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# 1 Volatile loss under a diffusion-limited regime in tektites: 2 evidence from tin stable isotopes

3 **John B. Creech<sup>1,2,\*</sup>, Frédéric Moynier<sup>1,3</sup>, and Christian Koeberl<sup>4,5</sup>**

4 *<sup>1</sup>Institut de Physique du Globe de Paris, Université de Paris, 1 Rue Jussieu, 75328 Paris  
5 cedex 05, France*

6 *<sup>2</sup>Department of Earth and Planetary Sciences, Macquarie University, NSW 2109, Australia*

7 *<sup>3</sup>Institut Universitaire de France, 75005, Paris*

8 *<sup>4</sup>Department of Lithospheric Research, University of Vienna, Althanstrasse 14, 1090 Vienna,  
9 Austria*

10 *<sup>5</sup>Natural History Museum, Burgring 7, 1010 Vienna, Austria*

11 *\*Email address: [john.creech@mq.edu.au](mailto:john.creech@mq.edu.au)*

## 12 **ABSTRACT**

13 Tektites are glasses derived from near-surface continental crustal rocks that were  
14 molten and ejected from the Earth's surface during hypervelocity meteorite impacts. They are  
15 among the driest terrestrial samples, although the exact mechanism of water loss and the  
16 behaviour of other volatile species during these processes are debated. Based on the  
17 difference in magnitude of the Cu and Zn isotopic fractionations in tektites, and the difference  
18 of diffusivity between these elements, it was suggested that volatile loss was diffusion-  
19 limited. Tin is potentially well suited to testing this model, as it has a lower diffusivity in  
20 silicate melts than both Cu and Zn, but a similar volatility to Zn. Here, we analysed the Sn  
21 stable isotopic composition in a suite of seven tektites, representing three of the four known  
22 tektite strewn fields, for which Zn and Cu isotopes were previously reported. Tin is enriched  
23 in the heavier isotopes ( $\geq 2.5\%$  on the  $^{122}\text{Sn}/^{118}\text{Sn}$  ratio) in tektites, correlated with the degree  
24 of Sn elemental depletion in their respective samples as well as with Cu and Zn isotope ratios,  
25 implying a common control. While the isotope fractionation of Sn, Cu and Zn is a result of

26 volatility, the magnitude of isotope fractionation is strongly moderated by their relative rates  
27 of diffusion in the molten tektite droplets. An Australasian Muong Nong-type tektite analysed  
28 has the least Sn depletion and Sn isotope fractionation, consistent with these samples being  
29 more proximal to the source and experiencing a shorter time at high temperatures.

30       Keywords: tektites; volatiles; tin; stable isotopes; impacts

## 31 **INTRODUCTION**

32       Tektites are distal impact glasses, typically up to a few cm in size, formed by rapid  
33 melting of terrestrial upper crustal rocks during hypervelocity impacts on the Earth's surface  
34 (e.g., Shaw and Wasserburg, 1982; Koeberl, 1986, 1990, 1994; Koeberl et al., 1996; Melosh  
35 and Artemieva, 2004). Tektites have chemical and isotopic compositions similar to that of the  
36 terrestrial upper crust, and thus are thought to have formed by fusion of the target  
37 rocks/sediments rather than from the impactor itself (e.g., Shaw and Wasserburg, 1982;  
38 Koeberl, 1986, 1994).

39       Although tektites are generally chemically and isotopically similar to the upper crust,  
40 they are characterized by extremely low water contents (Beran and Koeberl, 2010; 0.002–0.02  
41 wt.%; Koeberl, 1994), indicating a depletion of water (and potentially other volatiles) from  
42 their precursor rocks (typically sediments, with >1% water) while the tektites were molten.  
43 The mechanism for this depletion and, more broadly, the behaviour of volatile elements and  
44 molecules during tektite formation is not well understood. Volatilization is known to  
45 fractionate elemental and isotopic abundances of volatile elements, and isotopic fractionation  
46 of moderately volatile elements has been recorded, for example, in lunar samples (e.g., Cl,  
47 Ga, K, Zn, Rb; Sharp et al., 2010; Paniello et al., 2012; Kato et al., 2015; Boyce et al., 2015;  
48 Kato and Moynier, 2017; Pringle and Moynier, 2017; Wang and Jacobsen, 2016). Recently,  
49 even relatively refractory elements (e.g., Mg, Si, Ca, Ti) have been shown to have isotopic  
50 variations that are attributed to high-temperature volatility-related processes in the early Solar  
51 System (e.g., Richter et al., 2009; Pringle et al., 2014; Hin et al., 2017; Davis et al., 2018). As

52 such, developments of additional tools to investigate such processes are timely. Given that  
53 tektites were rapidly heated, molten, and quenched over timescales that are likely to capture  
54 kinetic effects related to volatility, stable isotopes of volatile elements are well-suited to  
55 studying tektite formation and the isotopic effects thereof.

56         So far, studies of isotopic fractionation in tektites have yielded equivocal results.  
57 Studies of Mg (Esat and Taylor, 1987, 1986), K (Herzog et al., 2008; Humayun and Koeberl,  
58 2004; Jiang et al., 2019), B (Chaussidon and Koeberl, 1995), and Li (Magna et al., 2011) in  
59 tektites found that vapour fractionation (at least for those elements) did not play an important  
60 role during tektite formation, and that isotope ratios reflect the compositions of their source  
61 materials. However, those elements lie towards the more refractory end of the moderately  
62 volatile elements. Among the more volatile of the moderately volatile elements, Cd is  
63 enriched in the heavy isotopes in one Muong Nong-type Australasian tektite, which could  
64 reflect evaporative loss of light Cd isotopes (Wombacher et al. 2003), although with only one  
65 sample, this finding is inconclusive. Moynier et al. (2010, 2009) reported Zn and Cu isotope  
66 compositions, respectively, for a suite of tektites from all strewn fields. These results were  
67 reaffirmed by a study of moldavites and sediments of the Ries crater (Rodovská et al., 2017).  
68 Both Cu and Zn show significant isotopic fractionation that correlates with their elemental  
69 depletions, consistent with preferential evaporative loss of light isotopes (Moynier et al.,  
70 2010, 2009), and the larger isotopic fractionation in Cu was attributed to the greater chemical  
71 diffusivity of Cu over Zn due to their differences in ionic charge (i.e.,  $Zn^{2+}$  vs.  $Cu^+$ ; Moynier  
72 et al. 2010). However, this model is based on only two elements, and further testing is  
73 required to validate this hypothesis.

74         Tin is amongst the more volatile of the moderately volatile elements, and is potentially  
75 more volatile than both Cu and Zn. No quantitative estimates exist for the temperature of  
76 volatilization of these elements in conditions relevant to tektite formation. Cosmochemical  
77 condensation temperatures, based on condensation from a gas of solar nebular composition at

78 a partial pressure of  $\sim 10^{-4}$  atm. (Lodders, 2003), are often used as a first order estimate of  
79 volatility. These would suggest the relative volatilities of these elements increase in the order  
80  $\text{Cu} \ll \text{Zn} \lesssim \text{Sn}$ . Recent vaporization experiments of basalts at higher oxygen fugacities  
81 suggest that the relative volatilities of these elements increase in the order  $\text{Zn} < \text{Cu} < \text{Sn}$  (Zn is  
82 getting relatively less volatile under oxidizing conditions, while Sn stays the most volatile  
83 element; Norris and Wood, 2017). However, another recent vaporization study under more  
84 oxidising conditions suggests that Zn remains more volatile than Cu (Sossi et al., 2019),  
85 although unfortunately Sossi et al. (2019) do not provide data for Sn. Here, we assume an  
86 order of volatility is maintained under tektite forming conditions whereby Sn and Zn have  
87 similar volatility, and Cu is less volatile than Sn and Zn.

88         Understanding the isotopic behaviour of Sn during evaporation will be central to  
89 understanding isotopic data in planetary materials and in impact-related samples. Tin isotope  
90 cosmochemistry is a new field, and so far, published data have been limited to conference  
91 abstracts (Bourdon et al., 2017; Creech and Moynier, 2017; Fitoussi et al., 2017) and one  
92 paper (Creech and Moynier, 2019). However, the discovery of a light isotopic composition of  
93 Sn in lunar samples compared to the Earth (Fitoussi et al., 2017) is unique amongst isotopic  
94 systems analysed so far. Tektites can be used as natural experiments to test the behaviour of  
95 Sn during evaporation.

96         Here, we report the first Sn stable isotope data in a suite of tektites from three of the  
97 four major tektite strewn fields in which Cu and Zn isotopic composition had been previously  
98 reported (Moynier et al., 2009, 2010). We use these data to investigate the processes leading  
99 to the chemical and isotopic compositions of tektites, and, in turn, the effect of volatility  
100 related processes on the isotopic composition of tin. By comparing the isotope fractionations  
101 of Cu, Zn, and Sn, as well as their relative diffusivities, we test the previous model of  
102 Moynier et al. (2009) of volatile loss in a diffusion-limited regime. These constraints on the  
103 isotopic behaviour of Sn during evaporation will be central to understanding isotopic data in

104 planetary materials and in impact-derived samples, and will assist in interpretation of future  
105 analyses of other cosmochemical samples, such as lunar rocks and chondrites.

## 106 **MATERIALS AND METHODS**

### 107 **Sample descriptions**

108 The Sn isotopic composition of a total of seven tektites were analysed in this study,  
109 comprising all samples from previous Zn and Cu work with sufficient remaining sample  
110 material for Sn isotope analysis. Details of the samples are summarized in Table 1. These  
111 tektite samples represent three of the four identified tektite strewn fields, comprising:

112 *Central European tektites:* Four moldavite tektites were studied, three of which are  
113 from the vicinity of Jankov (Jankov 1, Jankov 2, Jankov 3) and one is from the vicinity of  
114 Chlum (Chlum). Of these samples, three have been previously analysed for Zn isotope ratios  
115 by Moynier et al. (2009), and three for Cu isotope ratios by Moynier et al. (2010) – see Table  
116 1.

117 *North American tektites:* One North American tektite (BED 8402), which is from the  
118 bediasite sub-strewn field (Texas). This sample has been previously described in Weinke and  
119 Koeberl (1985), and analysed for Zn and Cu isotope ratios by Moynier et al. (2010, 2009).

120 *Australasian tektites:* Two Australasian tektites, both indochinites, were analysed in  
121 this study. These comprise one Muong Nong type tektite from Thailand (MN 8309) and one  
122 splash-form tektite from Hainan, China (HSF1). Both of the Australasian tektites have been  
123 analysed for major and trace element compositions by Koeberl (1992) or Moynier et al.  
124 (2009); Cu and Zn isotope data have also been published by Moynier et al. (2009, 2010).

125 For comparison with terrestrial compositions, isotopic and concentration data for four  
126 commonly used USGS terrestrial reference materials are also given in Table 1. Note that in  
127 the following discussion, a generic terrestrial upper crust  $\delta^{122/118}\text{Sn}$  (the per mil deviation of  
128 the  $^{122}\text{Sn}/^{118}\text{Sn}$  ratio from the IPGP\_Sn standard) composition is approximated as 0.00–  
129 0.50‰, with [Sn] of 0.65–2.50  $\mu\text{g g}^{-1}$ , based on data from Creech et al. (2017) and

130 Badullovich et al. (2017); the exact composition of the tektite source rocks may be further  
131 constrained in the future as Sn isotope data for more geological materials become available.

### 132 **Analytical methods**

133 Tin purification and isotopic analysis were conducted using the same methods as  
134 described by Creech et al. (2017; 2019). Powders comprising 55–380 mg of tektite samples  
135 were digested in a mixture of concentrated HF:HNO<sub>3</sub> in closed teflon beakers on a hotplate at  
136 120 °C. Dissolved tektite samples were evaporated to incipient dryness at 100 °C, taken up in  
137 a small volume of concentrated HCl and dried down again before being taken up in 2 mL of  
138 0.5 M HCl for loading on to columns. Biorad columns were prepared with 1.5 mL per column  
139 of TRU resin and cleaned prior to sample loading by repeated rinsing with 0.5 M HNO<sub>3</sub>, 0.5  
140 M HCl and milli-Q water. After loading, matrix was rinsed off the column with 4 mL of 0.5  
141 M HCl and 7 mL of 0.25 M HCl, and Sn was subsequently eluted in 10 mL of 0.5 M HNO<sub>3</sub>.

142 Prior to sample digestion, samples were doped with a <sup>117</sup>Sn–<sup>122</sup>Sn double-spike to  
143 give approximate proportions of 40% Sn from the double-spike and 60% from the sample, as  
144 detailed by Creech et al. (2017). Tin stable isotope measurements were carried out using a  
145 Thermo-Scientific Neptune Plus MC-ICPMS at the Institut de Physique du Globe de Paris,  
146 France, under analytical conditions identical to those described by Creech et al. (2017).  
147 Isotope results are reported in  $\delta^{122/118}\text{Sn}$  notation. Sufficient material was only available to  
148 measure each tektite sample once. Creech et al. (2017) reported an external reproducibility for  
149 this technique of (conservatively)  $\pm 0.065\text{‰}$  on  $\delta^{122/118}\text{Sn}$ , or ca. 0.016‰ on a per amu basis.  
150 Uncertainties on figures represent either the internal error of the respective analysis or the  
151 external reproducibility, whichever is larger.

152 Data reduction was conducted offline using the freely available double-spike data  
153 reduction tool IsoSpike ([www.isospike.org](http://www.isospike.org); Creech and Paul, 2015).

## 154 RESULTS

155 Tin isotopic and concentration data for tektites are presented in Table 1, along with  
156 literature data for Zn and Cu for the same tektite samples. Tektites are strongly fractionated  
157 towards heavy Sn isotopic compositions relative to those of terrestrial upper crust samples  
158 (Table 1; Fig. 1). Overall, the tektites span a range in  $\delta^{122/118}\text{Sn}$  of ca. 2 ‰, which is  
159 significantly larger than any natural Sn isotopic variation so far reported for geological  
160 samples. The single North American tektite has a  $\delta^{122/118}\text{Sn} = 1.40 \pm 0.18\text{‰}$ , whereas the  
161 Central European tektites span a range from  $\delta^{122/118}\text{Sn} = 1.35 \pm 0.03\text{‰}$  (Jankov 1) to  $2.53 \pm$   
162  $0.15\text{‰}$  (Chlum). The Australasian tektites are more variable; the splash form tektite from  
163 Hainan (HSF1) has a similar isotopic composition to the North American and Central  
164 European tektites ( $\delta^{122/118}\text{Sn} = 1.60 \pm 0.10\text{‰}$ ), while the Muong Nong-type tektite MN8309 is  
165 less fractionated, with a composition more similar to terrestrial upper crust ( $\delta^{122/118}\text{Sn} = 0.46 \pm$   
166  $0.18\text{‰}$ ; basalt, Badullovich et al., 2017).

## 167 DISCUSSION

168 We observe a general shift towards heavy isotope compositions in tektites that is  
169 correlated with decreasing Sn content (Table 1, Fig. 1). This observation is consistent with  
170 evaporative loss of the light Sn isotopes during the molten phase of tektite formation, as  
171 previously observed for Cu and Zn (Moynier et al., 2010, 2009; Rodovská et al., 2017). In  
172 Figure 1 we show calculated isotopic compositions based on mass-dependent fractionation by  
173 Rayleigh distillation, which describes the isotopic evolution of a system from which a phase  
174 is continuously extracted. The isotopic composition of the tektites is well reproduced by a  
175 Rayleigh distillation, following a fractional distillation relationship for Sn as given in  
176 equation (1):

$$177 \quad \frac{{}^{122}\text{Sn}/{}^{118}\text{Sn}}{{}^{122}\text{Sn}/{}^{118}\text{Sn}_0} = \left( \frac{[\text{Sn}]}{[\text{Sn}]_0} \right)^{\alpha-1} \quad (1)$$

178 where  $\alpha$  is the isotope fractionation factor, and the subscript 0 indicates initial values. The  
179 initial composition is an assumed terrestrial upper crust composition (note: on the scale of



180 isotopic fractionation observed in tektites, varying the initial composition within the range  
181 observed in terrestrial samples has little effect on the shape of projection, although it does  
182 change the relative position). In an ideal case, the kinetic fractionation factor,  $\alpha$ , is dependent  
183 on the square root of the ratio of the masses involved (i.e.,  $\alpha_{ideal} = \sqrt{(M^{118}Sn/M^{122}Sn)}$ ,  
184 where  $M^{118}Sn$  and  $M^{122}Sn$  are the masses of  $^{118}Sn$  and  $^{122}Sn$ , respectively; in the case of Sn,  
185  $\alpha_{ideal} = 0.983$ ), although in natural samples fractionation factors are found to deviate from this  
186 ideal case (e.g., Moynier et al., 2009, 2010; Day and Moynier, 2014; Wimpenny et al., 2019).  
187 Suitable natural samples can thus be used to derive empirical kinetic fractionation factors that  
188 may be useful in interpreting volatility related isotope effects in other settings (with the caveat  
189 that additional factors/reactions may also affect isotopic fractionation). Tektites are well  
190 suited for this purpose.

191 As shown in Figure 1, the Sn isotopic compositions of most tektites are very close to  
192 predicted compositions from fractional distillation, for  $\alpha$  values of 0.998 (Central European  
193 and North American tektites) or 0.999 (Australasian), and an initial composition of  
194  $\delta^{122/118}Sn_0 = 0.13\text{‰}$  and  $[Sn]_0 = 1.6 \mu\text{g g}^{-1}$ . This match between observed and predicted  
195 isotope fractionation behaviour supports that during kinetic evaporation of a silicate melt the  
196 lighter isotopes of Sn are preferentially lost in the vapour.

197 The Muong Nong-type tektite MN8309 is closest in composition to terrestrial upper  
198 crust. This is consistent with previous observations that these tektites are less depleted in  
199 volatiles than splash form tektites, contain mineral remnants of the target, and were possibly  
200 deposited closer to the site of impact, having experienced lower formation temperatures  
201 (Glass and Barlow, 1979; Koeberl, 1994, 1992). Muong Nong-type tektites were also found to  
202 be less isotopically fractionated for Zn and Cu (Moynier et al., 2010, 2009). These results  
203 reflect the temperature–time regime (i.e., likely shorter periods of melting and evaporation)  
204 experienced by this group of tektites.

205           The Sn isotope data from tektites show strong correlations with Cu and Zn isotope  
206 data (Table 1, Fig. 2). The correlations between these different isotope systems suggest that  
207 they have been fractionated by a common process. Linear regressions through the data yield  
208  $R^2$  values of 0.94 and 0.73 for Sn vs. Cu and vs. Zn, respectively (note that the moldavite  
209 from Chlum appears to be an outlier in Cu isotope space, and was excluded from that  
210 regression). Considering the similar relative mass differences of the respective isotope ratios  
211 (i.e., 3.13% for  $^{66}\text{Zn}/^{64}\text{Zn}$ , 3.17% for  $^{65}\text{Cu}/^{63}\text{Cu}$ , and 3.39% for  $^{122}\text{Sn}/^{118}\text{Sn}$ ), the slopes of the  
212 regressions in Figure 1 represent the relative rate of isotope fractionation for each element,  
213 which may result from e.g., volatility or diffusion rates, as discussed below. We find that the  
214 rate of Sn isotope fractionation in tektites is ca. 94% that of Zn isotopes (i.e., not significantly  
215 distinguishable within uncertainties), and ca. 17% that of Cu isotopes. If we assume  
216 volatilities decrease as  $\text{Sn} \geq \text{Zn} > \text{Cu}$ , as described above, this fractionation must be partially  
217 moderated by another factor.

218           In the model of Moynier et al. (2010, 2009), molten tektite droplets are viscously  
219 stirred, with differential evaporation occurring at the surface. While the evaporation is  
220 controlled by volatility, the availability of elements at the surface for evaporation is governed  
221 by the rate of diffusion, which depends on many factors (e.g., T, melt composition, ionic  
222 radii). Where the diffusion rate of an element is slow, a diffusion-limited regime  
223 predominates, and evaporative loss of the element is restricted. Thus, elements that diffuse  
224 slowly in a melt are expected to be less isotopically fractionated, than elements that diffuse  
225 quickly, and for a fixed evaporation flux, the relative isotopic fractionation between volatile  
226 elements will be controlled by the diffusivity.

227           In anhydrous melts, the relative diffusivities of these elements increase in the order Sn  
228  $< \text{Zn} \ll \text{K}, \text{Cu}$  (Behrens and Hahn, 2009; Zhang et al., 2010; Yang et al., 2016; Ni et al.,  
229 2017, 2018; see Supplementary Material for further discussion). Given the similar volatility  
230 of Sn relative to Zn, the larger relative fractionation in Zn is therefore likely a consequence of

231 the slower diffusion of Sn in molten droplets. The relatively large isotope fractionation  
232 observed in Cu isotopes (Moynier et al., 2010), despite its lower volatility than Sn (and  
233 possibly Zn), is largely due to the greater availability of Cu resulting from the faster diffusion  
234 of Cu<sup>+</sup>.

## 235 CONCLUSIONS

236 Tektites are enriched in the heavier isotopes of Sn, and the isotopic composition  
237 correlates with the degree of elemental Sn depletion in these samples. This is consistent with  
238 evaporative loss of light Sn isotopes in to the vapour phase during tektite formation. The  
239 observed Sn isotopic fractionation in tektites is consistent with calculations of fractional  
240 distillation of Sn, with corresponding kinetic fractionation factors ( $\alpha$ ) of ca. 0.998–0.999. A  
241 relationship is observed isotopic fractionation in Sn, Zn and Cu, suggesting a common  
242 fractionation process. The smaller relative isotopic fractionation of Sn compared to Zn,  
243 despite similar volatility, and the much larger isotopic fractionation of Cu reflects the  
244 availability of these elements at the droplet surface, which is limited by the rate of diffusion.

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## FIGURE CAPTIONS

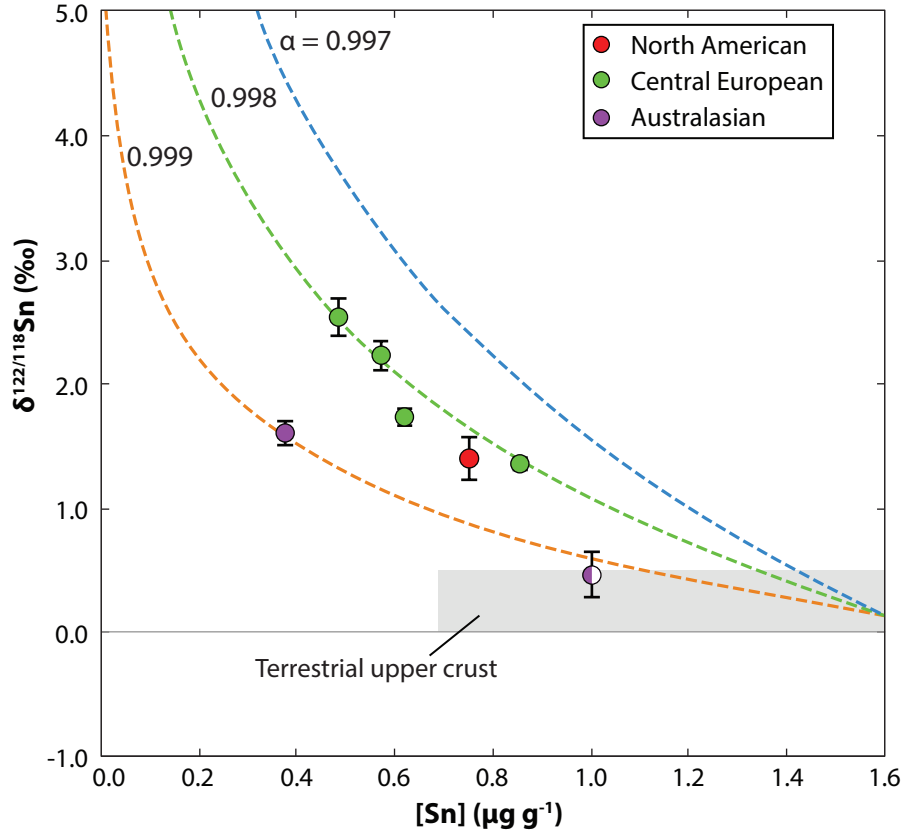
Figure 1. Sn isotopic composition versus Sn content in tektites, showing predicted isotopic fractionation (calculated using equation 1) from an initial terrestrial upper crust composition as Sn is lost from the reservoir. The different colour lines represent calculations using different hypothetical fractionation factors (i.e.,  $\alpha$  values), as indicated. The half-filled symbol indicates the Muong Nong type tektite, MN8309. Error bars reflect the internal error or the external reproducibility, whichever is larger. Where error bars are not visible, uncertainties are smaller than the data point.

Figure 2. Comparison between tektite data for Sn isotopes and a) Cu isotopes and b) Zn isotopes. Dashed lines represent linear regressions through the data, with  $R^2$  values as indicated on the figure. The half-filled symbol indicates the Muong Nong type tektite, MN8309. The upper-most symbol indicates the moldavite Chlum, which was excluded from the linear regression on plot (a) as it is an outlier in both Cu and Sn isotope space. Error bars reflect the internal error or the external reproducibility, whichever is larger. Where error bars are not visible, uncertainties are smaller than the data point.

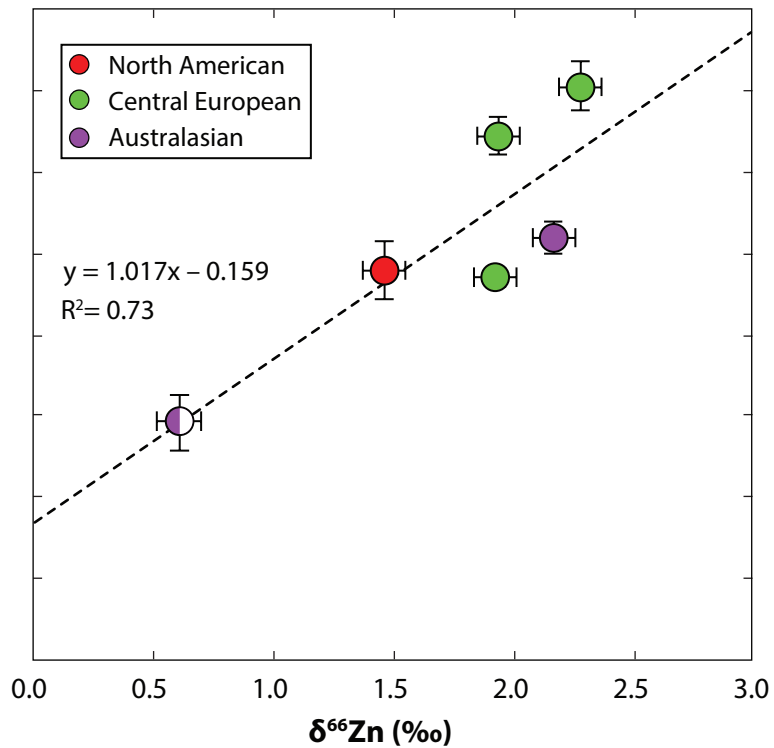
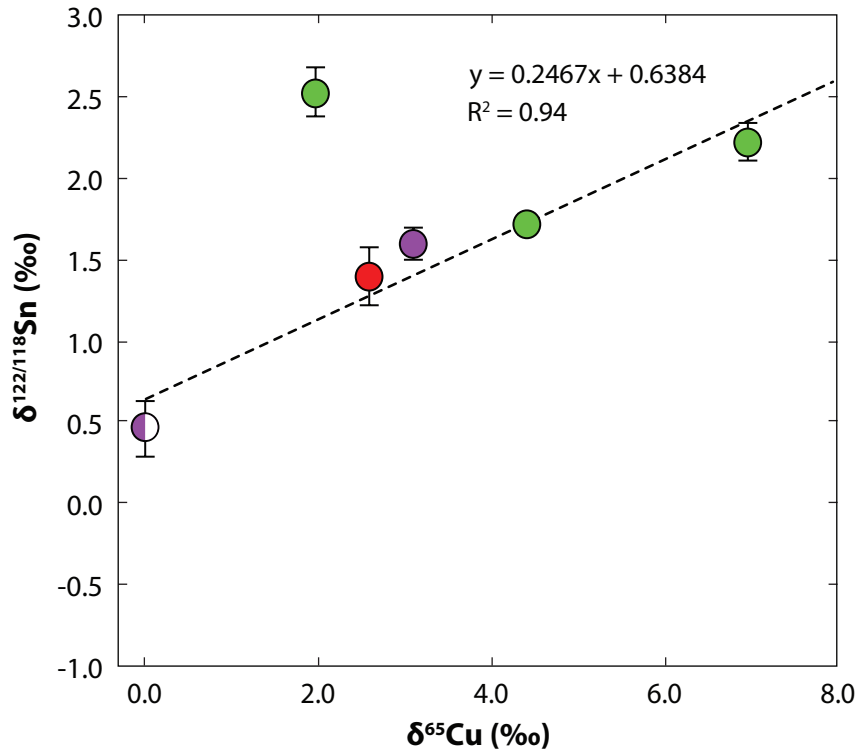
## TABLE CAPTION

Table 1. Tin, Cu, and Zn isotopic and concentration data for tektites and USGS reference materials.

Table 1 footnote. Tin isotope and concentration data for USGS reference materials are from Creech et al. (2017). Zinc isotope data are from Moynier et al. (2009), with the exception of AGV-2, which is from Moynier et al. (2010), and all Zn isotope data are relative to JMC-Lyon. Copper isotope data from Moynier et al. (2010), with the exception of BCR-2 which is from Chen et al. (2016), and all Cu isotope data are relative to NIST SRM 976. Cu and Zn concentration data for USGS reference materials are GeoReM recommended values.







Field	Type	Sample ID	$\delta^{122/118}\text{Sn}$ (‰)	$\pm$ (2 sd)	[Sn] ( $\mu\text{g g}^{-1}$ )	$\delta^{65/63}\text{Cu}$ (‰)	$\pm$ (2 sd)	[Cu] ( $\mu\text{g g}^{-1}$ )	$\delta^{66/64}\text{Zn}$ (‰)	$\pm$ (2 sd)	[Zn] ( $\mu\text{g g}^{-1}$ )
North American	Bediasite	BED 8402	1.400	0.180	0.754	2.60	0.10	6.5	1.46	0.09	26
Central European	Moldavite	Jankov 1	1.352	0.028	0.853				1.92	0.09	29
Central European	Moldavite	Jankov 2	2.230	0.120	0.570	6.98	0.10	0.1	1.93	0.09	23
Central European	Moldavite	Jankov 3	1.725	0.051	0.619	4.41	0.10	0.1			24
Central European	Moldavite	Chlum	2.530	0.150	0.486	1.99	0.10	0.4	2.27	0.09	19
Australasian	Hainan	HSF1	1.600	0.100	0.379	3.11	0.10	2.2	2.16	0.09	8
Australasian	Muong Nong	MN8309	0.460	0.180	1.000	0.02	0.10	11.4	0.61	0.09	74
BHVQ-2	Basalt		0.452	0.027	1.90	0.10	0.07	129.3	0.25	0.01	103.9
BCR-2	Basalt		0.404	0.650	2.36	0.20	0.09	19.7	0.27	0.06	129.5
GSP-2	Granodiorite		0.180	0.220	8.32			43.0			120.0
AGV-2	Andesite		0.310	0.120	2.07	0.10	0.11	51.5	0.25	0.09	86.7

n

Major element: Si	Al	Fe	Mg	Ca	Ti	K	Na	H <sub>2</sub> O	H <sub>2</sub> O Beran and Koeberl	Major element: SiO <sub>2</sub>	Al <sub>2</sub> O <sub>3</sub>	FeO	MgO	CaO	TiO <sub>2</sub>	K <sub>2</sub> O	Na <sub>2</sub> O
BED 8402	4.44	2.22	0.37	0.47	1.63	1.04	0.027			77.8	12.14	3.01	0.62	0.69	0.59	2.38	1.65
Jankov 1	5.54	1.34	1.17	1.78	3.01	0.35				0.529251	0.777305	0.603036	0.724691	0.599343	0.830148	0.741857	
Jankov 2	5.2	1.3	1.52	2.29	2.85	0.31				6.43	2.34	0.37	0.50	0.35	1.98	1.22	
Jankov 3																	
Chlum	3.84	1.17	1.49	2.45	2.39	0.2											
HSF1	4.41	2.81	0.93	1.15	1.79	0.82											
MN8309									0.024								
Major element: SiO <sub>2</sub>	Al <sub>2</sub> O <sub>3</sub>	FeO	MgO	CaO	TiO <sub>2</sub>	K <sub>2</sub> O	Na <sub>2</sub> O	H <sub>2</sub> O									
BED 8402	77.8	8.39	2.86	0.61	0.65	0.59	1.96	1.40	0.027								
Jankov 1	10.47	1.72	1.94	2.46	3.63	0.47											
Jankov 2	9.83	1.67	2.52	3.16	3.43	0.42											
Jankov 3																	
Chlum	7.26	1.51	2.47	3.38	2.88	0.27											
HSF1	8.33	3.62	1.54	1.59	2.16	1.11											
MN8309									0.024								
		0.529251	0.777305	0.603036	0.724691	0.599343	0.830148	0.741857									

Weinke and K Moynier et al. : Moynier et al. : Moynier et al. : Moynier et al. : Weinke and K Moynier et al. : Moynier et al. : Beran and Koeberl (2010)

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