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2 **Evidence of rapid phenocryst growth of olivine during ascent in basalts from the Big Pine volcanic**
3 **field: Application of olivine-melt thermometry and hygrometry at the liquidus**

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10 **Key Points:**

11 - Rapid phenocryst growth occurs during ascent in Mg-rich basalts (some carry mantle
12   xenoliths) from the Big Pine volcanic field, CA.

13 - The most Mg-rich olivine can be paired with the whole-rock composition to apply
14   olivine-melt thermometry/hygrometry at the liquidus.

15 - Large, high-quality data sets on the temperature and water content of basalts from various
16   tectonic settings can be obtained by this method.

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Abstract

The Quaternary Big Pine (BP) volcanic field in eastern California is notable for the occurrence of mantle xenoliths in several flows. This points to rapid ascent of basalt through the crust and precludes prolonged storage in a crustal reservoir. In this study, the hypothesis of phenocryst growth during ascent is tested for several basalts (13-7 wt% MgO) and shown to be viable. Phenocrysts of olivine and clinopyroxene frequently display diffusion-limited growth textures, and clinopyroxene compositions are consistent with polybaric crystallization. When the most Mg-rich olivine in each sample is paired with the whole-rock composition, resulting $^{Fe^{2+}-Mg}K_D$ (olivine-melt) values (0.31-0.36) match those calculated from literature models (0.32-0.36). Application of a Mg- and a Ni-based olivine-melt thermometer from the literature, both calibrated on the same experimental dataset, lead to two sets of temperatures that vary linearly with whole rock MgO wt%. Because the Ni-thermometer is independent of water content, it provides the actual temperature at the onset of olivine crystallization (1247-1097°C), whereas the Mg-thermometer gives the temperature under anhydrous conditions and thus allows ΔT (= T_{Mg}-T_{Ni} = depression of liquidus due to water) to be obtained. The average ΔT for all samples is ~59 °C, which is consistent with analyzed water contents of 1.5-3.0 wt% in olivine-hosted melt inclusions from the literature. Because the application of olivine-melt thermometry/hygrometry at the liquidus only requires microprobe analyses of olivine combined with whole-rock compositions, it can be used to obtain large global datasets of the temperature and water contents of basalts from different tectonic settings.

Plain Language Summary

Basaltic lavas are a window into their mantle source regions, which is why it is important to determine their temperatures and water contents. In this study, a new approach that allows these two parameters to be quantified is demonstrated for basalts from the Big Pine volcanic field, CA. They were targeted because many contain chunks of dense mantle rocks, which precludes storage in a crustal magma chamber and points to direct ascent from the mantle to the surface along fractures. Two hypotheses are proposed, tested, and shown to be viable in this study: (1) olivine crystallized in the basalts during ascent, and (2) the most Mg-rich olivine analyzed in each basalt represents the first olivine to grow during ascent. This enables the most Mg-rich olivine to be paired with the whole-rock composition in the application of olivine-melt thermometry and hygrometry. The results match those from published, independent studies. The success of this approach paves the way for the attainment of large, high-quality data sets for basalts from a wide variety of tectonic settings. This, in turn, may allow global variations in mantle temperature and volatile content to be mapped in greater detail and better understood.

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1 Introduction

High-Mg basalts are close approximations of partial melts of the mantle and have the potential to provide insights into mantle conditions at the time of melt segregation, including temperature, oxidation state, and water content. For example, numerous researchers have used high-Mg basalts to estimate mantle source temperatures (e.g., McKenzie & Bickle, 1988; Asimow et al., 2001; Putirka, 2005; Putirka et al., 2007; Herzberg & Asimow, 2008; Plank & Forsyth, 2016; Putirka et al., 2018). These efforts require a priori knowledge regarding: (i) mantle mineralogy, (ii) oxidation state, and (iii) the olivine-liquid Fe^{2+}-Mg exchange coefficient ($^{Fe2+-Mg}K_D$ (olivine-melt)) under melt equilibration/segregation conditions. A complementary approach is undertaken in this study, which is to constrain the temperature and water content of a basaltic melt *after* melt segregation, at the onset of olivine crystallization during ascent.

The primary question that is explored in this study is whether phenocryst growth in erupted basalts occurs during ascent along fractures, and if evidence of their rapid crystallization can be readily identified through a series of tests. If so, a second question is whether the most Mg-rich olivine in each basalt can be used as a close approximation of the first olivine to crystallize from a melt with a composition similar to that of the whole rock. This study explores these two hypotheses and, in cases where they are shown to be viable, the most Mg-rich olivine in each sample is paired with the whole-rock composition to apply olivine-melt thermometry and hygrometry at the onset of olivine growth. Additionally, for basalts that are closely co-saturated with clinopyroxene at the onset of phenocryst growth, clinopyroxene-melt thermobarometry is also applied. For those basalts that contain plagioclase phenocrysts, minimum constraints on melt water contents at the onset of plagioclase growth are obtained.

In order to test the hypotheses of phenocryst growth during ascent and whether olivine-melt thermometry can be applied to the most Mg-rich olivine analyzed in each sample, basalts from the Quaternary Big Pine (BP) volcanic field in eastern California were targeted for four key reasons. First, the BP basalts are young (many are <500 ka; Blondes et al., 2007, 2008; Vazquez & Woolford, 2015), fresh, and well characterized in terms of their isotopic (e.g. Beard & Glazner, 1995; Beard & Johnson, 1997; DePaolo & Daley, 2000; Blondes et al., 2008) and major- and trace-element compositions (e.g. Ormerod et al., 1991 Blondes et al., 2008; Gazel et al., 2012; Putirka et al., 2012). Second, the presence of mantle xenoliths in several flows (e.g., Wilshire et al., 1988; Beard & Glazner, 1995; Ducea & Saleeby, 1996, Lee et al., 2001) indicates that their ascent through the crust was sufficiently rapid to overcome xenolith settling velocities (e.g. Sparks et al., 1977; Spera, 1980, 1984), which is consistent with phenocryst growth during ascent and not during prolonged storage in a crustal reservoir. Third, Fe^{3+}/Fe^T (Fe^T = total iron) ratios in olivine-hosted melt inclusions in BP basalts are known from μ-XANES and V-partitioning analyses (Kelley & Cottrell, 2012; Gazel et al., 2012), which enables a test of whether the most Mg-rich olivine in each sample closely represents the first olivine to crystallize from the whole-rock liquid composition. Fourth, high-quality, published analyses of the water contents (1.5-3.0 wt%) in olivine-hosted melt inclusions in the BP basalts (Gazel et al., 2012) allow a test of the method applied in this study to constrain melt H_2O contents.

2 Geological Setting

The Big Pine volcanic field, which is located in Owens Valley, California along the eastern escarpment of the Sierra Nevada mountain range, sits along the western margin of the

Great Basin, which has been undergoing lithospheric thinning (i.e., Basin and Range extension) since 12 Ma (e.g., Henry & Perkins, 2001; Stockli et al., 2003; Colgan et al., 2006). Prior to extension, the region was affected by Mesozoic subduction of the Farallon slab beneath the North American Plate (e.g., Christiansen & Lipman, 1972; Lipman et al., 1972; Snyder et al., 1976). The BP volcanic field resides on Precambrian continental lithosphere, east of the boundary with the Panthalassan terrane to the west (Kistler, 1990).

The BP volcanic field contains >20 Quaternary monogenetic vents (Fig. 1) that erupted high-MgO ($\leq$ 13 wt%) basalt; several flows contain mantle xenoliths (Wilshire et al., 1988; Beard & Glazner, 1995; Ducea & Saleeby, 1996; Lee et al., 2001). Mantle xenolith localities from Wilshire et al. (1998) and Beard and Glazner (1995) are shown in Figure 1. The BP volcanic field has been active since 1.2 Ma (Blondes et al., 2007, 2008), with the youngest eruptions (17, 27 and 40 ka) identified using cosmogenic ^{36}Cl dating (Vazquez & Woolford, 2015). All BP basalts, regardless of age, have chondrite-normalized trace-element patterns that display enrichments in fluid-mobile elements (e.g. Ba), which is consistent with a subduction modified lithospheric mantle source (e.g., Ormerod et al., 1988, 1991; Blondes et al., 2008; Gazel et al., 2012; Putirka et al., 2012). Calculated depths of melt segregation in the literature range from 44-74 km (Putirka et al., 2012) to 30-80 km (Gazel et al., 2012).

3 Sample Locations and Petrography

Locations of the ten basalt samples collected and examined in this study are shown in Figure 1; all are from eruptive vents <500 ka (Blondes et al., 2007, 2008; Vazquez & Woolford, 2015). In this study, a mantle xenolith (4 cm diameter) was discovered in sample BP-11 (Fig. 1). An image of the BP-11 mantle xenolith is presented in Figure S1 (supplementary information). In two cases, two separate samples were obtained from different parts of the same flow (BP-11 and BP-24; BP-16 and BP-30; Fig. 1). The objective of sampling two different parts of the same flow was to evaluate any variation in composition (whole-rock and phenocryst assemblage).

Phenocrysts of olivine ($\leq$ 2.7 mm) and clinopyroxene ($\leq$1.8 mm) are present in basalts with >9 wt% MgO, with plagioclase sparse to absent. The only exception is BP-7, where clinopyroxene is only found as microlites ($\leq$100 μm). Conversely, olivine and plagioclase are the dominant phenocrysts in the four low-MgO (7-8 wt%) basalts, with clinopyroxene sparse to absent. Olivine is the only phenocryst to contain inclusions of chromite. Titanomagnetite is present in the groundmass of all samples and occurs as inclusions in rims of zoned (sector/oscillatory) clinopyroxene phenocrysts. Ilmenite is only found in three samples (BP-7, BP-20, BP-8), and is restricted to the groundmass.

Samples were evaluated for post-eruptive alteration/oxidation through thin section examination. Two criteria were employed: (1) the presence of any iddingsite along the rims of olivine, and (2) the presence of any alteration products in vesicles (for samples obtained along flow tops). Seven of the ten BP samples are extremely fresh, with no evidence of any iddingsite rims on olivine. The remaining three (BP-7, -8, and -23) have trace to minor occurrences of iddingsite on some olivine crystals. Unfortunately, the sample that contains the mantle xenolith (BP-11; taken from a vesicular flow top) has alteration products in its vesicles, despite its pristine olivine (i.e., no trace of iddingsite); alteration in vesicles was not visible in the other samples.

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154 **4 Whole-rock basalt compositions**
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156 4.1 Analytical methods
157
158 Samples were crushed and powdered in a tungsten carbide shatter box and analyzed for
159 major- and trace-elements. Efforts were made to avoid vesicle material in sample BP-11 when
160 powdering rock fragments. Major-and trace-element compositions were determined using
161 inductively coupled plasma mass spectrometry (ICP-MS) at Activation Laboratories, Ontario,
162 Canada. Samples were analyzed for bulk Ni content by single-element, pressed pellet X-ray
163 fluorescence (XRF) at Activation Laboratories, Ontario, Canada. For each sample, whole-rock
164 analyses of FeO wt% (Fe^{2+}) were obtained in duplicate by the Wilson (1960) titration method at
165 the University of Michigan. Additionally, four FeO titration analyses were made on USGS
166 (United States Geological Survey) standard W-2a to evaluate accuracy and reproducibility.
167
168 4.2 Results
169
170 All major-element oxide components are normalized to 100% and are presented in Table
171 1 with original totals and original loss on ignition (LOI). The BP basalts have variable MgO
172 contents (7-13 wt%; Table 1) and are normative in nepheline (i.e., classify as alkali olivine
173 basalts). Whole-rock Ni contents range from 76-343 ppm and the average reproducibility on
174 certified standards is $\pm$ 4 ppm. Note that the whole-rock compositions of the sample pairs taken
175 from the same flow are very similar (BP-11 and BP-24; BP-16 and BP-30; Table 1). The full
176 suite of analyzed trace-element concentrations for all samples are presented in Table S1
177 (supplementary information). A chondrite-normalized (McDonough & Sun, 1995) trace-element
178 diagram of the BP basalts compared to average Cascade arc basalt (Reiners et al., 2000) and an
179 average Hawaiian tholeiite (Hofmann & Jochum, 1996) is presented in Figure S2 (supplementary
180 information). All samples display an arc geochemical signature (i.e., enrichment in fluid-mobile
181 large ion lithophile elements, e.g. Ba) as previously noted in the literature (e.g., Ormerod et al.,
182 1988, 1991; Blondes et al., 2008; Gazel et al., 2012; Putirka et al., 2012).
183
184 The average ($\pm$1 s.d.) of four analyses of wt% FeO (ferrous iron concentration) for the
185 USGS standard (W-2a) is 8.44 ($\pm$ 0.22) wt%, in agreement with the certified value of 8.34 $\pm$ 0.09
186 wt%. The results of individual titrations on the BP samples and the standard (W-2a) are
187 presented in Table S2 (supplementary information). The average wt% FeO analyzed for each
188 sample is given in Table 1, together with resulting whole-rock Fe^{3+}/Fe^{T} ratios. Note that the
189 xenolith-bearing sample, BP-11, was obtained from a flow top with visible alteration products in
190 its vesicles. Although attempts were made to avoid altered material during crushing and
191 powdering, the Fe^{3+}/Fe^{T} value for BP-11 may reflect the effects of alteration, and is not
192 considered a magmatic value. Similarly, for BP-16 (also a highly vesicular sample), the
193 duplicate wt% FeO analyses were different outside of expected analytical error (1 s.d. = $\pm$ 0.51
194 wt% FeO), which points to heterogeneity that may reflect the presence of some vesicle alteration
195 material; its Fe^{3+}/Fe^{T} value was likely affected. For all other samples, the Fe^{3+}/Fe^{T} values based
196 on whole-rock analyses range from 0.17-0.30, with an average ($\pm$ 1σ) value of 0.24 ($\pm$ 0.04).
197 Therefore, within analytical error, the unaltered whole-rock Fe^{3+}/Fe^{T} ratios are close to those
198 obtained from olivine-hosted melt inclusions in the BP basalts by direct μ-XANES analyses

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199 (0.24-0.25; Kelley & Cottrell, 2012) and V-partioning calcualtions (0.23-0.30; Gazel et al.,
200 2012).
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202 **5 Mineral compositions**
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204   5.1 Electron microprobe analytical methods
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206   Compositional analyses of olivine, clinopyroxene, plagioclase, and Fe-Ti oxides were
207 obtained using a Cameca SX-100 electron microprobe at the Robert B. Mitchell Electron
208 Microbeam Analysis Lab at the University of Michigan. The microprobe standards used for
209 each element in each phase are presented in Table S3 (supplementary information).
210
211   For olivine, two beam conditions were used to analyze eight elements (Mg, Al, Si, Ca,
212 Cr, Mn, Fe, Ni). In the first set, an accelerating voltage of 15 kV, a focused beam of 20 nA, and
213 individual peak and background counting times of 20 seconds were used. The 1σ uncertainty for
214 each oxide based on counting statistics is $\pm$ 0.45 wt% SiO_2, $\pm$ 0.35 wt% MgO, $\pm$ 0.56 wt% FeO^T,
215 $\pm$ 0.05 wt% NiO, $\pm$ 0.03 wt% Al_2O_3, $\pm$ 0.03 wt% CaO, $\pm$ 0.04 wt% Cr_2O_3, and $\pm$ 0.11 wt% MnO.
216 Analytical transects across 574 different olivine crystals (30 to 105 per sample)($\leq$2.7 mm) were
217 performed. Data were obtained every 20 to 40 μm, yielding between 266 and 1328 olivine
218 analyses per sample.
219
220   In the second set of beam conditions, designed to increase the precision of the Ni
221 analyses, an accelerating voltage of 20 kV and a 40 nA current was used. Background and peak
222 counting times were 20 seconds for Al, Mn, Cr, and Ca, whereas peak and background counting
223 times of 30 and 20 seconds, respectively, were used for Si, Mg, Fe, and Ni. The 1σ uncertainty
224 for each oxide based on counting statistics is $\pm$ 0.36 wt% SiO_2, $\pm$ 0.24 wt% MgO, $\pm$ 0.24 wt%
225 FeO^T, $\pm$ 0.02 wt% NiO and less than <0.1 wt% for Cr_2O_3, Al_2O_3, MnO, and CaO. This second
226 set of beam conditions was used to reanalyze transects along two to three of the most Mg-rich
227 olivine crystals from each sample as well as olivine grains in the mantle xenolith from BP-11.
228 Approximately 50 analyses were obtained on olivine crystals in the mantle xenolith.
229 Additionally, a mantle xenolith olivine from the San Carlos volcanic field (AZ) was analyzed for
230 comparison. All olivine analyses were filtered to exclude those with totals outside of 99-101
231 wt%.
232
233   Analyses of clinopyroxene were obtained for the six BP basalts with >9 wt% MgO
234 (Table 1) (i.e., those with clinopyroxene phenocrysts). However, in sample BP-7, clinopyroxene
235 was only found as microlites (<100 μm) in the groundmass. Ten elements (Na, Mg, Al, Si, Ca,
236 Ti, Cr, Mn, Fe, Ni) were analyzed using an accelerating potential of 20 kV and a focused 20 nA
237 beam current with individual peak and background counting times of 20 seconds. Single
238 analytical transects were acquired for 357 individual clinopyroxene crystals (5-140 per sample).
239 Additional transects were collected for phenocrysts where complex chemical zonation (i.e.
240 oscillatory, sector, distinct cores) was observed via back-scattered electron (BSE) imaging. The
241 1σ uncertainty for each oxide based on counting statistics is $\pm$ 0.04 wt% Na_2O, $\pm$ 0.18 wt% MgO,
242 $\pm$ 0.13 wt% Al_2O_3, $\pm$ 0.40 wt% SiO_2, $\pm$ 0.24 wt% CaO, $\pm$ 0.06 wt% TiO_2, $\pm$ 0.04 wt% Cr_2O_3, $\pm$
243 0.03 wt% MnO, $\pm$ 0.12 wt% FeO^T, $\pm$ 0.03 wt% NiO. Analytical totals were filtered to exclude
244 those outside of 98.5–100.5 wt%. In one sample (BP-31), olivine inclusions in clinopyroxene

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245 were analyzed using the same beam conditions for clinopyroxene. Additional details on the
246 standards used for these olivine analyses are presented in Table S3 (supplementary information).
247
248   Plagioclase analyses were obtained on the four low-MgO basalt samples (Table 1). For
249 these analyses, a 15 kV accelerating potential with a focused 4 nA beam was used to analyze
250 nine elements (Mg, Na, Si, Al, Ti, Fe, Mn, Ca, K) with individual peak and background counting
251 times of 20 seconds. Transects of 90 plagioclase crystals (13 to 34 per sample)(185 μm – 1.4
252 mm) were performed, yielding between 237 and 415 analyses per sample. Plagioclase analyses
253 were filtered to exclude analytical totals outside 98.5 and 101 wt%. The 1σ uncertainty for each
254 oxide based on counting statistics is $\pm$ 0.23 wt% Na_2O, $\pm$ 0.06 wt% MgO, $\pm$ 0.75 wt% Al_2O_3, $\pm$
255 0.98 wt% SiO_2, $\pm$ 0.09 wt% K_2O, $\pm$0.52 wt% CaO, $\pm$ 0.08 wt% MnO, $\pm$ 0.14 wt% FeO^T, and $\pm$
256 0.06 wt% TiO_2.
257
258   Three samples from this study (BP-7, -8, -20) contain both ilmenite and titanomagnetite.
259 In all other samples, ilmenite is absent. For Fe-Ti oxide analyses, a focused beam with an
260 accelerating voltage of 15 kV and a 10 nA beam current was used. Nine elements (Si, Ti, Al, Fe,
261 V, Cr, Mn, Mg, Ca) were analyzed with individual peak and background counting times of 20
262 seconds. The ilmenite 1σ uncertainty based on counting statistics is $\pm$ 0.14 wt% MgO, $\pm$ 0.06
263 wt% Al_2O_3, $\pm$ 0.06 wt% SiO_2, $\pm$ 0.05 wt% CaO, $\pm$ 0.77 wt% V_2O_3, $\pm$0.09 wt% Cr_2O_3, $\pm$ 0.05
264 wt% MnO, $\pm$ 0.09 wt% FeO^T, and $\pm$ 0.83 wt% TiO_2. The titanomagnetite 1σ uncertainty based on
265 counting statistics is $\pm$ 0.13 wt% MgO, $\pm$ 0.10 wt% Al_2O_3, $\pm$ 0.06 wt% SiO_2, $\pm$ 0.05 wt% CaO, $\pm$
266 0.42 wt% V_2O_3, $\pm$0.09 wt% Cr_2O_3, $\pm$ 0.05 wt% MnO, $\pm$ 0.09 wt% FeO^T, and $\pm$ 1.10 wt% TiO_2.
267 Fe-Ti oxide analyses were filtered to discard those with $\geq$ 0.25 wt% SiO_2 and CaO, and totals
268 outside 98-101 wt%. FeO and Fe_2O_3 were calculated based on cation stoichiometry. Successful
269 analyses of both ilmenite and titanomagnetite were only obtained for BP-20.
270
271   5.2 X-ray intensity mapping methods (and BSE images)
272
273   X-ray intensity maps of phosphorous (P) and iron (Fe) were obtained for the most Mg-
274 rich olivine crystals in each of six BP basalts (BP-23, -31, -11, -19, -24, -7). A 20 kV
275 accelerating voltage and 200 nA beam current were utilized with dwell times of 200-300 μsec.
276 For comparison, similar maps were obtained on the olivine crystal from the San Carlos mantle
277 xenolith. In one of these samples (BP-31), X-ray intensity maps of Mg, as well as aluminum
278 (Al) and chromium (Cr) were obtained for a representative clinopyroxene phenocryst using a 15
279 kV accelerating voltage and 500 nA beam current with a dwell time of 200 μsec. Back-scattered
280 electron (BSE) images were obtained for all analyzed olivine and clinopyroxene crystals.
281
282   5.3 Olivine results
283
284   Individual analyses of olivine crystals in each BP sample (as well as those for the olivine
285 in the mantle xenolith in BP-11 and the San Carlos olivine) are given in Table S4 (supplementary
286 information) and summarized in histograms of mol% forsterite (Fig. 2). The results show a
287 continuous range in composition up to the most Mg-rich olivine in each sample (Table 2, Fig. 2).
288 Plots of analyzed NiO wt% and CaO wt% as a function of mol% forsterite (Fo) for each sample
289 are shown in Figures 3 and 4, respectively. Plots of analyzed CaO wt% (Fig. 4) show that the
290 dominant olivine population in each sample contains > 0.15 wt% CaO, whereas a sparse subset

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291   in two samples (BP-11, BP-24) contains ≤ 0.15 wt% CaO. For both NiO and CaO analyses, data
292   that span the highest three mol% Fo are fitted to a linear relationship for each sample, which is
293   used to calculate the NiO and CaO wt% in the most Mg-rich olivine. Application of the linear fit
294   reduces the scatter due to analytical error and allows a more accurate estimate of both the NiO
295   and CaO content in the most Mg-rich olivine (Table 2) in each sample.

297   In the BP-11 mantle xenolith, the average analyzed olivine composition is 88.8 ± 0.1 Fo
298   mol%, 0.32 ± 0.01 NiO wt% and 0.13 ± 0.004 wt% CaO (Table 2). Both the forsterite content
299   (Fig 2c) and the NiO wt% (Fig 3c) in olivines from the mantle xenolith overlap those observed in
300   olivine phenocrysts from BP-11, whereas the CaO contents are distinctly lower than those in the
301   phenocrysts (Fig. 4). Olivine crystals from BP mantle xenoliths from two previous studies
302   (Beard & Glazner, 1995; Lee et al., 2001), as well as olivine crystals from San Carlos mantle
303   xenoliths, are also lower in CaO relative to phenocrysts of olivine (Fig. 4c). The low-CaO
304   character of olivine from mantle peridotites has long been noted in the literature (e.g., Simkin &
305   Smith, 1970; Norman & Garcia, 1999; Housch et al., 2010). In this study, a filter of <0.15 wt%
306   CaO was applied to identify any low-Ca olivine in the BP samples (Fig. 4). The sparse
307   population of low-CaO olivine crystals (e.g., BP-11, BP-24; Fig 4) are interpreted as
308   disaggregated olivine xenocrysts from mantle xenoliths.

310   X-ray intensity maps of P and Fe for the most Mg-rich olivine crystal in each of six BP
311   basalts are shown in Figure 5. All samples contain olivine crystals that display complex P
312   zoning patterns. For comparison, a P map of an olivine from a mantle xenolith (San Carlos, AZ)
313   shows no zones of P enrichment (Fig 5n). BSE images of representative olivine phenocrysts
314   from six other BP basalts are shown in Figure 6.

316   5.4 Clinopyroxene results

318   All individual analyses of clinopyroxene from the six high-MgO (>9 wt%) BP basalts are
319   tabulated in Table S5 (supplementary information). Compositions for the most Mg-rich
320   clinopyroxene crystal in each sample are presented in Table 3. In Figure 7, plots of Na_2O (wt%)
321   vs Mg# show two distinct compositional trends, whereas plots of TiO_2 (wt%) vs. Mg# primarily
322   follow a continuous compositional range among the clinopyroxene phenocrysts in each of four
323   samples (BP-23, -31, -11, and -19). Similar data for BP-7 (no phenocrysts, only microlites) and
324   BP-24 (sampled from same flow as BP-11) are shown in Figure S3 (supplementary information).
325   The analyses in Figure 7 are given different symbols based on the texture observed for each
326   analytical spot along a transect: (1) non-spongy (non-vermicular), (2) spongy (vermicular), and
327   (3) sector- and/or oscillatory zoned. Examples of all three clinopyroxene textures from these
328   four samples are shown in BSE images (Fig. 8). In all cases, the clinopyroxene analyses with the
329   highest Na_2O content (and lowest TiO_2) content, have the non-spongy (non-vermicular) texture
330   (Fig. 7; Fig. 8a-d). The spongy (vermicular) texture is found across the compositional
331   continuum (Fig. 7; Fig. 8e-h). Importantly, most sector- and/or oscillatory zoned analyses plot in
332   the low-Na_2O and high-TiO_2 group (Fig. 7; Fig. 8i-l).

334   The relative abundance of the three clinopyroxene textures in each of the four basalt
335   samples is qualitatively illustrated in a histogram of Mg# (Fig. 9). BP-23 contains the highest
336   abundance of non-spongy, high-Na_2O and low-TiO_2 clinopyroxene, whereas BP-31 contains the

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most oscillatory- and sector-zoned (all low-Na_2O), with few analyses with the non-spongy texture. For one of the zoned clinopyroxene phenocrysts from BP-31 (cpx-1; Fig. 7c-d), X-ray intensity maps (Fig. 10) show that sector zoning strongly affects Al and Mg concentrations, but not Cr contents; in contrast, all three elements display marked oscillatory zonation in this same crystal. Note that analyses obtained on this single crystal (cpx-1) span the entire compositional range as those for all oscillatory- and sector-zoned clinopyroxene phenocrysts in BP-31 (Fig. 7c-d).

5.5 Plagioclase results

Individual plagioclase analyses of phenocrysts from the four low-MgO BP samples are tabulated in Table S6 (supplementary information). The analyses are summarized for each sample in histograms of mol% anorthite (An) in Figure S4 (supplementary information). The most calcic plagioclase composition in each sample is highlighted in Table S6. Each of the four samples contains a plagioclase phenocryst population that spans a continuous compositional range up to ~An_{75-76}.

5.6 Fe-Ti oxide results

Individual analyses of ilmenite and titanomagnetite crystals for BP-20 are presented in Table S7 (supplementary information). The Bacon and Hirschmann (1988) Mg/Mn equilibrium test was applied to all possible ilmenite-titanomagnetite pairs in order to discern those that could have co-crystallized together. The results indicate six viable equilibrium pairs, which are reported in Table S7. Temperature and oxygen fugacity values were calculated for each ilmenite-titanomagnetite pair that passed the Bacon and Hirschmann (1988) equilibrium test (Table S7) using the model of Ghiorso and Evans (2008). The thermometry results (980-840°C) are consistent with co-crystallization of both Fe-Ti oxides syn-eruptively in the groundmass. At the onset of ilmenite-titanomagnetite co-crystallization at ~980°C, the ΔNNO value was -0.45, which is consistent with a melt Fe^{3+}/Fe^T ratio intermediate between that at QFM (~0.16) and NNO (~0.21), namely ~0.19.

6 Evaluation of phenocryst growth during ascent

6.1 Mantle xenoliths in BP basalts

The BP volcanic field provides an opportunity to test the hypothesis of phenocryst growth during ascent due to the presence of mantle xenoliths in some of the basalt flows (e.g., Wilshire et al., 1988; Beard & Glazner, 1995; Ducea & Saleeby, 1996, Lee et al., 2001). The most important observation is that in order for a mantle xenolith to be carried to the surface, the ascent velocity of the host basalt must be sufficiently rapid to overcome the settling velocity of the mantle xenolith. Calculated minimum ascent rates based on Stoke's settling law (see supplementary information for details on calculation), prior to the onset of phenocryst growth (e.g., Sparks et al., 1977; Spera, 1984), range from 6 to 24 km/day (dependent on whether dissolved water contents of 1.5 or 3.0 wt% are used to calculate melt viscosity). The results are consistent with the hypothesis that phenocryst growth in the BP basalts occurred rapidly during ascent (Fig. 11), rather than slowly in a crustal magma chamber. To further test this hypothesis,

9

383   the compositions and textures of olivine and clinopyroxene phenocrysts in the BP basalts are
384   examined below.

385

386   6.2 Evidence of diffusion-limited growth textures in phenocrysts from BPVF basalts

387

388   *6.2.1 Olivine*

389

390   Many olivine phenocrysts in BP samples display melt diffusion limited growth textures
391   (e.g. hopper texture), as viewed in BSE images (Fig. 6). Diffusion-limited growth textures are
392   consistent with crystallization at an effective undercooling ($\Delta T_{UC} = T_{liquidus} - T_{melt}$) (e.g.,
393   Lofgren, 1974; Welsch et al., 2014; Shea et al., 2019). Undercooling may develop in basalts that
394   lose little dissolved water from the melt during initial ascent (i.e., were fluid undersaturated at
395   the time of melt segregation from their source, or underwent open-system degassing of a CO_2-
396   rich fluid phase). As illustrated in Figure 11, an ascending melt is expected to become super-
397   liquidus if its initial drop in temperature during ascent is less than that for the positive dP/dT
398   slope of its fluid-undersaturated liquidus. During continued ascent and cooling, the melt will
399   eventually cross its liquidus, but from a super-liquidus condition (i.e., free of pre-existing
400   nuclei), which will lead to a kinetic delay in nucleation. This, in turn, allows an undercooling
401   (ΔT_{UC}) to develop, where the melt (crystal-free) is cooler than the liquidus. Moderate ΔT_{UC}
402   values promote low nucleation rates and high crystal growth rates (e.g., Lofgren, 1974),
403   conditions that promote initial dendritic and/or hopper growth of sparse phenocrysts. The
404   complex P zoning observed in many of the BP olivine phenocrysts (Fig. 5) has been interpreted
405   as recording an initial period of rapid, diffusion-limited, dendritic growth (Milman-Barris et al.,
406   2008; Welsch et al., 2012, 2014; Shea et al., 2019). The lack of P zoning in the San Carlos
407   olivine (from mantle xenolith) is consistent with a different growth history that did not include
408   rapid crystallization from an undercooled liquid.

409

410   *6.2.2 Clinopyroxene*

411

412   Among the three distinct types of textures displayed by the clinopyroxene phenocrysts in
413   the BP basalts, two are indicative of diffusion-limited, rapid crystal growth. The vermicular (i.e.,
414   spongy) texture (Fig. 8 e-h) may reflect an initial dendritic architecture that is then partially
415   overgrown (Welsch et al., 2016). A similar texture is found in vermiform chromite crystals
416   formed during rapid crystal-growth experiments on basalts (Roeder et al., 2001). The sector-
417   zoning texture, found in most of the low-Na_2O group of phenocrysts in the BP basalts (Fig. 7, 8i-
418   l), also points to rapid crystal growth rates (e.g., Shimizu, 1990; Brophy et al., 1999; Ubide et al.,
419   2019). Experimental studies (e.g. Kouchi et al., 1983; Skulski et al., 1994) show that sector
420   zoning develops in clinopyroxene crystals at various undercooling ($\Delta T_{UC} = T_{liquidus} - T_{melt}$)
421   conditions, leading to growth rates that vary with ΔT_{UC}. For example, Kouchi et al. (1983)
422   report growth rates along the *010* sector that range from 1.6 x 10^{-7} cm/s (~138 µm/day) at ΔT_{UC}
423   of 13°C to 7.5 x 10^{-6} cm/s (~6.5 mm/day) at ΔT of 45 °C, consistent with growth of the
424   clinopyroxene phenocrysts in BP basalts over a period of days. Skulski et al. (1994) grew
425   phenocryst-sized (~220 µm) clinopyroxene crystals with sector zoning in a piston-cylinder run (1
426   GPa) within 24 hours.

427

10

428 The co-occurrence of oscillatory zonation (e.g., Cr, Al, Mg) with sector zoning has been
429 reported in several naturally-occurring clinopyroxene phenocrysts, often in mafic melts (e.g.,
430 Shimizu, 1990; Brophy et al., 1999; Ubide et al., 2019). There is no broad consensus on the
431 origin of the oscillatory zonation in clinopyroxene. Some studies point to magmatic recharge
432 events (e.g., Ubide et al., 2019), whereas others conclude that kinetic effects at the crystal-melt
433 interface play a dominant role (e.g., Shimizu, 1990), similar to the case presented for oscillatory
434 zoning in plagioclase (e.g., Haase et al., 1980; L'Heureux, 1993). The X-ray intensity maps for
435 the clinopyroxene phenocryst from BP-31 (Fig. 10) clearly show that both types of zonation
436 (sector and oscillatory) developed together, which broadly supports the role of various kinetic
437 effects (consistent with rapid growth) in their respective development.
438
439 6.3 Evidence of polybaric phenocryst growth of clinopyroxene
440
441 In the high-Mg BP basalts, the occurrence of two distinct populations of clinopyroxene
442 composition, reflected in plots of Na_2O vs. Mg# (Fig. 7), point to two distinct depth intervals for
443 phenocryst growth. The high-Na_2O population points to an elevated jadeite component (Fig. 7)
444 and thus a greater pressure of crystallization for the non-spongy (non-vermicular) clinopyroxene.
445 Mordick and Glazner (2006) also reported high Na_2O contents (0.86-0.48 wt%) in several
446 clinopyroxene crystals from the BP basalts, and noted that they must have crystallized at
447 relatively high pressures. In contrast, the oscillatory- and sector-zoned clinopyroxene, which
448 often occurs as rims on clinopyroxene crystals, with a sharp compositional contrast between core
449 and rim, is in the low-Na_2O group, which points to growth at low pressure. In some cases, the
450 core appears to have a rounded (resorbed) shape adjacent to the oscillatory-zoned rims (Fig. 8b,
451 c, d, i, k), which suggests an interval between the high-pressure and low-pressure growth of
452 clinopyroxene, where clinopyroxene was not stable and resorption was not kinetically inhibited
453 during ascent. Evidence of polybaric crystallization of clinopyroxene in a single basalt sample,
454 and even in a single phenocryst within a sample, has been reported in recent studies (e.g.,
455 Hammer et al., 2016; Mollo et al., 2020).
456
457 In Figure 11, the schematic illustration depicts a viable hypothesis to explain the
458 polybaric growth of clinopyroxene phenocrysts during ascent, consistent with available
459 information of phase relations. For example, it is well known from phase-equilibrium
460 experiments (e.g., Weaver et al., 2011) that the stability of clinopyroxene in basaltic melts is
461 strongly dependent on pressure and that the temperature of the clinopyroxene-in curve decreases
462 more strongly than that for olivine with decreasing pressure (e.g., MELTS; Ghiorso & Sack,
463 1995; Asimow & Ghiorso, 1998). It is therefore possible that there is a depth interval in the
464 lower-middle crust where the melt temperature is below the olivine- and clinopyroxene-in
465 curves, but that in the middle-upper crust (prior to significant loss of H_2O from the melt due to
466 degassing), the melt temperature may move above the clinopyroxene-in curve (Fig. 11). This
467 will be a depth interval that is marked by no clinopyroxene growth (and possibly some
468 resorption). Then, as the melt continues to ascend into the upper crust, significant loss of H_2O
469 will inevitably occur (due to degassing), causing the slope of the clinopyroxene-in (and olivine-
470 in) curve to change sign. The melt temperature will cross the clinopyroxene-in curve again,
471 leading to rapid growth of the oscillatory- and sector-zoned clinopyroxene under shallow, fluid-
472 saturated conditions prior to eruption.
473

11

474

7 Evidence that the most Mg-rich olivine represents the liquidus phase in each sample

476

477   If phenocryst growth during ascent is a viable hypothesis for the BP basalts (Fig. 11), it is
478 possible that the most Mg-rich olivine observed in each sample (Fig. 2) closely approximates the
479 first phase to crystallize from a liquid with the composition of whole-rock. In other words, if
480 phenocryst growth was rapid, there may not have been time for significant chemical re-
481 equilibration of the first (i.e., liquidus) olivine that initially grew from the melt. Thus, an
482 evaluation of whether the most Mg-rich olivine analyzed in each sample has a composition
483 expected for the liquidus olivine can be made.

484

485   To perform this assessment, hereafter referred to as the $^{Fe2+-Mg}K_D$(olivine-melt) liquidus test, a
486 plot of the most Mg-rich olivine phenocryst in each sample as a function of the whole-rock Mg#
487 is shown in Figure 12. For the calculation of Mg#, it is assumed that the Fe^{3+}/Fe^T in each sample
488 is 0.24 (average of μ-XANES analyses; Kelley & Cottrell, 2012). Superimposed on this plot are
489 isopleths of $^{Fe2+-Mg}K_D$ (olivine-melt) values:

490

491
$$^{Fe^{2+}-Mg}K_D(olivine - melt) = \left(\frac{X_{Fe^{2+}O}}{X_{MgO}}\right)^{olivine} \left(\frac{X_{MgO}}{X_{Fe^{2+}O}}\right)^{melt} \tag{2}$$

492

493 The results show that the most Mg-rich olivine analyzed in the BP basalts overlap $^{Fe2+-Mg}K_D$
494 (olivine-melt) isopleths that range from 0.31 to 0.36 (Table 2), which broadly match the range of
495 values calculated from various models in the literature (0.32-0.36; Table 3), including the
496 MELTS thermodynamic model (Ghiorso & Sack, 1995; Asimow & Ghiorso, 1998) and the
497 Putirka (2016) model. The K_D values from MELTS do not vary with oxidation state between
498 ΔQFM = -1 to +1, and the value at ΔNNO = 0 is reported in Table 2. For the Putirka (2016)
499 model, the effect of pressure on the K_D value is negligible at crustal depths. Pu et al. (2017)
500 showed that ~5-6 wt% dissolved H_2O leads to an increase in $^{Fe2+-Mg}K_D$ (olivine-melt) value by
501 ≤0.03, and thus a small increase in K_D of ~0.01 is possible for the Big Pine basalts and is also
502 evaluated (Table 3).

503

504   The results in Figure 12 and Table 2 strongly support the hypothesis that the most Mg-
505 rich olivine analyzed in each sample closely represents the first olivine to crystallize from a
506 liquid with the whole-rock composition (i.e., pass the $^{Fe2+-Mg}K_D$ (olivine-melt) liquidus test),
507 which is consistent with phenocryst growth of olivine during ascent (Fig. 11). Also shown in
508 Figure 12 are data from Pu et al. (2017) for 16 basaltic samples (>5 wt% MgO) from the
509 Mexican arc (TVF). Collectively, the results for both volcanic fields show that phenocryst
510 growth during ascent and preservation of the first olivine composition to crystallize from the
511 melt is not uncommon, even in samples without mantle xenoliths.

512

513   7.1. Consistency with Fe-Ti two-oxide oxybarometry and whole-rock FeO analyses

514

515   The approach taken in Figure 12 can be inverted to extract information on the melt
516 Fe^{3+}/Fe^T ratio at the onset of olivine crystallization for each of the BP samples, using $^{Fe2+-Mg}K_D$
517 (olivine-melt) values calculated for each sample from MELTS (Ghiorso & Sack, 1995; Asimow
518 & Ghiorso, 1998) and Putirka (2016) (Table 3). The results can then be compared to the Fe-Ti

12

519 two-oxide oxybarometry results for BP-20 (Table S7). Through application of Equation (2), the
520 only unknown in this calculation is the concentration of ferrous iron in the liquid (X_{FeO}^{liq}).
521 Resulting Fe^{3+}/Fe^T ratios are shown as histograms in Figure 13. Additional histograms
522 illustrating the effect of an increase in $^{Fe2+-Mg}K_D$ (olivine-melt) values of +0.01 applied to those
523 from MELTS and Putirka (2016) due to the effect of ~1.5 wt% dissolved H_2O (Gazel et al.,
524 2012) are shown for comparison (Figure 13b, d).
525
526   Among the four histograms (Fig. 13), the average (± 1σ) Fe^{3+}/Fe^T ratios range from 0.25
527 (± 0.04) to 0.22 (± 0.03), which overlap the average Fe^{3+}/Fe^T ratio (0.24 ± 0.01) from μ-XANES
528 (Kelley & Cottrell, 2012). Importantly, among the four samples with the lowest calculated
529 Fe^{3+}/Fe^T ratios (in all four histograms; Fig. 13), ilmenite was found in three of them, which is
530 consistent with the stabilization of ilmenite at slightly lower fO_2 conditions compared to rest of
531 the BP samples. For sample BP-20, the onset of two-oxide crystallization occurred at ΔNNO = -
532 0.45 (Fe^{3+}/Fe^T = .19) and thus overlaps the calculated Fe^{3+}/Fe^T values for the ilmenite-bearing
533 samples on the basis of olivine-melt oxybarometry. Because the average Fe^{3+}/Fe^T ratio in all
534 four histograms do not vary significantly from one another, the effect of using different models
535 to calculate the K_D value, and adding 0.01 to reflect the effect of >1.5 wt% H_2O (Gazel et al.,
536 2012; Pu et al., 2017) has little impact. The results in Figure 13 provide evidence that the most
537 Mg-rich olivine in each BP sample passes the $^{Fe2+-Mg}K_D$ (olivine-melt) liquidus test.
538
539   An additional test of whether the most Mg-rich olivine is the liquidus composition is
540 obtained by comparing the ferrous iron concentration in each sample calculated from olivine-
541 melt oxybarometry (Eq. 2) with those directly measured in the whole-rock sample by the Wilson
542 (1960) titration method. In Figure 14, four 1:1 plots of wt% FeO are shown based on the four
543 different sets of $^{Fe2+-Mg}K_D$ (olivine-melt) values presented in Figure 13. Also shown are results
544 from Crabtree and Lange (2012) and Waters and Lange (2016) on andesite, dacite and rhyolite
545 samples, where Fe-Ti oxide oxybarometry, instead of olivine-melt oxybarometry (Equation 2),
546 was used to constrain pre-eruptive wt% FeO concentrations. Samples BP-11 and -16, with
547 evidence of alteration and heterogeneity, are shown as open symbols. The error bars reflect ±2σ
548 analytical uncertainties in the titration results for each sample, and the propagated error in the
549 FeO concentration (± 0.29-0.41 wt%) from olivine-melt oxybarometer derived from an
550 uncertainty of ± 0.02 in the $^{Fe2+-Mg}K_D$ (olivine-melt) value. The results show a close adherence
551 to the 1:1 line in all four plots.
552
553 **8 Olivine-melt thermometry**
554
555   The collective data in Figures 12-14 confirm that the most Mg-rich olivine in each
556 sample closely represents the first olivine to crystallize from the whole-rock liquid composition.
557 This enables thermometry to be applied to this olivine-melt pairing in order to calculate
558 temperature at the onset of phenocryst growth in each sample.
559
560   Numerous olivine-melt thermometers based on the partitioning of Mg ($D_{Mg}^{ol/liq}$) have
561 been presented in the literature. In a review of these thermometers, Putirka (2008) showed that
562 the Beattie (1993) form of the Mg-based thermometer is the best at recovering experimental
563 temperature under anhydrous conditions and produces temperatures similar to those obtained by
564 the MELTS thermodynamic model (Ghiorso & Sack, 1995; Asimow & Ghiorso, 1998).

565 However, one of the limitations of this thermometer is that $D_{Mg}^{ol/liq}$ is strongly sensitive to
566 dissolved water in the melt (e.g., Almeev et al., 2007; Medard & Grove, 2008), which leads to
567 calculated temperatures that are too high if not corrected for the effect of H_2O. To address this
568 problem, Putirka et al. (2007) calibrated an olivine-melt thermometer based on $D_{Mg}^{ol/liq}$ that
569 includes a correction for H_2O in the melt. Its application requires that water contents in the melt
570 be known, which can be obtained from volatile analyses of olivine-hosted melt inclusions (e.g.,
571 Gazel et al., 2012).
572
573 Because information on melt water contents is not always known for all basalt samples,
574 Pu et al. (2017) developed an olivine-melt thermometer based on the partitioning of Ni ($D_{Ni}^{ol/liq}$),
575 which shows no dependence on dissolved water in the melt or pressure under crustal (< 1 GPa)
576 conditions (Pu, 2018). Thus, the Ni-thermometer provides the actual temperature at the onset of
577 olivine growth, because it requires no correction for H_2O content. In the Pu et al. (2017) study,
578 two olivine-melt thermometers (based on $D_{Mg}^{ol/liq}$ and $D_{Ni}^{ol/liq}$) were calibrated on the same set of
579 123 1-bar experiments (4-40 wt% MgO), and the resulting models recover the experimental
580 temperatures equally well. The average (± 1σ) temperature difference between the two
581 thermometers (i.e., T_{Mg}-T_{Ni}) is 1 (± 29) °C. Therefore, any systematic difference in calculated
582 temperatures between the two thermometers, when applied to the same basalt, cannot be
583 attributed to differences in their respective calibrations, but instead may reflect the role of
584 dissolved H_2O in the melt.
585
586 In this study, the Mg- and Ni-based olivine-melt thermometers from Pu et al. (2017) were
587 applied to all BP samples in Table 1. For the Mg-thermometer, values of $D_{Mg}^{ol/liq}$ were
588 constructed from the most Mg-rich olivine in each sample paired with a liquid of its whole-rock
589 composition and are reported in Table 2. Resulting temperatures from the Mg-thermometer
590 (T_{Mg}) range from 1309-1170 °C for melts that range from 13-7 wt% MgO (Table 2; Fig. 15).
591 Analytical uncertainties in $D_{Mg}^{ol/liq}$, derived from the microprobe and whole-rock analytical errors,
592 are relatively low (0.9-1.5%), and lead to propagated uncertainties in T_{Mg} of ± 4-5 °C. For the
593 Ni-thermometer, values of $D_{Ni}^{ol/liq}$ were constructed for the most Mg-rich olivine in each sample
594 based on a linear fit to the olivine data (wt% NiO vs. mol% Fo content) in Figure 3. This linear
595 fit reduces the analytical scatter in the NiO wt% data due to microprobe error and provides a
596 more accurate estimate of the NiO wt% in the most Mg-rich olivine in each sample. Resulting
597 temperatures (T_{Ni}) range from 1247-1097 °C and are systematically lower than those obtained
598 with the Mg-thermometer (Table 2; Fig. 15a). Analytical uncertainties in $D_{Ni}^{ol/liq}$, derived from
599 errors in the olivine (±0.02 wt% NiO) and the whole-rock (±4 ppm Ni) compositions, range from
600 5-11% and lead to propagated uncertainties in T_{Ni} of ± 12-23 °C. Pu et al. (2017) note that
601 additional error exists if the most Mg-rich olivine crystal is not used (e.g. not exposed/analyzed,
602 or lost to diffusive re-equilibration). Therefore the most Mg-rich olivine composition is
603 considered a minimum value. Calculated temperatures using both the Mg- and Ni-thermometers
604 are thus maximum values and ΔT (=T_{Mg} – T_{Ni}) is a minimum estimate. Pu et al., (2017)
605 calculate that a decrease in Fo by 1 mol% results in ΔT values that are 13-46°C too low for their
606 samples. Again, because $D_{Ni}^{ol/liq}$ is insensitive to dissolved water and variations in crustal
607 pressure (Pu, 2018), T_{Ni} values give the actual temperature at the onset of olivine growth.
608
609 **9 Olivine-melt hygrometry**

14

610

611   The thermometry results in Figure 15a show that temperatures based on $D_{Mg}^{ol/liq}$ (T_{Mg}) are

612 systematically higher than those based on $D_{Ni}^{ol/liq}$ (T_{Ni}) and both systematically decrease with

613 decreasing MgO content in the melt. The average temperature difference between the two

614 thermometers (T_{Mg}-T_{Ni}) is ~59 °C (Fig. 15a), which points to the presence of dissolved water in

615 the BP basalts. Similar differences in temperature (T_{Mg}-T_{Ni}) were obtained by Pu et al. (2017) on

616 a suite of Mexican arc basalts (Fig. 15b), where both the Mg- and Ni-thermometers were applied.

617 In that study, the average ΔT (=T_{Mg}-T_{Ni}) value was 114 °C, which is nearly twice as large as the

618 average value of 59°C from this study (Fig. 15). Since values of ΔT (=T_{Mg}-T_{Ni}) reflect the

619 depression of the olivine liquidus due to dissolved water in the melt, the expectation is that the

620 Mexican arc basalts were more hydrous than the BP basalts. This prediction is consistent with

621 water analyses in olivine-hosted melt inclusions, with 3.9-5.7 wt% in the Mexican basalts

622 (Johnson et al., 2008; 2009) and 1.5-3.0 wt% in the BP basalts (Gazel et al., 2012). The average

623 ΔT values of ~114°C and ~59°C can be converted into an average estimate of melt water content

624 of ~3.8 and 1.7 wt%, respectively, for the Mexican and BP basalts using the model of Medard

625 and Grove (2008).

626

627   Another assessment of melt water contents can be made through application of the

628 Putirka et al. (2007) Mg-based olivine-melt thermometer, which includes a correction for

629 dissolved water content. In the absence of any H_2O correction, application of the Putirka

630 thermometer to the BP basalts leads to calculated temperatures that closely match those obtained

631 by the Pu et al. (2017) Mg-thermometer (Table 2; Fig. 15). The next step is to input the water

632 contents (1.5-3.0 wt%) from Gazel et al. (2012) into the H_2O-corrected Mg-thermometer of

633 Putirka et al. (2007; Eq. 4) and calculate temperatures for all BP samples. A comparison of these

634 H_2O-corrected temperatures to those obtained with the Ni-thermometer of Pu et al. (2017) show

635 excellent agreement (Fig. 15). The results illustrate that the Ni-thermometer of Pu et al. (2017)

636 matches the results for the H_2O-corrected thermometer of Putirka et al. (2007) when the water

637 contents from Gazel et al. (2012) are employed. Note also that under anhydrous conditions,

638 when the respective Mg-thermometer of Putirka et al. (2007) and Pu et al. (2017) is each applied

639 to all ten BP basalts, the deviations in temperature are ≤4 °C (Table 2; Fig. 15a).

640

641   A similar assessment can be made on the 16 basalts (>5 wt%) from the Mexican arc from

642 Pu et al. (2017), where the Putirka et al. (2007) H_2O-corrected Mg-thermometer is applied (Fig.

643 15b.) When a range of melt H_2O content (4.0-6.5 wt%) consistent with measured values in

644 olivine-hosted melt inclusions (3.9-5.7 wt%; Johnson et al., 2008, 2009) is applied, temperatures

645 show agreement with those from the Ni-thermometer (Pu et al., 2017; Fig. 15b). Melt H_2O

646 contents measured in melt inclusions are minimum estimates and a slightly higher melt H_2O

647 content is needed to capture the range of calculated T_{Ni} values. This result shows that phenocryst

648 growth during ascent may be common, and that olivine-melt thermometry/hygrometry can be

649 applied at the liquidus in basalts that do not contain mantle xenoliths.

650

651   In summary, the internal consistency among four independent studies (Putirka et al.,

652 2007; Medard and Grove, 2008; Gazel et al., 2012; Pu et al., 2017) lends credibility not only to

653 the various thermometers that are employed in Figure 15, but also to the hypothesis that the T_{Ni}

654 temperatures are those at the onset of olivine crystallization during ascent of the basalt samples.

15

655 In other words, the pairing of the most Mg-rich olivine analyzed in each sample with the whole-
656 rock composition of each sample is a valid approach.

657

658 9.1 Correlation between $D_{Ca}^{ol/liq}$ and depression of olivine liquidus due to H_2O

659

660 Additional verification of the hydrous nature of the BP basalts is seen with an
661 examination of the partition coefficient for $D_{Ca}^{ol/liq}$ using the most Mg-rich olivine paired with the
662 whole-rock composition (proxy for liquid at onset of olivine crystallization). It has been
663 proposed in the literature (e.g., Gavrilenko et al., 2016) that $D_{Ca}^{ol/liq}$ values are a function of
664 dissolved water content, with a potential to be used as an olivine-melt hygrometer. Here, a
665 comparison is made between the average ΔT ($=T_{Mg}-T_{Ni}$) values, obtained from the Pu et al.
666 (2017) thermometers, and the average $D_{Ca}^{ol/liq}$ values obtained for three sets of basalts: (1) the BP
667 basalts from this study, (2) Mexican arc basalts (Pu et al., 2017), and (3) the suite of mid-ocean
668 ridge basalts (Allan et al., 1989) examined in Pu et al. (2017). The data are plotted in Figure 16
669 and reveal a strong inverse, linear correlation ($R^2=0.99$) among the averages from each volcanic
670 field. The highest $D_{Ca}^{ol/liq}$ values correspond to the mid-ocean ridge basalts from Allan et al.
671 (1989), which are nearly anhydrous (~0.1 wt% H_2O), whereas those samples with the lowest
672 $D_{Ca}^{ol/liq}$ values correspond to the Mexican arc basalts, which have the highest analyzed water
673 contents (≤ 5.7 wt% H_2O; Johnson et al., 2009). As expected, the BP basalts have $D_{Ca}^{ol/liq}$ values
674 in between those for the MORBs and arc basalts, consistent with melt water contents higher than
675 those for mid-ocean ridge basalts, yet lower than those for active arc basalts. Thus, the
676 $D_{Ca}^{ol/liq}$ values in Figure 16 provide independent corroboration that the application of both the
677 Mg- and Ni-thermometers from Pu et al. (2017) to basalts (with olivine compositions that pass
678 the $^{Fe^{2+}-Mg}K_D$ liquidus test) provide accurate constraints on temperature and melt water contents
679 at the onset of olivine crystallization.

680

681 9.2 Minimum water contents at onset of plagioclase phenocryst growth

682

683 A final evaluation of dissolved water contents in the melt can be made by applying the
684 plagioclase-liquid hygrometer (Waters & Lange, 2015) to the four samples than contain
685 plagioclase phenocrysts. This approach only constrains minimum melt water contents since the
686 temperature and melt composition at the onset of plagioclase crystallization is not known, only
687 those at the onset of olivine growth at the liquidus. Nonetheless, because using a temperature
688 that is too high and a melt composition that is too mafic both lead to melt water contents that are
689 systematically too low, minimum water contents can be obtained. Therefore, a pairing of the
690 most An-rich plagioclase to the whole-rock (liquid) composition at T_{Ni} was performed for each
691 sample. The results are illustrated in Figure S4 (supplementary information) and lead to
692 minimum H_2O contents of 1.5, 1.6, 1.4 and 0.9 wt%, respectively, for BP-16, -30, -8 and -20,
693 which are fully consistent with the melt inclusion analyses of Gazel et al. (2012).

694

695 **10. Pressure (depth) at onset of clinopyroxene growth**

696

697 Five of the BP samples (those with highest MgO content) contain clinopyroxene
698 phenocrysts. In each of these five samples, when the most Mg-rich clinopyroxene composition

16

699 (consistently from the non-spongy group; Table 3) is paired with the whole-rock composition,
700 the resulting $^{FeT-Mg}K_D$ (cpx-melt) value (using total iron; FeO^T) ranges from 0.26-0.31 (Table 4).
701 These $^{FeT-Mg}K_D$ values overlap the range expected for clinopyroxene-liquid equilibrium (0.27 $\pm$
702 0.03; Putirka, 2008), which shows that both clinopyroxene and olivine were closely co-saturated
703 at the onset of phenocryst growth.
704
705 The Neave and Putirka (2017) clinopyroxene-liquid barometer was applied to the most
706 Mg-rich clinopyroxene paired with the whole-rock liquid composition for each sample. This
707 geobarometer is dependent on temperature, and therefore the Ni-based temperature (T_{Ni}; Table 2)
708 at the onset of phenocryst growth was used. The results give the pressure at the onset of
709 clinopyroxene growth in these five samples, namely 0.73, 0.73, 0.89, 0.77 and 0.69 GPa for BP-
710 23, -31, -11, -19, and -24, respectively (Table 3). Note that the use of lower temperatures (below
711 liquidus values) and more evolved melt compositions (more differentiated than the whole-rock
712 composition) both lead to shallower calculated pressures, and therefore the pressures in Table 3
713 are maximum values. Thus, the trend of increasing Na_2O with decreasing Mg# (among the high-
714 Na_2O, non-spongy group; Fig. 7) does not preclude phenocryst growth during ascent. In other
715 words, the effect of lower melt temperature and more evolved melt composition (due to
716 progressive crystallization of olivine and clinopyroxene) will lead to higher Na_2O contents in
717 clinopyroxene during crustal ascent.
718
719 **11 Implications and Conclusions**
720
721 There are several key conclusions that can be drawn from this study. First and foremost,
722 there is abundant evidence to support the hypothesis that phenocryst growth of olivine (and
723 clinopyroxene) in the BP basalts occurred rapidly during ascent along fractures. Lines of
724 evidence include the presence of mantle xenoliths in several flows, diffusion-limited growth
725 textures in olivine and clinopyroxene, and polybaric crystallization of clinopyroxene
726 phenocrysts.
727
728 Second, when the most Mg-rich olivine analyzed in each BP basalt is paired with a liquid
729 composition that matches that for the whole rock, and a melt Fe^{3+}/Fe^T ratio of 0.24 (average from
730 μ-XANES analyses; Kelley & Cottrell, 2012) is used, the resulting $^{Fe2+-Mg}K_D$ (olivine-melt)
731 values match those predicted from various K_D models from literature. In other words, the most
732 Mg-rich olivine in each sample passes the $Fe^{2+-Mg}K_D$ (olivine-melt) liquidus test. This
733 observation points to rapid ascent of the host basalt such that there was insufficient time for re-
734 equilibration of the first olivine composition to crystallize from the melt. For those basalts that
735 also contain clinopyroxene phenocrysts, the most Mg-rich clinopyroxene also passes the $^{FeT-}$
736 $^{Mg}K_D$ (cpx-melt) liquidus test, suggesting that the two phases were closely co-saturated at the
737 onset of phenocryst growth in five of the high-MgO BP basalts.
738
739 The third major conclusion is that olivine-melt thermometry and hygrometery can be
740 applied to the most Mg-rich olivine in each sample paired with a liquid composition that matches
741 the whole-rock, for pairs that pass the $Fe^{2+-Mg}K_D$ (olivine-melt) liquidus test. Calculated
742 temperatures based on $D_{Ni}^{ol/liq}$ (Pu et al., 2017), which is independent of dissolved water in the
743 melt (Pu, 2018), are ~60°C lower, on average, than those based on $D_{Mg}^{ol/liq}$, which gives the
744 temperature under anhydrous conditions (T_{Mg}). According to the Medard and Grove (1998)

17

745 model, a depression of the olivine liquidus by ~60°C is expected for a melt water content of ~1.7
746 wt% H_2O, which matches the H_2O contents directly measured in olivine-hosted melt inclusions
747 (~1.5-3.0 wt%; Gazel et al., 2012). The accuracy of the temperatures based on the Ni-
748 thermometer (T_{Ni}), which range from 1273-1097°C, is further confirmed by application of the
749 H_2O-corrected Mg-thermometer of Putirka et al. (2007) using the H_2O contents analyzed by
750 Gazel et al. (2012). Finally, values of $D_{Ca}^{ol/liq}$ scale linearly and inversely with the magnitude of
751 the depression of the olivine liquidus ($\Delta T = T_{Mg} - T_{Ni}$) due to dissolved water in the melt, which
752 confirms previous work in the literature indicating that $D_{Ca}^{ol/liq}$ has the potential to be used as an
753 olivine-melt hygrometer (Gavrilenko et al., 2016).
754
755 A fourth conclusion is that the depth for the onset of phenocryst growth can be estimated
756 for the five basalts that were co-saturated with olivine and clinopyroxene at the onset of
757 phenocryst growth. When the olivine-liquidus temperatures are used to calculate pressure with
758 the clinopyroxene-melt barometer of Neave and Putrika (2017), results for the onset of
759 phenocryst growth range from 0.7-0.9 GPa.
760
761 The overall results from this study show that the hypothesis of rapid growth of olivine
762 (and clinopyroxene) phenocrysts during ascent and whether the most Mg-rich olivine analyzed in
763 each sample represents the first phenocryst to grow at the liquidus can both be tested in
764 individual basalt samples. This approach is best applied to samples where olivine is either the
765 liquidus or near-liquidus phase, and thus is likely best applied to samples ≥ 5 wt% MgO. To
766 perform the $^{Fe2+-Mg}K_D$ (olivine-melt) liquidus test, a large set of microprobe analyses of numerous
767 olivine crystals in each sample is required and an estimate of melt ferric-ferrous ratio is needed
768 (from V-partitioning between olivine-melt, μ-XANES analyses of olivine-hosted melt inclusions,
769 Fe-Ti oxide oxybarometry, and/or whole-rock FeO analyses of fresh, unaltered samples). For
770 those samples where the pairing of the most Mg-rich olivine composition with the whole-rock
771 composition passes the $^{Fe2+-Mg}K_D$ (olivine-melt) liquidus test, application of olivine-melt
772 thermometry and hygrometry can be applied at the liquidus, leading to key constraints on the
773 temperature and melt water content at the onset of phenocryst growth. Because this approach
774 only requires microprobe analyses of olivine analyses in thin sections (and whole-rock major
775 element and Ni contents), it can readily be applied to obtain large global data sets. To date, this
776 approach has been verified for basalts from the Big Pine volcanic field (this study) and the
777 Tancítaro volcanic field in the Mexican arc (Pu et al., 2017). The ready attainment of large
778 global data sets will allow the question of whether there are distinctly different temperatures and
779 water contents in basalts erupted within and between different tectonic settings to be more fully
780 addressed.

18

Acknowledgments, Samples, and Data

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1049 **Figure Captions**

1050 **Figure 1.** Map of the Quaternary Big Pine (BP) volcanic field adapted from Beard and Glazner
1051 (1995) and Vazquez and Woolford (2015). Gray regions are prominent basalt flows. Sample
1052 locations for this study are shown by solid circles. Triangles are localities of basalts with mantle
1053 xenoliths from Beard and Glazner (1995) and Wilshire et al. (1988). BP-11, shown in red (this
1054 study), contains a mantle xenolith. Cosmogenic ^{36}Cl ages and whole-rock ^{40}Ar/^{39}Ar ages are
1055 from Vazquez and Woolford (2015) and Blondes et al., (2008) respectively. The inset map
1056 shows the location of the Big Pine volcanic field along the western margin of the Basin and
1057 Range province.
1058
1059 **Figure 2.** Histograms of forsterite (Fo mol% = $X_{MgO}/(X_{MgO}+X_{FeO})$x 100) content of olivine
1060 crystals in 10 BP basalts. The most Fo-rich olivine compositions are reported in Table 2 (all
1061 analyses reported in Table S4). Whole-rock MgO wt%, maximum Fo analyzed, and number of
1062 analyses (n) are given for each sample. Fo contents for analyzed olivine in mantle xenolith (BP-
1063 11) are shown in red and overlap Fo contents of olivine phenocrysts. Fo contents for analyzed
1064 olivine inclusions in sector- or oscillatory-zoned clinopyroxene crystals in BP-31 are shown in
1065 green.
1066
1067 **Figure 3.** Wt% NiO in olivine vs Fo mol% = $(X_{MgO}/(X_{MgO}+X_{FeO})$ x 100) for olivine analyses
1068 that span the highest 3 mol% Fo in each sample. A linear fit to the data (yellow line) is used to
1069 calculate wt% NiO for the most Fo-rich olivine in each sample, which is used in the Ni-
1070 thermometer. BP-11 mantle xenolith olivines are shown in red for comparison. The most Fo-
1071 rich olivine composition and calculated NiO wt% are labeled on each plot and reported in Table
1072 2.
1073
1074 **Figure 4.** CaO wt% in olivine versus Fo mol% for olivines that span the top 3 mol% Fo in each
1075 sample. A linear fit to the top 3 mol% is shown on each plot in yellow and is used to calculated
1076 CaO wt% for the most Mg-rich olivine. Olivine phenocrysts (black) have systematically higher
1077 CaO wt% compared to mantle xenolith olivine analyzed in BP-11 (red). CaO wt% in olivine
1078 from other BPVF mantle xenoliths are shown in blue (Beard & Glazner, 1995) and green
1079 (Wilshire et al., 1988). Analyses of mantle olivine from San Carlos, AZ xenoliths are shown for
1080 comparison in yellow. All mantle olivine analyses have ≤ 0.15 wt% CaO (e.g., Simkin & Smith,
1081 1970; Norman & Garcia, 1999; Housh et al., 2010), which is used to identify possible olivine
1082 xenocrysts from mantle xenoliths (open symbols).

26

1083

1084 **Figure 5.** X-ray intensity maps of Fe (left) and P (right) for the most Fo-rich olivine analyzed,
1085 respectively, in six BP samples. Analytical spots and the associated Fo mol% are shown in
1086 green. X-ray maps of a mantle xenolith olivine from from San Carlos, Arizona are shown for
1087 comparison. Complex P zoning in olivine phenocrysts from BP basalts is consistent with an
1088 initial period of diffusion-limited rapid growth (e.g., Milman-Barris et al., 2008; Shea et al.,
1089 2019; Welsch et al., 2012; Welsch et al., 2014).

1090

1091 **Figure 6.** Representative back-scattered electron images of olivine phenocrysts that display
1092 diffusion-limited growth textures from six BPVF basalts. Fo content (mol %) spot analyses are
1093 shown in yellow.

1094

1095 **Figure 7.** Plots of Na_2O wt% vs. Mg# $=(MgO/(MgO+FeO^T)x100)$ (left) and TiO_2 wt% vs. Mg#
1096 (right) in clinopyroxene crystals from four BPVF samples (BP-23, -31, -11, 19). Additionally,
1097 three individual crystals are highlighted that collectively display examples of each textural
1098 category (one crystal can feature more than one texture). Individual analyses obtained across
1099 each crystal, with corresponding symbols that reflect the textural category for each analytical
1100 spot, are shown. Note that analyses from a single oscillatory- and sector-zoned crystal (cpx-1 in
1101 BP-31; Fig. 8) spans the entire compositional range as all zoned phenocrysts in this sample. Red
1102 crystals in each sample are the most Mg-rich and used to constrain depth at which cpx began to
1103 crystallize.

1104

1105 **Figure 8.** Representative BSE images of clinopyroxene crystals from each texture category
1106 including non-spongy (non-vermicular, often as cores; a, b, d, i and k), spongy (vermicular; as
1107 cores, f, and rims, h), and sector- and/or oscillatory-zoned crystals (sometimes rim only; b, d, f, i,
1108 and k). Two clinopyroxene crystals from BP-11 (cpx 7 and cpx 8) are microlites but are used
1109 here as examples of vermicular and sector- or oscillatory-zoned crystals. Individual analytical
1110 spots from these representative crystals are highlighted in Figure 7.

1111

1112 **Figure 9:** Histograms of Mg# in clinopyroxene crystals from four BP samples. Clinopyroxene
1113 analyses are given different colors according to texture to illustrate their relative abundance in
1114 each sample.

1115

1116 **Figure 10:** X-ray intensity maps of Mg, Cr and Al in a representative clinopyroxene crystal from
1117 BP-31 (cpx-1 in Figs. 7 and 8). Note that the compositional variation in this single crystal (Fig.
1118 7c-d) spans the entire range of all oscillatory and/or sector-zoned crystals in BP-31. Maps
1119 indicate that sector zoning strongly affects Al and Mg concentrations but not Cr contents. All
1120 three elements display oscillatory zoning.

1121

1122 **Figure 11.** A schematic diagram that illustrates how melt temperature (red) and liquidus
1123 temperatures of olivine (blue) and clinopyroxene (green) may vary during ascent, leading to
1124 development of undercooling and rapid phenocryst growth. Each BP basalt is expected to
1125 segregate from its source region (~30-80 km depth; Putirka et al., 2012; Gazel et al., 2012) fluid-
1126 undersaturated, which will cause its liquidus temperature to decrease during ascent. It is
1127 therefore plausible that initial ascent of basalt is super-liquidus (i.e., the drop in melt temperature
1128 during initial ascent is smaller than the drop in liquidus temperature). Eventually, the ascending

27

1129   melt cools sufficiently to cross its liquidus, but from a super-liquidus state, which leads to a
1130   kinetic delay in nucleation and crystal growth. This kinetic delay generates an undercooling
1131   ($\Delta T_{UC} = T_{liquidus} - T_{melt}$) and thus high crystal growth rates and low nucleation rates (e.g.,
1132   Lofgren, 1974), leading to growth of large, sparse phenocrysts with diffusion-limited growth
1133   textures. The onset of clinopyroxene phenocryst growth occurs between 26-33 km, on the basis
1134   of clinopyroxene-melt barometry (Table 4).

1135

1136   **Figure 12.** Plot of the most Mg-rich olivine analyzed in each BP sample versus the whole-rock
1137   Mg# (calculated with $Fe^{3+}/Fe^T = 0.24$, average of μ-XANES analyses; Kelley & Cottrell, 2012).
1138   Superimposed on this plot are isopleths of $^{Fe2+-Mg}K_D$ (olivine-melt). All BP samples from this
1139   study fall on K_D isopleths that range from 0.31-0.36 (expected values; Table 2) and thus pass the
1140 K_D liquidus test. Additionally, basalts (>5 wt% MgO) from the Mexican arc (TVF) from Pu et al.
1141   (2017) are shown for comparison, where whole-rock Mg# are also calculated using melt Fe^{3+}/Fe^T
1142 $= 0.24$. The most Mg-rich olivine in these basalts also pass the liquidus test, although none of
1143   the flows contain mantle xenoliths.

1144

1145   **Figure 13.** Histograms of melt Fe^{3+}/Fe^T values calculated using the most Mg-rich olivine
1146   analyzed in each sample, the whole-rock composition, and an $^{Fe2+-Mg}K_D$ (olivine-melt) value for
1147   that basalt liquid. Panels (a) and (c) demonstrate variation in calculated melt Fe^{3+}/Fe^T ratios
1148   when using $^{Fe2+-Mg}K_D$(olivine-melt) values calculated from MELTS ($\Delta NNO = 0$, anhydrous,
1149   Ghiorso & Sack, 1995; Asimow & Ghiorso, 1998) and the model of Putirka (2016), respectively.
1150   Panels (b) and (d) illustrate the effect of an increase in melt $^{Fe2+-Mg}K_D$ (olivine-melt) values of
1151   +0.01 applied to those from MELTS and Putirka (2016) due to the effect of dissolved melt H_2O
1152   (1.5 wt%, Gazel et al., 2012). Most Fe^{3+}/Fe^T values overlap the average analyzed value by μ-
1153   XANES in olivine-hosted melt inclusions (0.24 ± 0.01; Kelley & Cottrell, 2012). In all four
1154   histograms, three of the four lowest melt Fe^{3+}/Fe^T ratios are in samples that contain ilmenite.

1155

1156   **Figure 14.** Four 1:1 plots of post-eruptive wt% FeO analyzed by the Wilson (1960) titration
1157   method versus pre-eruptive wt% FeO obtained using olivine-melt oxybarometry at the liquidus,
1158   with the four $^{Fe2+-Mg}K_D$ (olivine-melt) values presented in Figure 13. Results from this study on
1159   BP basalts are compared to those from Crabtree and Lange (2012) and Waters and Lange (2016)
1160   on andesite, dacite, and rhyolite samples, where Fe-Ti oxide oxybarometry was used, instead of
1161   olivine-melt oxybarometry. Sample BP-11 and BP-16, with post-eruptive alteration products in
1162   its vesicles or heterogeneity between measurements are shown with open symbols. The 1:1 line
1163   is shown with dashed lines at ± 0.5 wt% FeO.

1164

1165   **Figure 15:** Calculated temperatures (˚C) from olivine-melt Mg- and Ni-thermometers (T_{Mg} and
1166 T_{Ni}, respectively) from Pu et al. (2017) versus whole-rock MgO (wt%) for (a) BP basalts and (b)
1167   Mexican arc basalts (TVF, Pu et al., 2017). Temperature error bars (Table 2) are from
1168   propagated analytical uncertainty and are within the symbol size for T_{Mg}. The average ΔT (=
1169 $T_{Mg} - T_{Ni}$) is approximately 59°C for the BP basalts and 115°C for the TVF arc basalts.
1170   Temperatures from the H_2O-corrected Mg-thermometer of Putirka et al. (2007) are shown for
1171   comparison. For the BP basalts, those corrected for 1.5 and 3.0 wt% H_2O (analyses from Gazel
1172   et al., 2012) overlap Ni-temperatures (which require no H_2O correction; Pu et al., 2017). For the
1173   TVF arc basalts, temperatures corrected for 4.0-6.0 wt% H_2O (consistent with olivine-hosted
1174   melt inclusion analyses from Johnson et al., 2008, 2009) overlap calculated T_{Ni} values.

28

1175

1176 **Figure 16:** Average ΔT ($=T_{Mg}-T_{Ni}$) vs. molar $D_{Ca}^{ol/liq}$ for basalts from three distinct volcanic

1177 fields (tectonic settings), including: (1) Mexican arc basalts (TVF; Pu et al., 2017), (2) mid-ocean

1178 ridge basalts (Allan et al., 1989; Pu et al., 2017), and (3) BP basalts from this study. A linear fit

1179 through the average of all three basalt groups shows a strong inverse correlation ($R^2 = 0.99$; y= -

1180 10027x + 225.83). Larger ΔT (and thus higher melt H_2O, due to its suppression of olivine

1181 liquidus; see text) correlates negatively with mol% D_{Ca}, where lower values are associated with

1182 higher melt water contents (e.g. Gavrilenko et al., 2016). Note that results for individual BP

1183 basalts (red) plot along the trend line within the $\pm 30°C$ dashed error bars.

1184

29

37.140°
118.33°   118.23°

N

BP-20
BP-23
273 ka

Sierra Nevada

200 km

395

37.056°

BP-7

91 ka
BP-31   761 ka

20 ka
BP-11   469 ka
62 ka   BP-24   BP-8
BP-19   16 ka
32 ka   40 ka   15 ka
17 ka
27 ka   5 km
36.972°
BP-30
17 ka   BP-16   ³⁶Cl ages: VW '15
                                    ⁴⁰Ar/ ³⁹Ar ages: B'08

(a) BP-23 — max Fo 91.5 mol%, calc NiO 0.42 wt%, y = 0.0561x − 4.715

(b) BP-31 — max Fo 90.9 mol%, calc NiO 0.35 wt%, y = 0.0303x − 2.4059

(c) BP-11 — max Fo 89.7 mol%, calc NiO 0.34 wt%, y = 0.0391x − 3.1656

(d) BP-19 — max Fo 90.0 mol%, calc NiO 0.31 wt%, y = 0.0236x − 1.8183

(e) BP-24 — max Fo 89.9 mol%, calc NiO 0.33 wt%, y = 0.0239x − 1.818

(f) BP-7 — max Fo 88.0 mol%, calc NiO 0.31 wt%, y = 0.0283x − 2.186

(g) BP-16 — max Fo 85.9 mol%, calc NiO 0.20 wt%, y = 0.0259 − 2.0195

(h) BP-30 — max Fo 85.8 mol%, calc NiO 0.20 wt%, y = 0.0258x − 2.018

(i) BP-8 — max Fo 86.4 mol%, calc NiO 0.26 wt%, y = 0.0324x − 2.5421

(j) BP-20 — max Fo 86.3 mol%, calc NiO 0.25 wt%, y = 0.0282x − 2.1815

(a) BP-23   calc. CaO 0.20 wt%
y = -0.0088x + 1.0085

(b) BP-31   calc. CaO 0.24 wt%
y = 0.0004x + 0.199

(c) BP-11   calc. CaO 0.22 wt%
y = -0.0036x + 0.5434

(d) BP-19   calc. CaO 0.22 wt%
y = 0.004x = 0.1358

(e) BP-24   calc. CaO 0.22 wt%
y = 0.0064x = 0.3508

(f) BP-7   calc. CaO 0.22 wt%
y = -0.0081x + 0.9316

(g) BP-16   calc. CaO 0.21 wt%
y = -0.0136x + 1.3782

(h) BP-30   calc. CaO 0.21 wt%
y = -0.0152x + 1.5165

(i) BP-8   calc. CaO 0.20 wt%
y = -0.0045x + 0.5845

(j) BP-20   calc. CaO 0.19 wt%
y = -0.0172x + 1.6759

BP23

(a) Fe — High/Low — 89.4, 91.2, 89.3 — 300 μm

(b) P — High/Low

BP31

(c) Fe — High/Low — 88.4, 90.5, 87.6 — 250 μm

(d) P — High/Low

BP11

(e) Fe — High/Low — 88.1, 89.7, 82.2 — 50 μm

(f) P — High/Low

BP19

(g) Fe — High/Low — 89.2, 90.0, 86.2 — 200 μm

(h) P — High/Low

BP24

(i) Fe — High/Low — 85.7, 89.9, 83.9 — 100 μm

(j) P — High/Low

BP7

(k) Fe — High/Low — 86.5, 88.0, 73.0 — 100 μm

(l) P — High/Low

San Carlos

(m) Fe — High/Low — average Fo = 90.8 ± 0.1 — 500 μm

(n) P — High/Low

non-spongy/non-vermicular · spongy/vermicular · oscillatory/sector zoned

(a) BP-23 cpx 28 — 200 μm
(e) BP-23 cpx 56 — 500 μm
(i) BP-23 cpx 20 — 200 μm

(b) BP-31 cpx 80 — 500 μm
(f) BP-31 cpx 22 — 200 μm
(j) BP-31 cpx 1 — 500 μm

(c) BP-11 cpx 31 — 100 μm
(g) BP-11 cpx 8* — 50 μm
(k) BP-11 cpx 7* — 50 μm

(d) BP-19 cpx 10 — 100 μm
(h) BP-19 cpx 61 — 250 μm
(l) BP-19 cpx 32 — 100 μm

Temperature

Pressure

cpx in

significant H₂O degassing

cpx out

ΔT $\Delta T_{UC} = T_{liquidus} - T_{melt}$
"undercooling"

cpx in (26-33 km)

olivine in

transition from
cpx/olv on liquidus

T_{melt}

Zone of melt segregation (30-80 km)

cpx in

cpx out

delayed onset of
phenocryst growth

kinetic delay: no growth

cpx in

olv in

superliquidus

(a) MELTS: $^{Fe2+-Mg}K_D$(oliv-melt)
Avg. Fe^{3+}/Fe^T = **0.23 ± 0.04**
0.24 ± 0.01 (KC '12)

(b) MELTS: $^{Fe2+-Mg}K_D$(oliv-melt) + 0.01
Avg. Fe^{3+}/Fe^T = **0.25 ± 0.04**
0.24 ± 0.01 (KC '12)

(c) Putirka, 2016: $^{Fe2+-Mg}K_D$(oliv-melt)
Avg. Fe^{3+}/Fe^T = **0.22 ± 0.03**
0.24 ± 0.01 (KC '12)

(d) Putirka, 2016: $^{Fe2+-Mg}K_D$(oliv-melt) + 0.01
Avg. Fe^{3+}/Fe^T = **0.24 ± 0.03**
0.24 ± 0.01 (KC '12)

(a) MELTS: $^{Fe2+-Mg}K_D$(oliv-melt)

post-eruptive reduction

post-eruptive oxidation

± 0.5

BP-16

BP-11

This study
Crabtree & Lange, 2012
Waters & Lange, 2016

Post-eruptive FeO wt%

Pre-eruptive FeO wt%

(b) MELTS: $^{Fe2+-Mg}K_D$(oliv-melt) + 0.01

post-eruptive reduction

post-eruptive oxidation

± 0.5

BP-16

BP-11

Post-eruptive FeO wt%

Pre-eruptive FeO wt%

(c) Putirka, 2016: $^{Fe2+-Mg}K_D$(oliv-melt)

post-eruptive reduction

post-eruptive oxidation

± 0.5

BP-16

BP-11

Post-eruptive FeO wt%

Pre-eruptive FeO wt%

(d) Putirka, 2016: $^{Fe2+-Mg}K_D$(oliv-melt) + 0.01

post-eruptive reduction

post-eruptive oxidation

± 0.5

BP-16

BP-11

Post-eruptive FeO wt%

Pre-eruptive FeO wt%

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(a) Big Pine (this study)

(b) TVF arc (Pu et al., 2017)