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Oxygen Isotope Thermometry Reveals high magmatic temperatures and short residence times in Yellowstone and other hot-dry rhyolites compared to cold-wet systems

ARTICLE *in* AMERICAN MINERALOGIST · MAY 2016

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1 **Revision 2:**

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3 **Oxygen isotope thermometry reveals high magmatic temperatures and short residence**
4 **times in Yellowstone and other hot-dry rhyolites compared to cold-wet systems**
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6
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13
14 **Abstract**

15 The eruption and storage temperatures of rhyolitic magmas are critical factors for understanding
16 the mechanisms of their eruption and petrogenesis. Temperatures are particularly important when
17 comparing the magmatic histories of hot-dry rhyolites from the Yellowstone-Snake River Plain
18 (YSRP) and Iceland to cold-wet rhyolites such as the Bishop Tuff. Here we employ mineral pair
19 oxygen isotope fractionations for estimating rhyolite temperatures independent of pressure and
20 other compositional factors. We report high precision oxygen isotope analyses of quartz,
21 pyroxene, magnetite, and zircon that we use to estimate crystallization and storage temperatures.
22 Temperatures for YSRP and Icelandic rhyolites are highest for quartz-magnetite and quartz-
23 clinopyroxene (~950°C), with lower quartz-zircon (850°C) temperatures that are similar to

24 estimates of zircon saturation. The magnitude and pattern of these temperatures is consistent with
25 crystallization from near-liquidus rhyolites. In contrast, oxygen isotope temperatures calculated
26 for the Bishop and other “cold-wet” type tuffs define low ~760°C temperatures for all three
27 mineral pairs consistent with prolonged mineral residence at near-solidus conditions.
28 Preservation of a down-temperature crystallization sequence of hot magnetite and clinopyroxene
29 with colder zircon in hot-dry YSRP and Icelandic rhyolites suggest <1000 year magma
30 residence, where magnetite does not have sufficient time to diffusively equilibrate oxygen in a
31 lower temperature melt. This is consistent with recently determined high precision U-Pb
32 crystallization ages zircons from the same units indicating magma generation shortly before
33 eruption.

34

35 **Keywords:** oxygen isotopes, thermometry, magma storage timescales, rhyolites, Yellowstone,
36 Snake River Plain

37

38

Introduction

39 The origin of rhyolite magmas has been a classic question for igneous petrology, and the
40 temperature of these magmas is a critical factor in understanding their petrogenesis. Generally
41 rhyolites can be divided into cold-wet and hot-dry types (e.g., Christiansen 2005). The cold and
42 wet types contain hydrous minerals, are often crystal-rich, and are typically associated with
43 convergent margins. Classic examples of this type include the Bishop and similar “monotonous
44 intermediate” tuffs, interpreted to originate from melt extraction from near-solidus magma
45 bodies stored as long-lived crystal mushes with >45% crystals (e.g., Bachmann et al. 2002;
46 Bachmann and Bergantz 2004; Hildreth et al. 2007). On the opposite end of the spectrum are the

47 hot and dry, plume and rift related-rhyolites, notably represented by the Yellowstone-Snake
48 River Plain (YSRP) of 15 Ma to present (e.g., Branney et al. 2008), and similar aphyric rhyolites
49 of Iceland. These rhyolites typically belong to bimodal basaltic-rhyolitic sequences and have
50 typically low crystallinity (<5-10%) with a paucity of hydrous mineral phases.

51 A great wealth of geothermometry estimates are available for YSRP rhyolites from
52 mineral-pair analysis (e.g., QUILF, two-pyroxene), zircon saturation, Ti-in-quartz and zircon,
53 and experimental and modeled phase equilibria, and these typically yield high temperatures of up
54 to 950°C (e.g., Cathey et al. 2004, 2009; Vazquez et al. 2009; Watts et al. 2011; Almeev et al.
55 2012). Many of these geothermometers, however, are highly sensitive to pressure, melt
56 composition, and activities of participating components. Furthermore, recent reformulation of the
57 Zr-saturation thermometer of Watson and Harrison (1983) by Boehnke et al. (2013) yields 30-
58 40°C lower zircon saturation temperatures in many systems. Some phase equilibria studies of
59 Bishop and Yellowstone compositions suggest lower, even <700°C, temperatures (Gardner et al.
60 2014; Befus et al. 2015; Befus and Gardner in review). Lower temperatures seem more
61 compatible with near-solidus mush storage and extraction models, and several authors now try to
62 apply this model for YSRP centers (e.g., Ellis et al. 2014, Stelten et al. 2015). These very
63 disparate estimates of YSRP rhyolite magma temperatures not only critically influence
64 petrogenetic implications, but also assumed eruption temperatures pertinent to eruption
65 mechanisms and volcanic hazards.

66 Here we present oxygen isotope thermometry between mineral pairs from YSRP centers
67 as compared to other rhyolites worldwide to provide a unique estimate of rhyolite temperatures
68 (Fig. 1a,b). Stable isotopic fractionations are independent of pressure, melt composition, and
69 phase equilibria, making them particularly valuable as independent tools for determining the

70 temperature of magmas. Clayton and Epstein (1958) determined that oxygen isotope
71 fractionation between coexisting minerals in igneous rocks could be used to estimate magmatic
72 temperatures. Despite this simple theoretical utility of oxygen isotope thermometers, their use in
73 magmatic systems has been limited. This is due to (1) earlier methods that relied on 10-30 mg of
74 material for analysis and resulted in challenging aggregation of pure mineral separates,
75 especially for crystal-poor YSRP rhyolites; (2) small fractionation factors at high temperatures
76 typical of most, and especially basaltic, igneous rocks coupled with insufficient analytical
77 precision of many earlier $\delta^{18}\text{O}$ determinations; and (3) postmagmatic retrogression of mineral-
78 mineral $\Delta^{18}\text{O}$ values during slow cooling that limited use especially in plutonic and metamorphic
79 rocks (Eiler et al. 1993). High-precision analyses by laser fluorination of single crystals or small
80 masses of mineral separates (0.7-2.2 mg), however, permits application of this geothermometer
81 for rapidly quenched and lower temperature (700-1000°C) silicic magmas. Application to the
82 highest temperature mafic melts (e.g., 1200°C) is still limited by extremely low fractionation
83 factors, since the high-temperature oxygen isotope fractionation between minerals is proportional
84 to $1/T^2$ (Clayton and Epstein 1958). Bindeman and Valley (2002) demonstrated that oxygen
85 isotope thermometry works for rhyolites and particularly well for rapidly quenched pumices
86 sampling the Bishop magma body, providing new and useful insight into its pre-eruptive
87 conditions.

88

89

Methods

90 Laser fluorination oxygen isotope analyses were carried out at the University of Oregon
91 using a 25 W CO_2 laser coupled with an MAT 253 10 kV gas source isotope ratio mass
92 spectrometer (IRMS) and purified BrF_5 reagent (for further details, see Loewen and Bindeman

93 2015). Analyzed mineral separates were pretreated with 48% HF for 3-5 minutes as needed to
94 remove alteration or adhered glass before ~1 mg of material was hand picked for analysis (see
95 electronic supplement for sample weight and raw analytical results). Quartz was selected as
96 single bipyramidal crystals. Clinopyroxenes were selected by color (bright green) and oblique
97 extinction. Magnetite was concentrated with a magnet and hand sorted for freshest grains (black,
98 shiny, and magnetic).

99 Oxygen isotope fractionation between coexisting solid equilibrium mineral pairs is
100 described by an experimentally determined A-factor and temperature (T, in Kelvin):

101
$$T = \sqrt{\frac{A \times 10^6}{1000 \ln \alpha}} \quad (1)$$

102 where $1000 \ln \alpha_{(\text{mineral 1} - \text{mineral 2})} \sim \Delta^{18}\text{O} = \delta^{18}\text{O}_{\text{mineral 1}} - \delta^{18}\text{O}_{\text{mineral 2}}$, for $\Delta^{18}\text{O} < 10$ (Bottinga and
103 Javoy 1973). Using this relationship, the difference in $\delta^{18}\text{O}$ between coexisting minerals in a
104 volcanic rock can be used to estimate a magmatic temperature. Magnetite, clinopyroxene, and
105 zircon have large A-factors relative to quartz, and, therefore, can be used to calculate a
106 reasonably precise temperature resolution given propagated uncertainty in $\delta^{18}\text{O}$ determinations.
107 We use A-factors determined using a calcite exchange medium from Chiba et al. (1989) for
108 $\Delta^{18}\text{O}_{\text{Qz-Cpx}}$ and $\Delta^{18}\text{O}_{\text{Qz-Mag}}$ of 2.75 and 6.29, respectively. It is important to note that for the
109 purpose of oxygen isotope thermometry in equation 1, clinopyroxene and magnetite are phases
110 of constant composition since cation substitution in these minerals plays an insignificant role in
111 ^{18}O - ^{16}O partitioning. The $\Delta^{18}\text{O}_{\text{Qz-Zrn}}$ A-factor is similar to $\Delta^{18}\text{O}_{\text{Qz-Cpx}}$, with a value of 2.64
112 proposed by Valley et al. (2003) and 2.33 by Trail et al. (2009). Here, we choose to use an

113 intermediate value of 2.55 corresponding to the observed $\Delta^{18}\text{O}_{\text{Qz-Zrn}}$ and temperature relationship
114 in the Bishop Tuff (Bindeman and Valley 2002).¹

115 As our 2σ uncertainty in a single mineral $\delta^{18}\text{O}$ analysis is approximately $\pm 0.08\text{‰}$ (this
116 study and Loewen and Bindeman 2015), the additive propagated uncertainty ($\sqrt{\sigma_1^2 + \sigma_2^2}$) of any
117 two analyses is $\pm 0.12\text{‰}$. This will result in an approximately $\pm 40^\circ\text{C}$ in $\Delta^{18}\text{O}_{\text{Qz-Cpx}}$ calculations
118 and $\pm 20^\circ\text{C}$ in $\Delta^{18}\text{O}_{\text{Qz-Mag}}$ calculations. Recycling of antecrystic zircon cores are common in some
119 YSRP units (e.g., Bindeman and Simakin 2014), therefore for units with known antecrystic
120 zircon core compositions we only consider published $\delta^{18}\text{O}$ zircon rim analyses measured by
121 SIMS. These analyses have higher individual uncertainties of at least $\pm 0.2\text{‰}$, therefore, these
122 calculated temperatures are $\pm 60^\circ\text{C}$. We report temperatures calculated by averages of multiple
123 mineral analyses from each sample thus reducing noise from analytical uncertainty.

124 Since the $\Delta^{18}\text{O}_{\text{melt-Qz}}$ changes very little over expected magmatic temperatures ($< 0.2\text{‰}$,
125 Fig. 1c), the $\Delta^{18}\text{O}_{\text{Qz-mineral}}$ is proportional to $\Delta^{18}\text{O}_{\text{melt-mineral}}$, with a single offset, where the mineral
126 pairs are clinopyroxene, magnetite, or zircon, all of which are much more sensitive to
127 temperature than quartz. Therefore, our temperature estimates should be accurate as long as the
128 phases crystallized from the same melt composition. Loewen and Bindeman (2015)
129 demonstrated that clinopyroxene in the Yellowstone Central Plateau Member lavas have
130 rhyolitic trace element compositions. We admit that the potential for disequilibrium exists if any
131 mineral phase has an ante- or xenocrystic origin, and for any one sample we cannot verify that

¹ Subsequent research has demonstrated that U-Pb dated zircons from the Bishop Tuff are identical to eruption age (Crowley et al. 2007) and thus their $\delta^{18}\text{O}$ appropriately reflects equilibrium with other phenocryst phases making the Bishop Tuff an ideal case study for testing isotopic fractionations.

132 the mineral separates analyzed are in equilibrium since mineral separation removes crystals from
133 their petrographic context.

134

135 **Results**

136 New $\delta^{18}\text{O}$ values are reported for clinopyroxene, magnetite, and quartz from the
137 Yellowstone, Heise, Picabo, and Bruneau Jarbidge rhyolites, two new samples from Krafla,
138 Iceland, and compiled data for other YSRP and worldwide units (Table 1; Fig. 1a). For 12 Ma to
139 recent YSRP rhyolites, the median temperature estimated from $\Delta^{18}\text{O}_{\text{Qz-Cpx}}$ is 943°C. The highest
140 ~1200°C temperatures recorded from Buffalo Lake and Liddy Hot Springs are beyond any
141 reasonable crystallization temperature for a rhyolite composition, and we suspect these may
142 represent disequilibrium between magmatic $\delta^{18}\text{O}_{\text{Cpx}}$ and antecrystic low $\delta^{18}\text{O}_{\text{Qz}}$ incorporated
143 during shallow rhyolite remelting. The median 950°C temperature estimated from $\Delta^{18}\text{O}_{\text{Qz-Mag}}$ is
144 very similar to that of clinopyroxene, although we note that on a sample-by-sample basis,
145 magnetite temperatures are on average 50°C higher than clinopyroxene. The median temperature
146 estimate from $\Delta^{18}\text{O}_{\text{Qz-Zrn}}$ is 847°C and was calculated using the average of previously published
147 SIMS data from zircon rims for each reported sample. On a sample-by-sample basis, zircon
148 temperatures are 70°C lower than clinopyroxene or magnetite temperatures. Average
149 temperatures calculated with $\Delta^{18}\text{O}_{\text{Qz-Zrn}}$ are remarkably similar to average whole rock zircon
150 saturation temperatures calculated following Watson and Harrison (1983).

151 Despite all these individual uncertainties we observe a very consistent overall pattern
152 between minerals. There are no apparent changes in temperature calculated with oxygen isotopes
153 over the 12 Ma YSRP eruptive history that we examine here, and a similar range of temperatures
154 is evident within any one eruptive period as across the entire province (Fig. 1a). In the youngest

155 Yellowstone volcanism (260 to 79 ka) there is no evidence for decreasing temperatures, as
156 described by Vazquez et al. (2009) using QUILF and Ti-in-quartz thermometry, but there may be
157 a slight increase similar to increases in zircon saturation temperatures (Loewen and Bindeman
158 2015). Similarly, we do not observe any eastward temperature decrease in YSRP rhyolites as
159 described in Nash et al. (2006).

160 The YSRP data are directly compared to identically determined oxygen isotope
161 temperatures reported for a number of classic cold-wet tuff deposits, including the Bishop, Cerro
162 Galan, Fish Canyon, lower Bandelier, and Toba tuffs from Bindeman and Valley (2002) and the
163 Kos Plateau (analyses from University of Oregon, reported in Bachmann 2010). Temperatures
164 for mineral pairs from these cold-wet systems are clearly lower and more restricted than YSRP
165 rhyolites (Table 1, Fig. 1b). For the Bishop Tuff, three $\Delta^{18}\text{O}_{\text{Qz-Cpx}}$ temperatures are 742, 809, and
166 829°C for the Bishop Tuff. The median temperature from $\Delta^{18}\text{O}_{\text{Qz-Mag}}$ is 750°C (n=17, range 715-
167 818°C) and is nearly identical to the median $\Delta^{18}\text{O}_{\text{Qz-Zrn}}$ temperature of 757°C (n=4, range 714-
168 782°C). These lower temperatures are also similar to estimated zircon saturation temperatures
169 (Gualda and Ghiorso 2013). Temperatures from other cold-wet rhyolites follow the same pattern,
170 with nearly identical magnetite and zircon temperatures of ~760°C, closely agreeing with zircon
171 saturation temperature estimates, and classical understanding of eutectic conditions in water-
172 saturated rhyolites. These similar temperatures derived from all minerals in these cold-wet
173 rhyolites contrast the elevated clinopyroxene and magnetite temperatures and relatively lower
174 zircon temperatures in YSRP rhyolites.

175 Icelandic rhyolites, from another hot-dry system in a rift-type setting have similarly
176 elevated temperatures as the YSRP (Fig. 1a), and especially high magnetite temperatures. In
177 addition to new analyses from Krafla, the average Icelandic values in Table 1 include the 1875

178 A.D. Askja, 1477 A.D. Torfajökull, 1362 A.D. Öräfajökull, 1158 A.D. Hekla, and 6.6 Ma
179 Austurhorn eruptions, previously reported in Carley et al. (2011) and Bindeman et al. (2012).
180 Recycling of antecrystic zircons is ubiquitous in these Iceland rhyolites so we could not compare
181 $\Delta^{18}\text{O}_{\text{Qz-Zrn}}$ temperatures (Bindeman et al. 2012; Bindeman and Simakin 2014).

182

183

Discussion

Elevated crystallization temperatures

185 Yellowstone-Snake River Plain oxygen isotope temperatures for clinopyroxene and
186 magnetite reflect the early crystallization of these minerals at elevated liquidus temperatures near
187 950°C, and clinopyroxene trace element compositions are consistent with origin from very
188 similar rhyolite melts (Loewen and Bindeman 2015). The average clinopyroxene and magnetite
189 temperatures are remarkably consistent with experimentally determined phase equilibria
190 temperatures in the Bruneau Jarbidge rhyolite from Almeev et al. (2012). These experiments
191 determined that the first minerals crystallizing on the liquidus were magnetite followed by
192 clinopyroxene, in agreement with our slightly higher $\Delta^{18}\text{O}_{\text{Qz-Mag}}$ temperatures. Lower $\delta^{18}\text{O}$
193 temperatures calculated for zircon are consistent with estimates of zircon saturation calculated
194 from either Watson and Harrison (1983) or Boehnke et al. (2013) formulations, but closer to the
195 former, based on either glass or whole rock Zr concentrations. These temperatures are broadly
196 similar to estimates from mineral-pair QUILF calculations although we do not observe similar
197 temperature decreases through time (e.g., Cathey and Nash 2004, 2009; Nash et al. 2006;
198 Vazquez et al. 2009).

199 In addition, our temperature estimates for the YSRP are remarkably similar to other hot-
200 spot/rift related rhyolites from Iceland (Fig. 1a). Only Lava Creek Tuff member A and the lower

201 Arbon Valley Tuff have magnetite temperature <900°C, with 710-750°C more in-line with cold
202 and wet rhyolites (Table 1). These two samples are both from early products of each of their
203 eruptive sequences and are some of the only YSRP rhyolites with hydrous minerals (amphibole
204 and biotite, respectively).

205 The high temperatures cannot be the result of secondary alteration or diffusive
206 retrogression of mineral phases (e.g., Eiler et al. 1993). Zircon and clinopyroxene are especially
207 resistant to low temperature alteration due to extremely small diffusion coefficients ($<1 \times 10^{-23}$
208 m^2/s) in wet or dry conditions at <700°C (Farver 1989; Watson and Cherniak 1997). Even for a
209 small 50 μm zircon crystal, it would take ~3 Ma to equilibrate with surroundings at 700°C in wet
210 conditions. Concern of the possibility for alteration and retrogression of magnetite oxygen
211 isotope values is more justified as diffusion coefficients are 2-5 orders of magnitude higher in
212 magnetite (e.g., Farver 2010). With respect to secondary alteration, the $\Delta^{18}\text{O}_{\text{Mag-water}}$ is -5 to -7‰
213 at subsolidus temperatures so water-rock reactions with isotopically light precipitation will only
214 decrease the $\delta^{18}\text{O}$ value of magnetite and result in a larger $\Delta^{18}\text{O}_{\text{Qz-Mag}}$ and therefore lower
215 apparent temperature estimate (Cole et al., 2004). The high temperatures calculated from our
216 magnetite analyses thus can only be achieved by and preserved from high temperature
217 equilibrium crystallization, unless mineral pairs crystallized from melts of different composition.
218 As we have discussed previously, the average temperatures for each mineral that we observe
219 across the YSRP are most likely the result of crystallization temperatures in-line with
220 experimental phase equilibria (Almeev et al. 2012).

221

222 **Timescales of crystal residence in magmas**

223 Preservation of high oxygen isotope temperatures in mineral phases constrains the time of
224 magma residence after crystallization. The extremely slow oxygen diffusion in zircon and
225 pyroxene means that the oxygen isotopic composition of these phases preserves their
226 crystallization temperature for $>10^7$ years even at magmatic temperatures (for review, see Farver
227 2010). Oxygen diffusion is much faster in magnetite with diffusivities 3-4 orders of magnitude
228 faster (Farver 2010). The high observed temperatures calculated from our magnetite analyses
229 could only result from initial crystallization at high temperature as even partial oxygen isotope
230 reequilibration of their rims during crystal residence in a lower temperature magma will quickly
231 increase $\Delta^{18}\text{O}_{\text{Qz-Mag}}$ and thus lower calculated temperature.

232 We can constrain the maximum residence time of magnetite at 850°C that will equilibrate
233 a rim large enough to bring the bulk crystal oxygen isotope value down to but not below the
234 observed 950°C average temperature. Using the experimentally determined diffusion coefficients
235 of Giletti and Hess (1988) for wet conditions and the dry empirically determined coefficients of
236 Sharp (1991), we estimate that the crystal residence time must be <100 years and $<10,000$ years,
237 respectively, for magnetite up to 400 μm -diameter magnetite crystallizing 40 μm -thick rims.
238 Diffusivities for water-bearing but not water-saturated YSRP magmas are likely between those
239 two endmembers, and therefore we consider <1000 years a reasonable estimate for maximum
240 magnetite residence.

241 High magnetite and clinopyroxene lower zircon temperatures, therefore, reflect two
242 distinctive thermal conditions for YSRP rhyolite magmas: (1) near-liquidus temperatures for
243 clinopyroxene and magnetite crystallization and (2) subsequent short <1000 year residence at
244 $\sim 100^\circ\text{C}$ lower storage at zircon saturation. The pattern of oxygen isotope temperatures from the

245 cold-wet systems starkly contrast with our results from the YSRP (Fig. 1). As with YSRP
246 samples, ~810°C temperatures calculated for pyroxene probably reflect crystallization of this
247 phase about 50°C above zircon saturation, and preservation of its oxygen isotope composition
248 due to very slow diffusion rates. Magnetite temperature estimates, however, are identical to
249 zircon temperatures. A simple explanation is that magnetite has equilibrated with the cool and
250 long-lived magma body proposed for these classic cold and wet rhyolites (e.g., Hildreth and
251 Wilson 2007), although magnetite may have also saturated at lower temperatures in these
252 systems.

253

254 **Implications**

255 Oxygen isotope thermometry strongly supports YSRP and Icelandic rhyolite
256 crystallization from high liquidus temperatures with short (several thousand years or less)
257 residence at lower, zircon-saturated temperatures, following a crystallization sequence in-line
258 with Almeev et al. (2012) phase equilibria experiments. This observation is incompatible with
259 rhyolite extraction from long-lived (10^5 - 10^6 years) crystal mushes that reside below their zircon
260 saturation temperature. Instead, the most simple explanation for YSRP rhyolite petrogenesis is
261 that they form from rapid, high temperature assembly of partial or wholesale melting of upper
262 crustal rocks generating a hot, near-liquidus melt (e.g., Bindeman and Simakin 2014; Loewen
263 and Bindeman 2015). This melt first crystallized high temperature phases such as magnetite and
264 clinopyroxene before cooling ~100°C and reaching zircon saturation. Storage at this lower
265 temperature is long enough to crystallize eruption-age zircon populations (e.g., Wotzlaw et al.
266 2014, 2015) but short enough to prevent diffusive reequilibration of magnetite to lower
267 temperature oxygen isotope values. Our results contradict potential extraction of such rhyolites

268 from any cold crystal mush achieved through purely mechanical separation without reheating
269 (e.g., Bachmann et al. 2004; Pistone et al. 2015; Stelten et al. 2015). In contrast, cold-wet
270 rhyolites display a limited range of lower oxygen isotope temperatures for all minerals.
271 Similarity of zircon and magnetite is entirely consistent with long residence in a cool, crystal-
272 rich, and more hydrous magma and application of the mush extraction models are more
273 appropriate explanations for their eruption. Our comparison of major silicic systems suggests
274 that aggregate patterns of temperatures calculated for different minerals can be useful in
275 differentiating pre-eruptive storage conditions and generation of rhyolite melts.

276

277

Acknowledgements

278 We appreciate discussions with Dylan Colón, Madison Myers, and John Valley during
279 development of this paper, and analytical assistance from Jim Palandri, editorial handling by Ian
280 Swainson, and constructive reviews from Renat Almeev, Jörn Wotzlaw, and an anonymous
281 reviewer. This study was funded by the National Science Foundation (grant number
282 EAR/CAREER-844772 to I.N.B.) and the Department of the Geological Sciences at the
283 University of Oregon.

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417 **Figure 1. (a)** Oxygen isotope mineral pair temperature calculations for hot-dry Yellowstone-
418 Snake River Plain (YSRP) and Icelandic rhyolites, and **(b)** the Bishop, Fish Canyon, Toba, Cerro
419 Galan, lower Bandelier, and Kos Plateau Tuffs, shown along-side zircon saturation temperatures
420 (references for compiled data in Table 1). **(c)** Calculated mineral-melt oxygen isotope
421 fractionation curves at different temperatures, with peak probability values plotted from kernel
422 density estimates of each mineral from hot-dry and cold-wet rhyolites. In hot-dry YSRP and
423 Iceland rhyolites, temperatures calculated with magnetite and clinopyroxene are generally very
424 similar, with the median quartz-magnetite 50°C higher than quartz-clinopyroxene where both
425 minerals are analyzed in the same sample. The median quartz-zircon is 70°C lower than quartz-
426 clinopyroxene. In cold-wet rhyolites, temperatures for all three minerals are lower and
427 remarkably similar. The average quartz-clinopyroxene is ~50°C higher (~810°C) than quartz-
428 magnetite and quartz-zircon (~760°C). Again, zircon saturation temperatures are very similar,
429 and almost 100°C lower than in hot-dry rhyolites. We note that two YSRP deposits, the Lava
430 Creek Tuff member A (LCT-A) and the lower Arbon Valley Tuff (AVT), return similar
431 temperatures to the cold-wet type rhyolites. These two units are the only YSRP units that we
432 have analyzed with hydrous minerals, biotite (bio) and hornblende (hbl), and are the earliest
433 deposits from their respective eruptive sequences.

Table 1. Oxygen isotope thermometry results for Yellowstone-Snake River Plain rhyolites compared to rhyolites worldwide.

Group	Unit	Sample	Age (Ma)	$\delta^{18}\text{O}$ (‰, SMOW)				Temperature (°C)			
				Cpx	Mag	Zrn	Qz	Qz-Cpx	Qz-Mag	Qz-Zrn	Zrn Sat. ^a
Yellowstone Central Plateau Member	Pitchstone Plateau	YL02-1	0.079	2.78	0.32	2.78 ^b	4.59	960	941	916	835 ^c
	Hayden Valley	08YS-6	0.102	2.94			4.64	997			
	West Yellowstone	Y-164,209	0.114	2.91	0.54	2.83 ^b	4.73	955	953	884	836 ^c
	Summit Lake	Y-74	0.124	3.06	1.17		4.86	962	1032		832 ^c
	Buffalo Lake	08YS-19a	0.160	3.07			4.39	1170			817 ^c
	West Thumb	06YS-3	0.176	2.86	0.91		4.52	1014	1047		
	Scaup Lake	YL96-9	0.257		-0.07	1.88 ^d	4.28		929	757	847 ^d
	South Biscuit Basin	YL96-2	0.260	2.13	0.00	2.32 ^{b,d}	4.49	806	911	811	850 ^d
Major Yellowstone Tuffs	Lava Creek Tuff-B	LCT-B	0.63	4.51	2.2	4.63 ^e	6.42	927	948	921	872 ^d
	Lava Creek Tuff-A	LCT-3a	0.63	3.26 [*]	-0.45	3.64 ^d	6.09	745 [*]	708	747	825 ^d
	Mesa Falls Tuff	MFT-1	1.30		0.95	3.64 ^{d,e}	5.52		900	890	800 ^d
	Huckleberry Ridge Tuff-C	HRT-C	2.08	6.68	4.06	5.58 ^{d,e}	7.61	921	1058	847	897 ^d
	Huckleberry Ridge Tuff-B	HRT-1	2.08	5.06	3.77	5.28 ^{d,e}	7.32	829	1058	844	863 ^d
Heise	Kilgore Tuff	2017-B	4.4	1.55		1.5 ^f	3.49	917		859	842 ^f
	Liddy Hot Springs	08HS-10	6.2	2.52		1.3 ^f	3.8 ^f	1193	1014	737	822 ^f
	Blacktail Creek	95-2001a	6.6	4.97	3.09	4.81 ^f	7.04	880	989	796	848 ^f
Picabo	Upper Arbon Valley	PC-14	10.4	6.61 ^g	3.88 ^g	6.38 ^h	8.41 ^h	963	905	848	835 ^h
	Lower Arbon Valley	PC-12	10.4		2.44 ^g		8.83 ^h		719		754 ^h
Bruneau Jarbidge (Cougar Point Tuff)	CPT-XVJ	ID-2	10.5	2.57	0.82		4.23	1013	1085		908 ⁱ
	CPT-XII	SB173	11.2	-1.92 ^j			-0.3 ^j	1024			905 ⁱ
	CPT-IX	ID-12	11.6	0.62 ^j			2.52 ^j	930			889 ⁱ
	CPT-V	ID-13	12.1	1.68 ^j			4.53 ^j	709			880 ⁱ
Median:								943	950	847	845
Worldwide tuff median values											
Bishop Tuff			0.76	5.79 ^j	2.37 ^k	5.84 ^j	8.33 ^k	809	750	757	735-765^l
Fish Canyon Tuff			28		2.22 ^k	5.75 ^j	8.27 ^k		757	762	779^m
Toba Tuff			0.074		3.61 ^k	6.62 ^j	9.41 ^k		769	719	762ⁿ
Cerro Galan			2.0		3.68 ^k		9.99 ^k		725		
Lower Bandelier Tuff			1.6		1.89 ^k		7.86 ^k		754		
Kos Plateau			0.16	6.61 ^{*o}	2.63 ^o		8.53 ^o	788[*]	760		730^o
Iceland								909^p	1075^p		864^p

434 ^a Zircon saturation following Watson and Harrison (1983). ^b Data from Watts et al. (2012). ^c Data from Loewen and Bindeman (2015). ^d Data from Bindeman et al. (2008). ^e Data from Wotzlaw et al.
 435 (2015). ^f Data from Watts et al. (2011). ^g Data from Drew et al. (2015). ^h Data from Drew et al. (2013). ⁱ Data from Cathey and Nash (2004). ^j Data from Bindeman and Simakin (2014). ^k Data are
 436 averages from Bindeman and Valley (2002). ^l Data from Gualda et al. (2013). ^m Data from Bachmann et al. (2002). ⁿ Data from Bindeman (2003). ^o Kos plateau data from Bachmann (2010). ^p Iceland
 437 data are from Carley et al. (2011) and Bindeman et al. (2012) for Öræfajökull, Askja, Hekla, Torfajökull, and Austurhorn; two new values for cpx are reported for Krafla; Average isotope ratios are not
 438 shown for Iceland due to large differences between eruptive centers. * *Italic* LCT-A analysis is hornblende instead of cpx, Kos Plateau analysis is biotite instead of cpx.

