 ka) BPVF basalts using phenocryst-melt thermobarometers and the volatile contents in phenocryst-hosted melt inclusions. Mordick & Glazner (2006) estimated deep cpx crystallization using the cpx-melt thermobarometer of Putirka et al. (1996). They examined four samples from younger BPVF vents, and calculated cpx phenocryst growth at $\sim 1.4 \pm 0.3$ GPa and $\sim 1250 \pm 100$ °C

(Mordick & Glazner, 2006). In contrast, Gazel et al. (2012) suggested shallower fractionation (~5 kbar) of olivine phenocrysts, based on volatile contents (CO_2 and H_2O) of melt inclusions trapped in the phenocrysts. These pressure constraints, if both accurate, would seem to indicate that cpx and olivine phenocrysts in BPVF basalts formed at very different depths. However, coexisting olivine and cpx in PC ultramafic xenoliths have similar Mg#, and olivine and cpx phenocrysts in PC basalts span similar ranges of Mg# and have the same average Mg#. These features indicate concurrent crystallization of olivine and cpx. Therefore, one of the pressure estimates described above is likely in error.

The use of the cpx-melt thermobarometer to calculate crystallization pressures presupposes that phenocrysts and host melts are in chemical equilibrium. However, as is true for the majority of PC phenocrysts, most of the samples investigated by Mordick & Glazner (2006) contain cpx with Mg#s that are out of equilibrium with their host basalts, even after whole rock compositions are corrected for phenocryst accumulation (Fig. 4). Furthermore, the crystallization pressures estimated by Mordick & Glazner (2006) positively correlate with the degree of disequilibrium between cpx and melt pairs. This suggests that there is a systematic error in pressure estimates introduced by the phase disequilibrium. The only cpx-melt pair that is close to equilibrium (sample B-1) results in the lowest estimated pressure (~1 GPa). The Putirka et al. (1996) thermobarometer is also calibrated for anhydrous melts, whereas BPVF magmas have relatively high water contents (1.5-3.0 wt.%; Gazel et al., 2012). Because water can lower the liquidus temperature of melts, failing to take into consideration the effects of water may lead to overestimation of crystallization temperatures. This in turn also results in overestimation of pressure, because temperature is an input in the cpx-melt barometer.

Therefore, the estimated cpx crystallization pressures of Mordick & Glazner (2006) are most likely too high.

The dominance of olivine phenocrysts in PC basalts also suggests relatively low fractionation pressures. Many experimental studies show that olivine fractionation in alkalic basalts primarily occurs at pressures <1 GPa (e.g., Green & Ringwood, 1967; Nekvasil et al., 2004). As pressure increases, the stability field of clinopyroxene expands at the expense of olivine and plagioclase (e.g., O'Hara, 1968). We estimated the pressure of PC magma fractionation by comparing phenocryst compositions with predictions of the MELTS thermodynamic model over a range of pressures, starting compositions and oxygen fugacities (Ghiorso & Sack, 1995). The details of the MELTS modeling are described in the Appendix. One representative set of results is shown in Fig. 6, and additional results are presented in Appendix Fig. A5. All models show that the compositional range where both olivine and clinopyroxene crystallize shrinks and shifts to higher Mg# with increasing pressure. Moreover, olivine with Mg# lower than ~80 does not form at pressures greater than 1 GPa. Phenocryst and xenolith olivine and cpx span a wide range of Mg# (73-89, Fig. A3). The presence of low Mg# olivine (<80) suggests magma fractionation at pressures <1 GPa. Furthermore, assuming isobaric rather than polybaric fractionation, the MELTS modeling results suggest the full range of observed olivine and cpx Mg# can be formed only at pressures of ~5-7 kbar (Fig. 6). These estimates are consistent with the pressure estimates derived from BPVF melt inclusions (Gazel et al., 2012), but lower than the pressure estimates based on the cpx-melt thermobarometer (Mordick & Glazner, 2006).

In summary, although there are significant uncertainties in the estimated fractionation pressures, the results of this study and of Gazel et al. (2012) are mutually consistent and suggest phenocrysts grew at pressures <1 GPa and most likely at ~5-7 kbar. The local crust thickness is ~36 km, which corresponds to a pressure of ~1 GPa (assuming average crustal density of 2,750 kg/m³) (Wang et al., 2002). The above pressure estimates therefore indicate that magmas ponded and fractionated within the middle to lower crust prior to eruption. Gazel et al. (2012) suggested that magmas may have stalled and ponded near the brittle-ductile transition (~20 km) within the crust, due to this rheological contrast. As discussed below, these magmas stored within the continental crust appear to have assimilated crustal material prior to eruption.

5.3 Crustal assimilation

Several lines of evidence indicate that the melt from which the phenocrysts and xenoliths derived most likely represents the “enriched” mixing endmember required by isotopic and trace element variations in PC basalts (see Section 5.1). The compositions of the melts from which PC cpx grew were calculated using measured cpx trace element abundances and estimated cpx-melt partition coefficients (see Appendix for more details). Most of the PC phenocryst and xenolith cpx appear to derive from parental melts that have higher incompatible trace element concentrations than any PC basalts (Fig. 2b). Moreover, the large variations in cpx trace element concentrations (up to a factor of 15) cannot be explained by fractional crystallization alone. Clinopyroxene HFSE and HREE concentrations are well correlated with cpx Mg#. These correlations are generally consistent with compositional variations derived from fractional crystallization in conjunction with increasing K_D values with

decreasing Mg# (e.g., Lu, Fig. 7a). In contrast, LILE and LREE concentrations do not correlate with Mg#, and pyroxenes spanning a narrow range in Mg# show much larger variations in the concentrations of these elements than can be explained by the effects of fractional crystallization (Fig. 7b). The LILE and LREE variations observed in these pyroxenes require pyroxene crystallization from melts with highly variable LILE and LREE concentrations .

In addition, PC ultramafic xenoliths have Sr, Nd, and Pb isotopic compositions that fall along the same array defined by PC basalts but extend to (and beyond in the case of Pb) the most enriched compositions of the basalts. In contrast, mantle peridotites from other BPVF vents show considerable isotopic scatter and for the most part have higher ϵ_{Nd} at a given $^{87}\text{Sr}/^{86}\text{Sr}$ than PC basalts and xenoliths (Fig. 5). Overall, PC phenocrysts and xenoliths appear to derive from melts that are isotopically more enriched than their host basalts, and characterized by LILE and LREE enrichment relative to HFSE and HREE. Therefore, the parental melt of PC phenocrysts and xenoliths resembles the “enriched” mixing component discussed in Section 5.1.

In principle, it is possible that the primitive and evolved melts derived from distinct mantle sources, and the chemically and isotopically more enriched compositions of the evolved melts are a mantle-derived primary feature. For example, the “enriched” melts may derive from partial melting of enriched mantle domains such as pyroxenite veins (Reiners, 2002; Blondes et al., 2008). Mantle pyroxenites span a wider range and extend to lower Mg# than mantle peridotites. Therefore, pyroxenite-derived melts could be both more enriched and have lower Mg# than peridotite-derived melts. However, this is an unlikely explanation for the observed range of phenocryst compositions. Correlations between Mg# and compatible element concentrations (e.g., Ni, Sc) in both PC phenocrysts and basalts are consistent with fractional

crystallization trends. Furthermore, as discussed above, olivine phenocrysts spanning a wide range in Fo content is consistent with phenocrysts formation at <1 GPa (within the continental crust). Finally, as discussed below, several lines of chemical and isotopic evidence suggest that crustal contamination played an important role in modifying the composition of the evolved melt that is the source of PC phenocrysts and ultramafic xenoliths.

Average continental crust has radiogenic $^{87}\text{Sr}/^{86}\text{Sr}$ and unradiogenic $^{143}\text{Nd}/^{144}\text{Nd}$ compared to both convecting upper mantle and primitive mantle. Continental crust is also characterized by significant HFSE depletion. For example, average continental crust has La/Nb of ~ 2.5 , N-MORB has La/Nb of ~ 1.0 , and ocean island basalts have La/Nb ≤ 1.2 (Rudnick & Gao, 2003; McDonough & Sun, 1995). Assimilation of continental crust therefore is expected to generate correlations between isotopic compositions and LILE/HFSE ratios, as has been observed in several previous studies (e.g., Davies & Macdonald, 1987; Rowe et al., 2011). PC basalts are variably depleted in HFSE, with La/Nb ranging from ~ 2 to 3.6. In addition, La/Nb ratios strongly correlate with $^{87}\text{Sr}/^{86}\text{Sr}$, $^{206}\text{Pb}/^{204}\text{Pb}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ (e.g., Fig. 8). These features are consistent with crustal contamination, although melt generation from subduction-modified mantle (especially if rutile is present as a trace phase) could also potentially produce isotopically enriched, HFSE-depleted melts.

The oxygen isotope compositions of PC phenocrysts and cumulate xenoliths are also consistent with crustal contamination. Olivine phenocrysts have $\delta^{18}\text{O}$ values ranging from +5.4 to +5.8 ‰. Olivines from PC cumulate xenoliths display a similar range of oxygen isotope composition ($\delta^{18}\text{O} = +5.5$ to +5.7 ‰). Although constraints on the oxygen isotope composition of the mantle beneath the BPVF are not available, Sierran mantle xenoliths from Big Creek

(~100 km west to the BPVF) have oxygen isotope compositions within the range of normal mantle (olivine $\delta^{18}\text{O} = +5.0$ to $+5.4$ ‰; Chin et al., 2014). PC olivines have $\delta^{18}\text{O}$ values that are slightly but consistently higher than normal mantle olivine. Continental crust typically has higher than mantle $\delta^{18}\text{O}$ values ($\delta^{18}\text{O} = -2$ to $+26$ ‰ with an average of ~ 9 ‰; e.g., Simon & Lecuyer, 2005). If the assimilated crust is assumed to have average $\delta^{18}\text{O}$ value of 9‰, the observed ~ 0.4 ‰ offset in $\delta^{18}\text{O}$ values relative to normal mantle peridotite can be generated by less than 20% crustal assimilation in the magma source of the phenocrysts and xenoliths.

Positive correlations have been observed between $\delta^{18}\text{O}$ and $^{87}\text{Sr}/^{86}\text{Sr}$ in many continental basalt suites (e.g., Puricutin, McBirney et al., 1987; Aeolian arc, Peccerillo et al., 2004). These correlations are generally attributed to crustal contamination. However, no correlation between olivine phenocrysts $\delta^{18}\text{O}$ and whole rock $^{87}\text{Sr}/^{86}\text{Sr}$ is observed in the PC basalts. This may reflect the fact that olivine phenocrysts derive primarily from the evolved (and contaminated) melt, whereas the whole rocks represent mixtures of evolved and primitive, uncontaminated melt. As a result, the Sr isotope composition of the melt mixture is dependent on the ratio of contaminated and primitive melt in the hybrid melt, but $\delta^{18}\text{O}$ values of the olivine phenocrysts, which formed from the contaminated melt prior to mixing, are not.

Finally, PC basalts also have radiogenic Os isotopic compositions ($^{187}\text{Os}/^{188}\text{Os} = 0.183\text{--}0.310$) compared to normal mantle peridotite ($^{187}\text{Os}/^{188}\text{Os} < 0.130$; e.g., Shirey & Walker, 1998; Lassiter et al., 2014) and most uncontaminated mantle-derived magmas ($^{187}\text{Os}/^{188}\text{Os} < 0.150$; e.g., Hauri & Hart, 1993; Lassiter & Hauri, 1998). Os isotope ratios do not systematically vary with stratigraphic position or correlate with other chemical and isotopic systems. Continental crust has high Re/Os ratios and therefore variable but high $^{187}\text{Os}/^{188}\text{Os}$ (e.g. Esser & Turekian,

1993). PC basalts have low Os concentrations (≤ 55 ppt). Previous studies have shown that Os-isotopes in evolved basalts with less than ~ 50 ppt Os from both continental and oceanic settings are extremely susceptible to crustal contamination, and decoupling between Os isotopes and other isotopic systems is common. For example, basalts from the Mexican Volcanic Belt extend to high $^{187}\text{Os}/^{188}\text{Os}$ ratios that are not correlated with Sr or Nd isotopes (Lassiter & Luhr, 2001). The decoupling between Os isotopes and other isotopic systems commonly observed may reflect the concentration of Os in trace phases such as sulfide inclusions within olivine phenocrysts.

Taken collectively, the features described above are all consistent with crustal contamination. Although we cannot exclude the possibility that the evolved melt derives from a distinct subduction-modified mantle source, it appears more likely that this melt fractionated and assimilated continental crust during magma storage. The compositional variations in PC basalts therefore likely reflect varying degrees of crustal contamination, as is suggested for many other monogenetic vents (e.g., McBirney et al., 1987; Schaaf et al., 2005; Erlund et al., 2010; Rowe et al., 2011).

5.4 Sills model

The above discussion indicates that the PC basalts contain a melt component that has experienced significant crustal storage and fractional crystallization. Therefore, the observed compositional variations in PC basalts are most readily explained by mixing of an evolved and crustally-contaminated melt with a more primitive melt. Signatures of the enriched and evolved component decrease over the course of the PC eruption, as suggested by decreasing incompatible trace element concentrations and the decreasing abundance of large olivine

phenocrysts and cumulate xenoliths. This temporal trend suggests the mixing of the two distinct melts is concurrent with the eruption. Here we propose a two-stage model, where an initial melt pulse stalls and fractionates in the crust, and the eruption is triggered by arrival of a later melt pulse which then mixes with the earlier component (Fig. 9).

In this model, an initial melt pulse stalled and formed one or more sills within the crust. According to pressure constraints, these sills most likely formed at ~20 km, near the brittle-ductile transition within the crust (Gazel et al., 2012). Magmas within the crustal sills fractionated and assimilated crustal material while cooling (DePaolo, 1981). Therefore, these magmas are characterized by high incompatible trace element concentrations, fractionated HFSE/LILE ratios and enriched isotopic compositions, and contain significant amounts of phenocrysts and cumulates. A later melt pulse then triggered the eruption. As the eruption began, magma from the sills mixed with the later, more primitive magma. Over the course of eruption, the sill component was gradually exhausted. As a result, the sill-component signatures decreased over time, reflected in decreasing basalt incompatible trace element concentrations, less enriched isotopic compositions and lower phenocryst and cumulate xenolith abundances.

The limited major element variation recorded in PC basalts places constraints on the proportion of evolved (low Mg#) melt mixed with primitive (high Mg#) melt. After correcting for large olivine phenocrysts, the Mg# of erupted PC melts varied from ~66 in the earliest PC melt to ~70 at the end of the eruption. We estimated the potential composition of the evolved sill component by calculating the liquid line of descent of primitive PC basalts using the MELTS program. Trace element abundances were estimated from the most enriched cpx, which has a

Ce concentration of ~54 ppm, and is estimated to derive from a melt with ~420 ppm Ce ($K_d = 0.13$). Modeling details are given in the appendix. A simple binary mixing model suggests addition of up to ~20% of the sill component could generate the relatively small variations in Mg# and large variations in incompatible trace element concentrations in the erupted melts (Fig. A6). This mixing process can also explain the variations in Sr-Nd-Pb isotopic compositions in PC basalts.

In summary, the compositional variations observed in the PC sequence appear to primarily derive from processes related to fractionation and crustal assimilation followed by magma mixing rather than mantle heterogeneity. Several other vents in the BPVF display disequilibrium between phenocrysts and host basalts similar to what is observed in the PC sequence (Fig. 4b). Therefore, pre-eruptive magma storage and crustal assimilation may have affected the compositions of BPVF basalts to a greater extent than has previously been recognized. In addition, phenocrysts-host rock disequilibrium is present in basalts from other localities, including Paricutin Volcano, Mexico (Rowe et al., 2011). Primitive major element compositions (e.g., high Mg#) do not guarantee that erupted basalts are unaffected by crustal storage and modification. The effects of crustal assimilation must be carefully considered before basalt compositions can be used to infer the evolution of the mantle source(s) at Big Pine or other monogenetic vent fields. Melt “crystal cargo” may provide the most effective way to discriminate signatures of magma ponding, fractionation and crustal assimilation. In addition, if crustal magma storage prior to eruption is widely applicable to monogenetic volcanoes, seismic activity or ground deformation related to magma injection may provide warning of impending eruptions.

6. Conclusions

- Correlations between trace element concentrations and isotopic ratios observed in Papoose Canyon (PC) basalts indicate binary mixing of two distinct melts, where the proportion of the chemically and isotopically more enriched component decreases over time.
- The major and trace element compositions of PC phenocrysts and ultramafic xenoliths suggest they derive from a melt that is more evolved and chemically more enriched than PC basalts.
- Pressure constraints suggest PC phenocrysts and ultramafic xenoliths fractionated at pressures of ~5-7 kbar, which corresponds to the middle crust, possibly near the brittle-ductile transition.
- Correlations between HFSE depletions and Sr-Nd-Pb isotopic compositions, high $\delta^{18}\text{O}$ in olivines, and radiogenic Os isotopic compositions in wholerocks suggest PC basalts have likely experienced crustal contamination.
- A deflating sills model is proposed to explain the observations and implications described above. An initial melt pulse stalled and formed one or more sills within the crust, where melt fractionated and assimilated crustal material. The eruption was triggered by the arrival of a later melt pulse, and the early, evolved melts were drained and mixed with the later, more primitive melts. The temporal-compositional trends therefore reflect exhaustion of the sill component over time.

- These results indicate that even “primitive” lavas may contain a significant signature of magma ponding, fractionation, and crustal contamination that is obscured within the bulk lava but is recorded in the melt “crystal cargo”.

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745 and Range, SW USA. *J. Geophys. Res.* 107 (B1). doi:10.1029/2001JB000209.

746 **Figures**

747 Figure 1. The modal abundance of large olivine phenocrysts (1-3.5 mm) in PC basalts
748 decreases over the course of eruption.

749

Figure 2. (a) Measured clinopyroxene REE concentrations of PC phenocrysts and xenoliths (normalized to primitive mantle; McDonough and Sun, 1995). (b) Calculated melt REE concentrations in equilibrium with the clinopyroxene assuming no trapped melt in the xenoliths. See Appendix for more details about K_d value estimation. Green field shows the compositional variations of bulk PC basalts.

Figure 3. Correlations between trace element concentrations and isotopic ratios of PC
basalts suggest binary mixing of two distinct components. (a) La and Sm concentrations are
linearly correlated. A representative partial melting curve is shown for comparison. This
partial melting curve is generated assuming batch melting of a typical mantle peridotite
that contains ~ 0.98 ppm La and ~ 0.46 ppm Sm (based on measured compositions of BPVF
mantle xenoliths; Lee et al., 2001). The bulk partition coefficients are calculated assuming
modal abundances of olivine:clinopyroxene:orthopyroxene:spinel=0.58:0.13:0.27:0.02.
Partition coefficients used are from McKenzie & O'Nions, 1991. (b) $^{87}\text{Sr}/^{86}\text{Sr}$ is linearly
correlated with the inverse of Sr concentration, which is consistent with binary mixing of
two isotopically distinct components. Literature data from Blondes et al. (2008).

Figure 4. Measured Mg# of PC phenocrysts are plotted versus whole rock Mg# of their host
basalts. Equilibrium fields are calculated from Fe^{2+}/Mg partition coefficient between
olivine or cpx and melt. (a) Olivine phenocrysts versus host basalts, (b) cpx phenocrysts
versus host basalts. PC literature data are from Blondes et al. (2008 and 2012) and Gazel et
al. (2012). BPVF literature data are from Mordick & Glazner (2006) and Gazel et al. (2012).
Open squares show wholerock data from Mordick & Glazner (2006) that are corrected for
olivine addition. Some olivine phenocrysts from Blondes et al. (2012) have extremely high
Mg# (>95), which they interpreted as the result of subsolidus oxidation after eruption, and
are therefore not relevant to this study.

777 Figure 5. The ultramafic xenoliths in early PC sequence have isotopic compositions
 778 overlapping but extending to more enriched compositions compared to PC basalts.
 779 Individual xenoliths are systematically more enriched than their host basalts. WGB –
 780 western great basin, B&R – basin and range, BPVF – Big Pine volcanic field (Rogers et al.,
 781 1995; Beard & Glazner, 1995; Reid & Ramos, 1996; Beard & Johnson, 1997; Kempton et al.,
 782 1991; Blondes et al., 2008).

783

784 Figure 6. MELTS modeling results of fractional crystallization using 14PC-2 as the starting
785 melt composition, 1.5 wt.% initial water content, and oxygen buffer initiated at FMQ+1, and
786 allowed to proceed unbuffered. The range of crystallized mineral compositions (Mg#)
787 varies with pressure. The range of olivine and clinopyroxene Mg# (73 – 88) observed in PC
788 phenocrysts and xenoliths is marked by vertical dotted lines. The minimum olivine Mg#
789 (73) suggests a maximum crystallization pressure of ~6.5 kbar, whereas the maximum
790 clinopyroxene Mg# (88) suggests a minimum crystallization pressure of ~5 kbar (marked
791 by horizontal dotted lines and the black vertical arrows).

792

Figure 7. (a) Lu concentrations in clinopyroxene from PC phenocrysts and xenoliths are negatively correlated with cpx Mg#. This trend is largely consistent with fractional crystallization. (b) Ce concentrations in these cpx show large variations that cannot be explained by fractional crystallization alone. The fractional crystallization curves are modeled assuming the initial melt has a composition resembling the least enriched (latest) PC lava ([Lu]=0.38 ppm, [Ce] = 71 ppm). The mass fraction and Mg# of fractional crystallization products of this initial melt was modeled using the MELTS software at 6 kbar, with an initial water content of 1.5 wt.%, initial oxygen buffer of QFM+1, and temperature steps of 10°C. Partition coefficients between cpx and melt was varied corresponding to the crystallization temperature of each step ($K_d^{Lu} = 0.32-0.49$, $K_d^{Ce} = 0.100-0.144$, see appendix for more information on K_d selection).

805 Figure 8. Correlation between La/Nb ratios and Sr isotope composition in PC basalts.

806

807 Figure 9. Deflating sills model: (a) One or more sills are created through a magmatic pulse
808 some time prior to eruption; (b) Melts ponding in the sills undergo fractional
809 crystallization and assimilate crustal rocks; (c) A second magma pulse triggers eruption.
810 As the eruption progresses, the sills deflate due to dropping pressure, which forces the
811 evolved and contaminated magma with phenocrysts and xenoliths to mix back into the
812 feeder conduit with more primitive melt.

814 Table 1. Compositions of PC basalts, phenocrysts and xenoliths.

	Mg# (wr)	Mg# ^{**} (ol corr.)	Mg# (ol, ave.)	Mg# (cpx, ave.)	⁸⁷ Sr/ ⁸⁶ Sr	¹⁴³ Nd/ ¹⁴⁴ Nd	²⁰⁶ Pb/ ²⁰⁴ Pb	²⁰⁷ Pb/ ²⁰⁴ Pb	²⁰⁸ Pb/ ²⁰⁴ Pb	¹⁸⁷ Os/ ¹⁸⁸ Os	Os (ppt)	δ ¹⁸ O (ol)
Basalts												
14PC-2	68.7	65.5	84.2		0.70614	0.51250	19.11	15.66	38.86	0.183	39	5.6
14PC-4	69.2	66.2	83.9	86.0	0.70610	0.51249	19.11	15.66	38.86	0.190	38	
14PC-5	69.8	67.2	83.4		0.70604		19.08	15.65	38.84	0.287	55	
PC-1	69.6	67.4	83.1	85.0	0.70600	0.51249	19.06	15.65	38.85	0.261	20	5.4, 5.6
PC-2	69.5	67.7	84.0	83.3	0.70591	0.51251	19.03	15.65	38.85			5.7
14PC-12	69.8	69.3	84.5		0.70575	0.51254	18.98	15.63	38.80	0.207	37	5.4
14PC-13	69.7	69.4	84.6		0.70559	0.51257	18.99	15.64	38.79			
PC-3	69.1	68.7	84.3	83.9	0.70581	0.51252	18.99	15.64	38.83			5.6, 5.8
14PCX-3 host basalt					0.70596		19.07	15.65	38.83			5.5
14PCX-6 host basalt					0.70600		19.09	15.66	38.85			5.5
14PCX-9 host basalt					0.70602		19.12	15.67	38.89			
14PCX-17 host basalt					0.70606		19.12	15.67	38.89			
Xenoliths												
14PCX-3			83.9	84.0	0.70625 [*]	0.51248 [*]	19.11 [*]	15.66 ^{**}	38.86 ^{**}			5.7
14PCX-6			84.5	83.4	0.70606 [*]	0.51248 [*]	19.08 [*]	15.66 ^{**}	38.85 ^{**}			5.5
14PCX-9				85.1	0.70629 [*]	0.51247 [*]	19.16 [*]	15.68 ^{**}	38.91 ^{**}			
14PCX-15			84.4	85.6	0.70591 [*]	0.51251 [*]	19.12 [*]	15.67 ^{**}	38.88 ^{**}			5.6
14PCX-17			86.3	85.3	0.70613 [*]	0.51248 [*]	19.14 [*]	15.68 ^{**}	38.93 ^{**}			

815 ^{*} Radiogenic isotopic compositions measured on clinopyroxene separates from ultramafic xenoliths.

816 ^{**} Mg# of PC melts are calculated by subtract cumulate olivine from whole rock. The average composition of PC olivine phenocrysts was used.

817 The amount of olivine cumulate subtracted is determined based on the least squares regression between large olivine phenocryst abundance

818 and relative stratigraphic position of samples (Fig. 1).

Appendix

1. Methods

1.1 Whole rock major and trace elements

Major and trace elements were analyzed on fresh whole rock samples using standard XRF and solution ICP-MS techniques at Washington State University. Duplicate measurements of major elements by XRF are reproducible within 1% (1 s.d.). Trace elements measured by ICP-MS are reproducible within 3% (1 s.d.). The secondary standard BHVO-2 is accurate within 1% for major elements, within 5% for most trace elements except Nb, Hf, Rb (10%) and Tb, Ta (15%).

1.2 Mineral major and trace elements

Olivine and clinopyroxene major element abundances were measured by EMPA on a JEOL-8200 microprobe at the University of Texas at Austin (UT Austin) using methods outlined in Gao et al. (2016). Fresh samples were crushed, and olivine and clinopyroxene grains with no visible alteration were handpicked under a binocular microscope and mounted in epoxy. A focused beam was used for olivine analyses, whereas a defocused beam (10 μ m) was used for cpx to reduce Na migration. Initial analyses demonstrated only minor core-rim variations [e.g. within-grain variation in Mg# ($100 \cdot \text{Mg}/(\text{Mg} + \text{Fe}^{2+})$) < 1]. Therefore, only core analyses were performed for remaining grains. Measurements of secondary standards were accurate to within 5% for Na, Ni and Cr, within 3% for Al, Si and Fe, within 1% for Mg and Ca, and 18% for MnO. Duplicates are reproducible within 5% (1 s.d.) for Fe and Al, within 2% for Na, Cr, Ca, Mg, Mn and Ni, within 1% for Si, and 10% for MnO.

Clinopyroxene trace element concentrations were measured via an Agilent 7500ce quadrupole with a New Wave UP-193FX laser system at UT Austin using methods outlined in Gao et al. (2016). BCR-2g was measured as a secondary standard, based on which all elements are accurate to within 10% and reproducible within 5% (1 s.d.). Three spots were analyzed for most grains, and the average of the analyses from a single grain and averaged BCR-2g are reported in Table A6.

1.3 Radiogenic and oxygen isotopes

For Sr, Nd, and Pb isotopes, ~100-150 mg cpx separates or ~20 mg phenocryst-free matrix chips were digested following the procedures outlined in Byerly & Lassiter (2012). AG1-X8, Sr-Spec and REE-spec/HDEHP resins were used to separate Pb, Sr and Nd, respectively. Sr, Nd, and Pb isotopes were analyzed on a TE Triton TIMS using methods outlined in Gao et al. (2016) and Byerly & Lassiter (2012). NBS 987, UT Ames Nd and NBS 981 standards were measured regularly during the course of this study. The average NBS 987 $^{87}\text{Sr}/^{86}\text{Sr}$ for the TIMS lab over the period during which analyses were performed was 0.710253 ± 15 (2 s.d.), consistent with previously reported values (e.g., 0.710248 ± 15 ; Thirlwall, 1991). The average $^{143}\text{Nd}/^{144}\text{Nd}$ of UT Ames Nd was 0.512072 ± 10 (2 s.d.), comparable to previously reported values (e.g., 0.512069 ± 14 ; Scher & Delaney, 2010). For Pb isotopes, instrumental mass fractionation was corrected using a Pb double spike. NBS 981 measured by the double spike method during the course of this study has average $^{206}\text{Pb}/^{204}\text{Pb} = 16.933 \pm 3$, $^{207}\text{Pb}/^{204}\text{Pb} = 15.488 \pm 3$, and $^{208}\text{Pb}/^{204}\text{Pb} = 36.688 \pm 11$ (2 s.d.). For comparison, the values reported by Todt et al. (1996) are $^{206}\text{Pb}/^{204}\text{Pb} = 16.9356 \pm 7$, $^{207}\text{Pb}/^{204}\text{Pb} = 15.4891 \pm 9$, and $^{208}\text{Pb}/^{204}\text{Pb} = 36.7006 \pm 34$ (2 s.d.). Total procedure blanks were less

than 150 pg, 75 pg and 15 pg for Sr, Nd and Pb, respectively. No standard or blank correction was applied to the reported data.

Osmium concentrations and Os isotopes were measured on whole rock powders of 5 samples. Fresh samples were cut with a diamond saw, and then carefully polished with aluminum oxide sandpaper to remove any saw marks. Samples were rinsed with DI water and dried before being wrapped in plastic (to avoid contact with metal) and crushed by hammer. Basalt chips were then powdered in an alumina ball mill. Rhenium and osmium were separated from whole rock powders following the procedures of Byerly & Lassiter (2012). Osmium concentrations were measured via isotope dilution using a mixed ^{185}Re - ^{190}Os spike. Osmium isotopes were measured with the TE Triton TIMS in negative ion mode. The UMD Os standard was measured regularly and the average measured $^{187}\text{Os}/^{188}\text{Os}$ was 0.11382 ± 0.00003 (2 s.d.), consistent with the value (0.1137920 ± 0.0000024) reported by Walker et al. (2005). Osmium blanks were ~ 0.3 pg/g.

Oxygen isotope compositions were measured on ~ 2 mg of olivine separates by laser fluorination following the methods of Sharp (1995) using a ThermoElectron MAT253 mass spectrometer. In-house quartz standard Lausanne-1 ($\delta^{18}\text{O} = 18.1\text{‰}$), garnet standard UWG-2 ($\delta^{18}\text{O} = 5.8\text{‰}$) (Valley et al., 1995), and olivine standard San Carlos ($\delta^{18}\text{O} = 5.2\text{‰}$) were measured to ensure accuracy and precision. $\delta^{18}\text{O}$ values are reported relative to SMOW, where the $\delta^{18}\text{O}$ value of NBS-28 is $+9.65\text{‰}$. Precision based on replicates of the standards is $\pm 0.1\text{‰}$ (1 s.d.). All the isotope analyses described above were performed at UT Austin.

2. MELTS modeling

The crystallization products of PC magma were constrained by modeling fractional crystallization over a range of pressures (1-11 kbar) using the MELTS thermodynamic model (Ghiorso & Sack, 1995). The effects of initial melt composition, water content, oxygen fugacity, and crystallization style (fractional versus batch) on modeling results were evaluated by varying these parameters over reasonable ranges. The compositions of both the earliest and the latest PC basalts were used as the initial model melt compositions. Initial water contents of 1.5 and 3.0 wt.% were used, consistent with the estimated water contents for primitive BPVF melts (Gazal et al., 2012). Oxygen fugacity was initiated at FMQ and FMQ+1, and allowed to proceed unbuffered in MELTS modeling. The effects of initial melt composition, water content, oxygen fugacity, and crystallization style (fractional versus batch) on modeling results were evaluated by varying one parameter at a time compared to the set of parameters reported in the main text. When using 14PC-13 as the initial melt, the Mg# of the first clinopyroxene decrease more significantly with decreasing pressure, compared to using 14PC-2 as initial melt. As a result, there is no pressure range where both clinopyroxene with high Mg# and olivine with low Fo can be both formed using 14PC-13 as the starting composition (Fig. A5-a). When 3 wt.% initial water content was used instead of 1.5 wt.% water, clinopyroxene crystallization at relatively low pressures is further suppressed, and the MELTS modeling results suggest crystallization of PC phenocrysts and xenoliths at ~7-7.5 kbar (Fig. A5-b). Batch crystallization of 14PC-2 with initial water content of 1.5 wt.% and oxygen fugacity of FMQ+1 suggests crystallization pressures of 5-7.5 kbar (Fig. A5-c). Finally, when initial oxygen buffer was set at FMQ, MELTS modeling suggests crystallization of PC magmas at ~7.5-8 kbar (Fig. A5-d). Therefore, although the fields

of olivine and clinopyroxene compositions shift slightly with different input parameters, all MELTS models suggest significant olivine crystallization (with Fo extending to 73) only occurs at crustal depths (<8 kbar).

3. Clinopyroxene-melt partition coefficients

Clinopyroxene-melt partition coefficients (K_d) are a function of melt composition, mineral composition, temperature and pressure. Therefore, choosing appropriate K_d values is important for modeling melting and crystallization processes. In general, K_d values for Sr increase and K_d values for HFSE decrease with increasing pressure. K_d values of most incompatible elements increase as temperature, melt Mg# and cpx Mg# decreases (e.g., Adam & Green, 2003; Bédard, 2014). K_d values of REE and HFSE also increase with increasing TiO_2 content in cpx (e.g., Adam & Green, 1994; Bédard, 2014).

Clinopyroxenes in PC basalts and ultramafic xenoliths span a range of compositions, which is consistent with progressive fractional crystallization (Mg# negatively correlates with TiO_2 contents). We assume that all PC phenocrysts and xenoliths were crystallized at relatively uniform pressures (5-7 kbar). MELTS modeling suggest that the majority of PC phenocrysts and ultramafic xenoliths crystallized at temperatures of 1090-1200°C (assuming 1.5 wt.% initial water content). The K_d values for high-Mg# cpx (88), which we estimate formed at ~1200 °C, were calculated using the BIGD program (Nielsen, 1992). The major element compositions of PC sample 14PC-2 and of the most magnesian clinopyroxene were used as inputs to the BigD model. The composition of the youngest PC basalt was assumed as melt composition, which has Mg# in chemical equilibrium with the high-Mg# cpx. Estimated K_d values for high-Mg# cpx

calculated using BigD are largely consistent with published K_d values for cpx in equilibrium with alkali basalt (e.g., Zack & Brumm, 1998), as well as with experimentally determined partition coefficients for cpx crystallized at similar pressure and temperature (Adam & Green, 1994; Green et al., 2000; Adam & Green, 2003).

Estimation of partition coefficients for more evolved cpx (Mg# = 78) is subject to greater uncertainty because the composition of the melt from which these cpx grew is more uncertain. Equilibrium crystallization of the more evolved cpx (Mg# = 78) requires a parental melt Mg# $\approx$ 42. A possible parental melt composition was estimated by calculating effects of fractional crystallization of 14PC-2 using the MELTs program as described above. When using compositions of this evolved melt, BIGD gives unreasonably high K_d values, e.g., $K_d^{\text{HREE}} > 2$. However, experiments run at similar pressure (5 kbar), temperature (1070 °C), cpx (Mg# = 74, $\text{TiO}_2 = 1.25$) and melt compositions (Mg# = 44) suggest $K_d^{\text{La}} = 0.1$, $K_d^{\text{Ce}} = 0.16$, $K_d^{\text{Nd}} = 0.29$, $K_d^{\text{Dy}} = 0.51$, $K_d^{\text{Er}} = 0.58$, and $K_d^{\text{Yb}} = 0.69$ (Mercer & Johnston, 2008). These partition coefficients were utilized for the low-Mg# cpx. For cpx with intermediate Mg# and estimated crystallization temperatures, we interpolated between the high-Mg# and low-Mg# K_d estimates. Estimated K_d values utilized in Figures 2 and 7 are reported in Table A7. Uncertainties in calculated K_d values may effect the exact position and slope of the calculated parental melt REE patterns shown in Figure 2. However, regardless of K_d value, the large variation in LREE abundance at roughly constant Mg# illustrated in Figure 7 requires cpx to crystallize from a range of melts variably enriched in LREE and LILE.

4. Binary mixing model

In order to constrain the proportion of the sill component mixed with primitive melts in the erupted PC basalts, we modeled binary mixing of a primitive melt and an evolved melt. The primitive melt is assumed to have a composition similar to the least enriched PC basalts (MgO = 9.32, FeO = 7.14 assuming $\text{FeO}/\text{FeO}^{\text{T}} = 0.85$, [Ce] = 71 ppm). To estimate the major element composition of the evolved sill component, we used the MELTS model to estimate the composition of melt generated by fractional crystallization, using a “primitive” melt (14PC-2) as the starting composition, using the same model parameters as above ($P = 6$ kbar, 1.5 wt.% H₂O, initial oxygen fugacity = FMQ). The melt composition in equilibrium with cpx with Mg# = 73 (the most evolved cpx phenocryst measured) was used to estimate the composition of the sill component (MgO = 2.31, FeO = 7.37). The most enriched cpx observed has Ce concentration of ~54 ppm, and therefore is in equilibrium with a melt having Ce concentration of ~420 ppm (using a $K_d = 0.13$, corresponding to the Mg# of this particular cpx). Binary mixing of this evolved, enriched melt and the primitive melt can generate the observed correlation between melt Mg# and Ce concentration by mixing a maximum of ~20% of the sill component with the primitive melt (Fig. A6).

Although this mixing model can account for the large variations in incompatible trace element abundance and modest variations in Mg# observed in the PC basalts, our estimated evolved melt component has higher SiO₂ than the primitive melt, and therefore this model predicts that SiO₂ should decrease rather than increase during the course of eruption (as the ratio of sill melt/primitive melt decreases with time). In addition, the model does not explain the slightly higher CaO in the early (more contaminated) melts. This suggests either that the

first melt pulse that ponded and fractionated within the crust initially had a lower SiO_2 content than the later pulse that triggered eruption, or that the process of assimilation caused SiO_2 to decrease rather than increase. The coupled low SiO_2 and high CaO required of the sill component based on the temporal major element trends may indicate assimilation of high-Ca, low-Si skarn material rather than more typical lower crust.

5. Figures

Figure A1. (a) Stacked basaltic eruption sequence was cut and exposed at the Papoose Canyon. (b) Schematic diagram showing our sampling strategy. We walked down the canyon valley, and collected lava flow samples from earlier to later Papoose Canyon eruption sequence. The dipping angle of actual PC lava strata varies from $\sim 30^\circ$ in the earliest strata to shallower angle in later strata. We calculated the relative stratigraphic height of each sample assuming a uniform dipping angle of 30° , which is also the typical static angle of repose for scoria cone (Carrigy, 1970). Given the fact that the dipping angle of actual PC lava strata appear to be shallower in later strata, the calculated stratigraphic thickness may be slightly overestimated, but the relative stratigraphic positions are still robust. Calculated relative stratigraphic positions are reported in Table A1.

Figure A2. Temporal-compositional variations of major and trace elements.

Figure A3. (a) Olivine and clinopyroxene phenocrysts in PC basalts span a wide range of Mg#, and are mostly out of equilibrium with their host basalts. (b) Olivine and clinopyroxene from PC ultramafic xenoliths span a similar range of Mg# as PC phenocrysts, which is lower than typical mantle peridotites.

Figure A4. Photomicrographs of PC ultramafic xenoliths show cumulate textures. (a) Frozen interstitial melts among grains under plane-polarized light. (b) Same area in (a) under cross-polarized light. (c) Partially melted grains and uneven grain boundaries. (d) Same area in (c) under cross-polarized light.

Figure A5. (a) Fractional crystallization of 14PC-13 with initial water content of 1.5 wt.% and FMQ+1 oxygen buffer. (b) Fractional crystallization of 14PC-2 with initial water content of 3.0 wt.% and FMQ+1 oxygen buffer. (c) Batch fractional crystallization of 14PC-2 with initial water content of 1.5 wt.% and FMQ+1 oxygen buffer. (d) Fractional crystallization of 14PC-2 with initial water content of 1.5 wt.% and FMQ oxygen buffer. The range of olivine and clinopyroxene Mg# (73 – 88) observed in PC phenocrysts and xenoliths is marked by vertical dotted lines. The minimum olivine Mg# suggests a maximum crystallization pressure, whereas the maximum clinopyroxene Mg# suggests a minimum crystallization pressure (marked by horizontal dotted lines and the black vertical arrows).

Figure A6. Binary mixing model suggests that mixing up to ~20% evolved melt into primitive melt can produce the observed compositional variations in PC basalts. Mg# of the modeled melt is calculated from the MgO and FeO contents of mixed melt. Mg# of PC samples are melt Mg# calculated from whole rock composition by subtracting cumulate olivine.

5. Tables

Table A1. Coordinates of samples. ELE – elevation, STRAT – calculated relative stratigraphic position.

	N	W	ELE (m)	STRAT (m)		N	W	ELE (m)
PC-1	37.01492	118.1684	1588.3	204	14PCX-6	37.01515	118.1615	1652.0
PC-2	37.01512	118.1697	1563.0	242	14PCX-7	37.01515	118.1615	1652.0
PC-3	37.01545	118.1731	1521.0	364	14PCX-8	37.01515	118.1615	1652.0
14PC-2	37.01515	118.1636	1638.0	18	14PCX-9	37.01592	118.1619	1676.4
14PC-4	37.01526	118.1645	1617.0	46	14PCX-10	37.01592	118.1619	1676.4
14PC-5	37.01528	118.1664	1597.2	119	14PCX-11	37.01592	118.1619	1676.4
14PC-8A	37.01478	118.1690	1586.5	231	14PCX-12	37.01658	118.1619	1676.4
14PC-9	37.01549	118.1703	1589.0	295	14PCX-13	37.01522	118.1629	1639.5
14PC-12	37.01537	118.1728	1527.0	353	14PCX-15	37.01522	118.1629	1639.5
14PC-13	37.01547	118.1731	1521.0	362	14PCX-16	37.01518	118.1635	1635.3
14QC-2	37.01493	118.1680	1596.2	193	14PCX-17	37.01510	118.1635	1636.2
14PCX-5	37.01515	118.1615	1652.0		14PCX-18	37.01493	118.1680	1596.2

Table A2. Phenocryst point counting results.

	14PC-4	14PC-5	14PC-8a	14PC-9	14PC-12	14PC-13
1-3.5mm (%)	4.1	3.8	3.4	2.2	0.9	0.6
<1mm (%)	0.7	2.8	1.6	3.9	1.8	1.2
Total (%)	4.8	6.5	4.0	6.1	2.7	1.8

Table A3. Major and trace element compositions of PC basalts and secondary standard BHVO2.

	14PC-2	14PC-4	14PC-5	14PCX-9L	14QC-2	PC-1	PC-2	14PC-12	14PC-13	PC-3	BHVO2
SiO ₂	45.83	45.66	46.17	46.48	46.58	46.60	47.00	47.42	46.96	47.20	49.88
TiO ₂	2.14	2.09	2.09	2.03	2.01	2.06	1.97	1.95	1.94	1.97	2.76
Al ₂ O ₃	15.76	15.83	15.82	15.91	16.09	16.33	16.10	16.24	16.18	16.29	13.61
FeO*	9.11	9.05	9.07	8.97	8.87	8.80	8.87	8.80	8.94	8.83	11.18
MnO	0.15	0.15	0.15	0.15	0.15	0.15	0.15	0.15	0.15	0.15	0.17
MgO	9.46	9.62	9.39	9.78	9.57	9.52	9.55	9.62	9.73	9.34	7.30
CaO	9.92	10.21	9.79	10.05	9.65	9.95	9.66	9.62	9.76	9.69	11.42
Na ₂ O	3.72	3.72	3.75	3.81	3.95	3.94	3.74	3.83	3.82	3.59	2.23
K ₂ O	2.04	2.09	1.95	1.95	1.75	1.89	1.85	1.68	1.40	1.68	0.52
P ₂ O ₅	1.10	1.02	0.94	0.93	0.86	0.89	0.77	0.71	0.69	0.73	0.26
Sum	99.23	99.44	99.12	100.05	99.47	100.13	99.67	100.03	99.58	99.47	99.32
LOI (%)	0.19	0.44	0.25	0.28	0.02	-0.17	-0.19	0.16	0.54	-0.02	0.00
Cs	0.28	0.27	0.28	0.29	0.34	0.31	0.31	0.28	0.25	0.28	0.10
Rb	28.5	24.8	22.7	27.1	26.4	24.3	23.6	26.9	20.2	23.0	8.9
Ba	1329.7	1230.7	1117.8	1102.6	1038.2	1064.7	927.9	842.8	780.8	866.3	128.2
Th	4.3	4.1	3.8	3.9	3.7	3.7	3.3	3.1	2.7	3.0	1.24
U	1.2	1.1	1.1	1.1	1.1	1.1	0.9	0.8	0.8	0.9	0.42
Ta	1.4	1.3	1.3	1.3	1.3	1.3	1.1	1.2	1.1	1.1	1.17
Nb	22.4	21.4	20.9	20.9	20.7	21.1	18.6	17.8	16.9	18.1	16.8
La	64.1	60.4	56.1	55.5	51.6	51.4	45.9	42.0	38.3	42.6	15.3
Ce	141.3	133.3	123.3	121.6	113.2	112.7	100.5	91.6	86.2	93.5	36.9
Pb	7.9	7.2	6.6	7.1	6.4	6.8	6.0	5.4	5.3	5.6	1.6
Pr	18.4	17.2	16.0	15.7	14.5	14.6	13.1	11.9	11.2	12.2	5.26
Sr	1678.9	1591.7	1503.5	1457.8	1371.0	1398.9	1253.3	1143.3	1083.2	1180.3	395.9
Nd	71.8	68.0	63.6	61.5	57.8	57.8	51.9	47.9	45.0	49.0	24.1
Hf	5.9	5.6	5.6	5.7	5.8	5.9	5.5	5.2	5.1	5.4	4.4
Zr	271.7	263.8	263.6	266.3	268.2	271.6	258.2	247.1	238.2	252.0	166.2
Sm	12.5	11.8	11.4	10.9	10.4	10.2	9.5	8.9	8.5	9.0	6.4
Eu	3.5	3.3	3.2	3.1	3.0	3.0	2.8	2.6	2.5	2.7	2.2
Gd	8.9	8.5	8.4	8.0	8.0	7.9	7.6	7.1	7.0	7.3	6.4
Tb	1.2	1.2	1.2	1.1	1.1	1.1	1.1	1.0	1.0	1.1	1.0
Dy	6.5	6.5	6.3	6.3	6.4	6.3	6.1	5.9	6.0	6.0	5.7
Ho	1.2	1.2	1.2	1.2	1.2	1.2	1.2	1.2	1.1	1.2	1.1
Er	3.1	3.0	3.0	3.0	3.1	3.1	3.0	3.1	3.0	3.0	2.6
Tm	0.42	0.41	0.41	0.40	0.43	0.44	0.42	0.42	0.41	0.42	0.35
Y	30.1	29.8	29.7	29.5	30.5	30.4	29.9	29.2	28.7	29.4	25.6
Yb	2.5	2.5	2.5	2.5	2.6	2.6	2.6	2.6	2.6	2.6	1.9
Lu	0.37	0.37	0.38	0.37	0.40	0.42	0.39	0.39	0.38	0.40	0.28
Sc	22.4	23.5	24.0	23.1	23.8	24.3	25.8	25.5	25.7	26.0	31.9
Ni	174.8	181.9	187.2	186.9	187.8	174.7	188.6	190.1	193.8	180.6	115.0
Cr	237.0	260.0	268.2	250.7	264.1	267.5	273.4	287.2	299.1	280.3	280.3
V	188.4	189.7	183.9	186.3	187.8	189.5	188.4	186.7	176.7	190.6	312.5
Ga	15.4	15.7	14.5	16.2	15.4	16.5	15.4	16.5	17.2	16.4	20.8
Cu	44.2	39.2	52.1	40.1	34.0	45.9	48.2	52.7	42.5	35.1	126.9
Zn	74.1	74.5	71.5	72.6	70.6	93.2	72.8	69.5	70.3	73.1	99.3

Table A4. Major element compositions of phenocryst and xenolith olivine. Olivine grains from individual samples are numbered in order. For example, PC-1-O1 stands for the first olivine grain from sample PC-1.

	SiO2	FeO	MnO	MgO	NiO	CaO	Totals	Mg#		SiO2	FeO	MnO	MgO	NiO	CaO	Totals	Mg#
<i>Phenocrysts</i>									12PC-2-O13	40.27	14.63	0.22	46.45	0.24	0.08	101.88	85.1
PC-1-O1	39.41	16.70	0.22	45.10	0.30	0.09	101.82	82.9	12PC-2-O14	40.15	13.19	0.16	47.13	0.24	0.08	100.95	86.5
PC-1-O2	39.97	13.49	0.19	46.83	0.32	0.10	100.89	86.2	12PC-2-O15	39.04	19.58	0.30	41.83	0.26	0.11	101.11	79.4
PC-1-O3	38.95	20.59	0.35	40.32	0.10	0.11	100.42	77.9	12PC-2-O16	39.74	14.24	0.22	46.38	0.22	0.09	100.89	85.4
PC-1-O4	38.86	21.05	0.31	40.90	0.08	0.12	101.32	77.8	12PC-2-O17	39.94	12.91	0.20	46.97	0.19	0.09	100.31	86.8
PC-1-O5	40.12	16.20	0.23	44.59	0.26	0.08	101.49	83.2	12PC-2-O18	40.12	13.53	0.20	47.76	0.26	0.08	101.95	86.4
PC-1-O6	39.59	13.30	0.22	46.67	0.18	0.12	100.07	86.3	12PC-2-O19	39.07	18.69	0.30	42.58	0.15	0.09	100.88	80.4
PC-1-O7	39.79	14.04	0.24	45.81	0.27	0.11	100.26	85.4	12PC-2-O20	40.34	12.79	0.20	47.26	0.49	0.12	101.20	86.9
PC-1-O8	40.14	15.77	0.20	45.65	0.18	0.12	102.05	83.9	14PC-4-O1	40.09	14.38	0.21	46.71	0.25	0.08	101.71	85.4
PC-1-O9	40.11	16.14	0.23	46.10	0.25	0.09	102.92	83.7	14PC-4-O2	40.44	13.05	0.16	47.53	0.28	0.09	101.55	86.8
PC-1-O10	40.51	14.01	0.23	47.03	0.25	0.09	102.12	85.8	14PC-4-O3	39.77	17.68	0.32	44.22	0.20	0.15	102.34	81.8
PC-1-O11	40.13	16.16	0.19	45.15	0.26	0.07	101.97	83.4	14PC-4-O4	38.64	21.89	0.35	40.63	0.05	0.10	101.67	77.0
PC-1-O12	38.83	20.00	0.27	41.87	0.07	0.17	101.22	79.0	14PC-4-O5	40.46	14.20	0.19	46.93	0.26	0.08	102.12	85.6
PC-1-O13	40.48	14.55	0.17	45.88	0.25	0.09	101.43	85.0	14PC-4-O6	40.14	13.77	0.19	47.08	0.28	0.09	101.55	86.0
PC-1-O14	39.74	16.59	0.21	45.14	0.14	0.10	101.92	83.0	14PC-4-O7	40.45	13.79	0.18	47.11	0.22	0.08	101.82	86.0
PC-2-O1	39.70	14.14	0.22	46.15	0.25	0.12	100.58	85.4	14PC-4-O8	40.44	13.28	0.22	47.23	0.25	0.10	101.53	86.5
PC-2-O2	39.25	18.49	0.27	43.20	0.15	0.20	101.56	80.8	14PC-4-O9	39.14	20.98	0.48	41.81	0.19	0.16	102.76	78.2
PC-2-O3	39.44	15.85	0.23	45.18	0.30	0.12	101.13	83.7	14PC-4-O10	39.98	15.11	0.22	45.74	0.16	0.08	101.29	84.5
PC-2-O4	39.24	18.41	0.43	42.90	0.22	0.25	101.45	80.7	14PC-4-O11	40.00	14.83	0.21	45.92	0.18	0.09	101.23	84.8
PC-2-O5	39.70	15.07	0.27	45.48	0.28	0.21	101.02	84.5	14PC-4-O12	39.78	15.30	0.27	45.40	0.25	0.07	101.08	84.2
PC-2-O6	39.37	18.55	0.35	43.34	0.16	0.11	101.88	80.8	14PC-4-O13	39.85	15.98	0.26	45.94	0.24	0.09	102.36	83.8
PC-2-O7	40.00	15.14	0.22	46.06	0.28	0.10	101.80	84.6	14PC-5-O1	39.37	16.58	0.25	44.43	0.18	0.10	100.91	82.8
PC-2-O8	39.99	14.47	0.24	46.55	0.27	0.12	101.63	85.3	14PC-5-O2	39.41	15.51	0.21	45.76	0.25	0.10	101.25	84.2
PC-2-O9	39.93	14.23	0.19	46.42	0.29	0.10	101.16	85.5	14PC-5-O3	40.57	14.89	0.21	45.89	0.17	0.10	101.83	84.7
PC-2-O10	40.20	14.61	0.22	46.59	0.24	0.11	101.96	85.2	14PC-5-O4	39.99	14.88	0.23	46.12	0.21	0.08	101.51	84.8
PC-2-O11	39.02	15.36	0.23	45.01	0.31	0.08	100.00	84.1	14PC-5-O5	39.86	16.63	0.26	45.21	0.18	0.10	102.24	83.0
PC-2-O12	40.02	14.96	0.20	45.01	0.29	0.08	100.57	84.4	14PC-5-O6	39.57	16.63	0.21	44.82	0.17	0.11	101.51	82.9
PC-2-O13	39.80	13.50	0.22	45.64	0.20	0.10	99.46	85.9	14PC-5-O7	39.83	16.87	0.25	44.88	0.24	0.11	102.19	82.7
PC-2-O14	39.06	18.08	0.24	43.93	0.14	0.10	101.54	81.4	14PC-5-O8	39.73	15.29	0.21	45.73	0.22	0.09	101.28	84.3
PC-2-O15	39.04	19.59	0.31	41.16	0.18	0.07	100.34	79.1	14PC-5-O9	38.95	20.42	0.25	41.90	0.18	0.15	101.85	78.7

Table A4 (continued).

	SiO2	FeO	MnO	MgO	NiO	CaO	Totals	Mg#		SiO2	FeO	MnO	MgO	NiO	CaO	Totals	Mg#
PC-2-O16	39.53	18.42	0.25	42.69	0.25	0.11	101.26	80.7	14PC-5-O10	39.91	16.40	0.25	44.87	0.27	0.10	101.81	83.1
PC-2-O17	39.56	13.69	0.20	45.48	0.25	0.08	99.26	85.7	14PC-5-O11	40.13	16.23	0.22	44.60	0.12	0.11	101.41	83.2
PC-2-O18	39.28	15.73	0.22	45.46	0.19	0.11	101.00	83.9	14PC-5-O12	40.01	15.00	0.20	46.20	0.27	0.09	101.76	84.7
PC-2-O19	40.22	15.03	0.22	46.87	0.29	0.09	102.72	84.9	14PC-5-O13	40.20	15.07	0.22	45.83	0.14	0.10	101.57	84.6
PC-2-O20	39.76	16.24	0.22	45.27	0.24	0.07	101.79	83.4	14PC-5-O14	39.33	16.58	0.25	45.12	0.23	0.12	101.63	83.0
PC-2-O21	39.85	15.58	0.22	45.22	0.19	0.11	101.18	83.9	14PC-5-O15	39.61	16.68	0.22	45.21	0.28	0.09	102.08	83.0
PC-2-O22	38.69	18.56	0.27	43.55	0.05	0.14	101.25	80.9	14PC-5-O16	38.89	20.39	0.28	42.17	0.08	0.14	101.95	78.8
PC-2-O23	39.36	16.84	0.28	44.33	0.27	0.04	101.13	82.6	14PC-5-O17	39.77	15.43	0.25	45.56	0.17	0.10	101.29	84.2
PC-2-O24	40.14	15.21	0.22	46.36	0.25	0.07	102.25	84.6	14PC-5-O18	39.63	16.94	0.28	45.25	0.19	0.11	102.39	82.8
PC-2-O25	39.95	13.56	0.21	47.46	0.28	0.08	101.54	86.3	14PC-5-O19	40.54	13.28	0.22	47.58	0.20	0.08	101.90	86.6
PC-2-O26	40.27	14.07	0.21	46.64	0.25	0.14	101.58	85.6	14PC-5-O20	39.88	14.35	0.23	46.78	0.24	0.09	101.56	85.4
PC-2-O27	39.95	14.99	0.24	46.05	0.29	0.09	101.62	84.7	14PC-12-O1	40.10	12.91	0.17	47.91	0.34	0.13	101.55	87.0
PC-2-O28	38.81	20.89	0.38	41.47	0.28	0.06	101.88	78.1	14PC-12-O2	39.53	15.75	0.25	44.96	0.19	0.09	100.77	83.7
PC-2-O29	39.70	14.70	0.21	46.25	0.18	0.09	101.12	85.0	14PC-12-O3	40.20	12.47	0.21	47.83	0.27	0.08	101.07	87.3
PC-2-O30	39.10	17.01	0.33	43.32	0.15	0.21	100.12	82.1	14PC-12-O4	39.02	17.95	0.28	43.20	0.11	0.13	100.69	81.2
PC-2-O31	40.72	11.99	0.20	48.77	0.35	0.22	102.24	88.0	14PC-12-O5	39.44	16.10	0.23	45.06	0.23	0.09	101.15	83.4
PC-2-O32	40.33	13.68	0.21	46.29	0.21	0.10	100.83	85.9	14PC-12-O6	39.45	15.88	0.25	44.99	0.18	0.11	100.85	83.6
PC-2-O33	40.18	15.87	0.24	45.47	0.25	0.10	102.12	83.8	14PC-12-O7	39.06	18.64	0.29	42.43	0.17	0.12	100.71	80.4
PC-2-O34	39.68	14.55	0.19	46.24	0.19	0.09	100.94	85.1	14PC-12-O8	39.99	14.52	0.23	45.87	0.27	0.10	100.98	85.0
PC-2-O35	39.75	14.43	0.25	46.27	0.24	0.09	101.01	85.2	14PC-12-O9	39.85	14.79	0.21	45.90	0.29	0.10	101.14	84.8
PC-2-O36	39.74	14.51	0.25	46.10	0.24	0.12	100.95	85.1	14PC-12-O10	39.75	13.60	0.18	46.69	0.25	0.09	100.57	86.1
PC-2-O37	39.24	17.45	0.26	44.37	0.16	0.09	101.56	82.1	14PC-12-O11	40.12	12.50	0.18	47.98	0.35	0.09	101.20	87.4
PC-2-O38	39.55	18.43	0.30	43.22	0.26	0.14	101.89	80.8	14PC-12-O12	39.80	16.01	0.22	45.14	0.25	0.10	101.52	83.5
PC-2-O39	40.78	12.68	0.20	47.94	0.23	0.22	102.04	87.2	14PC-12-O13	39.78	15.74	0.24	44.73	0.19	0.10	100.79	83.6
PC-2-O40	40.26	14.14	0.25	46.69	0.22	0.10	101.65	85.6	14PC-12-O14	38.72	16.02	0.21	44.29	0.17	0.09	99.51	83.3
PC-2-O41	39.31	12.56	0.20	47.24	0.22	0.09	99.64	87.1	14PC-12-O15	40.23	12.29	0.19	47.78	0.24	0.12	100.86	87.5
PC-2-O42	40.58	12.24	0.18	47.61	0.25	0.09	100.94	87.5	14PC-12-O16	39.02	18.69	0.28	43.17	0.16	0.12	101.44	80.6
PC-2-O43	39.70	12.30	0.16	47.26	0.27	0.12	99.81	87.4	14PC-12-O17	39.14	16.06	0.25	44.82	0.16	0.10	100.53	83.4
PC-2-O44	39.04	16.97	0.20	44.20	0.11	0.09	100.62	82.4	14PC-12-O18	39.99	12.40	0.18	47.70	0.23	0.10	100.60	87.4
PC-2-O45	40.48	14.77	0.24	45.91	0.30	0.08	101.78	84.8	14PC-12-O19	40.23	11.29	0.17	48.42	0.27	0.24	100.61	88.5
PC-2-O46	40.48	14.92	0.23	46.78	0.22	0.09	102.72	85.0	14PC-12-O20	39.77	15.01	0.21	45.92	0.27	0.09	101.27	84.6

Table A4 (continued).

	SiO2	FeO	MnO	MgO	NiO	CaO	Totals	Mg#		SiO2	FeO	MnO	MgO	NiO	CaO	Totals	Mg#
PC-2-O47	39.21	17.69	0.31	43.87	0.22	0.15	101.44	81.7	14PC-12-O21	40.59	12.27	0.23	48.25	0.22	0.07	101.64	87.6
PC-2-O48	39.99	16.15	0.26	45.61	0.19	0.15	102.36	83.6	14PC-12-O22	39.45	18.28	0.28	43.36	0.13	0.12	101.63	81.0
PC-2-O49	40.21	16.32	0.22	45.89	0.17	0.11	102.92	83.5	14PC-12-O23	39.87	14.92	0.22	45.83	0.21	0.12	101.17	84.7
PC-2-O50	39.28	18.89	0.27	43.39	0.10	0.09	102.02	80.5	14PC-13-O1	40.30	14.41	0.22	46.89	0.22	0.09	102.14	85.4
PC-2-O51	39.83	12.89	0.20	48.19	0.30	0.11	101.52	87.1	14PC-13-O2	39.99	17.65	0.27	44.03	0.18	0.11	102.23	81.8
PC-2-O52	39.71	16.55	0.21	45.44	0.23	0.10	102.23	83.2	14PC-13-O3	40.31	15.57	0.25	46.46	0.29	0.09	102.97	84.3
PC-2-O53	39.55	16.89	0.24	44.90	0.15	0.10	101.83	82.7	14PC-13-O4	40.50	14.07	0.24	47.01	0.25	0.09	102.16	85.7
PC-2-O54	40.37	16.42	0.28	45.62	0.32	0.08	103.08	83.3	14PC-13-O5	40.03	13.97	0.20	46.86	0.25	0.10	101.40	85.8
PC-2-O55	39.92	13.83	0.27	47.02	0.29	0.22	101.55	86.0	14PC-13-O6	39.75	18.04	0.25	44.32	0.11	0.11	102.58	81.6
PC-3-O1	39.53	14.47	0.20	46.31	0.23	0.07	100.81	85.2	14PC-13-O7	40.25	14.20	0.24	46.82	0.22	0.08	101.81	85.6
PC-3-O2	38.83	17.02	0.25	44.48	0.23	0.16	100.98	82.5	14PC-13-O8	39.72	16.64	0.27	44.83	0.18	0.07	101.70	82.9
PC-3-O3	39.95	14.44	0.24	46.35	0.32	0.10	101.40	85.2	14PC-13-O9	40.12	15.32	0.20	45.72	0.31	0.09	101.76	84.3
PC-3-O4	39.86	14.07	0.25	46.57	0.29	0.13	101.17	85.6	14PC-13-O10	40.72	14.37	0.21	46.86	0.19	0.09	102.43	85.4
PC-3-O5	38.38	21.66	0.55	40.03	0.17	0.29	101.07	76.9	14PC-13-O11	40.54	14.14	0.23	47.13	0.27	0.10	102.41	85.7
PC-3-O6	39.90	14.29	0.22	46.16	0.28	0.11	100.97	85.3	14PC-13-O12	39.82	17.23	0.28	44.67	0.24	0.11	102.35	82.3
PC-3-O7	40.33	13.20	0.20	47.23	0.29	0.12	101.38	86.6	14PC-13-O13	40.67	14.35	0.21	47.09	0.22	0.09	102.62	85.5
PC-3-O8	39.83	13.01	0.21	46.60	0.24	0.19	100.08	86.6	14PC-13-O14	39.85	15.30	0.24	46.05	0.24	0.09	101.77	84.4
PC-3-O9	39.17	13.88	0.18	47.35	0.23	0.09	100.90	86.0	14PC-13-O15	39.58	17.92	0.29	43.62	0.19	0.11	101.71	81.4
PC-3-O10	38.76	12.78	0.18	46.92	0.25	0.09	98.99	86.9	14PC-13-O16	39.87	15.94	0.26	45.64	0.18	0.11	102.01	83.7
PC-3-O11	39.30	13.96	0.19	46.97	0.18	0.09	100.69	85.8	14PC-13-O17	40.38	14.00	0.20	46.87	0.30	0.09	101.83	85.8
PC-3-O12	40.14	13.00	0.19	46.93	0.23	0.09	100.57	86.7	14PC-13-O18	39.31	16.04	0.22	44.74	0.17	0.11	100.59	83.4
PC-3-O13	39.54	13.77	0.16	46.98	0.19	0.08	100.71	86.0	14PC-13-O19	40.16	14.96	0.22	46.45	0.26	0.10	102.15	84.8
PC-3-O14	40.01	14.26	0.17	47.39	0.27	0.08	102.17	85.7	14PC-13-O20	39.57	14.15	0.17	46.76	0.31	0.07	101.03	85.6
PC-3-O15	39.31	15.94	0.20	44.88	0.14	0.06	100.52	83.5									
PC-3-O16	39.75	15.61	0.22	45.80	0.24	0.11	101.74	84.1	<i>Xenoliths</i>								
PC-3-O17	39.44	16.46	0.26	44.83	0.22	0.11	101.32	83.1	14PCX-1-O1	39.84	13.05	0.21	46.80	0.28	0.12	100.30	86.6
PC-3-O18	39.23	17.19	0.37	43.57	0.15	0.10	100.61	82.0	14PCX-2-O1	37.42	16.13	0.22	45.30	0.16	0.10	99.34	83.5
PC-3-O19	39.49	15.99	0.27	45.12	0.26	0.15	101.27	83.6	14PCX-2-O2	37.32	16.44	0.23	45.26	0.12	0.10	99.48	83.2
PC-3-O20	39.67	13.78	0.16	46.53	0.29	0.09	100.52	85.9	14PCX-2-O3	37.85	16.26	0.21	45.35	0.22	0.10	99.99	83.4
PC-3-O21	39.29	13.04	0.18	47.11	0.29	0.12	100.02	86.7	14PCX-2-O4	38.04	16.01	0.23	45.36	0.12	0.08	99.84	83.6
PC-3-O22	39.74	11.91	0.18	47.46	0.25	0.10	99.64	87.8	14PCX-3-O1	37.86	15.56	0.23	45.71	0.23	0.08	99.66	84.1

Table A4 (continued).

	SiO2	FeO	MnO	MgO	NiO	CaO	Totals	Mg#		SiO2	FeO	MnO	MgO	NiO	CaO	Totals	Mg#
PC-3-O23	39.63	14.38	0.17	46.82	0.20	0.09	101.28	85.4	14PCX-3-O2	37.42	15.85	0.23	45.66	0.23	0.09	99.48	83.8
PC-3-O24	39.38	13.70	0.26	46.26	0.22	0.08	99.90	85.9	14PCX-3-O3	37.66	15.65	0.22	45.32	0.21	0.10	99.16	83.9
PC-3-O25	39.85	16.60	0.30	44.27	0.17	0.06	101.26	82.8	14PCX-4-O1	37.75	15.98	0.22	45.32	0.10	0.11	99.49	83.6
PC-3-O26	39.56	13.84	0.26	46.50	0.27	0.17	100.60	85.8	14PCX-4-O2	37.77	15.78	0.26	46.20	0.17	0.09	100.27	84.1
PC-3-O27	39.23	13.65	0.22	46.48	0.28	0.08	99.94	86.0	14PCX-4-O3	38.27	16.05	0.24	45.21	0.13	0.10	100.01	83.5
PC-3-O28	39.47	15.23	0.23	45.97	0.17	0.10	101.18	84.5	14PCX-5-O1	39.79	14.25	0.19	46.15	0.24	0.10	100.71	85.4
PC-3-O29	39.68	14.25	0.22	46.55	0.32	0.13	101.15	85.5	14PCX-5-O2	39.62	14.46	0.22	45.40	0.20	0.08	99.98	85.0
PC-3-O30	40.22	17.39	0.25	44.50	0.22	0.10	102.68	82.2	14PCX-5-O3	39.83	14.20	0.21	45.88	0.18	0.09	100.39	85.3
PC-3-O31	38.90	17.16	0.26	44.40	0.15	0.10	100.97	82.3	14PCX-6-O1	37.55	15.09	0.23	45.97	0.11	0.08	99.02	84.6
PC-3-O32	39.34	17.75	0.29	44.43	0.23	0.07	102.12	81.8	14PCX-6-O2	37.54	15.35	0.22	45.64	0.18	0.07	99.00	84.3
PC-3-O33	39.87	15.24	0.21	46.07	0.20	0.10	101.71	84.5	14PCX-6-O3	37.89	14.94	0.21	45.76	0.22	0.10	99.12	84.6
PC-3-O34	39.90	13.10	0.18	47.74	0.25	0.07	101.23	86.8	14PCX-7-O1	38.19	14.92	0.22	45.83	0.19	0.08	99.44	84.7
PC-3-O35	40.49	14.35	0.22	46.92	0.29	0.09	102.36	85.5	14PCX-7-O2	38.69	15.29	0.20	45.51	0.17	0.08	99.94	84.3
PC-3-O36	40.70	14.97	0.18	46.63	0.18	0.09	102.75	84.9	14PCX-7-O3	38.82	14.46	0.19	46.37	0.20	0.09	100.13	85.2
PC-3-O37	40.04	17.89	0.24	44.67	0.16	0.09	103.09	81.8	14PCX-7-O4	37.76	14.34	0.19	46.01	0.18	0.10	98.58	85.2
PC-3-O38	40.65	14.61	0.26	47.20	0.25	0.09	103.06	85.3	14PCX-8-1-O1	38.53	14.69	0.20	46.36	0.21	0.10	100.09	85.0
PC-3-O39	39.96	13.53	0.19	47.26	0.23	0.09	101.27	86.3	14PCX-8-1-O2	38.54	14.58	0.21	46.16	0.22	0.07	99.78	85.1
PC-3-O40	39.86	15.88	0.21	46.02	0.21	0.07	102.25	83.9	14PCX-8-1-O3	37.47	14.42	0.20	45.86	0.14	0.08	98.17	85.1
PC-3-O41	40.15	13.99	0.18	47.76	0.17	0.09	102.34	86.0	14PCX-8-1-O4	38.92	14.41	0.22	46.30	0.13	0.09	100.06	85.3
PC-3-O42	40.51	15.19	0.23	46.52	0.23	0.09	102.78	84.6	14PCX-10-O1	38.13	13.36	0.20	47.63	0.23	0.10	99.66	86.5
PC-3-O43	38.66	21.36	0.27	41.24	0.08	0.16	101.77	77.7	14PCX-10-O2	38.87	13.48	0.19	47.60	0.21	0.10	100.45	86.4
PC-3-O44	39.28	20.45	0.42	41.89	0.13	0.15	102.32	78.7	14PCX-10-O3	38.59	13.09	0.18	47.91	0.20	0.10	100.07	86.8
PC-3-O45	40.17	14.05	0.22	47.41	0.33	0.08	102.26	85.9	14PCX-10-O4	38.53	13.26	0.20	47.63	0.24	0.10	99.96	86.6
PC-3-O46	39.90	13.70	0.19	47.45	0.28	0.09	101.61	86.2	14PCX-11	37.89	20.33	0.27	42.61	0.14	0.12	101.36	79.0
PC-3-O47	40.64	13.63	0.18	48.02	0.28	0.09	102.84	86.4	14PCX-13-1-O1	36.14	16.42	0.23	44.72	0.17	0.09	97.76	83.1
12PC-2-O1	40.51	13.27	0.19	47.54	0.31	0.10	101.93	86.6	14PCX-13-1-O2	37.81	16.55	0.22	45.45	0.20	0.08	100.32	83.2
12PC-2-O2	39.90	17.17	0.22	44.63	0.20	0.08	102.20	82.4	14PCX-13-2-O1	37.86	15.32	0.24	46.16	0.21	0.07	99.86	84.4
12PC-2-O3	38.90	18.63	0.26	42.99	0.14	0.10	101.03	80.6	14PCX-15-O1	37.94	15.21	0.24	46.15	0.24	0.09	99.86	84.5
12PC-2-O4	39.76	13.93	0.19	46.68	0.22	0.09	100.87	85.8	14PCX-15-O2	38.55	15.55	0.22	46.72	0.27	0.08	101.40	84.4
12PC-2-O5	39.60	15.04	0.21	45.96	0.20	0.10	101.12	84.6	14PCX-15-O3	37.97	15.38	0.23	46.35	0.26	0.09	100.30	84.4
12PC-2-O6	39.82	15.59	0.20	45.58	0.23	0.13	101.55	84.0	14PCX-15-O4	37.77	15.45	0.21	46.20	0.24	0.08	99.96	84.3

Table A4 (continued).

	SiO2	FeO	MnO	MgO	NiO	CaO	Totals	Mg#		SiO2	FeO	MnO	MgO	NiO	CaO	Totals	Mg#
12PC-2-O7	40.40	10.91	0.16	49.36	0.23	0.09	101.16	89.1	14PCX-16-O1	34.93	24.45	0.42	38.11	0.04	0.12	98.07	73.7
12PC-2-O8	39.45	20.01	0.27	42.01	0.18	0.10	102.02	79.1	14PCX-17-O1	38.07	14.16	0.20	47.09	0.22	0.08	99.82	85.7
12PC-2-O9	39.79	15.48	0.23	45.73	0.23	0.08	101.54	84.2	14PCX-17-O2	38.69	13.22	0.20	47.64	0.16	0.07	99.99	86.6
12PC-2-O10	40.37	13.56	0.20	47.08	0.33	0.08	101.63	86.2	14PCX-17-O3	40.35	12.79	0.21	47.31	0.17	0.07	101.06	86.9
12PC-2-O11	39.29	18.98	0.28	42.58	0.16	0.10	101.39	80.2	14PCX-17-O4	40.17	13.78	0.22	46.76	0.19	0.09	101.37	85.9
12PC-2-O12	39.92	15.61	0.22	45.45	0.24	0.09	101.53	84.0	14PCX-17-O5	39.91	13.42	0.20	46.89	0.20	0.04	100.72	86.3

Table A5. Major element compositions of phenocryst and xenolith clinopyroxene.

	SiO2	TiO2	Al2O3	FeO	Cr2O3	MnO	MgO	CaO	Na2O	Totals	Mg#		SiO2	TiO2	Al2O3	FeO	Cr2O3	MnO	MgO	CaO	Na2O	Totals	Mg#
<i>Phenocrysts</i>												14PCX-4-P4	51.46	0.69	5.57	5.76	0.83	0.14	17.52	17.69	1.07	100.77	84.6
PC-1-P1	50.10	0.73	7.23	4.55	0.73	0.16	15.28	20.57	1.02	100.38	85.8	14PCX-5-P1	49.51	0.75	6.89	4.67	0.70	0.12	15.55	19.96	0.82	99.08	85.7
PC-1-P2	49.98	0.70	6.78	4.52	0.79	0.12	15.70	18.53	0.77	97.89	86.2	14PCX-5-P2	50.83	0.68	6.49	4.76	0.60	0.14	15.91	19.89	0.99	100.27	85.8
PC-1-P3	50.22	0.56	6.80	5.07	0.83	0.11	19.38	16.16	0.67	99.79	87.3	14PCX-5-P3	50.21	0.70	6.33	4.75	0.61	0.16	15.66	19.98	0.85	99.30	85.6
PC-1-P4	50.54	0.65	6.92	4.49	0.87	0.10	17.14	18.09	0.68	99.48	87.3	14PCX-5-P4	49.93	0.82	6.69	4.46	0.72	0.13	15.21	20.65	0.91	99.60	86.0
PC-1-P5	51.72	0.59	5.50	4.88	0.94	0.15	16.07	20.24	0.82	100.96	85.6	14PCX-6-P1	49.66	0.97	7.55	5.93	0.44	0.15	15.13	20.16	0.85	100.84	82.1
PC-1-P6	51.58	0.35	5.79	6.21	0.17	0.19	14.64	21.11	1.00	101.03	80.9	14PCX-6-P2	49.86	0.80	7.26	6.14	0.28	0.12	15.68	19.87	0.72	100.75	82.1
PC-1-P7	51.44	0.45	5.45	5.69	0.13	0.16	14.54	19.74	0.96	98.55	82.1	14PCX-6-P3	49.51	1.01	7.66	4.90	0.63	0.12	15.20	19.69	0.77	99.48	84.8
PC-2-P1	52.31	0.56	5.39	5.18	0.76	0.15	15.24	21.59	1.00	102.21	84.1	14PCX-6-P4	48.92	0.90	7.63	4.99	0.50	0.15	15.13	19.69	0.96	98.87	84.5
PC-2-P2	51.99	0.39	5.05	5.44	0.63	0.16	15.00	21.52	0.94	101.13	83.2	14PCX-7-P1	50.38	0.72	6.50	5.79	0.34	0.14	15.91	19.97	0.74	100.51	83.2
PC-2-P3	51.83	0.49	5.12	5.27	0.96	0.16	15.09	21.26	1.04	101.20	83.8	14PCX-7-P2	50.17	0.77	6.92	5.67	0.38	0.13	15.74	19.97	0.85	100.61	83.3
PC-2-P4	50.81	0.54	6.45	5.79	0.50	0.17	14.55	20.96	1.07	100.87	81.9	14PCX-7-P3	50.32	0.63	6.25	5.15	0.34	0.11	16.04	19.60	0.78	99.29	84.9
PC-2-P5	51.71	0.54	5.48	5.48	0.70	0.16	14.94	21.29	0.94	101.23	83.1	14PCX-7-P4	49.96	0.72	6.70	4.81	0.29	0.13	15.49	19.24	0.80	98.22	85.3
PC-2-P6	51.69	0.52	5.14	5.01	0.67	0.13	15.29	21.39	0.91	100.75	84.6	14PCX-8-1-P1	49.92	0.65	6.99	5.18	0.73	0.13	15.81	19.55	0.61	99.54	84.6
PC-2-P7	53.44	0.36	4.68	6.36	0.55	0.19	18.45	16.97	0.84	101.90	83.9	14PCX-8-1-P2	49.03	0.99	7.21	4.57	0.86	0.11	15.41	20.26	0.87	99.28	85.9
PC-2-P8	52.78	0.44	5.32	5.45	0.72	0.18	15.12	21.41	0.98	102.38	83.3	14PCX-8-1-P3	49.61	0.73	7.25	5.22	0.50	0.12	15.68	20.02	0.71	99.86	84.4
PC-2-P9	52.40	0.51	5.67	6.66	1.03	0.15	16.66	18.66	0.81	102.58	81.8	14PCX-8-1-P4	49.86	0.68	6.23	5.41	1.01	0.13	15.62	19.77	0.99	99.76	83.9
PC-3-P1	49.85	0.48	5.64	5.48	0.48	0.17	15.10	20.89	0.97	99.05	83.2	14PCX-8-2-P1	47.62	1.74	8.06	7.78	0.02	0.14	15.24	17.70	0.86	99.18	77.9
PC-3-P2	51.46	0.31	4.57	5.13	0.69	0.13	15.62	20.99	0.96	99.86	84.6	14PCX-8-2-P2	46.86	1.91	9.00	7.45	0.01	0.16	13.89	18.88	0.98	99.13	77.0
PC-3-P3	51.46	0.45	5.16	5.49	0.55	0.15	15.49	21.12	0.97	100.84	83.5	14PCX-8-2-P3	47.47	1.79	8.74	6.84	0.01	0.17	14.01	18.63	1.06	98.66	78.7
PC-3-P4	51.33	0.53	5.12	5.36	0.90	0.16	15.25	21.32	0.83	100.81	83.7	14PCX-9-P1	51.36	0.73	6.07	4.80	0.24	0.13	16.45	19.95	0.76	100.52	86.0
PC-3-P5	51.20	0.34	4.61	5.54	0.52	0.18	15.35	20.84	0.95	99.53	83.3	14PCX-9-P2	51.54	0.68	5.71	5.48	0.25	0.12	17.00	18.83	0.89	100.55	84.8
PC-3-P6	51.74	0.53	5.54	5.33	0.49	0.17	14.95	21.16	0.95	100.87	83.5	14PCX-9-P3	50.64	0.85	6.09	5.25	0.28	0.12	15.90	19.89	0.84	99.89	84.5
PC-3-P7	50.97	0.57	5.48	4.70	0.56	0.17	14.84	21.28	0.90	99.46	85.0	14PCX-10-P1	50.37	0.60	5.64	4.13	1.15	0.14	16.56	19.62	0.80	99.04	87.8
PC-3-P8	51.85	0.44	4.99	6.38	0.55	0.19	18.59	17.54	0.68	101.22	84.0	14PCX-10-P2	50.44	0.47	5.20	4.29	1.29	0.10	16.19	19.77	1.07	98.84	87.2
PC-3-P9	50.40	0.56	5.18	5.00	0.71	0.11	15.01	21.56	0.88	99.42	84.4	14PCX-11-P1	48.50	1.52	8.34	6.90	0.04	0.16	14.43	18.41	1.02	99.28	79.0
PC-3-P10	51.35	0.37	4.94	5.33	0.42	0.16	15.35	21.01	0.93	99.86	83.8	14PCX-12-1-P1	50.82	0.59	5.93	4.95	0.53	0.14	16.22	19.84	1.23	100.27	85.5
PC-3-P11	50.54	0.53	4.89	5.29	0.36	0.09	15.22	21.14	0.85	98.92	83.8	14PCX-12-1-P2	51.29	0.54	8.91	4.53	0.53	0.13	13.25	18.87	1.44	99.50	84.1
14PC-4-P1	51.54	0.67	6.17	4.52	1.09	0.12	16.21	20.28	0.75	101.34	86.6	14PCX-12-2-P1	50.70	0.74	7.30	4.78	0.77	0.15	16.65	19.31	0.78	101.19	86.3
14PC-4-P2	49.92	1.00	8.61	4.75	0.59	0.14	14.66	20.72	0.93	101.35	84.8	14PCX-12-2-P2	49.55	0.87	8.21	4.77	0.76	0.12	14.89	20.17	0.90	100.28	84.9
14PC-4-P3	50.69	1.24	5.35	5.41	0.21	0.12	15.11	22.88	0.57	101.58	83.4	14PCX-12-2-P3	49.85	0.85	7.34	4.91	0.42	0.14	16.22	19.16	0.68	99.60	85.6

Table A5 (continued).

	SiO2	TiO2	Al2O3	FeO	Cr2O3	MnO	MgO	CaO	Na2O	Totals	Mg#		SiO2	TiO2	Al2O3	FeO	Cr2O3	MnO	MgO	CaO	Na2O	Totals	Mg#
14PC-4-P4	50.20	0.76	7.22	4.61	0.75	0.09	15.85	20.35	0.74	100.60	86.1	14PCX-12-2-P4	50.24	0.78	7.12	5.02	0.87	0.13	15.73	19.09	1.02	100.06	84.9
14PC-4-P5	52.05	0.52	5.51	4.36	1.17	0.13	16.59	20.53	0.68	101.58	87.3	14PCX-13-1-P1	50.77	0.57	5.77	4.80	0.93	0.13	16.42	20.02	0.70	100.11	86.0
14PC-4-P6	51.09	0.77	6.87	4.55	1.02	0.12	16.47	19.93	0.69	101.51	86.7	14PCX-13-1-P2	51.68	0.45	4.92	4.24	0.86	0.14	15.82	20.89	0.90	99.98	87.0
14PC-4-P7	51.57	0.57	6.14	4.38	1.04	0.10	15.84	20.96	0.75	101.37	86.7	14PCX-13-1-P3	49.80	1.09	6.41	4.86	0.90	0.11	15.11	22.34	0.47	101.10	84.8
14PC-4-P8	51.03	0.64	6.01	4.99	0.97	0.13	15.90	19.42	0.90	100.13	85.1	14PCX-13-2-P1	50.17	0.86	6.84	4.73	0.81	0.12	15.55	19.52	1.08	99.67	85.5
14PC-4-P9	51.89	0.56	5.57	4.77	1.18	0.14	17.31	18.83	0.76	101.04	86.7	14PCX-13-2-P2	51.12	0.60	6.05	4.56	0.97	0.13	15.41	19.96	1.05	99.83	85.9
14PC-4-P10	51.41	0.67	6.52	4.87	0.78	0.15	17.03	19.41	0.72	101.62	86.3	14PCX-13-2-P3	48.62	1.51	8.25	5.59	0.23	0.13	14.38	19.63	1.08	99.43	82.3
14PC-4-P11	52.85	0.46	5.39	4.80	1.06	0.13	17.57	19.16	0.82	102.24	86.8	14PCX-13-2-P4	49.60	1.02	7.60	4.76	0.60	0.10	15.32	20.20	0.94	100.22	85.3
<i>Xenoliths</i>												14PCX-15-P1	50.03	0.71	6.67	4.83	1.06	0.12	15.77	20.40	0.97	100.52	85.5
14PCX-1-P1	50.23	0.82	6.76	5.01	0.85	0.10	16.84	17.90	0.80	99.43	85.8	14PCX-15-P2	48.61	1.44	5.92	5.12	0.57	0.12	15.06	22.19	0.37	99.47	84.1
14PCX-2-P1	49.16	1.12	8.41	6.30	0.27	0.16	14.87	19.81	0.86	100.97	80.9	14PCX-15-P3	50.12	0.54	6.45	4.64	1.07	0.12	16.86	19.08	0.64	99.57	86.7
14PCX-2-P2	48.71	1.27	8.05	5.67	0.24	0.14	14.37	19.58	0.90	99.00	82.0	14PCX-15-P4	49.85	0.66	6.20	4.73	1.03	0.11	16.20	19.82	0.66	99.31	86.0
14PCX-2-P3	49.91	1.01	7.55	6.55	0.09	0.14	15.30	19.52	0.93	101.09	80.8	14PCX-16-P1	47.31	2.00	9.47	7.40	0.02	0.22	12.87	19.24	1.22	99.69	75.8
14PCX-2-P4	49.05	0.93	8.31	7.02	0.16	0.15	15.11	19.59	0.69	101.04	79.5	14PCX-16-P2	48.63	1.02	12.12	6.29	0.02	0.16	11.05	16.15	4.25	99.75	76.0
14PCX-3-P1	49.12	1.09	8.16	5.40	0.35	0.14	14.43	20.13	0.96	99.84	82.8	14PCX-16-P3	46.96	1.86	9.72	8.33	0.00	0.23	12.21	19.37	1.20	99.83	72.5
14PCX-3-P2	49.85	0.97	5.92	4.96	0.42	0.12	15.68	21.76	0.53	100.26	85.0	14PCX-17-P1	50.75	0.60	5.77	4.58	1.03	0.10	15.59	20.87	0.89	100.19	86.0
14PCX-3-P3	48.89	0.91	7.97	5.32	0.25	0.13	15.14	20.33	0.83	99.81	83.7	14PCX-17-P2	51.12	0.53	5.06	4.96	0.88	0.11	16.17	19.88	0.90	99.57	85.4
14PCX-3-P4	49.08	1.05	6.36	5.23	0.52	0.13	15.62	20.34	0.78	99.18	84.3	14PCX-18-P1	51.35	0.48	5.29	4.60	1.01	0.12	16.37	20.42	0.80	100.51	86.5
14PCX-4-P1	49.14	1.17	8.22	6.23	0.11	0.17	15.39	18.98	0.84	100.24	81.6	14PCX-18-P2	50.19	0.75	7.33	5.43	0.44	0.10	15.63	19.45	0.78	100.14	83.8
14PCX-4-P2	48.80	1.35	8.00	5.54	0.20	0.16	14.71	19.24	0.95	99.06	82.7	14PCX-18-P3	50.78	0.63	6.44	5.65	0.85	0.13	18.40	16.36	0.67	99.97	85.4
14PCX-4-P3	48.45	1.73	8.11	6.16	0.24	0.15	14.87	19.58	0.93	100.30	81.3	14PCX-18-P4	49.65	1.15	8.27	5.49	0.46	0.14	14.94	20.02	1.06	101.22	83.0

Table A6. Clinopyroxene trace element compositions of PC phenocrysts and xenoliths. The average of repeatedly measured secondary standard BCR-2 is also reported.

	PC-1-1	PC-1-4	PC-3-1	PC-3-2	PC-3-3	14PCX-6-1	14PCX-6-2	14PCX-8-2-1	14PCX-8-2-2	14PCX-9-1	14PCX-9-2	BCR-2
Nb	0.0	0.0	0.4	0.1	0.3	0.2	0.2	0.2	0.2	0.1	0.2	11.7
La	1.2	1.1	17.3	15.3	11.5	8.4	3.6	3.7	3.6	6.8	6.1	23.1
Ce	5.4	5.4	54.2	53.2	34.1	36.3	14.6	17.0	16.6	25.7	25.2	50.4
Pb	0.0	0.0	0.7	0.5	0.6	0.3	0.3	0.2	0.2	0.2	0.2	10.3
Sr	64.5	65.0	228.9	209.0	208.2	163.4	110.2	136.1	140.8	235.0	205.5	321.3
Nd	6.8	6.4	31.5	34.9	20.8	34.5	14.4	19.1	19.9	20.3	19.4	26.6
Hf	0.8	0.8	1.1	1.5	0.8	4.5	1.2	1.8	1.9	1.4	1.3	4.3
Zr	22.6	22.8	47.5	54.6	28.0	126.1	30.3	40.4	44.7	46.9	44.3	169.0
Sm	2.1	2.1	5.6	6.4	4.2	8.0	4.1	5.7	6.2	4.2	4.1	6.2
Eu	0.8	0.8	1.6	1.8	1.1	2.2	1.2	1.8	1.9	1.2	1.2	1.8
Gd	2.3	2.3	3.9	4.4	2.8	6.2	3.5	5.1	5.6	3.2	3.1	6.0
Dy	2.5	2.3	3.1	3.2	2.3	5.0	3.0	4.5	4.9	2.5	2.4	6.0
Er	1.3	1.2	1.6	1.5	1.1	2.5	1.5	2.2	2.3	1.3	1.2	3.4
Y	12.3	11.7	14.6	14.9	10.5	22.7	13.3	20.0	21.5	11.5	11.0	31.0
Yb	1.1	1.2	1.4	1.4	0.9	2.1	1.2	1.8	1.8	1.0	1.0	3.1
Lu	0.2	0.2	0.2	0.2	0.1	0.3	0.2	0.2	0.2	0.1	0.1	0.5

Table A7. K_d value used for melt composition calculation.

cpx Mg#	T (°C)	La	Ce	Nd	Sm	Eu	Gd	Dy	Er	Yb	Lu
77	1090	0.080	0.146	0.282	0.384	0.466	0.463	0.507	0.520	0.504	0.493
79	1110	0.076	0.139	0.268	0.365	0.444	0.440	0.481	0.494	0.478	0.468
81	1130	0.069	0.126	0.241	0.328	0.398	0.394	0.430	0.441	0.428	0.419
84	1150	0.068	0.123	0.235	0.319	0.388	0.384	0.419	0.430	0.417	0.408
86	1170	0.062	0.112	0.214	0.290	0.352	0.348	0.380	0.390	0.378	0.371
87.5	1190	0.060	0.108	0.205	0.277	0.337	0.333	0.363	0.372	0.362	0.355
88	1200	0.058	0.106	0.201	0.271	0.329	0.325	0.355	0.364	0.354	0.347

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