

25 mantle wedge reacting with the mantle en route. Due to their re-equilibration during
26 ascent, these hydrous magmas ultimately record the conditions in the hot, shallow nose of
27 the mantle wedge at the end of their mantle ascent path rather than the conditions at their
28 point of origin as often interpreted. When the mantle residue for this process is lherzolite,
29 calc-alkaline basalt is generated. When the mantle residue is harzburgite to dunite, either
30 high-Mg primitive andesite or high-MgO liquid is generated, depending on the H₂O
31 content. A different type of primitive arc magma, specifically nominally anhydrous arc
32 tholeiite, is generated by near-fractional decompression melting at or near the anhydrous
33 lherzolite solidus in the upwelling back limb of corner flow at ~25-10 kbar and is focused
34 into the same region of the shallow mantle wedge as the hydrous melts. The similarity in
35 the terminus of the mantle ascent paths for both wet and dry primitive arc magmas likely
36 explains their eruption in close spatial and temporal proximity at many arcs. The
37 conditions of last mantle equilibration for primitive arc tholeiites generated by
38 decompression melting also imply that the convecting mantle extends to 10 kbar (~30
39 km) or less below most arcs. The range of mantle-melt equilibration conditions calculated
40 here agrees well with the range of temperatures predicted for the shallow mantle wedge
41 beneath arcs by geodynamic models, although it suggests some subduction zones may
42 have higher maximum temperatures at shallower depths in the wedge than originally
43 predicted. Primitive hydrous arc magmas also constrain natural variation on the order of
44 200-250°C in the maximum temperature in the hot shallow nose of the mantle wedge
45 between arcs. Thus the new primitive magma thermobarometry presented here is useful
46 for understanding melt migration processes and the temperature structure in the

47 uppermost part of the mantle wedge, as well as the origin of different primitive magma
48 types at arcs.

49 **Introduction**

50 Substantial work has advanced our knowledge of the underlying processes that
51 give rise to volcanic arcs since the advent of plate tectonic theory. This includes an
52 understanding of the first order processes that produce arc magmas such as models of the
53 volatile flux from the subducting plate into the overlying mantle (e.g., Poli and Schmidt,
54 1995, 2002; Hacker 2008; van Keken et al. 2011) and its effect on mantle melting
55 behavior (Kushiro et al. 1968; Green, 1973; Mysen and Boettcher 1975; Kawamoto and
56 Holloway 1997; Grove et al. 2006; Till et al. 2012b). Many different types of mantle-
57 derived arc magmas have been observed, including high-Mg andesites, calc-alkaline
58 basalts, high alumina olivine tholeiites, boninites and sanukitoids. But questions remain
59 as to what is ultimately responsible for producing each variety of primitive arc magma
60 and why different arcs have different abundances of these magmas. Thermobarometry of
61 the reconstructed primary parental magmas for these primitive liquids provides a
62 powerful means to answer these questions, as well as to determine where in the mantle
63 these primitive liquids are sourced. In addition, thermobarometry of primary arc liquids
64 provides observational constraints on the temperature at a given pressure in the mantle
65 that can be used in conjunction with dynamical models of mantle flow (e.g., Kelemen et
66 al. 2003). Outstanding questions about melt flow in the mantle wedge (e.g., reactive
67 porous flow vs. diapric or channelized flow: Navon and Stolper 1987; Grove et al. 2002)
68 can also for example be addressed through the study of primitive magmas and the
69 pressures and temperatures they record.

70 The opportunity to answer these questions on a global-scale requires a
71 compilation of existing thermobarometry of primitive arc magmas. However, the
72 methodology and assumptions vary between past studies. For example, the process of
73 determining a primitive magma's liquid line of descent in order that the primary liquid
74 composition can be estimated and used with thermobarometers remains somewhat of an
75 art form and thus the methodology and assumptions vary between past studies (e.g.,
76 Leeman et al. 2005; Lee et al. 2009; Till et al. 2012a; Kimura et al. 2014). Similarly, the
77 different thermobarometers used in past studies incorporate different assumptions about
78 the mantle residue composition (e.g., harzburgite: Mitchell and Grove 2015 vs. lherzolite:
79 Till et al. 2012a vs. olivine + orthopyroxene in residue: Lee et al. 2009). In addition, there
80 are a multitude of methods for reconstructing the volatile content of primary arc liquids,
81 specifically the H₂O content (e.g., Sisson and Grove 1993; Baker et al., 1994; Wade et al.
82 2008; Kelley et al. 2010; Ruscitto et al. 2010; Waters and Lange, 2015; Mitchell and
83 Grove 2015), which has a significant effect on the pressure and temperature returned by
84 thermobarometry. As a result, a direct comparison of existing thermobarometric
85 estimates is problematic.

86 In order to provide a global-scale comparison, this paper compiles the results of
87 published studies of mantle-melt thermobarometry conducted on erupted natural
88 primitive arc lavas, tephra, matrix glass compositions, and melt inclusions. The methods
89 used in these studies are reviewed, such that the effect of the methodology on the
90 calculated pressure and temperature can be quantified and potential pitfalls identified.
91 These results are then used to recommend a series of best practices for calculating
92 primary liquid compositions from erupted primitive magmas and conducting the

93 associated thermobarometry. These best practices are subsequently applied to a subset of
94 the published primitive arc rock and melt inclusion compositions where there is adequate
95 information in the source publication to carry out the recommended methods. These
96 recalculated pressures and temperatures are then used to address overarching questions in
97 arc magma genesis, such as the processes that govern the production of the most common
98 types of primitive arc magmas, as well as provide observational constraints on the
99 thermal structure and melt flow in the mantle wedge below arcs.

100 **Literature Review**

101 A literature review of primitive arc liquid thermobarometry is presented here and
102 restricted to thermobarometry that provides constraints on the mantle origin and
103 evolution of arc magmas. Primary magmas are melts that have not been chemically
104 modified in any manner since they segregated from their source region. In reality, all
105 magmas experience some processing en route to the surface, which consists of crystal
106 growth (fractional crystallization) and/or mixing of the magma with new materials
107 (assimilation) or other magmas (magma mixing or recharge), such that they are instead
108 called primitive magmas. The P-T constraints reviewed here are from natural primitive
109 magmas or melt inclusions generated in the mantle wedge in the range between arc fronts
110 and back-arcs.

111 Two methods of constraining the pressures and temperatures using natural
112 primitive arc magmas are included in this literature review of mantle-melt
113 thermobarometry (Fig. 1). The first method is thermometric and barometric calculations
114 for primitive arc-related magmatic liquids, where bulk rock, matrix glass or melt

115 inclusion compositions are used with thermobarometers such as Mitchell and Grove
116 (2015), Till et al. (2012a), Lee et al. (2009), or Putirka (2007, 2008) following
117 calculations to reverse fractional crystallization, post-entrapment crystallization etc.
118 These studies are designed to calculate the temperature and pressure at which the liquid
119 was in equilibrium with a peridotitic mantle (rather than the pressure or temperature of
120 crystal fractionation, for example). The second method of constraining primitive liquid
121 pressure and temperature in the literature compilation is experimental location of the
122 multiple saturation points for primitive arc magmas. For these cases, a primitive magma
123 or a synthetic oxide mixture of the same composition is used as the starting composition
124 to determine the phase relationships for this sample over the portion of pressure-
125 temperature space relevant to the upper mantle below volcanic arcs. The mineral phase
126 boundaries are thus located, such that the point in pressure-temperature space where a
127 melt is in equilibrium with a peridotitic mineral assemblage at given H₂O content, or the
128 *multiple saturation point*, can be determined. These experiments are also used to
129 calibrate many of the thermobarometers used in the first method. For the types of
130 experiments covered in this review, the peridotitic assemblages at the multiple saturation
131 point are either lherzolithic (olivine + clinopyroxene + orthopyroxene + aluminous phase
132 (plagioclase, spinel, garnet)) or harzburgitic (olivine + orthopyroxene).

133 Thermobarometric studies in the literature were filtered for inclusion in this
134 compilation with the following requirements: 1) that the samples used for
135 thermobarometry were natural rock compositions erupted in an arc setting past or present,
136 2) that the samples were “primitive” as identified by the authors, and 3) that their
137 pressure and temperature of melting or melt segregation from the mantle (or melt

138 entrapment in the case of melt inclusions) were estimated with thermobarometry or
139 experimentation (Table 1). The samples used for mantle-melt thermobarometry in these
140 studies have a wide compositional variation, wider than what is usually considered
141 primitive (Fig. 2) as no filters based on composition were applied for inclusion in the
142 literature compilation in Figure 1. The source publications for melt inclusion with <57
143 wt% SiO₂, Mg# > 0.5 and locations from arc settings in the GEOROC database were also
144 queried for studies that met the criteria listed above. This review only includes melt
145 inclusion pressure-temperature studies where the author identified the melt inclusion
146 composition as reflecting a primitive liquid composition and both a temperature and
147 pressure of mantle melt equilibration were determined. A few of the samples in the
148 Cascades have been utilized for pressure-temperature determinations by multiple studies,
149 and are represented as distinct points for each study in the literature compilation in Figure
150 1, such that not every plotted point is a unique sample. In total the literature compilation
151 includes 638 independent estimates of the temperature and/or pressure of melt
152 segregation for natural primitive arc magmas from 35 references (Table 1; electronic
153 supplement Table S1).

154 Published thermobarometric estimates for primitive arc magmas and melt
155 inclusions compiled directly from the literature vary between ~6-50 kbar and ~1000-
156 1600°C (n=638) (Figure 1; electronic supplement Table S1). The Cascades (n=410) and
157 Izu-Bonin-Marianas (IBM) (n=131) arcs have the largest number of samples investigated
158 to date and span the complete range of pressures and temperatures in the literature
159 compilation with the exception of several experimental samples from Japan that suggest
160 lower temperatures of mantle equilibration (<1150°C).

161 **Review of the Methods for Mantle-Melt Thermobarometry in the Literature**

162 **Compilation**

163 Here the methods employed to arrive at the pressure-temperature estimates in
164 Figure 1 are reviewed in order to quantify the effect of different methodologies and
165 identify best practices for an internally consistent set of new pressure-temperature
166 calculations (presented in “Recalculation of Pressures & Temperatures for Common Arc
167 Magma Types” section below).

168 **Calculating the Primary Liquid Composition through Reverse Fractional**
169 **Crystallization**

170 To obtain the P-T conditions of mantle partial melting, a melt inclusion, bulk rock
171 or matrix glass composition first must be adjusted for crystal fractionation, until it is in
172 equilibrium with an assumed mantle olivine composition and/or mantle mineral
173 assemblage (i.e., lherzolite, harzburgite or dunite). In addition, calculations to adjust for
174 post-entrapment crystallization and/or diffusion may be necessary to return the
175 composition of primitive melt inclusions to their original primary composition.

176 The fractional crystallization paths of mid-ocean ridge basalts (MORB) have been
177 well established through experimental and petrologic studies (e.g., Tormey et al. 1987;
178 Grove et al. 1992; Yang et al. 1996). MORBs tend to fractionate olivine and plagioclase
179 and the critical variable in reconstructing fractionation paths is pressure. Calculations to
180 adjust for fractional crystallization become more complex for primitive arc magmas and
181 other primitive magmas erupted in continental settings, as their phase assemblages and
182 sequences of crystallization are dependent on other variables besides pressure. The key is

183 to have a set of lavas related by fractional crystallization that can be used to identify 1)
184 the proportion of the phases that crystallized, 2) the degree of crystallization of the
185 magma with a given phase assemblage, and if applicable 3) the switching point between
186 sets of co-crystallizing phase assemblages (i.e., when the magma leaves a cotectic) (see
187 full review of reverse fractionation methods in the supplementary materials). An
188 alternative approach is to restrict thermometry and barometry to primitive samples that
189 have experienced a minimal amount of crystal fractionation, such that they have only
190 experienced olivine crystallization, which is more straightforward to adjust for. For
191 example, for MORB, FeO* increases and CaO drops sharply when plagioclase joins
192 olivine and a plot of FeO* or CaO vs. MgO can reveal samples that fall on an olivine
193 control line vs. those that experienced multiphase fractionation.

194 Failure to adequately approximate the fractionation path, and in particular
195 ignoring the requirement of the final liquid plotting on a mantle multiple saturation point,
196 can have a significant effect on the adjusted liquid composition and thus the calculated
197 source pressure and temperature. For example, if a polybaric, near fractional MORB
198 primary melt experienced multiphase fractionation (olivine, followed by
199 olivine+plagioclase) but is then adjusted for olivine-only fractionation, the calculated
200 “primary melt” has artificially high MgO + FeO* and low Al₂O₃ contents and thus the
201 calculated mantle conditions are 100-150°C hotter and up to 13 kilobars higher than the
202 actual pressure and temperature of generation for the parental MORB melt (see Fig. 6 of
203 Till et al. 2012a for a worked example). A variety of approaches to reverse fractionational
204 crystallization have been used for the primitive liquids with thermobarometry in the

205 literature collected in Figure 1, with the dominant approach being the addition of only
206 olivine back to the liquid composition.

207 H₂O also plays an important role in calculations to reverse fractionational
208 crystallization as it affects the order and composition of the fractionating phases. The
209 body of work on arc magma fractionation paths suggests that the onset of plagioclase
210 crystallization is the most important variable controlling the liquid line of descent of arc
211 magmas. Experimental studies of the liquid line of descent illustrate that increasing the
212 water contents at a constant pressure will cause the temperature of plagioclase
213 crystallization to go down relative to the liquidus, thereby reducing the proportion of
214 plagioclase crystallized compared to the ferromagnesian silicates and increasing the SiO₂
215 content of the residual liquid, as well as the relative temperature of magnetite or FeTi
216 oxide crystallization (e.g., Sisson and Grove 1993; Grove et al. 2003; Hamada and Fuji
217 2008; Tatsumi and Suzuki 2009; Parman et al 2011; Blatter et al. 2013). Although H₂O
218 suppresses plagioclase crystallization at any oxygen fugacity, the amount of H₂O required
219 to evolve the liquid to a calc-alkaline composition increases as oxygen fugacity
220 decreases. Thus increasing the H₂O content at a given *f*O₂ drives the liquid more directly
221 to the calc-alkaline field, essentially decreasing the concavity of the fractionation path as
222 demonstrated in experimental studies such as Hamada and Fuji (2008) and summarized in
223 Figure 3. Sulfur degassing can also cause a liquid to become more reduced during ascent
224 (Anderson and Wright, 1972; Kelley and Cottrell 2012; Moussallam et al. 2014), which
225 alters the Fe²⁺ content appropriate for the primary melt temperature and pressure
226 calculations (Brounce et al. 2014). The composition of the minerals that crystallize from
227 hydrous magmas also change, for example the Ca/Na of the feldspar varies as a function

228 of melt H₂O content (e.g., Sisson and Grove, 1993), which will affect the
229 thermobarometers that consider Ca and Na contents.

230 Although the theory for reverse fractional crystallization calculations for MORB
231 and nominally anhydrous arc basalts discussed above still applies to wetter primitive arc
232 magmas, there is not a well-accepted parameterization for the cumulative changes H₂O
233 causes in the liquid line of descent. The cutoff H₂O content above which MORB-like
234 fractionation paths no longer apply is not clear, but MORB-like fractionation schemes are
235 likely most applicable over the range of H₂O contents relevant for MORB genesis and the
236 higher the magmatic H₂O content, the more fractionation behavior will deviate from
237 MORB-like behavior. Kimura et al. (2014) and Kimura and Araskin (2014) are the only
238 studies in the literature review in addition to the experimental studies on primitive arc
239 magma pressures and temperatures that adjust the composition of the fractionating phases
240 according to H₂O content. Therefore, future work is required to parameterize the discrete
241 effects of H₂O during fractionation, such that we can ultimately calculate hydrous
242 primary arc magma compositions accurately.

243 **Calculating Pressure and Temperature Using Primary Magma**
244 **Compositions.**

245 Many experiments have demonstrated that the composition of the melt in
246 equilibrium with a mantle peridotite (lherzolite or harzburgite) varies as a function of
247 pressure, temperature, mantle source composition, and volatile content (e.g., Tatsumi et
248 al. 1981; Takahashi and Kushiro, 1983; Kushiro 1990; Kinzler and Grove 1992a; Parman
249 and Grove 1998; Gaetani and Grove 1998; Till et al. 2012a; Grove et al. 2013; Mitchell

250 and Grove 2015). Thus primary liquids composition can be used as a purely empirical
251 geothermometer as long as the appropriate restrictions in bulk composition and saturating
252 phases are made (Helz & Thonber 1987; Grove and Juster 1989). It follows that the
253 temperatures and pressures reported in the literature (Fig. 1) are a function of the models
254 that were used to calculate them. A full review and evaluation of the different
255 thermometers and barometers that have been applied to calculated primary arc magmas is
256 beyond the scope of this study (the reader is referred to existing reviews of thermometers
257 for volcanic rocks in Putirka 2008 and for melts in equilibrium with a harzburgite in
258 Mitchell and Grove 2015). Instead here the focus is to determine the spread of
259 temperatures and pressure in the literature compilation attributable to the use of different
260 thermobarometers, as well as make recommendations regarding the most reliable
261 thermobarometers for primitive arc liquids.

262 The spread of temperatures that can be attributed to the choice of thermometer
263 alone is on the order of $\leq 75^{\circ}\text{C}$ overall, and more likely to be $\leq 35\text{-}40^{\circ}\text{C}$ in most cases (see
264 electronic supplement). Recently published thermometers calibrated on the largest and
265 most robust databases of experiments such as Putirka et al. (2008) eqn. 4, Lee et al.
266 (2009), and Till et al. (2012a) tend to agree to within 35°C on average, which is within
267 the error of these thermometers (SEE Putirka= 52°C ; Lee= $\sim 40^{\circ}\text{C}$ (3%) and Till= 11°C).
268 These three thermometers also include a term to consider the effect of H_2O , which makes
269 them the most reliable and practical for the application to the continuum of dry to wet arc
270 magmas at present. However, Lee et al. (2009) explicitly states that their melt barometer
271 is “not intended for hydrous, unusually fertile or depleted mantle compositions that might
272 characterize subduction-modified mantles” because it is based on the activity of SiO_2 in

273 the melt, which varies with source mineralogy and water content, as well as pressure.
274 Therefore although calibration of Lee et al (2009) utilized a wide range of ultramafic melt
275 equilibria experiments where olivine and orthopyroxene were stable in the residue, it is
276 not well suited to melt generated in the garnet stability field, as the relationship between
277 melt SiO₂ and pressure diminishes above ~2.5 GPa when garnet is present in the residue
278 (T. Plank, personal communication, 2014), nor water-saturated systems, which also alter
279 the silica activity of the melt. Till et al. (2012a) and Grove et al. (2013) are updated
280 calibrations using the methodology of Kinzler and Grove (1992b), which require a
281 lherzolite residue (plagioclase or spinel lherzolite for Till et al. (2012a), and garnet
282 lherzolite for Grove et al. (2013)) but have the additional advantage over Kinzler and
283 Grove (1992a) of being calibrated to consider metasomatized as well as variably depleted
284 lherzolite. These lherzolite thermometers reproduce experimental liquids with the
285 smallest average absolute errors of all the thermometers examined, only 11°C for Till et
286 al. (2012a) and 24°C for Grove et al. (2013), and are thus extremely well suited to
287 application in arc environments, provided the liquid appears to be in equilibrium with a
288 lherzolite residue.

289 In the case where primary arc liquids are in equilibrium with Fo90-91 olivine but
290 are not in equilibrium with a lherzolite residue (e.g., Fig. 4b,c), the liquids may instead be
291 saturated with a harzburgite or dunite residue, natural examples of which are found in
292 exhumed sub-arc mantle sections and xenoliths (e.g., Kelemen et al., 1995; Pearce and
293 Parkinson 1993; Morishita et al. 2011; Pirard et al. 2013). How to identify liquids in
294 equilibrium with each of these potential residues is discussed in the supplementary
295 material (see electronic supplement Fig. S1 & S2). Mitchell and Grove (2015) developed

296 a thermobarometer similar to Till et al (2012a) and Grove et al. (2013) but specifically for
297 primitive liquids in equilibrium with harzburgite residues. The Lee et al (2009) and
298 Putirka et al. (2007) thermometers significantly under predict the temperatures of
299 experimental melts in equilibrium with harzburgite compared to Mitchell and Grove
300 (2015) and Till et al. (2012a), which are within error. With the addition of water to the
301 melt, the thermometers of Till et al (2012a) and Mitchell and Grove (2015) continue to be
302 within error of the experimental temperature, whereas the Lee et al. (2009) and Putirka et
303 al. (2007) are $\sim 100^\circ\text{C}$ hotter. A number of thermometers based on olivine-liquid
304 equilibria exist, including a new model that includes the effect of oxygen fugacity by
305 Putirka (2016), which can be used for liquids in equilibrium with dunite.

306 Differences in the assumed mantle $f\text{O}_2$ and thus the appropriate Fe^{2+} content of
307 the primary melt will also affect the temperatures calculated. Although a range of
308 approximately QFM-3 to QFM+2 has been estimated for the upper mantle (Frost and
309 McCammon 2008 and references therein), and QFM to NNO for oceanic basalts (Putirka,
310 2016), work on primitive arc magmas suggests a more oxidized range of oxygen
311 fugacities (\sim QFM+1 to QFM+2 or even MH) is appropriate for the mantle in subduction
312 zones (Brounce et al. 2014; Putirka 2016). The lower oxygen fugacities are thought to be
313 found in drier/back-arc regions and the higher fugacities in the hydrated mantle due to the
314 role of the subduction fluids/melts in oxidizing mantle wedge environments (e.g., Kelley
315 and Cottrell 2009; Brounce et al. 2014).

316 The H_2O contents for samples included in the literature compilation in Figure 1
317 have been estimated using a range of techniques and this is likely the variable for which
318 the quality of the constraints is the most heterogeneous. For example, some studies have

319 carefully determined the H₂O content of each individual sample through ion probe or
320 FTIR analyses of melt inclusions (e.g., Kelley et al. 2010; Ruscitto et al. 2010), where
321 others utilized mineral chemistry and hygrometry (e.g., Mullen & McCallum 2014), or
322 made comparison to experimental liquid lines of descent (e.g., Leeman et al. 2005, 2009;
323 Kimura et al. 2006), and others choose to assign all samples from a given subduction
324 zone the same water content (e.g., Lee et al. 2009). Even if the H₂O content of a primary
325 melt has been effectively estimated, the calculated effect of this H₂O on mantle
326 equilibration temperature and pressures differs between models and may vary
327 systematically with melt compositions. The thermobarometers of Putirka et al. (2007),
328 Lee et al. (2009) and Till et al. (2012a) include a calibration for the effect of H₂O on the
329 calculated liquid temperature for lherzolite melts. The Till et al. (2012a) calibration
330 results in the largest effect of H₂O on the liquid temperature, followed by Putirka et al.
331 (2007) and then Lee et al. (2009). The temperature difference predicted by these models
332 can be up to 67°C at 5 wt% H₂O for a calc-alkaline basalt, which is greater than the
333 uncertainty of these thermometers (further discussed in the electronic supplement). The
334 effect of H₂O on the liquid temperature can therefore introduce an equivalent or often
335 larger spread in temperatures in the literature compilation than the choice of which
336 thermometer was used (see electronic supplement). The effects of H₂O on pressure are
337 compared for the Lee et al. (2009) and Till et al. (2012) barometers in the electronic
338 supplement. The Lee et al. (2009) barometer has a larger model dependence on H₂O than
339 Till et al. (2012a) and the calculated pressure of equilibration is 2.5 kbar (or ~8.5 km)
340 higher with Lee et al. (2009) than Till et al. (2012a) for a calc-alkaline basalt with 5 wt%
341 H₂O. This variability in the pressure calculations as a function of H₂O content is

342 equivalent to or smaller than the range of pressures produced by using different
343 barometers or types of reverse fractionation calculations on the same sample.

344 Thus large variations in pressure and temperature in Figure 1 may be attributed to
345 different assumptions about H₂O content. And it follows that our ability to calculate the
346 pressures and temperatures at which primitive arc magmas are sourced in the mantle and
347 thus interpret mantle wedge processes is the most limited by the consideration of H₂O,
348 specifically by differences in how the primary magmatic H₂O-contents are estimated (or
349 not), the lack of consideration of H₂O's effects on the reverse fractionation crystallization
350 calculations in existing studies, and the differences between models for the effects of H₂O
351 on mantle equilibration pressure and temperature.

352 **Recalculation of Pressures & Temperatures for Common Primitive Arc Magma** 353 **Types**

354 **Methods**

355 A representative subset of the primitive arc samples with thermobarometry in the
356 literature review in Figure 1 and 2 (n=208 of 638 in literature review) were selected for a
357 reassessment of their reverse fractional crystallization calculations and a recalculation of
358 their pressures and temperatures of mantle equilibrium using an internally consistent set
359 of methods and the latest thermobarometric tools that consider H₂O (Table 2). Samples
360 were chosen whose source literature provided substantial information about the sample's
361 (or suite of samples') petrography and mineral compositions, oxygen fugacity and where
362 possible H₂O content. A wholesale recalculation for all samples used in the literature
363 compilation is not possible because many source publications contain insufficient

364 information to conduct the reverse fractionation calculations and/or estimate H₂O
365 content. However, the samples included in the recalculation are representative of the full
366 range of sample compositions in Figure 2 and calculated pressures and temperatures in
367 the literature compilation in Figure 1. Once investigated the samples fall into three
368 categories: 1) calc-alkaline basalts and low-K tholeiites, 2) primitive andesites, and 3)
369 primitive boninites, picrites and other high-MgO liquids (Fig. 4, 5), and the methods for
370 the recalculation of each category is discussed below (also see the worked example in the
371 supplementary material). Here we distinguish high-MgO liquids with <52 wt. % SiO₂
372 and >15 wt% MgO from primitive andesite with >52 wt% SiO₂ and <15 wt% MgO (av.
373 9-10 wt% MgO). In addition, three samples appear to have formed in the presence of
374 significant amounts of CO₂ or be the result of melting pyroxenite as demonstrated by
375 their major element compositions projected onto pseudo-ternary (see diagrams Figures 9
376 and 11, respectively and associated text in Grove et al., 2013) and are excluded from the
377 pressure-temperature calculations.

378 The subset of calc-alkaline and tholeiitic basaltic lavas or melt inclusions used for
379 the recalculations are from the Cascade (Leeman et al. 2005; Rowe et al. 2009; Till et al.
380 2013), Lesser Antilles (Pichavant, et al. 2010) and Mariana arcs (Kelley et al. 2010) and
381 are similar to the majority of the rock types in the literature with pressure-temperature
382 estimates. These samples tend to have experienced olivine, olivine +plagioclase,
383 olivine+clinopyroxene or olivine+plagioclase+clinopyroxene fractionation based on the
384 mineralogy of the samples and/or the liquid compositions and required calculations to
385 reverse between 4 to 26% crystallization (Fig. 6). These minerals were added back to the
386 whole rock composition (usually in proportions reflecting the modal proportions in the

387 sample), treating all Fe as FeO* and using partition coefficients appropriate for the H₂O
388 contents of the sample (e.g., varying the Ca-Na plagioclase-liquid K_D as a function of
389 H₂O) (see worked example in electronic supplement). The H₂O content of these samples
390 were determined via SIMS or FTIR on melt inclusion or matrix glass by several studies
391 and estimated using the methods listed in Table 2 for the remaining samples. The major
392 element compositions of these liquids are consistent with forming from a lherzolite
393 residue, such that the reverse fractionation calculations aims at returning them to a
394 lherzolite multiple saturation point and equilibrium with Fo90 mantle olivine (Fig. 4a).
395 These adjusted primary liquids were used with the lherzolite thermometers and
396 barometers of Till et al. (2012a) and Grove et al. (2013), which includes the effect of
397 H₂O.

398 The subset of primitive andesites used for the recalculations are those studied in
399 Mitchell and Grove (2015) and are from the Setouchi volcanic belt (Tatsumi & Ishizaka
400 1982; Tatsumi et al. 1983), the Cascades (Baker et al. 1994; Grove et al. 2002, Mitchell
401 & Grove, 2015), Kamchatka (Bryant et al. 2010), and the Trans-Mexico volcanic belt
402 (Weaver et al. 2011, Weber et al., 2011). These samples have high Mg#’s (0.71-0.76)
403 and their compositions are consistent with liquids in equilibrium with a harzburgite
404 residue as predicted by the model of Mitchell and Grove (2015) without any reverse
405 fractionation calculations (Fig. 4b, 5). Given the degrees of freedom, two of the three
406 variables of pressure, temperature and H₂O contents can be calculated for these liquids
407 using the thermometer, barometer and/or hygrometer for liquids in equilibrium with
408 harzburgite from Mitchell and Grove (2015). Here the thermometer and hygrometer were

409 employed and pressure was assumed based on the thickness of modern crust in these
410 locations, which is on average 30 km (~10 kbar).

411 The subset of boninites, picrites and other high MgO magmas (>13.5 wt% MgO)
412 used for the recalculations are dredge samples from the Bonin forearc (Li et al. 2013) and
413 Mariana (Bloomer and Hawkins 1987), subaerial samples from the Marianas (Dietrich et
414 al., 1978) and the New Britain arc (Cameron et al., 1983) and melt inclusions from the
415 Tonga trench (Sobolev and Danyushevsky 1994) and eastern Kamchatka (Kamenstky et
416 al. 1995). They represent the samples with the highest Mg#’s and high MgO contents in
417 the literature compilation (Fig. 2), as well as the samples that record the highest
418 temperatures and pressures of mantle equilibration in the literature compilation (Fig. 1).
419 Their high Mg#’s of 0.74-0.82 suggest these liquids were in equilibrium with Fo90-94
420 olivine with a Fe-Mg K_D of 0.3 or that K_D values were higher than 0.3 to be in
421 equilibrium with Fo90 olivine. As such, it was not necessary to adjust the liquids for
422 fractionation in order to return these samples to equilibrium with the mantle. When
423 plotted in pseudoternary space, these liquids exhibit lower plagioclase and clinopyroxene
424 components than the mantle lherzolite multiple saturation points suggesting they may be
425 in equilibrium with either a harzburgite or dunite residue instead (Fig. 4c, 5, S2). As the
426 liquids fall along the olivine-orthopyroxene saturation boundaries experimentally
427 determined by Wagner and Grove (1998), they may be the product of either harzburgite
428 melting or lherzolite melts that re-equilibrated with harzburgite in the lithosphere as they
429 ascended (Grove et al. 2013; Wagner and Grove 1998). A few of the samples with the
430 lowest clinopyroxene mineral component values plot in regions of liquids in equilibrium
431 with a dunite (see Fig. S1, S2) produced experimentally at very low melt-rock ratios (5 to

432 20% melt) at high temperatures 1220-1260°C found in the hot core of the wedge
433 (Mitchell and Grove, 2016). Given the similarity of the majority of the high-Mg liquids
434 to those in equilibrium with harzburgite, the Mitchell and Grove (2015) harzburgite
435 thermometer and hygrometer were applied to all of them. For samples with H₂O contents
436 measured via ion probe in melt inclusions or matrix glass, the measured values agree with
437 those estimated via the Mitchell and Grove (2015) hygrometer.

438 **Results**

439 The results of the recalculations show that the calc-alkaline and low-K tholeiitic
440 basaltic samples were last in equilibrium with a lherzolite residue at intermediate
441 pressures and temperatures of arc magma genesis between ~1130-1390°C at 8.5-19 kbar
442 at 0-6 wt% H₂O (Fig. 7). The primitive andesites that were last in equilibrium with a
443 harzburgite residue at intermediate to high H₂O contents and represent the lowest
444 temperatures and intermediate to low pressures of last equilibration between ~1075-
445 1260°C with H₂O contents of 3.2-7.2 wt% H₂O at 10 kbar. If pressures of last
446 equilibration were in fact slightly higher at ~15 kbar, the liquids represent mantle
447 equilibration temperatures of 1090-1270°C at 6-10 wt% H₂O. The primitive boninites,
448 picrites and other high-MgO liquids were last in equilibrium with a harzburgite residue
449 on average and represent the highest temperatures over the entire range of pressures. Four
450 of these samples are olivine-hosted melt inclusions (Kamenetsky et al. 1995; Sobolev and
451 Danyushevsky 1994) and record temperatures of 1400-1450°C at 18 to 19 kbar using the
452 H₂O content of the melt inclusions determined by ion probe (0.60-1.4 wt% H₂O). The
453 other seven high MgO liquids, including boninite from the IBM arc (Bloomer and

454 Hawkins 1987; Dietrich et al. 1978; Li et al. 2013), record shallower conditions of last
455 mantle equilibration of 1310-1385°C at 14-10 kbar at 1.0-1.6 wt% H₂O.

456 Overall the recalculated fractionation paths and thermobarometry yielded lower
457 temperatures ($\Delta T = -1-198^\circ\text{C}$, $n=166$) than the published literature estimates with only a
458 few samples from the Cascades yielding higher temperatures ($\Delta T = +22-327^\circ\text{C}$, $n=7$)
459 (Table 2). The recalculated pressures also tend to be lower than the literature values
460 ($\Delta P = -1-21$ kbar, $n=168$) with a few being higher ($\Delta P = +0.5-19$ kbar, $n=5$). These lower
461 temperatures and pressures largely result from using multiple phases that reflect the
462 sample's mineralogy to adjust for fractional crystallization, as well as the use of modern
463 thermobarometers for lherzolite and harzburgite residues (Till et al. 2012a; Grove et al.
464 2013; Mitchell and Grove 2015). This exercise reinforces the large effect of a reverse
465 fractionational crystallization calculation scheme consistent with the rock mineralogy and
466 H₂O content, as well as imposing the criteria that the major element composition of the
467 fractionation-adjusted liquid match the composition of experimentally-determined liquids
468 in equilibrium with the appropriate mantle residue (e.g., lherzolite vs. harzburgite), in
469 addition to the appropriate mantle olivine forsterite content. Because the recalculated
470 samples included those with the maximum and minimum pressures and temperatures
471 reported in the literature, these recalculations suggest the range of pressures and
472 temperatures recorded by primitive arc magmas is in fact much smaller than suggested in
473 Figure 1. Instead arc primitive arc magmas likely only record last pressures and
474 temperatures of equilibration between 1050-1450°C at 8-19 kbar, rather than up to
475 1600°C at 50 kbar as reported in the literature (Fig. 7).

476

Discussion

477 Now that an assessment of the variability in the literature compilation due to
478 methodology has been made and a subset of the primary arc magma compositions have
479 been used to recalculate their conditions of mantle equilibration using internally
480 consistent methodology (hence forward referred to as the ‘recalculated compilation’),
481 there is the opportunity to interrogate the new P-T calculations for what they reveal about
482 the underlying mantle processes.

483 **Mantle Melting Processes**

484 In part, the pressure and temperature variations in the recalculated compilation are
485 the result of the multiple mantle melting processes. The tholeiitic arc basalts are thought
486 to be generated by adiabatic decompression melting of nominally anhydrous, hot mantle
487 being advected into the mantle wedge during corner flow (e.g., Grove et al. 2002; Sisson
488 and Bronto 1998). The calc-alkaline lavas on the other hand likely result from hydrous
489 flux melting, where a slab-derived H₂O-rich component initiates melting at the vapor-
490 saturated lherzolite solidus at the base of the mantle wedge and the buoyant melt ascends
491 into the hot core of the mantle wedge (e.g., Grove et al. 2003; Till et al. 2012b). As it
492 rises, this melt equilibrates with the hotter mantle, dissolving mantle minerals to increase
493 the melt fraction and lower the melt H₂O-content. Based on the calculations of the likely
494 mantle residue composition here and in previous studies, the high-MgO arc primitive
495 magmas and primitive arc andesites are likely the result of these same processes causing
496 melting of harzburgite rather than lherzolite.

497 Once the first melt is formed, the continuation of mantle melting can be
498 represented by two end-member categories; that where there is equilibration between the

499 melt and solid at all times so that the bulk composition is fixed (i.e., equilibrium or batch
500 melting) and that where the liquids are extracted as soon as they form so that the bulk
501 composition of the residual solid changes (i.e., fractional melting). The major and trace
502 element composition of MORBs reveal that they are the product of polybaric near-
503 fractional melting (e.g., Johnson et al. 1990; Langmuir et al. 1992) and record the
504 *average pressure and temperature of the mantle melting column* (e.g., Kinzler and Grove
505 1992b). In contrast, a number of detailed studies suggest that anhydrous to damp
506 primitive magmas erupted in arc settings appear to be the product of batch melting (e.g.,
507 Bartels et al. 1991; Bacon et al. 1997; Kent and Elliott 2002; Kelley et al. 2010; Till et al.
508 2012a), such that they only record *the last pressure and temperature of equilibration with*
509 *the mantle* rather than the average pressure and temperature as in near-fractional melting.
510 This hypothesis is further tested here using the forward mantle melting model of Behn
511 and Grove (2015), which is built on the formulation of Kinzler and Grove (1992a, 1992b,
512 1993) and Kinzler (1997) for MOR-melting and incorporates new experiments from Till
513 et al. (2012a) on metasomatized and depleted mantle melting so as to make the model
514 appropriate for melting beneath arcs. ~1-10% isobaric batch melting of a depleted Hart
515 and Zindler (1986) mantle (HZ-Dep1 in Table 1a of Kinzler and Grove (1992b)) in the
516 spinel lherzolite field at 10-20 kbar overall reproduces the major element composition of
517 the calc-alkaline basalts used for the recalculations (Fig. 8b). The tholeiitic basalts with
518 recalculated pressures and temperatures can be fit by incremental batch melting of the
519 same mantle composition with 90% melt extraction at each step (i.e., near fractional
520 melting) between 20-9 kilobars, dF/dP of 1% per kilobar and an adiabatic gradient of
521 1.5°C per kilobar over a range of mantle potential temperatures between ~1450-1300°C.

522 They can also be fit by batch melting curves like those that fit the calc-alkaline basalts
523 (Fig. 8a). Therefore the calc-alkaline basalts reviewed here, and perhaps also the
524 tholeiitic basalts, record their last pressure and temperature of mantle equilibration (i.e.,
525 the conditions of melt extraction), not the initiation of melting. These equilibration
526 conditions are commonly misinterpreted as indicating shallow and hot melting beneath
527 arcs. Instead this interpretation reinforces prior observations by studies such as Kelley et
528 al. (2010), Weaver et al. (2011), and Till et al. (2013) that primitive arc magmas tend to
529 re-equilibrate near Moho depths as they rise from their deeper points of origin. This is
530 further illustrated when the recalculated compilation is compared to geodynamics models
531 of the temperature distribution within the mantle at modern subduction zones in the
532 following section.

533 In the case of melt inclusions, the recorded temperatures and pressures reveal the
534 conditions at which the primitive melt was trapped by the host mineral. For the majority
535 of cases in this study the host mineral is olivine, with the remainder being clinopyroxene.
536 Therefore, they record the pressures and temperatures during olivine or clinopyroxene
537 crystallization. In the literature compilation and the recalculated compilation, the
538 temperatures and pressures of melt inclusion formation overall appear to be similar to
539 those where primitive magmas are extracted from the mantle.

540 **Geodynamic Models**

541 Geodynamic models with increasing complexity have been applied to solid-state
542 mantle convection in the mantle wedge. Vertical paths through a suite of modern models
543 are compared to the recalculated magma thermobarometry in Figure 9. The recalculated

544 pressures and temperatures for damp to wet magmas tend to match the thermal structure
545 of the Kelemen et al. (2003) models, while those from the nominally anhydrous magmas
546 reflect higher temperatures at a given pressure than predicted by any model. Kelemen et
547 al. (2003) compared the much more limited set of petrologic constraints on the pressure-
548 temperature conditions for arc magmas and sub-arc crust available at the time to existing
549 geodynamic models, and geared their modeling efforts toward reproducing the natural
550 observations.

551 The comparison of the recalculated magmatic pressure-temperature compilation
552 to the thermal structures predicted by dynamic models suggest that the magmas
553 experience thermal equilibration in the hottest shallowest nose of the mantle wedge
554 before they are extracted (Fig. 10). Syracuse et al. (2011) predict maximum temperatures
555 in the hot core of the wedge that vary between 1200°C or 1275°C (depending on the
556 location of full coupling between the mantle and slab) to 1459°C at different subduction
557 zones with an average of $\sim 1400 \pm 54$ °C. This suggests re-equilibration in the hot shallow
558 nose of the mantle wedge at different arcs could lead to a natural variation in magmatic
559 temperatures of ~ 200 -250°C. While the range of recalculated pressures and temperatures
560 broadly match Syracuse et al. (2011), in some cases the magmas are warmer than the
561 model predictions at specific arcs. For example, magmatic temperatures are as warm as
562 1400°C below the Cascades and 1450°C below the Kamchatka & Tonga arcs, while the
563 model predicts maximum temperatures beneath Cascadia of 1285-1312°C and 1300°C
564 below the Kamchatka & Tonga arcs. This may be in part due to the limitations of
565 modeling mantle wedge thermal structures in two dimensions.

566 Shown for comparison in Figure 9 are the maximum pressure-temperature
567 conditions for subduction zone blueschists and eclogites and thermal models for the
568 subducting slab (Penniston-Dorland et al. 2015). The peak thermobarometric conditions
569 recorded in exhumed metamorphic rocks are on average 100-300°C warmer than the
570 models, and the greatest discrepancies occur at <2 GPa. Penniston-Dorland et al. (2015)
571 argue that the omission of significant shear heating (up to 250°C at 35 km depth) and the
572 exothermic hydration reactions within the overlying mantle just above the slab-wedge
573 interface (<200° at 1 GPa for a flux of ~0.1 kg H₂O/m²/yr from slab) are two of the most
574 significant potential causes of this discrepancy. While the physics of heating up a slab
575 are simpler than predicting the temperature of convecting mantle in the wedge, these
576 features could also account for some of the discrepancy between the hottest primitive arc
577 magma samples and the dynamic models. However the dimensionalization of
578 temperature in the dynamic models and the prescribed boundary conditions are likely
579 more significant factors controlling the discrepancies between the petrologic estimates
580 and geodynamic models. For example, geodynamic models may underestimate the
581 temperatures possible at the shallowest depths because of the prescribed lithospheric
582 thickness in the models (e.g., 45-55 km for van Keken et al. (2002) and Kincaid and
583 Sacks (1997)). Results from Till et al (2013) for the southern Cascadia subduction zone
584 suggest the continental lithosphere must be ≤35 km thick. The observation that the
585 warmer, driest arc basalts in the recalculated compilation require adiabatic
586 decompression melting of asthenospheric mantle supports the interpretation that the
587 convecting mantle extends to an average depth of ~30 km or less (~10 kbar) at arcs, even
588 at arcs with an overriding continental plate.

589 In addition, strong focusing mechanisms that direct fluids and melts to hot and
590 shallow regions beneath the arc may help explain the abundance of shallow, hot arc
591 magmatic temperatures. Wilson et al. (2014) develop models that incorporate strong
592 temperature-dependent rheologies in the slab and the wedge, and a physically reasonable
593 model of fluid flow that includes the interaction of fluid transport with solid rheology in
594 the form of compaction pressure. Similarly, Wada and Behn (2015) examine the effects
595 of grain size on fluid flow in the mantle wedge. These models are able to reproduce the
596 localization of fluids and melts to the subarc region with this more realistic permeability
597 and solid viscosity structure. Although these models do not predict temperatures as warm
598 as many of the magmatic temperatures in the recalculated compilation, they suggest a
599 mechanism to explain the clustering of the P-T points at the shallowest pressures. These
600 focusing mechanisms combined with the petrologic observations of shallow last
601 conditions of mantle equilibration also provide an explanation for why wet and dry arc
602 magmas are erupted in close spatial and temporal proximity at many arcs such as the
603 Cascades (e.g., Till et al. 2013; Carlson et al. submitted).

604 **Melt Flow Mechanisms**

605 In addition to the various melting mechanisms for primitive arc magmas, the
606 rising melt flow behavior also controls the ultimate pressure and temperature recorded.
607 The maintenance of equilibrium between the melt and the mantle requires reactive porous
608 flow as a mechanism for transporting the melts rather than diapiric or channelized flow
609 (e.g., Navon and Stolper 1987; Grove et al. 2002). Reactive porous flow can be
610 approximated as Darcy flow where permeability exerts the main control on the melt
611 ascent rate and whether or not the melts can achieve thermal and chemical equilibrium.

612 Provided permeability is sufficient for the fluids to outpace subduction, small degree
613 batch melts will re-equilibrate with the hotter overlying mantle, dissolving silicate
614 minerals and diluting the H₂O content as they rise (Grove et al. 2002). Alternatively, if
615 melt flow occurs as diapiric or channelized flow, it can be approximated by Stoke's flow
616 and the size of diapirs is the main control on whether thermal and chemical equilibration
617 with the surrounding mantle will occur. Modeling suggests that for diapirs large enough
618 to escape subduction flow, the ascent rate is too rapid for thermal equilibration with the
619 surrounding mantle to occur (Grove et al. 2002).

620 Thermal gradients in the mantle wedge can be up to 30-40°C/km based on
621 geodynamic models (Cagnioncle et al. 2007; Syracuse et al. 2010). Therefore in the
622 reactive porous flow model required for batch melting (Fig. 8), a difference in the
623 pressure of last equilibration of 10 vs. 15 kbar (~15 km) equates to a difference in the
624 temperature of last equilibration of up to 450°C (Fig. 10). This difference in last
625 equilibration conditions is similar to the difference between the coolest primitive
626 andesites and the hottest boninites, including notably within the Kamchatka arc where
627 these conditions are recorded within the same arc. Geochemical and isotopic modeling
628 suggests >90% of the major element abundances in primitive hydrous arc magmas can be
629 explained as a product of flux melting and ascent via reactive porous flow (e.g., Grove et
630 al. 2002), which is also supported by the composition of mantle xenoliths and field
631 observations from the roots of arcs (e.g., Kelemen et al., 1992; Bouihol et al., 2009).
632 Therefore, a likely explanation for the calc-alkaline basalts and primitive andesites with
633 temperatures lower than the anhydrous peridotite solidus is that these magmas re-
634 equilibrate as they rise, and record the decrease in temperature during ascent out of the

635 hot nose of the wedge (dashed teal ascent path in Fig. 10). Variability within the
636 pressures and temperatures recorded by these hydrous magmas can be attributed to
637 variations in subduction zone thermal structure through time, along strike, or between
638 arcs (e.g., Carlson et al. submitted).

639 Alternatively, if channelized flow and a fluid adiabat of 1°C/km are assumed
640 (Nisbet 1982), a 5 kbar difference in a magma's last pressure of equilibration equates to
641 only a ~15°C difference in the temperature. There is evidence that the incompatible trace
642 element budget of hydrous arc magmas is contributed from a fluid and/or melt
643 component present in the mantle that rises via adiabatic diapiric or channelized flow and
644 does not re-equilibrate (e.g., Grove et al. 2002; Pirard and Hermann 2015). However, if a
645 hydrous magma forms at or near the water-saturated solidus at 30 kbar, adiabatic ascent
646 to 10 kbar would lower magmatic temperatures on the order of 60°C and induce
647 crystallization (see Grove et al. (2011) Fig. 2). As no magmas in the recalculated
648 compilation record temperatures of $\leq 1000^\circ\text{C}$ at any pressure, these magmas either
649 crystallize before they reach the surface, or water-rich magmas do not rise via diapiric
650 flow. Instead the recalculated compilation suggests adiabatic ascent may only be possible
651 for the nominally anhydrous tholeiitic magmas (Fig. 8).

652 Models that include compaction pressure such as Wilson et al. (2014) represent an
653 intermediate melt/fluid flow mechanism on the continuum between reactive porous and
654 channelized flow models, which could also be responsible for intermediate temperature
655 magma types. Alternatively, reactive porous flow may operate until there is a change in
656 mantle permeability that causes a transition to channelized flow (e.g., Kelemen et al.
657 1997, Aharonov et al. 1997). Future work is required to determine if these hypotheses

658 about melt flow are robust. The recalculated compilation provides a powerful set of
659 pressure, temperature and compositional observations to test any proposed model.

660 **Summary of Thermobarometry and Mantle Origins for Four Types of Primitive**
661 **Arc Magmas**

662 Low-K tholeiitic arc basalts represent nominally anhydrous lherzolite melts and
663 record high average temperatures (~1300-1390°C) between 10-20 kbar, which are
664 generated by adiabatic decompression melting in the back limb of corner flow (Fig. 10).
665 These magmas tend to follow more tholeiitic liquid lines of descent in the crust similar to
666 MORB's due to their low H₂O and more reducing fO_2 . Thus the appropriate reverse
667 fractionation adjustment for these samples are the easiest to predict. These samples are
668 the best suited to modern lherzolite thermobarometers (e.g., Till et al. 2012a; Lee et al.
669 2009; Putirka et al. 2008) which yield temperatures of origin within 30°C on average,
670 making their pressures and temperatures of mantle equilibration the most reliable.

671 Calc-alkaline arc basalts record comparatively lower average temperatures (1100-
672 1300°C) over the same pressure interval due to melting lherzolite in the presence of
673 higher H₂O contents (>1 wt%). The lower temperature samples at a given pressure likely
674 represent melts generated at or near the H₂O-saturated solidus, which rise through the
675 mantle via reactive porous flow (Fig. 10). Higher temperature calc-alkaline basalts (1250-
676 1300°C) may be the result of the same process or adiabatic ascent from an H₂O-
677 undersaturated solidus due to lower H₂O contents. The higher H₂O and fO_2 of calc-
678 alkaline basalts result in their more complex and variable fractionation paths in the crust.
679 Thus caution is required when reverse-fractionating these samples and the more

680 information about the suite of samples, their mineral contents, mineral compositions and
681 H₂O contents, the better the chance of an accurate fractionation adjusted composition.
682 These samples are also appropriate for use with the lherzolite thermobarometers but will
683 yield different results depending on the thermobarometer chosen because of the different
684 calibrations for the effect of H₂O on their mantle equilibration pressure and temperature.

685 Primitive high-Mg andesites record the lowest temperatures at a given pressure in
686 the recalculated compilation and are generated by 20-30% melting of harzburgite residue
687 that has been enriched by alkalis during metasomatism over a range of H₂O contents (0-7
688 wt%). The primitive nature of the samples in the recalculated compilation is such that
689 they do not require any reverse fractionation calculations to be in equilibrium with the
690 mantle, although liquid lines of descent for these primitive rock types have been studied
691 by Grove et al (2003) and can be used to restore these samples to liquids in equilibrium
692 with the harzburgite. The composition of harzburgite-derived mantle melts and their
693 associated pressures, temperatures and H₂O contents have been experimentally calibrated
694 by Mitchell and Grove (2015). If these liquids are used with lherzolite thermometers or
695 similar, they will yield higher temperatures, in error by up to 250°C.

696 High-MgO arc magmas, such as boninites and picrites, record the highest
697 temperatures for a given pressure in the recalculated compilation and their major element
698 compositions are consistent with either harzburgite melting or lherzolite melts that re-
699 equilibrated with harzburgite or dunite as they ascended (Grove et al. 2013; Wagner and
700 Grove 1998; Mitchell and Grove, 2016) at relatively low H₂O contents. Their pressures
701 and temperatures are consistent with thermal re-equilibration in the hottest nose of the
702 mantle wedge and in some cases reactive porous flow to somewhat shallower conditions.

703 These samples also tend to be so primitive that no reverse fractionation calculations are
704 required to be in equilibrium with the mantle. Little to no experimental work has been
705 done to constrain their fractionation paths. These samples will also yield a wide range of
706 temperatures when used with lherzolite thermometers, which vary by almost 200°C. The
707 harzburgite thermobarometer of Mitchell and Grove (2015) appears to be the most
708 appropriate given the composition of the liquids examined here, and yields temperatures
709 at the lower end of the range.

710 **Implications**

711 The collection of published mantle pressure-temperature constraints from
712 primitive arc magmas and the recalculations presented here provides observational
713 constraints for shallow mantle processes at subduction zones relevant to a variety of
714 disciplines. For example, in addition to the petrologic and geochemical perspective on
715 how much of the range of calculated magmatic pressures and temperatures at arcs is
716 “real” and likely melt generation models, the recalculated compilation provides
717 observational constraints for geodynamic models of the thermal structure and melt flow
718 at subduction zones. The magmatic pressures and temperatures also provide the
719 opportunity to further quantify the effect of melt on seismic velocity, attenuation, and
720 electrical conductivity in the upper mantle below arcs and to continue to evolve three
721 dimensional maps of melt distribution at subduction zones.

722 A number of opportunities for future petrologic and geochemical study also arise.
723 This paper focuses on the what can be learned from the major element composition of arc
724 magmas and a logical next step is to interrogate the trace element and isotopic

725 compositions of the magmas in the literature and recalculated compilation to further
726 interrogate and differentiate between magmas formed by decompression melting and flux
727 melting. As discussed in the review of reverse fractional crystallization calculations
728 above, there is a need to develop robust parameterizations for the compositional effects of
729 variable H₂O contents on the liquid line of descent for primitive arc magmas over the
730 range of pressures and oxygen fugacities at which they crystallize in the arc lithosphere.
731 In addition, the recalculated magmatic pressure-temperature compilation can ideally be
732 used to understand the extent to which variations in the conditions of mantle equilibration
733 can be attributed to variations between arcs with different subduction parameters (slab
734 dip, convergence rate etc.), as well as variations along strike within an arc.

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1159

1160 **Figure & Table Captions**

1161 **Figure 1. Compilation of primitive arc magma thermobarometry in the literature.**

1162 The pressure and temperatures compiled here were calculated for primitive arc liquids by
1163 the studies listed in Table 1. The points are color-coded based on the technique used by
1164 each study to calculate the temperature and pressure of origin in the mantle and the water
1165 content. The anhydrous peridotite solidus is from Hirschman (2000) and the H₂O-
1166 saturated peridotite solidi from Till et al. (2012b). Inset is same plot color coded
1167 according to subduction zone where each sample is found. The bulk of the data is from
1168 three subduction zones: 410 are from the Cascades, 131 are from the Izu-Bonin-Marianas
1169 and 21 are from Japan (n=638 total).

1170 **Figure 2. Harker diagrams.** Compositions of primitive arc samples taken directly from
1171 studies in Table 1 and Figure 1, all of which are reported prior to reverse fractional
1172 crystallization calculations or any post entrapment crystallization corrections in the case
1173 of melt inclusions. Samples with red, blue, green and pink colored symbols were used to
1174 for the new pressure and temperature calculations.

1175 **Figure 3. Summary of the liquid line of descent for a hydrous calc-alkaline basalt.**

1176 Arrows represent the liquid lines of descent for high MgO basalt following olivine
1177 fractionation at different conditions based on the experiments of Hamada and Fuji (2008).
1178 The bottom two plots represent experiments conducted at 2 kbar, and the top two
1179 illustrate experimentally-determined liquid evolution at 7 kbar. Dark blue arrow
1180 represents experiments with the higher H₂O content in each plot. Comparison amongst
1181 the four plots illustrates differences in the liquid line of descent due to changes in oxygen
1182 fugacity, H₂O content and pressure.

1183 **Figure 4. Pseudo-ternary projections for representative arc calc-alkaline and**

1184 **tholeiitic arc basalts, primitive andesites and high MgO liquids. a)** Pseudo-ternary
1185 projections depicting the compositions of representative tholeiitic and calc-alkaline arc
1186 basalts prior to reverse fractional crystallization calculations. These samples are shown
1187 along with the location of a melt in equilibrium with plagioclase (blue y's), spinel (green
1188 y's) and garnet (red y's) lherzolite, also known as the lherzolite "multiple saturation
1189 points", over a range of pressures from Till et al. (2012a) and Grove et al. (2013), which
1190 are plotted in all three diagrams. These liquids in a) are relatively silica-understaturated
1191 and plot towards the plagioclase apex because they originated from a lherzolite residue.
1192 **b)** Pseudo-ternary projections depicting the compositions of representative primitive arc
1193 andesites and **c)** high MgO arc magmas such as boninites and picrites illustrated prior to
1194 any reverse fractional crystallization calculations. Overall both the liquid types in b) & c)
1195 are relatively silica-saturated and have a comparatively lower plagioclase component
1196 relative to the more typical arc basalts because they are in equilibrium with a more
1197 depleted mantle residue (i.e., harzburgite or in the case of a few high-MgO liquids
1198 dunite), rather than lherzolite. Samples plotted in all three figures are those used in the
1199 new calculations of pressures and temperatures with the exception of the black circles in
1200 c) that included to illustrate the range of high MgO primitive arc magma compositions in
1201 the literature.

1202 **Figure 5. Recalculated Primitive Arc Magma Compositions.** Composition of samples
1203 included in the new pressure and temperature calculations compared to the calc-alkaline
1204 vs. tholeiitic fields of Miyashiro (1974). Arrow indicates progressive depletion of the
1205 mantle residue that sources the primitive magmas (lherzolite>harzburgite>dunite).

1206 **Figure 6.** Comparative amount of crystal fractionation to be reversed prior to the new
1207 pressure and temperature calculations for the tholeiitic and calc-alkaline primitive basalts.
1208 Black curves represent the composition of 1-20% isobaric batch melts for a depleted Hart
1209 and Zindler (1986) mantle composition at 10, 15, 20 kbar as predicted by the forward
1210 lherzolite melting model of Till et al. (2012a) as modified by Behn and Grove (2015).
1211 These curves illustrate the composition of primary nominally anhydrous melts prior to
1212 crystal fractionation. Crystal fractionation shifts the composition of these melts to the
1213 right along a vector whose direction is determined by the combination of olivine ±
1214 plagioclase ± clinopyroxene crystallization appropriate for that liquid. Samples on the
1215 right side of the plot have experienced more crystal fractionation than those on the left.

1216 **Figure 7. Recalculated temperatures and pressures with subset of the literature**
1217 **compilation.** a) Pressure-temperature plot color coded by primitive arc magma type. The
1218 new pressures and temperatures were calculated following internally consistent methods
1219 as described in the Methods section. The anhydrous peridotite solidus in a) is from
1220 Hirschman (2000) and the H₂O-saturated peridotite solidi from Till et al. (2012b).
1221 Pressures for the high-Mg andesites are all the same because the pressure was assumed to
1222 be that of the base of the av. arc crust (30 km, ~10 kbar on average) in order to accurately
1223 assess their temperatures of mantle equilibration using the Mitchell and Grove (2015)
1224 harzburgite-liquid hygrometer and thermometer. b) Recalculated pressure and
1225 temperature contoured (by color and bubble size) for the H₂O content used in the
1226 recalculation. Samples in the recalculation are limited amongst other criteria to a range of
1227 representative compositions and to those with H₂O contents were measured via SIMS or
1228 FTIR or adequate information to estimate H₂O via hygrometry with the exception of 4
1229 tholeiites which are similar in composition to other nominally anhydrous tholeiitic
1230 magmas with H₂O estimates (see details in methods for recalculations in Table 2).
1231 Symbols with bold outlines and arrow illustrates the shift in P-T that results from
1232 estimating the P and T of the sample with 0 wt% H₂O (left) vs. 4.6 wt% H₂O (on the right
1233 at end of arrow).

1234 **Figure 8. Comparison of arc basalts to forward modeling of batch vs. near-**
1235 **fractional mantle melting.** a) Comparison of tholeiitic basalts used for the pressure and
1236 temperature recalculations to incremental batch melts of a depleted Hart and Zindler
1237 mantle composition with 90% melt extraction and dF/dP = 1% per kilobar and an
1238 adiabatic gradient of 1.5°C per kilobar using the forward mantle model of Till et al.
1239 (2012a) as modified by Behn and Grove (2015) Gray batch melting curves shown for
1240 comparison as described in B. b) Comparison of the calc-alkaline basalts used for
1241 pressure and temperature recalculations to isobaric batch melting curves for a depleted
1242 Hart and Zindler (1986) mantle composition at 10, 15 and 20 kbar as predicted by the
1243 forward lherzolite melting model of Till et. al (2012a) as modified by Behn and Grove
1244 (2015). c) Comparison of all rock types to the batch melting and near-fractional melts of
1245 spinel lherzolite. The calc-alkaline basalts are consistent with 1-10% batch melts of a

1246 depleted mantle at 10-20 kbar. The tholeiites can be modeled by either batch melting at
1247 average higher pressures and extents of melting or by near-fractional melting between 20-
1248 9 kbar.

1249 **Figure 9. Comparison to Geodynamic Models.** P-T paths from a selection of modern
1250 thermal models of subduction zones are compared to the thermobarometry dataset.
1251 Colored lines represent the temperature conditions at vertical slices through the mantle
1252 wedge. The Kincaid and Sacks (1997) curve is from their model for fast subduction of a
1253 thin plate assuming a mantle potential temperature of 1400°C. The van Keken et al.
1254 (2002) curve is from their models with a non-linear mantle viscosity and compares the
1255 structure of a “warm” (Cascadia) vs. “cold” (Japan) subduction zone. The Kelemen et al.
1256 (2003) curves compare slices through the model at two different distances from the
1257 trench, one closer to the trench where the slab is at 100 km depth (representing the hottest
1258 conditions at the shallowest depth from the models) and one further from the trench with
1259 the slab at depths of 150 km. Black solid and dashed lines are temperature conditions for
1260 the slab surface from a suite of relevant models compared to the gray field of
1261 thermobarometry constraints for the slab from exhumed metamorphic rocks as
1262 summarized in Penniston-Dorland et al. (2015). In both the magmatic thermobarometry
1263 reviewed here and the slab thermobarometry, the models tend to only reproduce the
1264 cooler petrologic observations.

1265 **Figure 10. Summary of mantle processes that form primitive arc magmas reviewed**
1266 **in this paper.** a) 2D schematic cross section of subduction zone with isotherm locations
1267 from Kelemen et al. (2003). b) Pressure-temperature diagram with the recalculated
1268 pressure and temperatures for primitive arc magma in Table 2 shown in colored squares
1269 (which all fall within the gray field in panel a). The anhydrous peridotite solidus is from
1270 Hirschmann (2000) and the H₂O-saturated peridotite solidus from Till et al. (2012b).
1271 Hydrous arc magmas are first formed at the locations of the water-saturated mantle
1272 solidus just above the subducting lithosphere at pressures of ~20-30 kbar and then rise via
1273 reactive porous flow (teal dashed arrows) into the hot core of the mantle wedge where
1274 their melting extent increases and the water content decreases due to continued re-
1275 equilibration with the mantle. As they continue to rise into the cooler top portion of the
1276 mantle wedge via reactive porous flow their melt fraction decreases. The recalculated
1277 pressures and temperatures (gray field in a. and b.) reflect the conditions during their re-
1278 equilibration in the top half of the mantle wedge immediately prior to their extraction
1279 from the mantle. If hydrous magmas rise via channelized flow, they would record much
1280 lower temperatures at a given pressure than any magma in the recalculations done here or
1281 the literature compilation in Figure 1. When the mantle residue for this process is
1282 lherzolite, calc-alkaline basalts are generated. Alternatively, when the mantle residue is
1283 harzburgite, depending on the H₂O content either high-Mg andesites (higher av. H₂O
1284 content, lower av. temperatures) or high-Mg liquids (lower av. H₂O content, higher av.
1285 temperatures) are generated. Nominally anhydrous arc tholeiites are generated by near-
1286 fractional decompression melting at or near the anhydrous lherzolite solidus in the
1287 upwelling back limb of corner flow at ~20-10 kbar. These melts are focused into the
1288 same region of last mantle equilibration as the hydrous melts. Adiabatic ascent of these
1289 magmas (black dashed arrows) preserves the higher temperatures of their formation
1290 rather than the lower temperatures in the top half of the wedge. Some relatively dry high-

1291 Mg liquids may also be generated via this process when the mantle residue is
1292 harzburgitic.

1293 **Table 1. Published studies that include arc mantle-melt thermobarometry compiled**
1294 **in Figure 1.**

1295 **Table 2. Studies and methods used for the recalculation of primitive arc magma**
1296 **pressures and temperatures in “Recalculation of Pressure and Temperature for**
1297 **Common Arc Magma Types” section and Figure. 7.**