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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7   Matthew W. Loewen[1,*] and Ilya N. Bindeman[1]

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9   [1]Department of Geological Sciences, 1272 University of Oregon, Eugene, OR 97403, USA

10   [*]Corresponding Author: email: loewenm@uoregon.edu, ph: 1-503-302-3762

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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

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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.

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35   **Keywords:** oxygen isotopes, thermometery, magma storage timescales, rhyolites, Yellowstone,

36   Snake River Plain

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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

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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 $\delta^{18}O$ determinations; and (3) postmagmatic retrogression of mineral-

78   mineral $\Delta^{18}O$ 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

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

102   where 1000 ln $\alpha_{(mineral\ 1 - mineral\ 2)} \sim \Delta^{18}O = \delta^{18}O_{mineral\ 1} - \delta^{18}O_{mineral\ 2}$, for $\Delta^{18}O$ <10 (Bottinga and

103   Javoy 1973). Using this relationship, the difference in $\delta^{18}O$ 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 $\delta^{18}O$ determinations.

107   We use A-factors determined using a calcite exchange medium from Chiba et al. (1989) for

108   $\Delta^{18}O_{Qz-Cpx}$ and $\Delta^{18}O_{Qz-Mag}$ 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 $\Delta^{18}O_{Qz-Zrn}$ A-factor is similar to $\Delta^{18}O_{Qz-Cpx}$, 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

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113   intermediate value of 2.55 corresponding to the observed $\Delta^{18}O_{Qz\text{-}Zrn}$ and temperature relationship

114   in the Bishop Tuff (Bindeman and Valley 2002).[1]

115   As our 2σ uncertainty in a single mineral $\delta^{18}O$ analysis is approximately ±0.08‰ (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 ±0.12‰. This will result in an approximately $\pm40°C$ in $\Delta^{18}O_{Qz\text{-}Cpx}$ calculations

118   and $\pm20°C$ in $\Delta^{18}O_{Qz\text{-}Mag}$ 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 $\delta^{18}O$ zircon rim analyses measured by

121   SIMS. These analyses have higher individual uncertainties of at least ±0.2‰, therefore, these

122   calculated temperatures are $\pm60°C$. We report temperatures calculated by averages of multiple

123   mineral analyses from each sample thus reducing noise from analytical uncertainty.

124   Since the $\Delta^{18}O_{melt\text{-}Qz}$ changes very little over expected magmatic temperatures (<0.2‰,

125   Fig. 1c), the $\Delta^{18}O_{Qz\text{-}mineral}$ is proportional to $\Delta^{18}O_{melt\text{-}mineral}$, 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

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[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 $\delta^{18}O$ 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 $\delta^{18}O$ 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 $\Delta^{18}O_{Qz-Cpx}$ 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 $\delta^{18}O_{Cpx}$ and antecrystic low $\delta^{18}O_{Qz}$ incorporated

143   during shallow rhyolite remelting. The median 950°C temperature estimated from $\Delta^{18}O_{Qz-Mag}$ 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 $\Delta^{18}O_{Qz-Zrn}$ 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 $\Delta^{18}O_{Qz-Zrn}$ 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 $\Delta^{18}O_{Qz\text{-}Cpx}$ temperatures are 742, 809, and

166   829°C for the Bishop Tuff. The median temperature from $\Delta^{18}O_{Qz\text{-}Mag}$ is 750°C (n=17, range 715-

167   818°C) and is nearly identical to the median $\Delta^{18}O_{Qz\text{-}Zrn}$ 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

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178   A.D. Askja, 1477 A.D. Torfajokull, 1362 A.D. Öræfajökull, 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    $\Delta^{18}O_{Qz-Zrn}$ temperatures (Bindeman et al. 2012; Bindeman and Simakin 2014).

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183   **Discussion**

184   **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 $\Delta^{18}O_{Qz-Mag}$ temperatures. Lower $\delta^{18}O$

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

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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 $\Delta^{18}O_{Mag-water}$ is -5 to -7‰

213   at subsolidus temperatures so water-rock reactions with isotopically light precipitation will only

214   decrease the $\delta^{18}O$ value of magnetite and result in a larger $\Delta^{18}O_{Qz-Mag}$ 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 $\Delta^{18}O_{Qz-Mag}$ 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   ~100°C lower storage at zircon saturation. The pattern of oxygen isotope temperatures from the

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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

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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.

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286   **References**
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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)	δ¹⁸O (‰, SMOW) Cpx	Mag	Zrn	Qz	Temperature (°C) 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 [e]
	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 [e]
	Summit Lake	Y-74	0.124	3.06	1.17		4.86	962	1032		832 [e]
	Buffalo Lake	08YS-19a	0.160	3.07			4.39	1170			817 [e]
	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**	**762 [n]**
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]**

[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. (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 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 data are from Carley et al. (2011) and Bindeman et al. (2012) for Öræfajökull, Askja, Hekla, Torfajokull, and Austurhorn; two new values for cpx are reported for Krafla; Average isotope ratios are not 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.

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