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1 **Collisional and alteration history of the CM parent body**

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3 Lionel G. Vacher<sup>a\*</sup>, Yves Marrocchi<sup>a</sup>, Johan Villeneuve<sup>a</sup>, Maximilien J. Verdier-Paoletti<sup>b</sup> and  
4 Matthieu Gounelle<sup>c,d</sup>

5  
6 <sup>a</sup>CRPG, CNRS, Université de Lorraine, UMR 7358, Vandoeuvre-les-Nancy, F-54501, France

7 <sup>b</sup>Department of Terrestrial Magnetism, Carnegie Institution of Washington, 5241 Broad

8 Branch Road NW, Washington, DC 20015, USA

9 <sup>c</sup>IMPMC, MNHN, Sorbonne Universités, UMR CNRS 7590, 57 rue Cuvier, 75005 Paris,

10 France

11 <sup>d</sup>Institut Universitaire de France, Maison des Universités, 103 boulevard Saint-Michel, 75005  
12 Paris, France

13  
14 \*Corresponding author: lvacher@crpg.cnrs-nancy.fr

15  
16 **Keywords:**

17 Clasts, calcite, aragonite, oxygen and carbon isotopes, impacts, X-ray tomography

18

19 **ABSTRACT**

20

21 Boriskino is a little studied CM2 chondrite composed of millimeter-sized clasts of  
22 different lithologies and degrees of alteration. Boriskino thus offers a good opportunity to  
23 better understand the preaccretionary alteration history and collisional evolution that took  
24 place on the CM parent body. The least altered lithology displays  $^{16}\text{O}$ -poor Type 1a calcite  
25 and aragonite grains ( $\delta^{18}\text{O} \approx 30\text{-}37\text{‰}$ ,  $\delta^{17}\text{O} \approx 15\text{-}18\text{‰}$  and  $\Delta^{17}\text{O} \approx -2$  to  $0\text{‰}$ , SMOW) that  
26 precipitated early, before the establishment of the petrofabric, from a fluid whose isotopic  
27 composition was established by isotopic exchange between a  $^{16}\text{O}$ -poor water and  $^{16}\text{O}$ -rich  
28 anhydrous silicates. In contrast, the more altered lithologies exhibit  $^{16}\text{O}$ -rich Type 2a and  
29 veins of calcite ( $\delta^{18}\text{O} \approx 17\text{-}23\text{‰}$ ,  $\delta^{17}\text{O} \approx 6\text{-}9\text{‰}$  and  $\Delta^{17}\text{O} \approx -4$  to  $-1\text{‰}$ , SMOW) that  
30 precipitated after establishment of the deformation, from transported  $^{16}\text{O}$ -rich fluid in  
31 preexisting fractures. From our petrographic and X-ray tomographic results, we propose that  
32 the more altered lithologies of Boriskino were subjected to high intensity impact(s) (10-30  
33 GPa) that produced a petrofabric, fractures and chondrule flattening. Taking all our results  
34 together, we propose a scenario for the deformation and alteration history of Boriskino, in  
35 which the petrographic and isotopic differences between the lithologies are explained by their  
36 separate locations into a single CM parent body. Based on the  $\delta^{13}\text{C}$ - $\delta^{18}\text{O}$  values of the  
37 Boriskino Type 2a calcite ( $\delta^{13}\text{C} \approx 30\text{-}71\text{‰}$ , PDB), we propose an alternative  $\delta^{13}\text{C}$ - $\delta^{18}\text{O}$  model  
38 where the precipitation of Type 2a calcite can occur in an open system environment with  
39 the escape of  $^{13}\text{C}$ -depleted  $\text{CH}_4$  produced from the reduction of C-bearing species by  $\text{H}_2$   
40 released during serpentinization or kamacite corrosion. Assuming a mean precipitation  
41 temperature of  $110^\circ\text{C}$ , the observed  $\delta^{13}\text{C}$  variability in T2a calcite can be reproduced by the  
42 escape of  $\approx 15\text{-}50\%$  of dissolved carbon into  $\text{CH}_4$  by Rayleigh distillation.

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## 1. INTRODUCTION

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The CM carbonaceous chondrites are samples of dark primitive C-type asteroids (Clark et al., 2010; Cloutis et al., 2011) accreted closed to the snowline (Morbidelli et al., 2016) and that contain a high proportion of water (i.e. water/rock ratio = 0.4; Marrocchi et al., 2018), mainly in hydroxyl form in the matrix (Rubin et al., 2007). Their mineralogy and petrography suggest they have undergone brecciation and low temperature hydrothermal alteration, which could have been induced by low velocity impacts and/or by heat released by the decay of the  $^{26}\text{Al}$ , respectively (Dufresne and Anders, 1962; Metzler et al., 1992; Zolensky et al., 1997; Nakamura, 2005; Rubin et al., 2007; Nakato et al., 2008). CM chondrites experienced different degrees of aqueous alteration, from moderate (CM2) to high (CM1), as reflected by their varying proportion of secondary phases, including Fe-, Mg- and S-rich phyllosilicates (e.g., sulfide-hydroxide/cronstedtite association, hereafter referred as TCI), Fe-Ni sulfides, and carbonates (McSween, 1979; Bunch and Chang, 1980; Tomeoka and Buseck, 1985; Zolensky et al., 1997; Brearley, 2006; Rubin et al., 2007; Howard et al., 2009, 2015; Hewins et al., 2014; Lee et al., 2014; Pignatelli et al., 2016, 2017; Rubin and Ma, 2017).

Carbonates are ubiquitous in CMs and represent a minor proportion of secondary mineral assemblages (i.e., up to 2-3 vol.%; Marrocchi et al., 2014; Lee et al., 2014). Their mineralogy provides information about the chemical evolution of the fluid during aqueous alteration. For example, Ca-carbonates occur as aragonite and calcite in most CM2 chondrites, whereas dolomites are only found in the more altered CM chondrites (Johnson and Prinz, 1993; Riciputi et al., 1994; Browning and Bourcier, 1998; Benedix et al., 2003; De Leuw et al., 2010; Lee et al., 2012; Lee et al., 2014; Tyra et al., 2016; Vacher et al., 2017). Furthermore, their oxygen isotopic compositions (i.e.,  $\delta^{17,18}\text{O}$ ) can be used to decipher their precipitation temperature and the origin and evolution of the fluid. Based on the O-isotopic

70 compositions of Ca-carbonates, it has been proposed (i) that they precipitated at varying  
71 temperatures, in the range 0-300°C (Clayton and Mayeda, 1984; Benedix et al., 2003; Guo  
72 and Eiler, 2007; Alexander et al., 2015; Verrier-Paoletti et al., 2017a), (ii) that aqueous  
73 alteration processes occurred essentially in a closed system with a static fluid interacting with  
74 anhydrous silicates (DuFresne and Anders, 1962; Clayton and Mayeda, 1999; Benedix et al.,  
75 2003) and (iii) that the least altered CMs (i.e., Paris and Maribo) accreted a significant  
76 proportion of <sup>17,18</sup>O-rich water inherited from the outer Solar System (Horstmann et al., 2014;  
77 Vacher et al., 2016). Carbon isotopic composition of carbonates ( $\delta^{13}\text{C}$ ) also reflects aqueous  
78 alteration and can constrain the origin of dissolved carbon into the fluid. From bulk and *in situ*  
79 C-isotopic compositions of CM carbonates, it has been postulated that carbonates precipitated  
80 from C-bearing inorganic molecules (i.e., CO or CO<sub>2</sub>; Alexander et al., 2015) or soluble  
81 organic matter (SOM) (Vacher et al., 2017). C-isotopes can also be affected by different  
82 fractionation mechanisms, such as escape of C-bearing gas (e.g., CH<sub>4</sub>, Guo and Eiler, 2007)  
83 during, for example, venting event (Telus et al., 2017).

84 Carbonate precipitation in response to stress events, such as impacts or  
85 accretional/gravitational compaction (Cain et al., 1986; Fujimara et al., 1983) has previously  
86 been reported based on the presence of different petrographic features: e-twinning calcite,  
87 calcite veins and replacement of chondrules by calcite grains (Bunch and Chang, 1980;  
88 Benedix et al., 2003; Lee and Ellen, 2008; Lindgren et al., 2011; Tyra et al., 2012; Lee et al.,  
89 2014). Calcite veins show <sup>16</sup>O-rich isotopic compositions that suggest late precipitation from  
90 evolved solutions (Lindgren et al., 2017). This assumption is consistent with a late carbonate  
91 precipitation event, during which the evolved fluid would have been mobilized from different  
92 region of the parent body or by dehydration of existing phyllosilicates (Rubin, 2012; Lee et  
93 al., 2013). In addition, it has been proposed that impacts could create open fracture networks  
94 and microcracks in CM chondrites that would facilitate the circulation of fluids (DeCarli et

95 al., 2001; Rubin, 2012). Impact-induced formation of fractures and/or cracks has been  
96 reported in shock experiments involving carbonaceous chondrites (Nakamura et al., 1995;  
97 Tomeoka et al., 1999; Nakamura, 2000). These experiments also highlighted the formation of  
98 foliation petrofabrics of the chondritic particles (i.e., chondrules, calcium-aluminum-rich  
99 refractory inclusions, clumps of tochilinite-cronstedtite intergrowths, olivine grains and  
100 flattened aggregates) and the flattening of chondrules with the increase in impact intensity  
101 (Nakamura et al., 1995; Tomeoka et al., 1999; Nakamura, 2000). Preferential orientations  
102 have been observed in CM and CV chondrites with different degrees of foliation intensity  
103 (Rubin, 2012; Lindgren et al., 2015; Hanna et al., 2015). However, impact events are not the  
104 only processes able to create preferential orientation of particles; lithostatic compaction may  
105 also produce flattened chondrules and CAIs and crystallographic orientation of phyllosilicates  
106 and carbonates (Fujimura et al., 1983; Cain et al., 1986; Zolensky et al., 1997; Lee and Ellen,  
107 2008; Lindgren et al., 2015). This latter process was likely involved in the establishment of  
108 the CM and CV petrofabrics as attested by (i) the occurrence of preferential orientation in  
109 poorly altered CM chondrites without fractures and carbonate veins (Lindgren et al., 2015;  
110 Hanna et al., 2015) and (ii) the lack of evidence for strong shock characteristics in some  
111 meteorites (Scott et al., 1992). In addition, CM chondrites were probably highly porous prior  
112 to their deformation (i.e., 30-50%, Hanna et al., 2015), allowing the impact energy to be  
113 dissipated from the collapse of the pore spaces (Suttle et al., 2017). This inconsistency  
114 between a low shock-stage and evidence of compaction in CM chondrites (Lindgren et al.,  
115 2015) highlights that the relationship between aqueous alteration and stress deformation  
116 remains poorly understood.

117 Most of the CM chondrites are breccias and exhibit clasts with various degrees of  
118 aqueous alteration (Rubin and Wasson, 1986; Metzler et al., 1992; Lindgren et al., 2013;  
119 Bischoff et al., 2017). These clasts provide important constraints on the post-accretionary

120 alteration history and collisional activity of the CM chondrites (Lindgren et al., 2013). A  
121 recent report of the carbonaceous chondrite Boriskino, a CM2 chondrite that fell in Russia in  
122 1930, describes the presence of centimetric clasts with contains clasts with a variety of  
123 lithologies (Verdier-Paoletti et al., 2017b), suggesting that this chondrite may provide a good  
124 opportunity to better understand the link between aqueous alteration and the deformation  
125 history of the CM chondrites that is not well investigated yet. The present study therefore  
126 aims to (i) understand the preaccretional alteration history of CM chondrites by characterizing  
127 clasts in CM2 Boriskino that exhibit different degrees of alteration, (ii) retrace the isotopic  
128 evolution of the fluid from the O & C-isotopic composition of Ca-carbonates, and  
129 (iii) identify the mechanisms that induced preferential orientation and aqueous alteration in  
130 CM chondrites. From a survey of two Boriskino polished sections, we adopt a multi-pronged  
131 approach composed of petrographic observations, preferential orientation measurements on  
132 chondritic particles, mineralogical identifications, and O and C-isotopic measurements on Ca-  
133 carbonates. We focus our attention on bulk Boriskino samples by performing X-ray computed  
134 tomography to map flattening chondrules and measure their preferential orientation.

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## 2. MATERIALS AND METHODS

### 2-1 SEM imaging, characterizations and petrofabric analyses

Petrographic observations of two polished sections of the Boriskino CM chondrites (#3788-3 and #3788-4) from the Muséum National d'Histoire Naturelle of Paris (MNHN) were made at the Centre de Recherches Pétrographique et Géochimiques (CRPG) using a scanning electron microscope (SEM) JEOL JSM-6510 equipped with an energy dispersive X-ray detector (Bruker-AXS XFlash, silicon drift detector: SDD). The SEM observations were performed with a beam current of 3 nA at 15 kV. The entire two polished sections were mapped to produce the Back-Scattered Electron (BSE) mosaics. The modal abundance of phases for each lithology was determined from the BSE mosaics using JMicroVision software.

Using the BSE mosaics of the two polished sections, the petrofabric characteristics of each clast were determined following the method of Rubin (2012). In order to compare the median orientation of the petrofabrics within each polished section (i.e., #3788-4), we measured the tilt angle between an arbitrary reference (i.e., the horizontal = 0°) and the primary axis length of each particle (i.e., chondrule, TCI clump and olivine grain). The tilt angles were measured using Adobe Illustrator. We then estimated the standard deviation ( $c$ ) of the petrofabric(s), the high maximum frequency ( $a$ ) and the mean value of the Gaussian curve ( $b$ ) for each clast using a Gaussian fit on the data set ( $y = a \times e^{\left(\frac{x-b}{c}\right)^2}$ ).

157 **2-2 Raman spectroscopy**

158

159 Raman spectra were collected using a LabRAM HR spectrometer (Horiba Jobin Yvon)  
160 equipped with a  $600 \text{ g}\cdot\text{mm}^{-1}$  grating and an Edge filter. The confocal hole aperture and slit  
161 aperture were  $500 \text{ }\mu\text{m}$  and  $100 \text{ }\mu\text{m}$ , respectively. The excitation beam was produced by a  
162  $457.94 \text{ nm}$  blue laser at a power of  $\sim 20 \text{ mW}$ , focused on the sample using a  $\times 100$  LWD IR  
163 objective (Olympus). The number of accumulations and the acquisition time were  $2 \times 5 \text{ s}$ . The  
164 signal-to-noise ratio (S/N) was less than 1% of the signal. Calcite and aragonite were  
165 identified on the basis of their minor Raman bands; calcite has specific bands at  $282$  and  $713$   
166  $\text{cm}^{-1}$  while aragonite has bands at  $207$  and  $704 \text{ cm}^{-1}$  (White, 2009).

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168 **2-3 Carbon and Oxygen isotopes**

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170 The two polished sections were initially coated with carbon to measure oxygen  
171 isotopes. Oxygen isotopic compositions of Ca-carbonates were measured using a CAMECA  
172 ims 1280 HR2 at CRPG (Nancy, France).  $^{16}\text{O}^-$ ,  $^{17}\text{O}^-$ , and  $^{18}\text{O}^-$  ions produced by a  $\text{Cs}^+$  primary  
173 ion beam ( $\approx 15 \text{ }\mu\text{m}$  spot size,  $5 \text{ nA}$ ) were collected in multi-collection mode using three  
174 Faraday cups. A normal-incidence electron gun was used for charge compensation. In order to  
175 remove  $^{16}\text{OH}^-$  interference on the  $^{17}\text{O}^-$  peak and achieve maximum flatness on the top of the  
176  $^{16}\text{O}^-$  and  $^{18}\text{O}^-$  peaks, the entrance and exit slits of the central Faraday cup ( $80$  and  $173 \text{ }\mu\text{m}$ ,  
177 respectively) were adjusted to obtain a Mass Resolving Power ( $\text{MRP} = M/\Delta M$ ) of  $\approx 7000$  for  
178  $^{17}\text{O}^-$  on the central Faraday cup.  $^{16}\text{O}^-$  and  $^{18}\text{O}^-$  were measured on L'2 and H1 (slit 1,  $\text{MRP} \approx$   
179  $2500$ ). Acquisition times were set so as to obtain counting statistics on the order of  $\pm 0.2 \text{ ‰}$   
180 ( $1\sigma$ ) for  $\delta^{18}\text{O}$  and  $\delta^{17}\text{O}$ . We analysed three terrestrial standards: (i) quartz from Brazil ( $\delta^{18}\text{O} =$   
181  $9.6 \text{ ‰}$ ,  $^{16}\text{O} = 1.8 \times 10^9 \text{ Cps}$ ), (ii) forsterite from San Carlos ( $\delta^{18}\text{O} = 5.5 \text{ ‰}$ ,  $^{16}\text{O} = 1.7 \times 10^9$

182 Cps), and (iii) calcite from Mexico (55.6 wt% of CaO and 0.03 wt% of MgO,  $\delta^{18}\text{O} = 23.6 \text{ ‰}$   
183 and  $^{16}\text{O} = 2.3 \times 10^9$  Cps, with a reproducibility 0.3‰,  $n = 26$ ) to define the Instrumental Mass  
184 Fractionation (IMF) for the three oxygen isotopes and correct for instrumental mass  
185 fractionation for calcite.  $^{17}\text{O}/^{16}\text{O}$  and  $^{18}\text{O}/^{16}\text{O}$  isotope ratios were normalized using the  
186 Standard Mean Ocean Water (SMOW). IMF for both calcite and aragonite was determined  
187 from the same Mexico calcite standard because carbonate identification by Raman  
188 spectroscopy was performed after isotopic measurements. It should be noted that the matrix  
189 effect between calcite and aragonite is negligible (i.e.,  $\approx 1\text{‰}$ , Lécuyer et al., 2012) regarding  
190 the precision of oxygen measurements by SIMS. In addition, the low Fe, Mg and Mn content  
191 in Boriskino's Ca-carbonate (i.e.,  $< 1 \text{ wt\%}$ , Verdier-Paoletti et al., 2017b) is not expected to  
192 produce significant IMF (Rollion-Bard et al., 2007). Because IMF values of standards have  
193 been observed to derived linearly over the time, we calculated the IMF value for each sample  
194 analysis by interpolated its value during time (i.e., number of measurements). Typical  
195 measurement errors ( $2\sigma$ ), which took into account the errors in each measurement as well as  
196 the external reproducibility of the standard, were estimated to be  $\approx 0.7 \text{ ‰}$  for  $\delta^{18}\text{O}$ ,  $\approx 0.6 \text{ ‰}$   
197 for  $\delta^{17}\text{O}$ , and  $\approx 0.8 \text{ ‰}$  for  $\Delta^{17}\text{O}$  (where  $\Delta^{17}\text{O}$  represents the deviation from the TFL:  $\Delta^{17}\text{O} =$   
198  $\delta^{17}\text{O} - 0.52 \times \delta^{18}\text{O}$ ).

199 In order to avoid carbon contamination inherited from previous carbon coatings (i.e.,  
200 after the SEM and O-isotopes), sample surfaces were polished with liquid cerium oxide and  
201 ultrasonically cleaned in ethanol for 5 min (Hanon et al., 1998). The samples were then coated  
202 with gold ( $\approx 50 \text{ nm}$ ) in preparation for carbon isotope analysis using a CAMECA ims 1280  
203 HR2 ion microprobe at CRPG-CNRS (Nancy, France). A  $\text{Cs}^+$  primary Gaussian beam of 5 nA  
204 was focused to produce a spot size of  $\approx 20 \text{ }\mu\text{m}$ . A normal-incidence electron gun was used for  
205 charge compensation. The microprobe was specifically tuned to obtain a MRP of  $\approx 5000$  so  
206 that the  $^{12}\text{CH}^-$  contribution to  $^{13}\text{C}$  could be eliminated. The settings of the entrance and

207 multicollector slits (slit 2, Rollion-Bard et al., 2007) and the field aperture were 70  $\mu\text{m}$ , 240  
208  $\mu\text{m}$  and 3000  $\mu\text{m}$ , respectively. Carbon isotopes were measured in multi-collection mode  
209 using two off-axis faraday cups, C and H1, with  $10^{11}$  ohm resistors for  $^{12}\text{C}$  and  $^{13}\text{C}$ ,  
210 respectively. Before each measurement, a presputtering step of 120s with a raster of 20 x 20  
211  $\mu\text{m}$  was applied in order to remove any remaining traces of the carbon coating. The  
212 acquisition time was 5s long and repeated over 40 cycles and the typical internal error for  
213 each measurement was  $\approx 0.8\text{‰}$  ( $2\sigma$ ) on  $\delta^{13}\text{C}$ . A terrestrial calcite from New Caledonia's  
214 carbonatite (55 wt% of CaO and 0.4 wt% of MgO,  $\delta^{13}\text{C} = -5.39 \text{‰}$ ,  $^{12}\text{C} = 1.4 \times 10^7$  Cps,  
215 Rollion-Bard et al., 2007, with a reproducibility of 0.2 (n = 15) and 0.6  $\text{‰}$  (n = 8) for sessions  
216 1 and 2, respectively) was used as a standard to correct for Instrumental Mass Fractionation  
217 (IMF) for both calcite and aragonite. As for O-isotopes, C-isotopes could be affected by  
218 matrix effect. However, the effect of chemical compositions and the aragonite/calcite IMF ( $\approx$   
219 2 $\text{‰}$ , Lécuyer et al., 2012) is supposed to be negligible considering: (i) the precision on SIMS  
220 C-isotopic measurements (i.e.,  $\approx 1 \text{‰}$ ) and (ii) the large range of  $\delta^{13}\text{C}$  shown by Ca-  
221 carbonates (i.e., 20-90 $\text{‰}$ ). Isotopes ratios were normalized to Pee Dee Belemnite (PDB).  
222 Typical measurement errors ( $2\sigma$ ), which took into account the errors in each measurement as  
223 well as the external reproducibility of the standard, were estimated to be  $\approx 0.8 \text{‰}$  for  $\delta^{13}\text{C}$ .

224

#### 225 **2-4 X-ray computed tomography (XCT)**

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227 A bulk sample of Boriskino with a mass of 328 mg was subjected to X-ray computed  
228 tomography (XCT) imaging to provide a full 3D dataset of the sample (**Fig. 1a**). The CT  
229 imaging was made on the AST-RX platform of the Muséum National d'Histoire Naturelle of  
230 Paris (MNHN) using a GE Sensing and Inspection Technologies Phoenix|x-ray v|tome|x  
231 L240-180 CT scanner. We used the microfocus RX source at 240 kV/320 W, detector 400  $\times$   
232 400 mm, with a matrix of 2024 pixels (pixel size: 200  $\times$  200 microns). The settings

233 parameters were as follows: voltage = 80 kV; current = 230  $\mu$ A; exposure: 500 ms; and  
234 isotropic voxel size = 8  $\mu$ m. Data were reconstructed using datavox reconstruction software  
235 (Phoenix|x-ray, release 2.0), then exported into a 16-bit TIFF image stack of 1740 virtual  
236 slices in transverse view.

237 In the full 3D data set of the sample, the grey level of each slice corresponds to the X-  
238 ray attenuation, which is related to density and the atomic number of the material. The darkest  
239 pixels represent the least attenuating material or the lowest density (e.g., matrix and  
240 chondrules) while the brightest pixels represent the most attenuating material or the highest  
241 density (e.g., metal and sulphide grains). Thus, based on the grey level of each image, we  
242 observed 61 elliptic dark-toned objects that we interpret as type I chondrules (**Fig. 1b**). We  
243 painted all voxels into these chondrules on each step of 5 slices for the biggest objects and on  
244 each slice for the smallest object with the Avizo<sup>TM</sup> software. Each chondrule was then  
245 modeled in 3D volume by interpolation of the painted voxels. Next, an ellipsoid was fitted to  
246 each chondrule by merit function using Blob3D software package (Ketcham, 2005a,  
247 Ketcham, 2005b). The orientation, axial length, aspect ratio and location of each ellipsoid  
248 were measured (**Table 2**). Plunges, trends and eigenvectors ( $v_1$ ,  $v_2$  and  $v_3$ ) of the raw data  
249 were calculated and plotted on a stereogram with Stereonet software from the direction  
250 cosines given by the Blob3D results. The eigenvalues ( $\lambda_1$ ,  $\lambda_2$  and  $\lambda_3$ ) were also determined in  
251 order to estimate whether the orientation of the ellipsoids were non-random and defined a  
252 petrofabric, according to the method of (Woodcock and Naylor, 1983). In this method, two  
253 parameters must be determined: (i) the shape parameter (K), which ranges from K = 0 (girdle  
254 distribution) to K >1 (cluster distribution); and (ii) the strength parameter (C), which ranges  
255 from C  $\approx$  0 (weak petrofabric) to C > 4 (strong petrofabric).

256

### 3. RESULTS

#### 3.1 Petrography of Boriskino

The two polished sections of Boriskino are composed of millimeter-sized clasts in direct contact with each other, which can be distinguished by their petrography (**Fig. 2**). Among these clasts, we defined three lithologies (labeled 1 to 3) characterized by varying contents of (i) phyllosilicate, (ii) metallic iron, (iii) sulfide, and (iv) different petrographic types of Ca-carbonates (**Fig. 3, Table 1**).

Lithology 1 is present in both sections of Boriskino and represents 88% and 24% of the sample's surface of section #3 and #4, respectively (**Fig. 2a**). This lithology is characterized by a moderate proportion of matrix and chondrules ( $\approx 60.15$  vol.% and  $\approx 30$  vol.%, respectively), a high proportion of Fe-Ni sulphide grains ( $\approx 8.6$  vol.%) and a low abundance of Ca-carbonates ( $\approx 1.2$  vol.%) and metallic Fe-Ni ( $\approx 0.05$  vol.%) (**Table 1; Fig. 3a**).

Lithology 2 is only present in section #4 and corresponds to 26% of the sample's surface, where it is found in two separated clasts. This lithology shows comparable proportions of matrix, chondrules and Ca-carbonate relative to lithology 1 ( $\approx 62.3$  vol.%,  $\approx 34.7$  vol.% and  $\approx 1.4$  vol.%, respectively) but a lower proportion of Fe-Ni sulphide (1.6 vol.%) and a lack of Fe-Ni metal (**Table 1; Fig. 3b**).

Lithology 3 is present in section #4 and represents 22% of the sample's surface, and possibly in section #3 although it is difficult to assess properly due to its low surface area (i.e., 5%, **Fig. 2**). This lithology differs from the others in that it is characterized by a high abundance of matrix and Ca-carbonate ( $\approx 68.5$  vol.% and  $\approx 2.4$  vol.%, respectively). It has a similar proportion of chondrules to lithology 1 ( $\approx 23.9$  vol.%), a moderate proportion of Fe-Ni sulphide (5.2 vol.%) and a high proportion of Fe-Ni metal ( $\approx 0.5$  vol.%) (**Table 1; Fig. 3c**).

282 **3.2 Ca-carbonate grains**

283  
284 Twenty Ca-carbonate grains were observed in the three different lithologies of Boriskino  
285 (**Fig. 2**). They mainly occur as calcite in all three lithologies (n = 17, **Table 3**), but a few  
286 grains of aragonite were also identified in lithology 3 (n = 4, **Table 3**) (**Fig. 2 & 4**). Based on  
287 the mineralogical classification of CM carbonates (Tyra et al., 2007; Tyra et al., 2012; Lee et  
288 al., 2014; Vacher et al., 2017), they can be classified into two types:  
289 (i) Type 1a grains, which are surrounded by a serpentine/tochilinite rim, and (ii) Type 2a  
290 grains, which have replaced primary minerals with inclusions of Fe-Ni sulphide inclusions.

291

292 3.2.1 Type 1a Ca-carbonates (*represented in figures as green circles for calcite and*  
293 *diamonds for aragonite*)

294

295 Type 1a Ca-carbonates (hereafter T1a) are scattered throughout the matrix of lithology 3  
296 (**Fig. 2a & 2b**). They mostly occur in the form of subhedral or anhedral single or  
297 polycrystalline grains with typical sizes in the range from a few tens to hundreds of  
298 micrometers (**Fig. 4b**). They are systematically partially or totally surrounded by a  
299 serpentine/tochilinite rim with an average thickness of a few micrometers (**Fig. 4a & 4b**).  
300 These rims show irregular boundaries at the contacts with the host-grains. The proportions of  
301 T1a calcite and aragonite grains are similar, with a fraction of  $\approx 66\%$  of calcite (n = 5) and  $\approx$   
302  $44\%$  of aragonite (n = 4). T1a Ca-carbonates have  $\delta^{18}\text{O}$  values ranging from 29.8 to 36.8‰  
303 (mean of 33.9‰,  $\sigma = 2.4\%$ ),  $\delta^{17}\text{O}$  values from 15.2 to 18.3‰ (mean of 16.7‰,  $\sigma = 1\%$ ) and  
304 variable  $\Delta^{17}\text{O}$  values, ranging from -1.8 to -0.4‰ (mean of -1‰,  $\sigma = 0.4\%$ ) (**Fig. 5, Table 3**).  
305 Significant variation in C isotopes is also observed, with  $\delta^{13}\text{C}$  values between 18.3 and 91.8‰  
306 (mean of 55.8‰,  $\sigma = 25.7\%$ ).

307

308 3.2.2 Type 2a calcites (*red circles in figures*)

309

310 The Type 2a calcites (hereafter T2a) were all found in lithologies 1 and 2, with the  
311 exception of grain CC16, which was in lithology 3 (**Fig. 2, Table 3**). They occur as  
312 polycrystalline grains of hundreds of micrometers in size containing Fe-Ni sulphide  
313 inclusions of a few tens of micrometers in size (**Fig. 4c**). According to their location and  
314 morphology, T2a calcites can be divided into two groups: (i) polycrystalline aggregates  
315 located in the matrix, exhibiting phyllosilicate and a high proportion of Fe-Ni sulphide (**Fig.**  
316 **4c**); and (ii) polycrystalline grains that occur in association with phyllosilicates, olivine grains  
317 and Fe-Ni sulphide inclusions (**Fig. 7a & 7b**) inside flattened chondrules surrounded by a  
318 fine-grained rim (hereafter FGR) (**Fig. 7a & 7b**). T2a calcites show  $\delta^{18}\text{O}$  values between 17.2  
319 and 22.5‰ (mean of 19‰,  $\sigma = 1.4\%$ ),  $\delta^{17}\text{O}$  values between 6.2 and 8.8‰ (mean of 7.4‰,  $\sigma$   
320 = 0.8‰) and heterogeneous  $\Delta^{17}\text{O}$  values, ranging from -3.5 to -0.6‰ (mean of -2.5‰,  $\sigma =$   
321 0.8‰) (**Fig. 5, Table 3**). T2a calcites have variable C-isotopic compositions with  $\delta^{13}\text{C}$  values  
322 ranging from 30 to 70.7‰ (mean of 50.1‰,  $\sigma = 12.4\%$ ) (**Fig. 6, Table 3**).

323

324 3.2.3 Veins of calcite (*blue triangles in figures*)

325

326 Calcite veins were only identified in lithology 1 (in the #3, **Fig. 2a**) where they occur as  
327 polycrystalline grains of hundreds of micrometers in length and tens of micrometers in width  
328 (**Fig. 4d**). These veins are in direct contact with the matrix, are free of phyllosilicate, and can  
329 contain micrometer-sized Fe-Ni sulphide inclusions. Due to their irregular form and their low  
330 thickness, only one successful O-isotopic measurement was carried out, revealing a  $\delta^{18}\text{O}$   
331 value of 16.9‰ ( $2\sigma = 0.6\%$ ), a  $\delta^{17}\text{O}$  value of 6.7‰ ( $2\sigma = 0.5\%$ ) and a  $\Delta^{17}\text{O}$  value of -2‰

332 ( $2\sigma = 0.7\%$ ). Unfortunately, no C-isotopic composition was successfully measured due to the  
333 narrow width of the veins.

334  
335

### 3.3 Petrofabric analyses and aspect ratio of type I chondrules

#### 336 3.3.1 *Polished sections*

337

338 Lithology 1 in section #3 contains several sets of ~200-300  $\mu\text{m}$  fractures and ~100-200  
339  $\mu\text{m}$  veins of calcite orientated in planes near the petrofabric plane (i.e., within  $30^\circ$  of the  
340 median azimuth), but also randomly oriented (**Fig. 8a**). Lithologies 1 and 3 in section #3  
341 define a strong petrofabric with 46% and 41%, respectively, of the particle long-axes within  
342  $10^\circ$  of the median azimuth (**Fig. 8a & 8d**). In contrast, lithologies 1 and 2 in section #4  
343 express a weak petrofabric with 26% and 21%, respectively, of the particle long-axes within  
344  $10^\circ$  of the median azimuth (**Fig. 8b & 8c**).

345

#### 346 3.3.2 *Bulk Boriskino*

347

348 The orientation of the primary (R1) and tertiary (R3) axis lengths of 61 elliptic type I  
349 chondrules were plotted on stereonet in order to determine their degree of preferred  
350 alignment (**Fig. 9a & 9b**). According to the method of Woodcock and Naylor (1983), the  
351 primary ellipsoid axis orientations of type I chondrules have a girdle distribution, with a shape  
352 parameter (K) of 0.6 and a moderately weak foliation fabric strength, with a strength  
353 parameter (C) of 1.3 (**Fig. 9a**). The R3 ellipsoid axis orientations of type I chondrules have a  
354 cluster distribution, with a shape parameter (K) of 1.9 and a moderately weak foliation fabric  
355 strength, with a strength parameter (C) of 1.3 (**Fig. 9b**). Both sample distributions are non-  
356 random at the 99% confidence level ( $S_1/S_3 = 3.77$  and  $S_1/S_3 = 3.61$  for R1 and R3,  
357 respectively; Woodcock and Naylor, 1983). The diameters and volumes of best-fit chondrule

358 ellipsoids are variable and range from ~82 to 531  $\mu\text{m}$  (primary axis length) and from 2 to 213  
359  $\mu\text{m}^2$ , respectively (**Table 2**). Their aspect ratios are also heterogeneous, ranging from 1.11 to  
360 3.34 (mean value of 1.87,  $\sigma = 0.46$ , **Table 2**).

## 4. DISCUSSION

### 4.1 Oxygen isotopes and sequence of Ca-carbonate precipitation

Boriskino contains two petrographic types of Ca-carbonate depending on the lithology considered: T1a Ca-carbonates (calcite and aragonite) and T2a calcites. T1a Ca-carbonates are  $^{16}\text{O}$ -poor (mean  $\delta^{18}\text{O}$  and  $\delta^{17}\text{O}$  values of 33.9‰ and 16.7‰, respectively) with a mean and occur as small subhedral grains dispersed throughout the matrix of lithology 3 (**Fig. 5, Table 2**). In contrast, the T2a grains and calcite veins in lithologies 1 and 2 exhibit  $^{16}\text{O}$ -rich compositions compared to T1a grains (mean  $\delta^{18}\text{O}$  and  $\delta^{17}\text{O}$  values of 19‰ and 7.4‰ respectively, **Fig. 5, Table 2**). Some of the T2a calcites were observed in association with olivine grains and phyllosilicates located within chondrules that are surrounded by FGRs (**Fig. 7a & 7b**). Based on these petrographic and isotopic features, it thus appears that Boriskino's Ca-carbonates define two distinct populations: a first population, made up of T1a calcite and aragonite, with mean  $\delta^{18}\text{O}$  and  $\Delta^{17}\text{O}$  values of 33.9‰ ( $\sigma = 2.4\text{‰}$ ) and -1‰ ( $\sigma = 0.4\text{‰}$ ) respectively; and a second population, composed of T2a calcite and calcite veins, with  $\delta^{18}\text{O}$  and  $\Delta^{17}\text{O}$  values of 18.9‰ ( $\sigma = 1.5\text{‰}$ ) and -2.5‰ ( $\sigma = 0.8\text{‰}$ ) respectively (**Fig. 5b**). Taken together, the O-isotopic compositions of Boriskino's Ca-carbonates define a trend with  $\delta^{17}\text{O} = 0.62 (\pm 0.05) \times \delta^{18}\text{O} - 4.3 (\pm 1.3)$  ( $2\sigma$ ,  $R^2 = 0.97$ ;  $\text{MSWD} = 11.6$ ; **Fig. 5a**) that does not fall on the mass-dependent fractionation line (TFL; **Fig. 5a**). This trend is closed, within errors, from those reported in previous studies in different CM chondrites with:

- (i)  $\delta^{17}\text{O} = 0.62 (\pm 0.09) \times \delta^{18}\text{O} - 4.1 (\pm 2.8)$  ( $2\sigma$ ,  $R^2 = 0.95$ ,  $\text{MSWD} = 0.26$ , Tyra et al., 2012)

- 386 (ii)  $\delta^{17}\text{O} = 0.62 (\pm 0.04) \times \delta^{18}\text{O} - 4.08 (\pm 1.1)$  ( $2\sigma$ ,  $R^2 = 0.99$ , MSWD = 1.7,  
 387 Benedix et al., 2003)
- 388 (iii)  $\delta^{17}\text{O} = 0.66 (\pm 0.05) \times \delta^{18}\text{O} - 4.7 (\pm 1.5)$  ( $2\sigma$ ,  $R^2 = 0.87$ , MSWD = 3.3,  
 389 Verdier-Paoletti et al., 2017a)
- 390 (iv)  $\delta^{17}\text{O} = 0.65 (\pm 0.03) \times \delta^{18}\text{O} - 5.4 (\pm 1.2)$  ( $2\sigma$ , Horstmann et al., 2014).
- 391 (v)  $\delta^{17}\text{O} = 0.70 (\pm 0.03) \times \delta^{18}\text{O} - 5.65 (\pm 1.25)$  ( $2\sigma$ , Lindgren et al., 2017)

392  
 393  
 394

The trend with a slope of  $0.62 \pm 0.05$  defined by the Boriskino carbonates (and other  
 395 CMs) cannot result from precipitation of the carbonates from a common fluid (i.e., with  
 396 identical  $\Delta^{17}\text{O}$  values) along a temperature gradient, as this would have produced a slope of  
 397 0.52 (**Fig. 10**). Instead, it implies that the main process controlling the O-isotopic  
 398 compositions of Ca-carbonates relies on variable degrees of isotopic exchanges between a  
 399  $^{16}\text{O}$ -poor fluid (i.e., a CM primordial water with  $\delta^{18}\text{O} = 55 \pm 13\%$  and  $\delta^{17}\text{O} = 35 \pm 9\%$ ;  
 400 Fujiya, 2018) and  $^{16}\text{O}$ -rich anhydrous silicates (Clayton and Mayeda, 1999). In addition, the  
 401 trend defined by the Boriskino Ca-Carbonates is, within error, parallel to the CM Water line  
 402 (i.e., 0.69, Verdier-Paoletti et al., 2017a), and thus supports a mean carbonate precipitation  
 403 temperature of  $\approx 110^\circ\text{C}$ . However, this isotopic exchange model does not exclude the  
 404 possibility that small temperature variations could have occurred during the course of  
 405 alteration, generating a shift in carbonate O-isotopic compositions along a slope of 0.52 for a  
 406 given degree of isotopic exchange (**Fig. 10**). In such a scenario, the  $^{16}\text{O}$ -poor population of  
 407 Ca-carbonates would have precipitated early in Boriskino's alteration history, while the  $^{16}\text{O}$ -  
 408 rich population would have formed later from more evolved fluids (Fujiya et al., 2016). These  
 409 two stages of precipitation are also supported by (i) the identification of several aragonite  
 410 grains in population 1 (**Fig. 4a**) – aragonite is expected to have predated calcite in some other  
 411 least altered CM chondrites (Sofe, 2013; Lee et al., 2013, 2014; Lindgren et al., 2017; Vacher

412 et al., 2017), and (ii) the occurrence of chondrule-hosted calcite in population 2 – calcites  
413 which are frequently found in other highly altered CM chondrites and are inferred to have  
414 formed under extensive episodes of aqueous alteration (Tyra et al., 2012; Lee et al., 2013,  
415 2014; Fujiya et al., 2016). Determining the sequence of precipitation of Boriskino’s Ca-  
416 carbonates within each population is difficult to assess however. T1a calcites and aragonites  
417 are commonly surrounded by Fe-, S- and Si-rich phyllosilicates (**Fig. 4a & 4b**), suggesting  
418 that the precipitation of T1a Ca-carbonates predates the formation of phyllosilicates, as  
419 already observed in other CM chondrites (Browning and Bourcier, 1998; Lee et al., 2013;  
420 Fujiya et al., 2016; Pignatelli et al., 2017, 2018; Vacher et al., 2017). Nevertheless, our  
421 isotopic and petrographic results do not allow us to discriminate the formation sequence of  
422 T1a calcite and aragonite (**Fig. 5 & 6; Table 3**). In population 2, the T2a calcites were  
423 identified in lithologies 1 and 2. These lithologies exhibit high proportions of phyllosilicates  
424 compared to lithology 3 (**Table 1**) and show systematic evidence of mafic silicate  
425 replacement by serpentine (**Fig. 3a, 3b and 7**). To the contrary, T2a calcites and calcite veins  
426 have not been affected by phyllosilicate replacement, as demonstrated by the absence of  
427 serpentine rims around these grains (**Fig. 4c, 4d and 7**). Taken together, these results indicate  
428 that T2a and veins of calcite precipitated after the formation of phyllosilicates. However,  
429 based on our results, no order of precipitation for T2a grains and calcite veins can be inferred.  
430

431  
432

## 4.2 A complex deformation history arising from high pressure impact(s)

433

434 Our 2D measurements of aligned particles in the three lithologies of Boriskino confirmed  
435 that lithologies 1 (section #3) and 3 define a strong petrofabric whereas lithologies 1 (section  
436 #4) and 2 exhibit a weak petrofabric (**Fig. 8**). Conceptually, it might appear surprising that,  
437 from one section to another, the level of aligned particles can show dramatically different  
438 values within a given lithology. However, similar results have already been reported from  
439 SEM-BSE images of several faces of a 3D block of Murchison (Lindgren et al., 2015).  
440 Significant differences in the level of chondrule alignment were reported, based on the  
441 orientation of the sides of the block (20%, 40% and 50% of chondrule long axes within 10° of  
442 the median azimuth for +z, -y and +x sides, respectively; Lindgren et al., 2015). Because the  
443 Boriskino lithology 1 occurs in two different sections, clasts from lithology 1 could show  
444 different spatial orientations, depending on their location on the sample or the orientation of  
445 the section cutting and thus, display varying petrographic intensities. Nevertheless, the  
446 strongest fabric value measured in lithology 1 (i.e., 46%, **Fig. 8a**) is probably the one with the  
447 closest value to the actual maximum. Consequently, we assume that lithology 1 defines a  
448 strong petrofabric despite variations of its particle orientations.

449 The 3D bulk analyses of type 1 chondrules revealed that Boriskino exhibits a moderately  
450 weak foliation fabric ( $C = 1.3$ ) and a high degree of chondrule flattening ( $1.87 \pm 0.46$ ). This  
451 result is consistent with the two XCT studies of Murchison that showed moderately weak  
452 petrofabric (R1) and high aspect ratio ( $C = 1.46$ , aspect ratio =  $1.75 \pm 0.39$ , Lindgren et al.,  
453 2015;  $C = 1.02$ , aspect ratio =  $1.54 \pm 0.22$ , Hanna et al., 2015). A large number of studies  
454 have reported chondrule flattening and/or occurrence of petrofabrics in (i) ordinary chondrites  
455 (hereafter OCs) (Cain et al., 1986; Stöffler et al., 1991; Scott et al., 1992; Gattacceca et al.,  
456 2005; Rubin and Swindle, 2011), (ii) CV chondrites (Nakamura et al., 2000; Watt et al., 2006;

457 Rubin, 2012) (iii) fined-grained micrometeorites (hereafter FgMMs) (Suttle et al., 2017) and  
458 (iv) CM chondrites (Tomeoka et al., 1999; Rubin, 2012; Lindgren et al., 2015; Hanna et al.,  
459 2015). Chondrule flattening or petrofabrics are generally interpreted to be the result of  
460 impact-induced preferential orientations (Sneyd et al., 1988; Stöffler et al., 1991; Scott et al.,  
461 1992; Tomeoka et al., 1999; Nakamura et al., 2000; Gattacceca et al., 2005; Rubin and  
462 Swindel, 2011; Rubin, 2012; Lindgren et al., 2015; Hanna et al., 2015; Forman et al., 2016;  
463 Suttle et al., 2017) rather than a consequence of lithostatic compaction (Martin and Mills,  
464 1980; Fujimara et al., 1983; Cain et al., 1986; Zolensky et al., 1997; Lee and Ellen, 2008).

465 Shock metamorphism evidence in OCs was used to propose a comprehensive petrographic  
466 classification of their degree of shock (S1 to S6, Stöffler et al. 1991). Based on shock  
467 classification, Gattacceca et al. (2005) demonstrated a relationship between the anisotropy of  
468 magnetic susceptibility (AMS) and the silicates's shock features in OCs. Collisional evidence  
469 was also reported in CV chondrites with (i) the occurrence of S3-S4 petrofabrics (Rubin,  
470 2012) and shock features in olivine and pyroxene grains (Nakamura et al., 1992) and (ii) a  
471 strong anticorrelation between their porosity content and their level of shock stage  
472 (Gattacceca et al., 2005). More recently, the study of a large number of micrometeorites  
473 showed that around 40-80% of FgMMs display petrofabrics, biaxial fabrics and brittle  
474 deformation features, indicating that at least some fabrics are impacted at shock pressures >  
475 0.08 GPa. (Suttle et al., 2017). Based on oxygen isotopic compositions as well as textural and  
476 geochemistry features, it appears that micrometeorites are related to CM chondrites (Kurat et  
477 al., 1994; Brownlee et al., 1997; Genge et al., 1997; Suavet et al., 2010; Gounelle et al.,  
478 2005). Although CM chondrites contain shock compaction features similar to FgMMs,  
479 (Lindgren et al., 2011; 2015; Rubin, 2012; Hanna et al., 2015), they do not contain high grade  
480 shock features (i.e., > S1-S2; Scott et al., 1992). This contradiction might be related to the  
481 capability of porous CMs to absorb and dissipate impact energy due to collapse of the pore

482 space (Suttle et al., 2017). A such process has been demonstrated in Murchison with (i) the  
483 observation of squeezed chondrules resulting from a component of noncoaxial shear induced  
484 by impact and (ii) the loss of  $\approx 10\text{-}31\%$  of chondrule bulk porosity after shock compactions  
485 (Hanna et al., 2015). In Boriskino, we also observed squeezed chondrules within lithology 1  
486 (#4, **Fig. 7c**). Other evidence of impact-induced CM petrofabrics has been highlighted in  
487 different studies, including the regolith-brecciated nature of CMs (Metzler et al., 1992; Rubin,  
488 2012) and multiple episodes of deformation recorded by a calcite vein in the LON 94101 CM  
489 (Lindgren et al., 2011).

490 Shock recovery-experiments were performed on CM2 and CV3 chondrites and  
491 successfully reproduced the flattened shapes of chondrules and their petrofabrics (Nakamura  
492 et al., 1995; 2000; Tomeoka et al., 1999). These experiments were carried out at various peak  
493 pressure intensities on Murchison (4-49 GPa) and Allende (11-21GPa) and demonstrated a  
494 good correspondence between the aspect ratio of CM and CV chondrules and impact  
495 pressures (**Fig. 11**). From these experimental correlations, it is possible to quantitatively  
496 estimate the intensity of *a single impact* that induced the chondrule aspect ratio in Boriskino.  
497 The mean XCT chondrule aspect ratio of  $1.87 \pm 0.46$  (**Table 2**) corresponds to impacts with  
498 intensities ranging from 18 to  $> 30$  GPa and 8-31 GPa, taking into account the experiment  
499 results obtained on Murchison and Allende, respectively (**Fig. 11**; Tomeoka et al., 1999;  
500 Nakamura et al., 2000). They are consistent with those determined in the CMs MET 01072,  
501 Murchison and LAP 031166 (Lindgren et al., 2015; Hanna et al., 2015) (**Fig. 11**). Taken  
502 together, our results suggest that some of the CMs probably underwent high intensity impacts.  
503 Nevertheless, shock pressure estimations should be treated with caution as they represent the  
504 effect of only a single impact on the parent body. Because Boriskino represents the final  
505 assemblage of several clasts of different lithology and deformation degree, it seems unlikely  
506 that each clast experienced the same collisional history. In addition, the pressure estimates

507 determined from shock-recovery experiments on Murchison suggest a high intensity of shock  
508 pressure, with pressures that may have exceeded 30 GPa and which should have induced  
509 melting of the matrix and chondrules (Tomeoka et al., 1999). However, no evidence for  
510 melting has been observed in Boriskino.

511 To summarize, there are several lines of evidence for shock features in Boriskino,  
512 including (i) a strong 2D petrofabric in different lithologies, (ii) flattened chondrules, (iii)  
513 calcite veins and fractures subparallel to the petrofabric, (iv) squeezed chondrules, and (v)  
514 millimeter clasts of various petrographies in direct contact with each other. Based on these  
515 observations, we suggest that some of the clasts that make up the Boriskino meteorite (i.e.,  
516 clasts composed of lithologies 1 and 2) were subjected to high intensity impact(s) (i.e., 10-30  
517 GPa) that produced petrofabrics, chondrule flattening, and fractures.

518

### 4.3 Relationship between aqueous alteration and impacts

519

520

521 Understanding the relative timing between deformation and aqueous alteration in  
522 chondrites is crucial, because it can provide constraints on the conditions of parent-body  
523 alteration (Brearley, 2006). It is commonly accepted that aqueous alteration of CM chondrites  
524 took place on their parent body(-ies) due to heat released by the decay of  $^{26}\text{Al}$  (Dufresne and  
525 Anders, 1962; Zolensky et al., 1997). Shock compaction has also been advanced as a principal  
526 alteration process in CM chondrites (Greenwood et al., 1994; Rubin, 2012; Lindgren et al.,  
527 2015; Hanna et al., 2015). Based on a strong correlation between the degree of alteration and  
528 the intensity of impact compaction (**Fig. 12**), Rubin (2012) proposed that random collisions  
529 may induce the formation of fractures and favor the mobilization of liquid water by  
530 dehydration of phyllosilicates or the melting of ice. This is also supported by the occurrence  
531 of dolomite veins with similar orientation to their respective petrofabrics in the highly altered  
532 CM2 QUE 93005 and CM1 SCO 06043 (Lindgren et al., 2015). Other reported petrographic  
533 evidence for impact-induced alteration includes (i) serpentine veins parallel to the foliation  
534 fabric in Murchison (Hanna et al., 2015) and (ii) millimeter lenses rich in phyllosilicates, Ni-  
535 bearing sulfide and Ca-phosphate grains subparallel to fractures and elongated chondrules in  
536 the CM1 MET 01070 (Rubin et al., 2007; Lindgren et al., 2012). The latter suggests the  
537 precipitation of secondary minerals from transported fluid in preexisting fractures (Rubin,  
538 2012). In Boriskino, lithologies 1 and 2 exhibit petrographic features consistent with a high  
539 degree of alteration (i.e., petrologic subtype expected to be 2.1) as demonstrated by the  
540 (i) partial alteration of mafic silicates into chondrules (**Fig. 3a & 3b**), (ii) low metallic Fe-Ni  
541 content ( $\leq 0.05$  vol.%, **Table 1**) and (iii) occurrence of  $^{16}\text{O}$ -rich T2a grains (**Fig. 5a**, Rubin,  
542 2012). Lithologies 1 and 2 also define strong and weak petrofabrics (i.e., 26 and 46% for  
543 lithology 1, respectively, **Fig. 8a and b**) that probably result from different spatial

544 orientations (also expected for the lithology 2 due to its weak petrofabric) and/or separate  
545 section's positions on the cutting plane. However, considering the strongest fabric measured  
546 as the closest value to the actual maximum (i.e., 41%, **Fig. 8a**), the relationship between  
547 degree of alteration *vs* strength of the petrofabric for the lithology 1 is consistent with the  
548 correlation defined by Rubin (2012) and thus favour models of impact-induced alteration  
549 (**Fig. 12**). In addition, the occurrence of large calcite veins subparallel to the petrofabric (**Fig**  
550 **4a and 8a**) and the replacement of primary elongated chondrules by T2a calcite (**Fig. 7a & b**)  
551 also support this type of alteration process. For the lithology 2, the lack of calcite vein and the  
552 weak intensity of its fabric are more problematic, but according to petrographic observations  
553 (i.e., T2a grains and replacement of elongated chondrules), we assumed similar alteration  
554 process than for the lithology 1.

555 In contrast, lithology 3 presents characteristics consistent with a lower degree of alteration  
556 (i.e., presumed to be petrologic subtype 2.5-2.6; Rubin, 2015), due to: (i) unaltered mafic  
557 silicates into chondrules (**Fig. 3c**), (ii) a significant abundance of Fe-Ni metal beads (i.e., ~0.5  
558 vol.%, **Table 1**) and (iii) the presence of <sup>16</sup>O-poor T1a calcite and aragonite. Furthermore,  
559 lithology 3 does not show any petrographic features characteristics of impact-induced fluid  
560 circulation within fractures (e.g., vein of carbonate). Instead, this lithology appears to contain  
561 undistorted Ca-carbonates wrapped by phyllosilicates (**Fig. 13a**), stretched clumps of  
562 phyllosilicates, and elongate sulphide grains parallel to the petrofabric (**Fig. 13b**). These  
563 observations suggest that secondary minerals precipitated before the formation of the  
564 petrofabric and that aqueous alteration predates the deformation. Furthermore, by reporting  
565 the petrofabric intensity (i.e., 41%) as function of the expected CM subtype of the lithology 3  
566 (**Fig. 12**), it is observed that the percentage of particles with long axes within 10° is too high  
567 compared to the predicted value from the correlation of Rubin (2012) (i.e., ≈ 15%, **Fig. 12**).  
568 The strong petrofabric observed in the lithology 3 probably results on later episodes of

569 deformation that have not affected alteration but only increased the preferential orientation of  
570 particles in the lithology 3 (**Fig. 12**).

571 Our results demonstrate that Boriskino is a clastic polymict breccia, composed of various  
572 lithologies, in direct contact with each other, that display different degrees of alteration and  
573 deformation histories. Based on our petrographic and isotopic results, we propose a  
574 deformation and alteration history for Boriskino based on a single CM parent body, where the  
575 different clasts of the breccia chondrite come from different regions of the asteroid (**Fig. 14a**)  
576 (Scott and Taylor, 1983; Lindgren et al., 2013). At time  $t_1$  in our scenario, aqueous alteration  
577 processes were initiated through the melting of ice induced by  $^{26}\text{Al}$  decay within the parent  
578 body (**Fig. 14a**). Hence, the  $^{16}\text{O}$ -poor primitive fluid interacts with the  $^{16}\text{O}$ -rich anhydrous  
579 components of the parent body to precipitate isolated aragonite and calcite grains in the  
580 matrix as observed in lithology 3. Then, as alteration progresses, the chemical composition of  
581 the fluid changes, stopping the precipitation of T1 carbonate and initiating phyllosilicate  
582 formation. We do not exclude the possibility that these events occurred in lithologies 1 and 2  
583 as well, but to date, no evidence for these events has been observed. Next, at time  $t_2$ , one or  
584 several high intensity impacts induced the formation of fractures and petrofabrics and the  
585 stretching of chondrules, principally in lithologies 1 and 2 due to their location close to the  
586 impact zone (**Fig. 14b**). Subsequently, at time  $t_3$ , the establishment of cracks and fractures  
587 favored the circulation of evolved  $^{16}\text{O}$ -rich fluids within the parent body (**Fig. 14c**). This late  
588 fluid flow would have allowed the formation of phyllosilicate and T2 carbonate in chondrules  
589 and the precipitation of veins of calcite in fractures (**Fig. 14c**). Finally, at time  $t_4$ , a new  
590 episode of low intensity impact(s) generated the release of all Boriskino's clasts from their  
591 initial locations, making possible the mixing of the different altered lithologies together to  
592 form the clastic polymict breccia Boriskino (**Fig. 14d**).

593

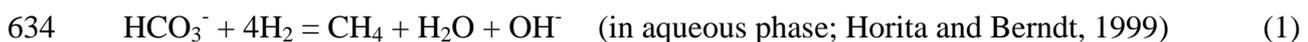
#### 594 4.4 Carbon and oxygen isotopes

595

596 The *in situ*  $\delta^{13}\text{C}$ - $\delta^{18}\text{O}$  compositions of Ca-carbonates in Boriskino are consistent with  
597 those reported in other CM carbonates (**Fig. 15**; Fujiya et al., 2015, 2016; Tyra et al., 2016;  
598 Telus et al., 2017; Vacher et al., 2017). Our results show that Boriskino's Ca-carbonates  
599 exhibit two different populations with distinct  $\delta^{18}\text{O}$  values and significant variations in  $\delta^{13}\text{C}$   
600 values (i.e.,  $\delta^{18}\text{O} \approx 35\text{‰}$ ,  $\delta^{13}\text{C} \approx 20\text{-}90 \text{‰}$  and  $\delta^{18}\text{O} \approx 20\text{‰}$ ,  $\delta^{13}\text{C} \approx 30\text{-}70\text{‰}$  for T1a and T2a,  
601 respectively). Furthermore, these Ca-carbonates do not show positive or negative  $\delta^{13}\text{C}$ - $\delta^{18}\text{O}$   
602 correlations, whether for all the dataset or within each population (**Fig. 15**). Because the  
603 fractionation factor of C isotopes between calcite and  $\text{HCO}_3^-$  in the fluid is small at 0-300°C  
604 (i.e.,  $1000\ln\alpha_{\text{calcite-HCO}_3^-} \approx 1\text{-}2 \text{‰}$ ; Deines et al., 1974; Luo and Wang, 2009), the  $\delta^{13}\text{C}$  values  
605 of Ca-carbonates are useful for (i) tracking the source of carbon involved in their precipitation  
606 (Grady et al., 1988; Alexander et al., 2015; Fujiya et al., 2015; Vacher et al., 2017) and (ii)  
607 identifying fractionation mechanisms, such as equilibrium or kinetic isotope fractionation  
608 (e.g., production of  $\text{CH}_4$  or loss of C-bearing gas from the fluid during serpentinization or  
609 degassing events, respectively; Guo and Eiler, 2007; Fujiya et al., 2016; Telus et al., 2017).

610 The C and O-isotopic compositions of Boriskino's Ca-carbonates can be explained in  
611 numerous ways. Their  $\delta^{13}\text{C}$ - $\delta^{18}\text{O}$  values can be reproduced by the equilibration of a CO gas  
612 ( $\delta^{13}\text{C}_{\text{CO}} = -33\text{‰}$ ) with different mole fraction of  $\text{CO}_2$  in a  $^{16}\text{O}$ -rich aqueous fluid ( $\delta^{18}\text{O} = 5\text{‰}$ )  
613 at low temperature ( $\approx 0\text{-}30^\circ\text{C}$  and  $\approx 80\text{-}120^\circ\text{C}$  for T1a and T2a, respectively; Alexander et al.,  
614 2015) (**Fig. 16a**). They can also be the result from different fluids derived from variable O-  
615 isotopic equilibration and contributions of different  $^{12}\text{C}$ - and  $^{13}\text{C}$ -rich water-soluble organic  
616 compounds (SOM) which demonstrated similar range of  $\delta^{13}\text{C}$  values (i.e., -10 to 60‰, **Fig.**  
617 **16a**, Vacher et al., 2017). However, these two models consider a closed-system alteration  
618 without the escape of C-bearing species. As demonstrated by our petrographic results, T2a

619 calcites occur in lithologies that have probably undergone multiple episodes of shock  
620 compactions causing the formation of cracks and fractures. Under such conditions, the closed-  
621 system is merely an approximation; highly diffusive gas such as C-bearing molecules (e.g.,  
622 CO<sub>2</sub> or CH<sub>4</sub>) likely escaped rapidly from the shocked regions. Such a process has been put  
623 forward to explain the occurrence of two distinct isotopic groups of calcite in CM MET  
624 01070, characterized by (i) high δ<sup>18</sup>O and variable δ<sup>13</sup>C compositions, and (ii) low δ<sup>18</sup>O and  
625 high δ<sup>13</sup>C values (**Fig. 15**; Telus et al., 2017). The authors interpreted this shift in δ<sup>13</sup>C values  
626 as the result of a venting event, in agreement with the petrographic evidence of fluid flow in  
627 MET 01070 (Rubin et al., 2007). Based on a negative δ<sup>13</sup>C-δ<sup>18</sup>O correlation observed in CM  
628 carbonates, it has been also proposed that carbonate precipitation was accompanied by the  
629 production of <sup>13</sup>C-depleted CH<sub>4</sub> and its escape from the CM parent body (Guo and Eiler,  
630 2007). However, as highlight in terrestrial environment, abiotic methane is not produced  
631 directly by serpentinization, but through the reduction of various C-bearing (e.g., CO, CO<sub>2</sub> or  
632 HCO<sub>3</sub><sup>-</sup>) species by H<sub>2</sub> during Fischer Tropsch type (FTT) or Sabatier processes  
633 (Proskurowski, 2010; Etiopie and Sherwood Lollar, 2013):



635 In CM chondrites, H<sub>2</sub> is believed to form during the serpentinization of olivine/pyroxene  
636 minerals (Pignatelli et al., 2017) and/or the corrosion of kamacite grains (Palmer and Lauretta,  
637 2011). Reaction kinetics of FTT or Sabatier processes are theoretically much efficient at high  
638 temperature than at lower temperatures (<100°C). Nevertheless, CH<sub>4</sub> production could be  
639 significant at low temperature due to the action of different catalyses, such as chromite,  
640 magnetite or Fe-Ni alloys (Horita and Berndt, 1999; Foustoukos and Seyfried, 2004) which  
641 occur in CM chondrites (Brearley, 2006; Hewins et al., 2014).

642 Because T2a calcites have precipitated after the formation of fractures and petrofabrics  
643 in the lithologies 1 and 2, we propose that the variability in the δ<sup>13</sup>C values can reflect the

644 release of  $^{13}\text{C}$ -depleted  $\text{CH}_4$  produced by FFT or Sabatier reactions through cracks induced by  
 645 high intensity impact(s). We also considered that escape of  $\text{CH}_4$  has been limited for T1a  
 646 carbonates because they probably formed prior serpentine (Lee et al., 2014; Vacher et al.,  
 647 2017) and thus, they are no include in the model. Furthermore, escape of  $\text{CO}_2$  is not taking  
 648 account in this model, because carbon dioxide is preferentially release in acid environment  
 649 (e.g., at pH = 4-6 in terrestrial evaporating Dead Sea brines, Stiller et al., 1985) whereas  
 650 alteration in CM chondrites appears to occur in alkaline environment (Palmer and Lauretta,  
 651 2011; Zolotov, 2014; Pignatelli et al., 2016; 2017). Assuming an aqueous fluid with dissolved  
 652  $\text{HCO}_3^-$ ,  $\text{CH}_4$  and  $\text{CO}_2$  at isotopic equilibrium in an open system model, the amount of can be  
 653 calculated using a Rayleigh-type isotopic equation:

$$654 \quad R_{\text{HCO}_3^-} = R_{\text{HCO}_3^-_0} \times f^{(\alpha_{\text{HCO}_3^-/\text{CH}_4})^{-1}} \quad (2)$$

655 where  $R_{\text{HCO}_3^-}$  is the isotopic ratio of the  $\text{HCO}_3^-$  after a certain amount of Rayleigh distillation,  
 656  $R_{\text{HCO}_3^-_0}$  is the initial isotopic ratio of  $\text{HCO}_3^-$ ,  $f$  is the fraction of phase  $\text{HCO}_3^-$  remaining, and  
 657  $\alpha_{\text{HCO}_3^-/\text{CH}_4}$  is the equilibrium fractionation factor between  $\text{HCO}_3^-$  and  $\text{CH}_4$ . By incorporating  
 658 the delta notation (i.e.,  $\delta^{13}\text{C}$ ) instead of the isotopic ratio ( $^{13}\text{C}/^{12}\text{C}$ ) in Eqn. (1), we obtain the  
 659 following equation:

$$660 \quad \delta^{13}\text{C}_{\text{HCO}_3^-} = \left[ \left( 1000 + \delta^{13}\text{C}_{\text{HCO}_3^-_0} \right) \times f^{(\alpha_{\text{HCO}_3^-/\text{CH}_4})^{-1}} \right] - 1000 \quad (3)$$

661 Hence, the estimates of the C-isotopic compositions of the fluid (i.e., the  $\delta^{13}\text{C}_{\text{HCO}_3^-}$  values) for  
 662 different amounts of  $\text{HCO}_3^-$  escape are dependent on (i) the initial C-isotopic composition of  
 663 the fluid (i.e., the  $\delta^{13}\text{C}_{\text{HCO}_3^-_0}$  value) and (ii) the equilibrium fractionation factor between  
 664  $\text{HCO}_3^-$  and  $\text{CH}_4$ , which is a direct function of the temperature. This calculation is also linked  
 665 to a fundamental assumption, that the oxygen isotopes of the water constitute an infinite  
 666 reservoir and thus, the O-isotopic composition of the fluid is not affected by the Rayleigh  
 667 distillation.

668 Interestingly, the *in situ*  $\delta^{13}\text{C}$ - $\delta^{18}\text{O}$  values of CM2 and CM1 Ca-carbonate define two  
669 similar trends (except for the  $\delta^{13}\text{C}$ - $\delta^{18}\text{O}$  values of MET 01070) with their C-isotopic  
670 compositions converging toward a low range of  $\delta^{13}\text{C}$  values for a high degree of O-isotopic  
671 equilibration (i.e.,  $\delta^{13}\text{C}_{\text{calcite}} \approx 20\text{-}30\text{‰}$ ) (**Fig. 15**). Furthermore, by including the  $\delta^{13}\text{C}$ - $\delta^{18}\text{O}$   
672 values of the type 1 and type 2 calcites from the CM2 Nogoya (Fujiya et al., 2016), these two  
673 groups exhibit constant  $\delta^{13}\text{C}$  values for different O-isotopic compositions (excluding 2 points  
674 from the type 2) at  $\delta^{13}\text{C}_{\text{calcite}} \approx 30\text{‰}$  ( $\sigma \approx 4\text{‰}$ ) (**Fig. 16b**). As the lowest  $\delta^{13}\text{C}$  value of the T2a  
675 calcite in Boriskino displays similar C-isotopic composition to the Nogoya calcites and the  
676 two CM tendencies (i.e.,  $\delta^{13}\text{C}_{\text{calcite}} = 30\text{‰}$  for the CC9 grain), we assume that this value  
677 represents the initial  $\delta^{13}\text{C}_{\text{HCO}_3^-}$  of the fluid. Interestingly, this value is consistent with the  
678 range of C-isotopic values of the SOM reported by Sephton and Botta (2005) in Murchison.  
679 Thus, by combining the carbon equilibrium fractionation factors of  $\text{HCO}_3^-/\text{CO}_2$  (5-125°C,  
680 Mook et al., 1974) and  $\text{CO}_2/\text{CH}_4$  (0-700°C, Bottinga, 1969), we can determine the  
681 equilibrium fractionation factor of  $\text{HCO}_3^-/\text{CH}_4$  for a given temperature. Assuming a mean  
682 temperature of precipitation for T2a calcites of 110°C, the equilibrium fractionation factor of  
683  $\text{HCO}_3^-/\text{CH}_4$  is 0.9535 at 110°C ( $1000\ln\alpha_{\text{CH}_4\text{-HCO}_3^-} = -47.6\text{‰}$ ). Hence, the observed  $\delta^{13}\text{C}$   
684 variabilities in T2a calcite in Boriskino (i.e.,  $\delta^{13}\text{C}_{\text{calcite}} = 38\text{-}65\text{‰}$ , **Table 3**) can be explained  
685 by a loss of  $\approx 15\text{-}50\%$  of dissolved carbon from the fluid into  $\text{CH}_4$ . We note that this estimate  
686 is largely temperature dependent, and different results will be found at other temperatures  
687 (Fujiya et al., 2016).

688

## 689 5. CONCLUSIONS

690

691 Brecciation is common in CM chondrites and understanding the origin of individual  
692 lithologies recorded among clasts in breccia can provide constraints on the preaccretionary

693 alteration history and collisional activity that took place on their parent body. From a  
694 petrographic and isotopic survey of Ca-carbonates in the clastic breccia CM2 Boriskino, we  
695 have investigated and traced the relative timings of deformation and aqueous alteration of the  
696 CM parent body. The principal results of this paper are:

697 (1) Boriskino is composed of millimeter clasts, in direct contact with each other, that  
698 display various lithologies characterized by different degrees of alteration and  
699 deformation histories.

700 (2) The least altered lithology (lithology 3) contains Type 1a Ca-carbonates that are  
701 surrounded by phyllosilicates. Type 1a Ca-carbonates are  $^{16}\text{O}$ -poor and precipitated  
702 early in Boriskino's alteration history. This is interpreted as the result of isotopic  
703 exchange between a  $^{16}\text{O}$ -poor fluid and  $^{16}\text{O}$ -rich anhydrous silicates in a closed-system  
704 model. Petrographic observations suggest that secondary minerals formed before the  
705 establishment of the petrofabric, and thus that alteration predated the deformation.

706 (3) The more altered lithologies (lithologies 1 & 2) are composed of  $^{16}\text{O}$ -rich Type 2a Ca-  
707 carbonates (calcites) and veins of calcite lying subparallel to the petrofabric. These  
708 calcites precipitated after the establishment of the deformation, from transported  $^{16}\text{O}$ -  
709 rich fluid in preexisting fractures and cracks.

710 (4) Based on strong evidence of shock features in Boriskino, we propose that some of the  
711 clasts that make up the Boriskino meteorite were subjected to high pressure impact(s)  
712 (i.e., 10-30 GPa) that generated strong petrofabrics, chondrule flattening and fractures.

713 (5) Taking all our results together, we propose a reconstruction of the deformation and  
714 alteration history of Boriskino in which the different lithologies are explained by their  
715 separate positions into the CM parent body. Assuming multiple shock compaction  
716 events, the occurrence of a strong petrofabric, calcite veins and elongated chondrules  
717 in the more altered lithologies could be explained by their location close to the impact

718 zone, promoting a late circulation of fluid flow into fractures.

719 (6) Based on the C-isotopic composition of Boriskino Ca-carbonates, we propose that the  
720 formation of T2a calcite probably can take place in an open system environment with  
721 a loss of  $^{13}\text{C}$ -depleted  $\text{CH}_4$  produced from the reduction of C-bearing species by  $\text{H}_2$   
722 released during serpentinization or kamacite corrosion (Guo and Eiler, 2007).  
723 Considering a mean precipitation temperature of Ca-carbonates of  $110^\circ\text{C}$ , an escape of  
724  $\approx 15\text{-}50\%$  of dissolved carbon into  $\text{CH}_4$  by Rayleigh distillation is able to reproduce  
725 the range of C-isotopic compositions observed in T2a calcites.

726

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728

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736

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- 1015
- 1016

1017 **FIGURE CAPTIONS**

1018

1019 **Figure 1:** (a) Full 3D volume rendering of the bulk sample of Boriskino using X-ray  
1020 computed tomography (XCT). The dashed line represents the location of (b). (b) XCT slice  
1021 #1005 from the full volume. The elliptical dark-toned objects are interpreted as type I  
1022 chondrules.

1023

1024 **Figure 2:** Back-scattered electron (BSE) mosaics of the two polished sections of the CM2  
1025 Boriskino (sections #3 & #4). (a) BSE mosaic of Boriskino section #3 in which the different  
1026 lithologies are marked (numbered white circles) and delimited (dashed lines). Analyzed  
1027 Type1a aragonites are represented by green diamonds, Type 1a calcites by green points, Type  
1028 2a calcites by red circles and calcite veins by purple triangle. (b) BSE mosaic of Boriskino  
1029 section #4 in which Ca-carbonate grains and different lithologies are indicated and outlined.  
1030 The arbitrary orientation of two sections used for petrofabric determination is represented by  
1031 the axis at the bottom left of the figure (the horizontal = 0°).

1032

1033 **Figure 3:** BSE mosaics of delimited regions of the three lithologies in section #4. (a) BSE  
1034 mosaic of lithology 1 (number in white circle) in contact with lithology 2 (delimited by the  
1035 dashed line). The Type 2a calcite grain is outlined with a red dashed line. Sph: sulphide, Phy:  
1036 phyllosilicate. (b) BSE mosaic of lithology 2 in contact with lithology 1. (c) BSE mosaic of a  
1037 specific region of lithology 3 in contact with lithology 1 and other lithologies. The Type 1a  
1038 aragonite grains are represented by green diamonds.

1039

1040 **Figure 4:** (a) BSE image of Type 1a aragonite (A) surrounded by a fine rim of phyllosilicate  
1041 (Phy) in lithology 3 (white circle, #4). (b) BSE image of Type 1a calcite (C) surrounded by a

1042 fine rim of phyllosilicate in lithology 3 (#4). (c) BSE image of polycrystalline aggregates of  
1043 Type 2 calcite (C), exhibiting phyllosilicate (Phy) and sulphide (Sph), located in the matrix of  
1044 lithology 1 (#3). (d) BSE image of a calcite vein, hundreds of micrometers in length and ten  
1045 of micrometers in width, in lithology 1 (#3). The large white arrows indicate the mean  
1046 orientations of the petrofabrics of the three lithologies in section #4 as determined in Fig. 7b,  
1047 7c and 7d. The holes on the Fig. a-c show positions of ion probe spots.

1048

1049 **Figure 5:** (a) Oxygen three-isotope plot for Type 1a aragonite (green diamonds), Type 1a  
1050 calcite (green circles), Type 2a calcite (red circles) and the calcite vein (purple triangle) from  
1051 this study (errors are given in  $2\sigma$ ). The dashed line represents the linear regression of our  
1052 dataset ( $\delta^{17}\text{O} = 0.62 (\pm 0.05, 2\sigma) \times \delta^{18}\text{O} - 4.3 (\pm 1.3, 2\sigma)$ ). The black star denotes the mean  
1053 O-isotopic composition of CM anhydrous silicates (Anhydrous silicate<sup>1</sup>: Clayton and Mayeda,  
1054 1984). (b)  $\Delta^{17}\text{O}$  vs  $\delta^{18}\text{O}$  plot of the populations 1 (green points) and 2 (red and purple points)  
1055 of Ca-carbonates.  $\Delta^{17}\text{O}$  represents the deviation from the TFL:  $\Delta^{17}\text{O} = \delta^{17}\text{O} - 0.52 \times \delta^{18}\text{O}$ .  
1056 The O-isotopic compositions of the two populations are represented by the grey square  
1057 (population 1) and the white and grey square (population 2). TFL: Terrestrial Fractionation  
1058 Line,  $\delta^{17}\text{O} = 0.52 \times \delta^{18}\text{O}$ . CMW: CM Water line,  $\delta^{17}\text{O} = 0.69 \times \delta^{18}\text{O} - 2.12$  (Verdier-Paoletti  
1059 et al., 2017a). CCAM: Carbonaceous Chondrite Anhydrous Mineral line,  $\delta^{17}\text{O} = 0.94 \times \delta^{18}\text{O} -$   
1060 4.2.

1061

1062 **Figure 6:**  $\delta^{13}\text{C}$  values of Type 1a aragonite (green diamonds), Type 1a calcite (green circles)  
1063 and Type 2a calcite (red circles) from this study (errors are given in  $2\sigma$ ). The mean values of  
1064 the Type 1 and Type 2 Ca-carbonates are represented by the green and the red square,  
1065 respectively. The *in situ* data from the literature (grey circles) are also shown in the figure  
1066 (CM2.7 Paris<sup>1</sup>: Vacher et al., 2017; CM2.5 Murchison<sup>2</sup>: Fujiya et al., 2015; CM2.2-2.3

1067 Nogoya<sup>3</sup>: Fujiya et al., 2016; CM1 ALH 83100<sup>4</sup>, ALH 84034<sup>4</sup>, MET 01070<sup>4</sup>: Telus et al.,  
1068 2017; CM1 ALH 84049<sup>5</sup>: Tyra et al., 2016).

1069

1070 **Figure 7:** a) BSE image of Type 2a polycrystalline calcite (C) grains located in a flattened  
1071 chondrule surrounded by a fine-grained rim (FGR) and accompanied by phyllosilicate (Phy)  
1072 and sulphide inclusions (Sph) in lithology 2. The solid white line represents the direction of  
1073 the long axis of the fitted chondrule ellipse. The double white arrow indicates the mean  
1074 orientation of the petrofabric of this lithology. b) BSE image of another example of Type 2a  
1075 polycrystalline calcite (C) grains located in a flattened chondrule in lithology 1 (section #3).  
1076 c) BSE image of a squeezed type I chondrule located in lithology 1 (section #4). The red  
1077 dashed line delimits a Type 2a polycrystalline grain located in the matrix. The matrix that  
1078 surrounds the squeezed chondrule is accompanied by numerous patches of phyllosilicate  
1079 (Phy) and sulphide (Sph) grains.

1080

1081 **Figure 8:** Histograms showing the deviation of particles (i.e., chondrule, TCI clump and  
1082 olivine grain) from the median azimuth of each lithology according to the method of Rubin  
1083 (2012). The blue area represents the Gaussian fits ( $y = a \times e^{\frac{(x-b)^2}{c}}$ ) of the data set. (a)  
1084 Lithology 1 in section 3 (#3) has a strong peak near the median azimuth materialized by a  
1085 high maximum frequency ( $a = 24\%$ ), a low standard deviation ( $c = 20.5^\circ$ ) and 46% of the  
1086 particles within  $10^\circ$  of the median azimuth angle ( $R^2$  of the Gaussian fit = 0.98). The solid  
1087 purple lines and the black dashed lines indicate the mean orientations of the calcite veins and  
1088 the fractures, respectively identified in this lithology. The solid black line corresponds to the  
1089 median azimuth value normalized to  $0^\circ$  using the Rubin (2012) method. (b) Lithology 1 in  
1090 section 4 (#4) has a weak peak near the median azimuth revealed by an intermediate  
1091 maximum frequency ( $a = 13\%$ ), an intermediate standard deviation ( $c = 27^\circ$ ) and 26% of the

1092 particles within  $10^\circ$  of the median azimuth angle ( $R^2$  of the Gaussian fit = 0.63). The solid  
1093 black line corresponds to the median azimuth value normalized to  $0^\circ$ , but for comparing the  
1094 mean directions of the three lithologies in section 4, the mean direction values determined  
1095 from the orientation of the section (see Fig. 2) are indicated. (c) Lithology 2 in section 4 (#4)  
1096 also has weak peak near the median azimuth demonstrated by a low maximum frequency ( $a =$   
1097  $10\%$ ), a high standard deviation ( $c = 34^\circ$ ) and 21% of the particles within  $10^\circ$  of the median  
1098 azimuth angle ( $R^2$  of the Gaussian fit = 0.48). (d) Lithology 3 in section 4 (#4) has a strong  
1099 peak near the median azimuth confirmed by a high maximum frequency ( $a = 22\%$ ), a low  
1100 standard deviation ( $c = 21^\circ$ ) and 41% of the particles within  $10^\circ$  of the median azimuth angle  
1101 ( $R^2$  of the Gaussian fit = 0.98).

1102

1103 **Figure 9:** Orientation plot of fitted ellipsoid from bulk X-ray computed tomography analyses  
1104 of a type I chondrule in Boriskino. (a) Projection of the primary axes (R1) of the fitted  
1105 ellipsoid (black circles) and the mean value (V1) and pole value (V3) of the data set (red  
1106 diamonds). The blue color gradient represents the density contour values of the data set.  $C =$   
1107 the strength parameter,  $K =$  the shape parameter and  $n =$  the number of extracted objects. (b)  
1108 Projection of the tertiary axes (R3) of the fitted ellipsoid (black circles) and the mean value  
1109 (V1) and pole value (V3) of the data set (blue diamonds). The red color gradient represents  
1110 the density contour values of the data set.

1111

1112 **Figure 10:** Schematic diagram representing the main processes controlling the oxygen  
1113 isotopic compositions of Ca-Carbonates. The main trend, with a slope in the range 0.62-0.70  
1114 (dashed line, see text), results from O-isotopic exchange between  $^{17,18}\text{O}$ -rich CM primordial  
1115 water (blue point, Fujiya, 2018) and anhydrous silicates (black star, Clayton and Mayeda,  
1116 1984). For a fluid with a given  $\Delta^{17}\text{O}$  value, a thermal gradient can occur during carbonate

1117 precipitation, generating a shift of carbonate O-isotopic compositions along a slope of 0.52  
1118 for a given degree of isotopic exchange. TFL: Terrestrial Fractionation Line, CCAM:  
1119 Carbonaceous Chondrite Anhydrous Mineral line.

1120

1121 **Figure 11:** Relationship between intensity of a *single* shock pressure and mean chondrule  
1122 aspect ratio in an experimental shock on the CM Murchison<sup>1</sup> (solid black line and white  
1123 circle; Tomeoka et al., 1999) and on the CV Allende<sup>2</sup> (black dotted line and black circle;  
1124 Nakamura et al., 2000). The mean X-ray computed tomography (XCT) aspect ratio of the  
1125 Boriskino type I chondrules (yellow square) and Murchison<sup>3</sup> dark-toned XCT objects (purple  
1126 square; Hanna et al., 2015) is reported on the CV and CM trends. The light grey rectangle that  
1127 cuts the CM trend represents the end of the correlation and the effects of the CM shock  
1128 experiments above 30 GPa (Tomeoka et al., 1999).

1129

1130 **Figure 12:** Relationship between degree of alteration *vs* percentage of particles with long  
1131 axes within 10° of the median azimuth. Data from the lithology 1, 2 and 3 are represented by  
1132 red circles, red square and green circle, respectively. Data and associated correlation from  
1133 Rubin (2012) are shown in blue circles and black dashed line, respectively. Percentages of  
1134 chondrules alignment with long axis within 10° from Lindgren et al. (2015) are also  
1135 represented (yellow circles).

1136

1137 **Figure 13:** (a) BSE image showing an interpreted pre-deformation Type 1a Ca-carbonate  
1138 (CC) grain isolated in the matrix and a stretched phyllosilicate (Phy), set in the petrofabric of  
1139 lithology 3 (section #4). The double white arrows represent the deformation of the matrix  
1140 around the Ca-carbonate grain, as highlighted by the elongation of the phyllosilicate. (b) BSE

1141 image of stretched clumps of phyllosilicates (Phy) and elongate sulphide (Sph) grains parallel  
1142 to the petrofabric of lithology 3 (#4).

1143

1144 **Figure 14:** Schematic representation of the thermal and deformation history of the CM parent  
1145 body and the expected locations of the Boriskino lithologies at different times. (a)  $t_1$ :  
1146 precipitation of the Type 1a calcite and aragonite of lithology 3 by the decay of the  $^{26}\text{Al}$ . (b)  
1147  $t_2$ : formation of fractures and cracks by high intensity impact(s). (c)  $t_3$ : circulation of fluid  
1148 flow within the newly formed fractures in lithologies 1 and 2. (d)  $t_4$ : release and mixing of  
1149 Boriskino's clasts from the CM parent body by a new set of low intensity impact(s).

1150

1151 **Figure 15:**  $\delta^{13}\text{C}$  vs.  $\delta^{18}\text{O}$  values for Type 1a aragonite (green diamonds), Type 1a calcite  
1152 (green circles) and Type 2a calcite (red circles) from this study (errors are given in  $2\sigma$ ). We  
1153 also show the *in situ*  $\delta^{13}\text{C}$ - $\delta^{18}\text{O}$  data of the Type 1 (light grey circles) and Type 2 (dark grey  
1154 circles) calcite from the CM2.2-2.3 Nogoya<sup>1</sup> (Fujiya et al., 2016). Because of the large  
1155 amount of data, the ranges of Ca-carbonate  $\delta^{13}\text{C}$ - $\delta^{18}\text{O}$  values reported by Fujiya et al. (2015),  
1156 Fujiya et al. (2016), Telus et al. (2017) and Vacher et al. (2017) are shown as light grey  
1157 (CM2) and dark grey (CM1) areas for simplicity. The  $\delta^{13}\text{C}$ - $\delta^{18}\text{O}$  values of the CM1 MET  
1158 01070 have also been isolated from the literature dataset by black dashed lines (Telus et al.,  
1159 2017). The  $\delta^{13}\text{C}$ - $\delta^{18}\text{O}$  values of CC6, CC7, CC8 and CC15 grains are given as mean values  
1160 and errors associated were calculated using the error propagation law.

1161

1162 **Figure 16:** (a)  $\delta^{13}\text{C}$  vs.  $\delta^{18}\text{O}$  values for Type 1a aragonite (green diamonds), Type 1a calcite  
1163 (green circles) and Type 2a calcite (red circles) from this study (errors are given in  $2\sigma$ ). The  
1164 grey squares represent the *in situ*  $\delta^{13}\text{C}$ - $\delta^{18}\text{O}$  data of the Type 1 Ca-carbonates in the CM2.5  
1165 Murchison<sup>1</sup> (Fujiya et al., 2015) and CM2.7 Paris<sup>1</sup> (Vacher et al., 2017). We also show the *in*

1166 *situ*  $\delta^{13}\text{C}$ - $\delta^{18}\text{O}$  data of the Type 1 (light grey circles) and Type 2 (dark grey circles) calcite  
1167 from the CM2.2-2.3 Nogoya<sup>1</sup> (Fujiya et al., 2016). The white triangle lines represent the  
1168 predicted isotopic compositions of CM carbonates as a function of temperature and the mole  
1169 fraction of  $\text{CO}_2$  in a  $\text{CO}$ - $\text{CO}_2$ - $\text{H}_2\text{O}$  system ( $\delta^{13}\text{C}_{\text{CO}} = -5\text{‰}$  and  $\delta^{18}\text{O}_{\text{H}_2\text{O}} = 5\text{‰}$ ) ( $\text{CO}$ - $\text{CO}_2$  gas<sup>1</sup>:  
1170 Alexander et al., 2015). The white triangles on the model mark  $10^\circ\text{C}$  intervals. The blue  
1171 rectangle represents the predicted range of  $\delta^{13}\text{C}$ - $\delta^{18}\text{O}$  values of Ca-carbonates that  
1172 precipitated from varying equilibration between  $^{17,18}\text{O}$ -rich water with  $^{12}\text{C}$ - and  $^{13}\text{C}$ -rich  
1173 soluble organic matter at  $110^\circ\text{C}$  (Vacher et al., 2017). The  $\delta^{13}\text{C}$ - $\delta^{18}\text{O}$  values of CC6, CC7,  
1174 CC8 and CC15 grains are given as mean values and errors associated were calculated using  
1175 the error propagation law. (b)  $\delta^{13}\text{C}$  vs.  $\delta^{18}\text{O}$  values of the Type 2a calcite (red circles) from  
1176 this study and Type 1 (light grey circles) and Type 2 (dark grey circles) calcite from the CM  
1177 Nogoya<sup>2</sup> (Fujiya et al., 2016) (errors are given in  $2\sigma$ ). The grey dashed line represents the  
1178 initial  $\delta^{13}\text{C}_{\text{HCO}_3^-}$  value of the fluid used in the Rayleigh distillation calculations (see Eq.(2),  
1179  $\delta^{13}\text{C}_{\text{HCO}_3^-} = 30\text{‰}$ ). Percent estimates of dissolved carbon escape from the fluid by Rayleigh  
1180 distillation are also shown for a mean temperature of  $110^\circ\text{C}$ . The  $\delta^{13}\text{C}$ - $\delta^{18}\text{O}$  values of CC6  
1181 and CC8 grains are given as mean values and errors associated were calculated using the error  
1182 propagation law.

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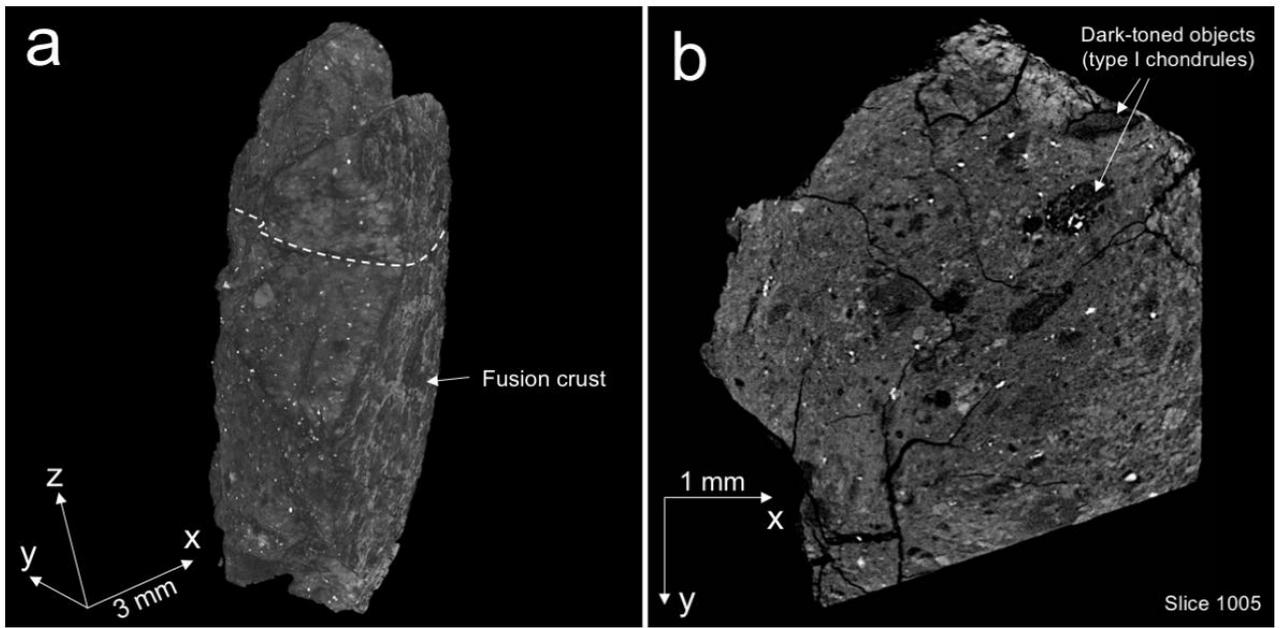
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1187 **FIGURES**

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**Figure 1**

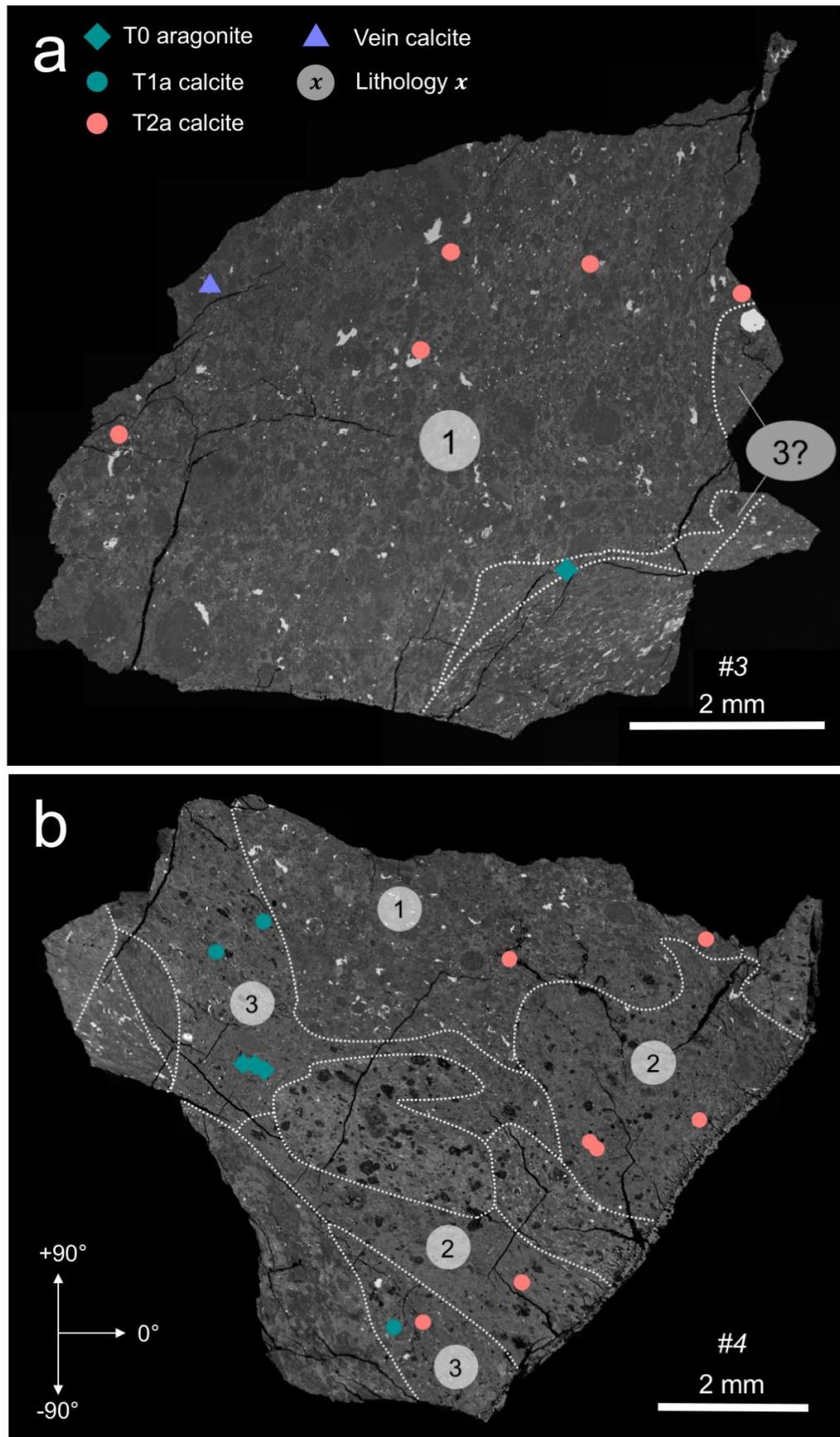
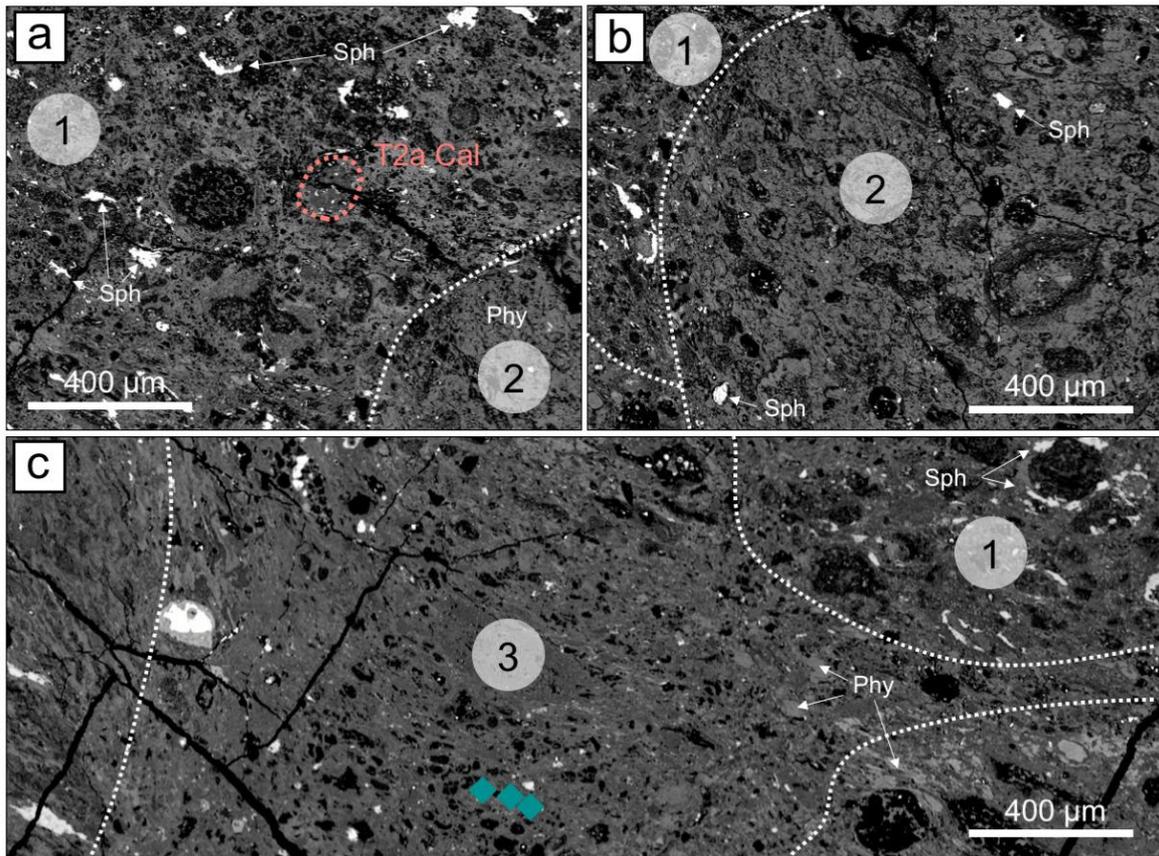


Figure 2

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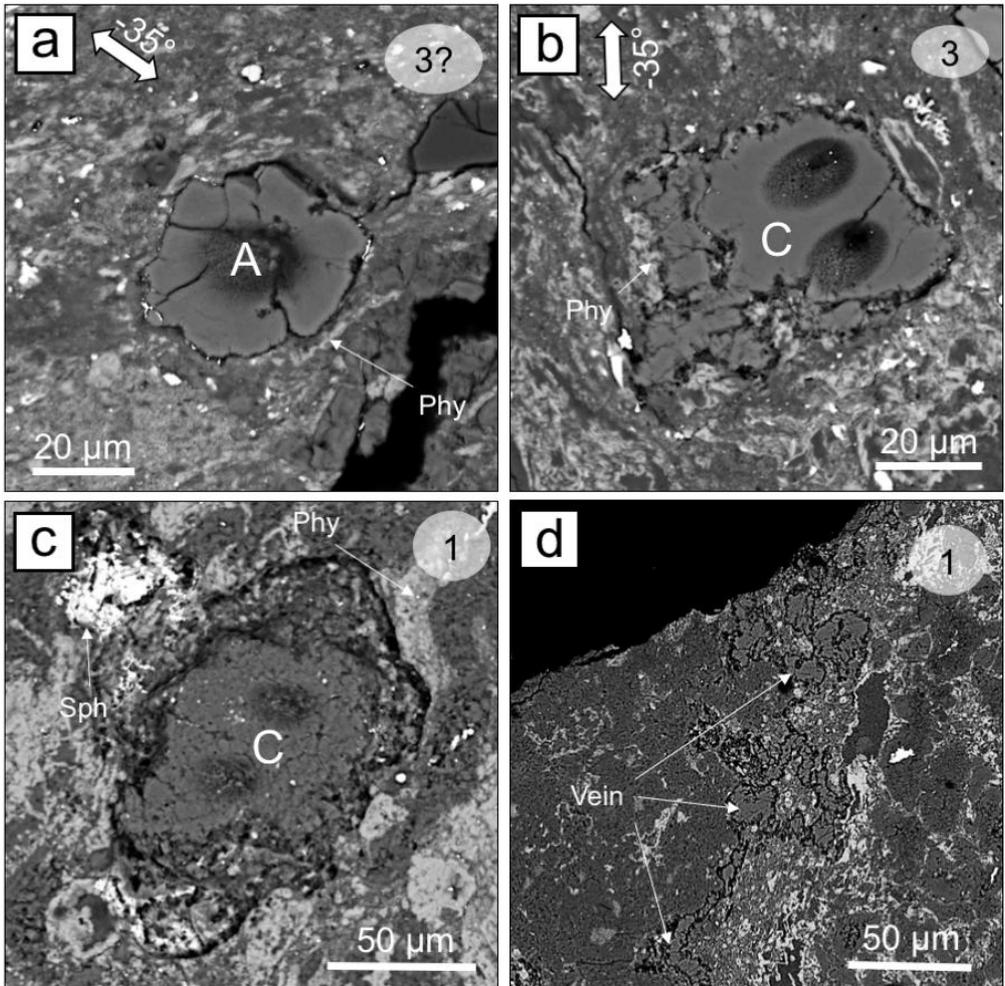
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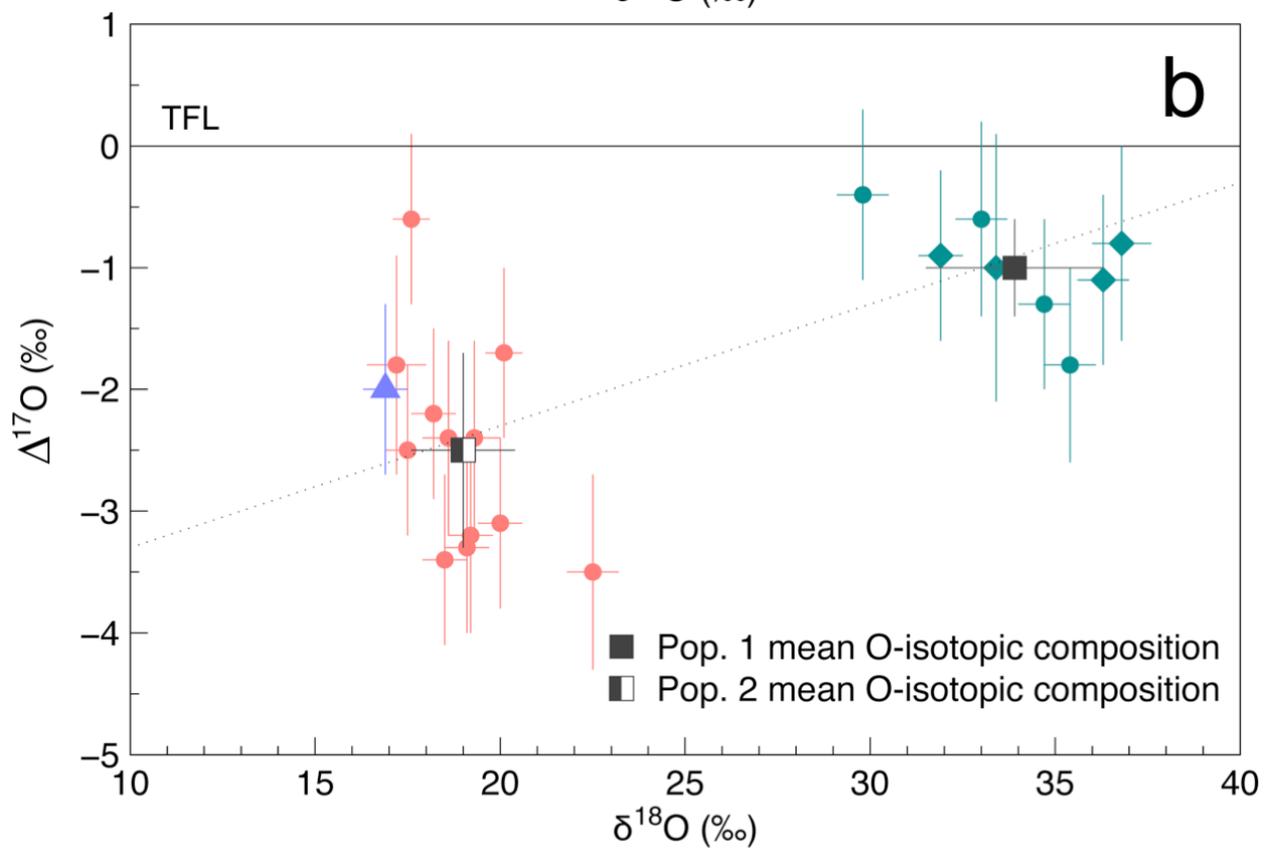
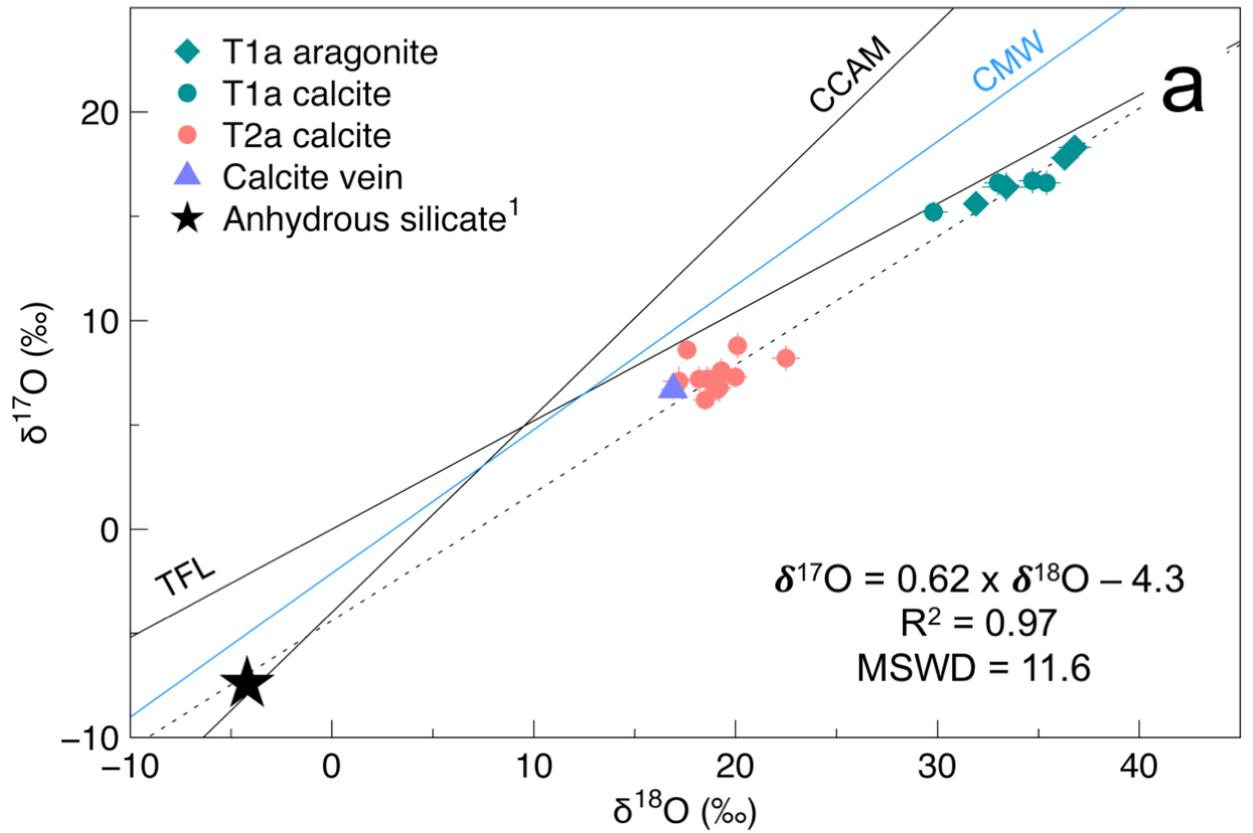
Figure 3



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Figure 4

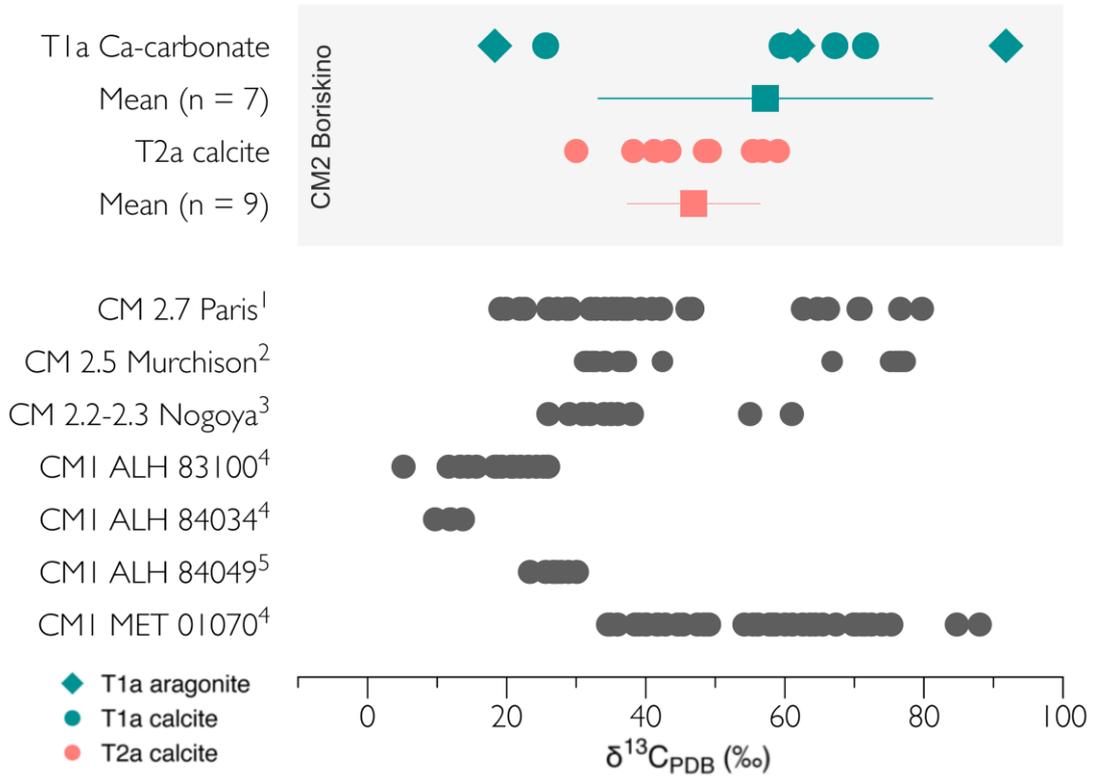


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Figure 5

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Figure 6

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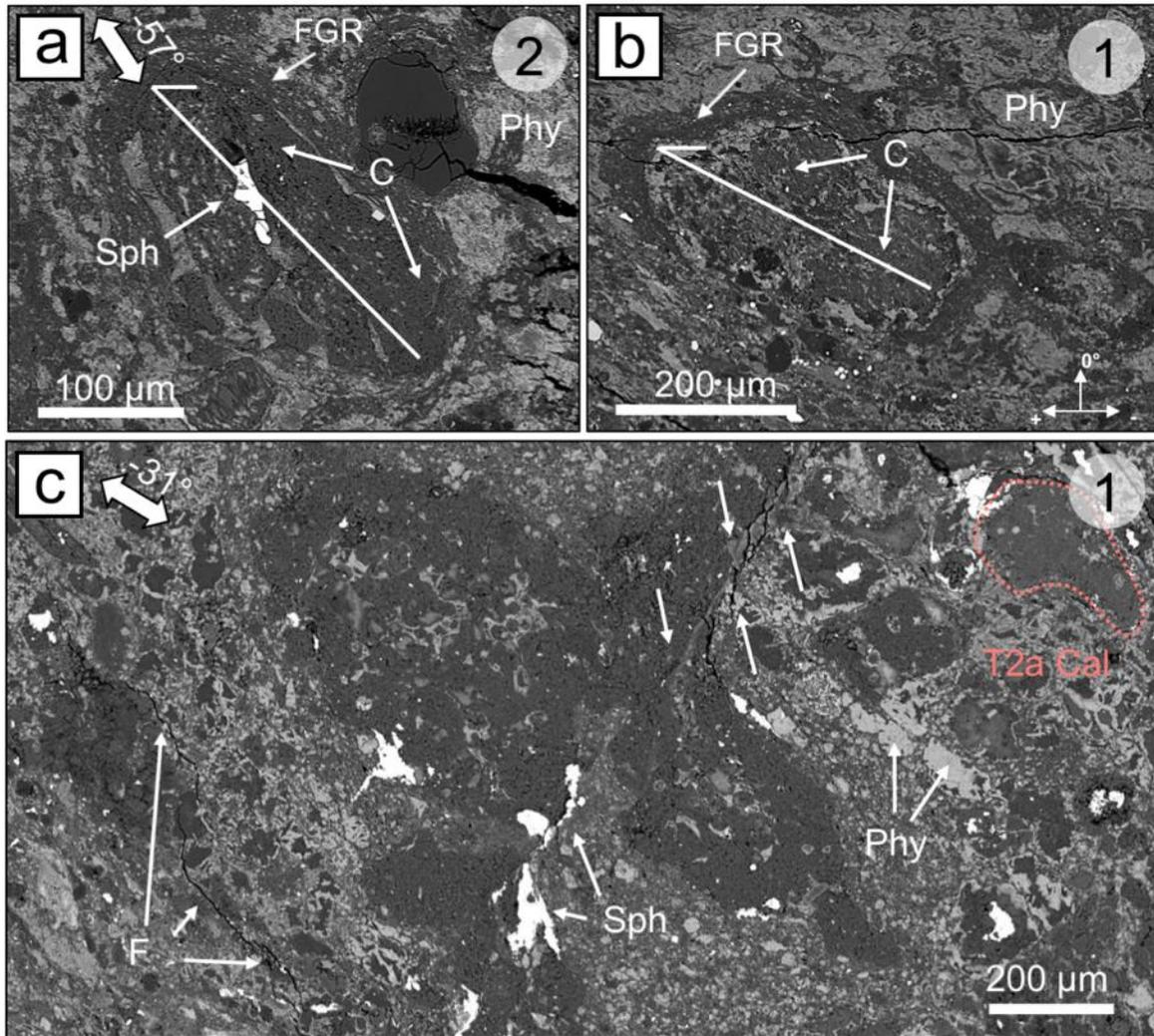
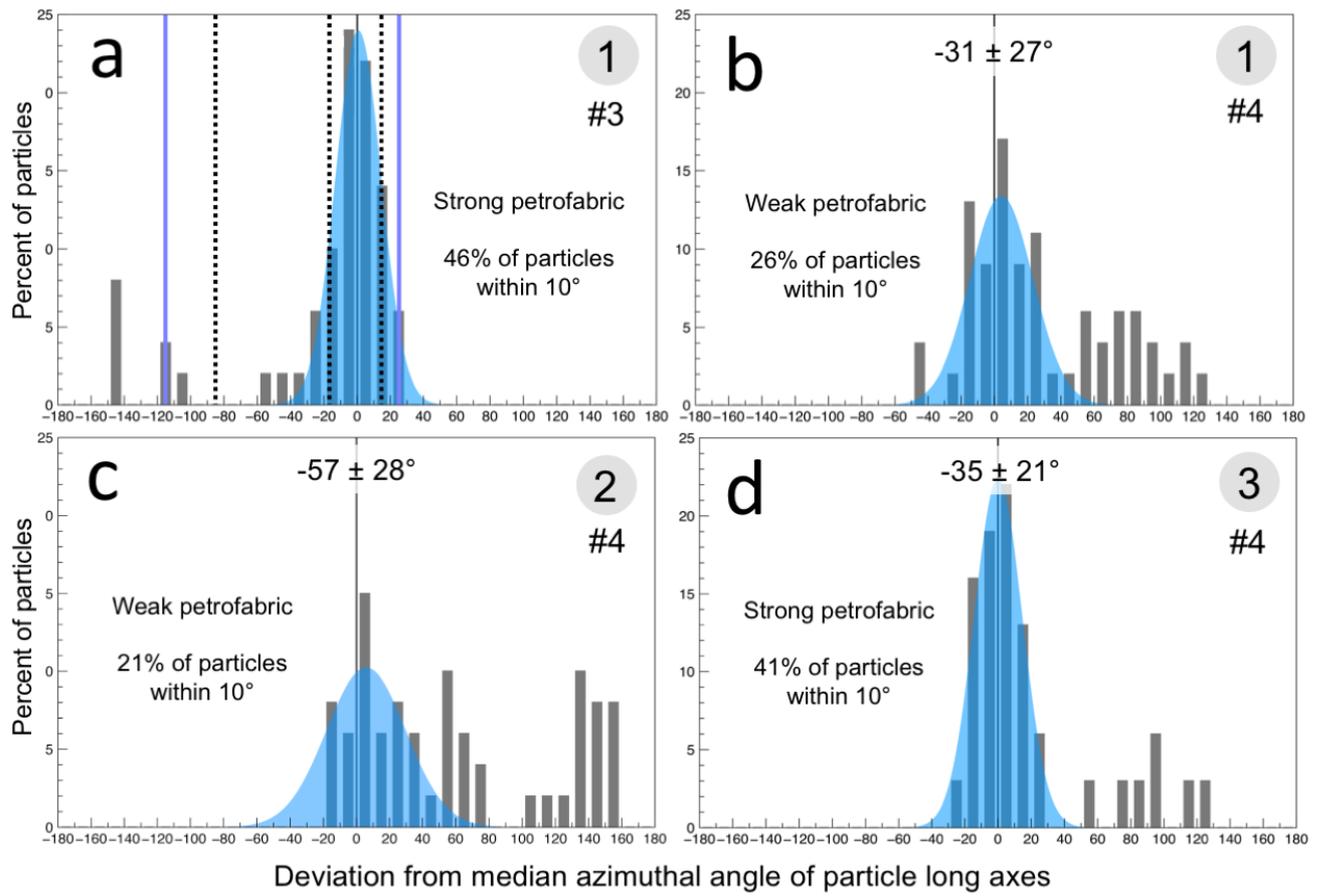


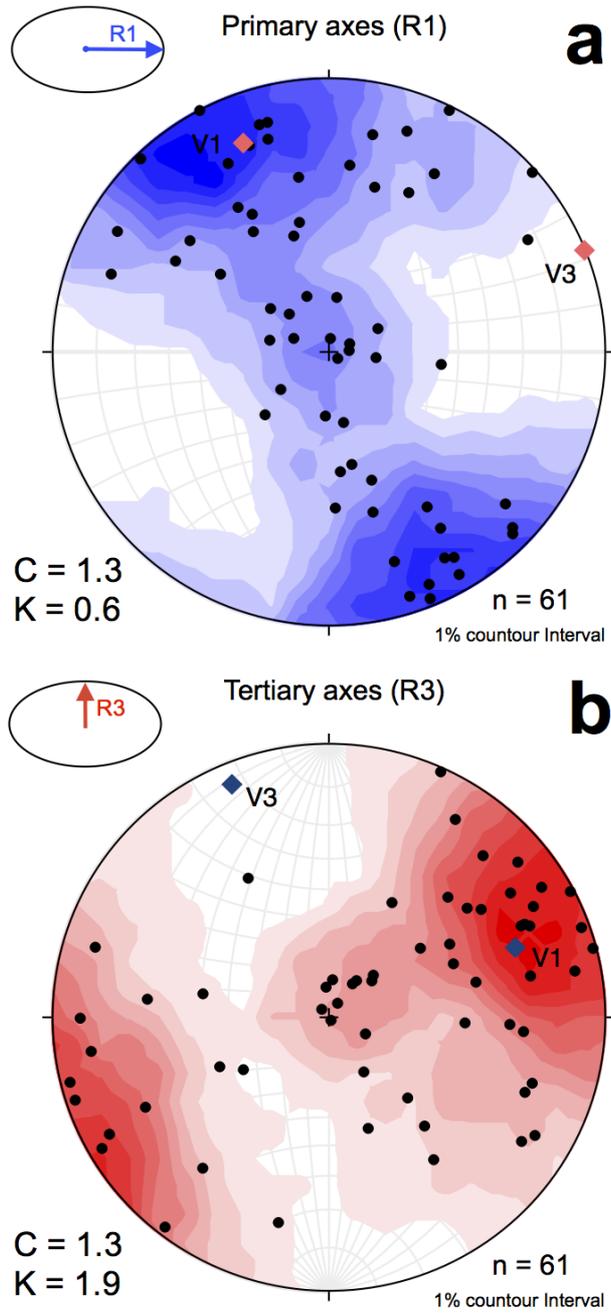
Figure 7

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**Figure 8**

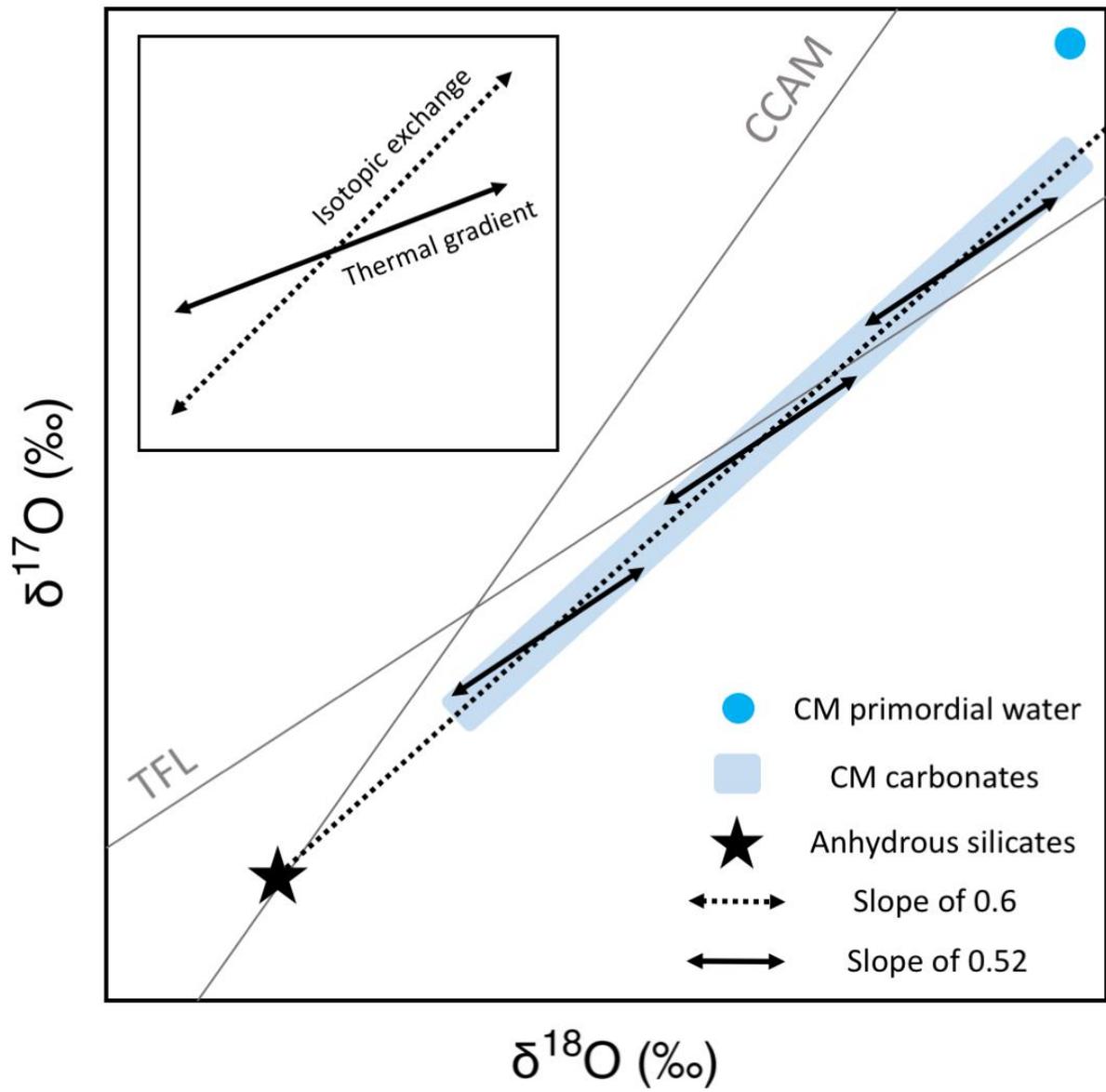
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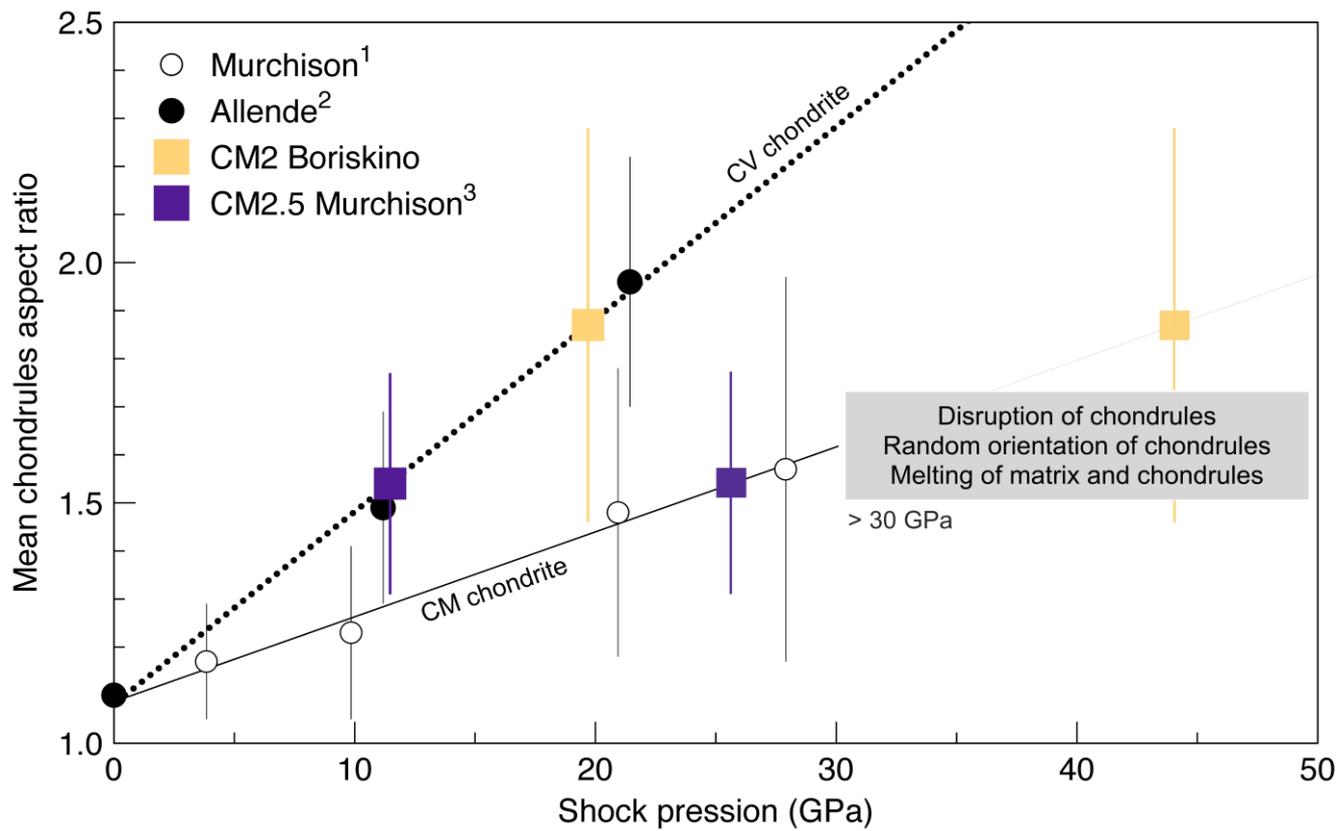
Figure 9



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Figure 10

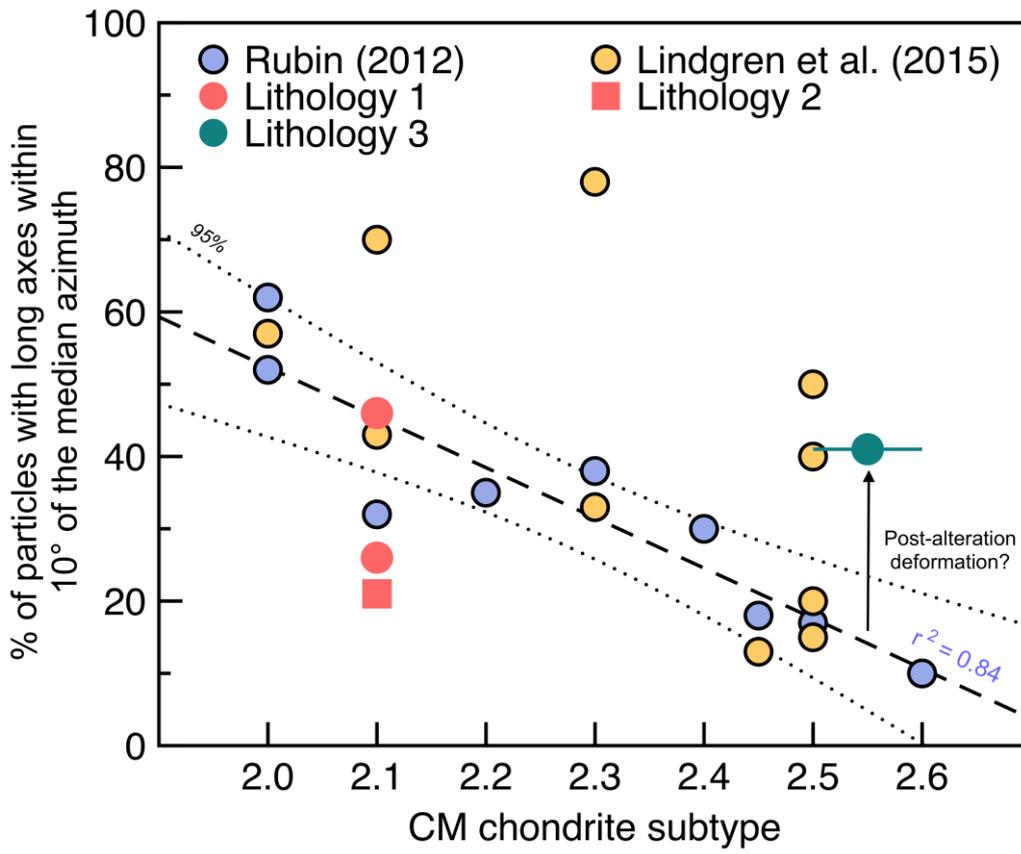


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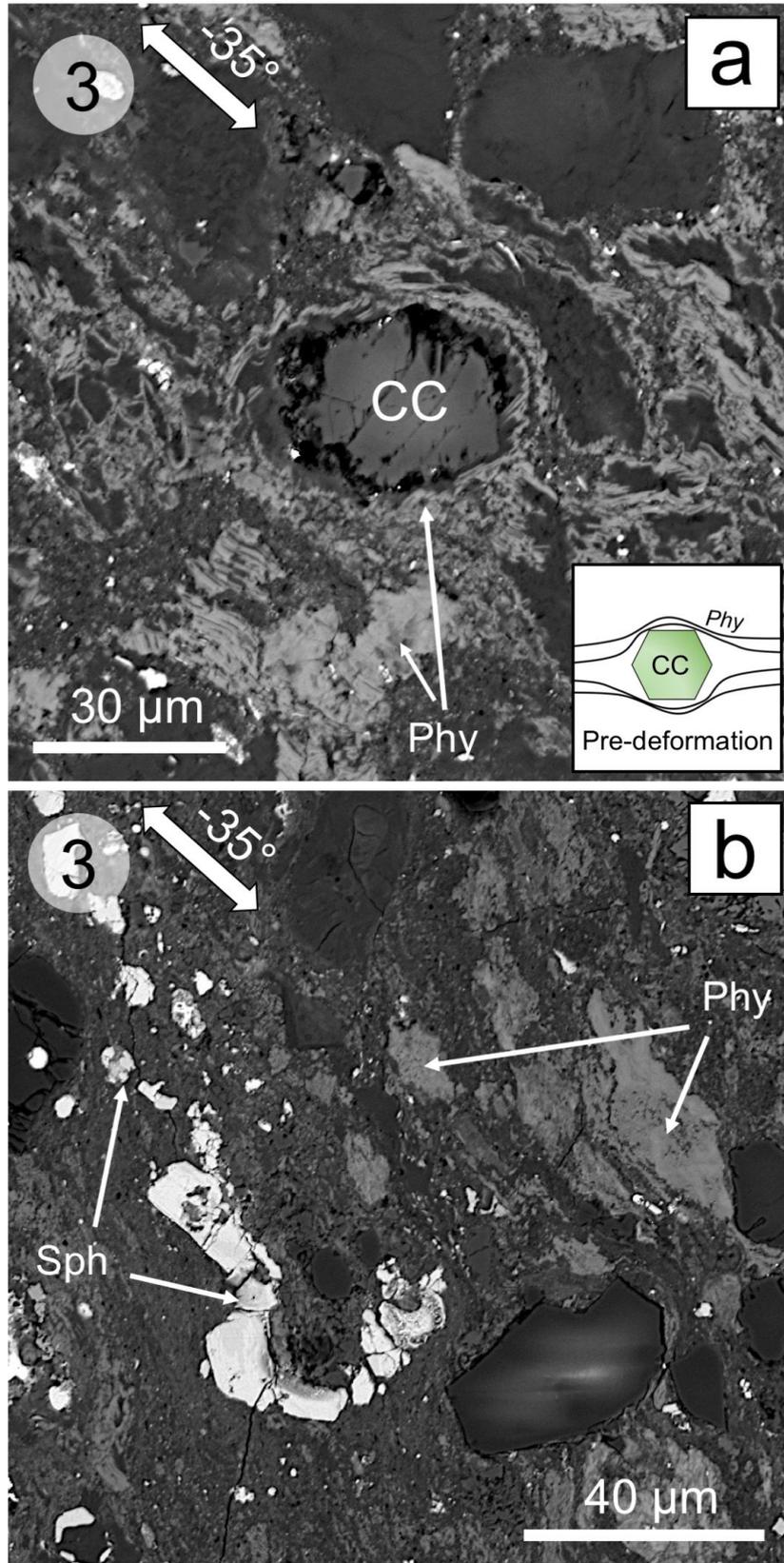
**Figure 11**



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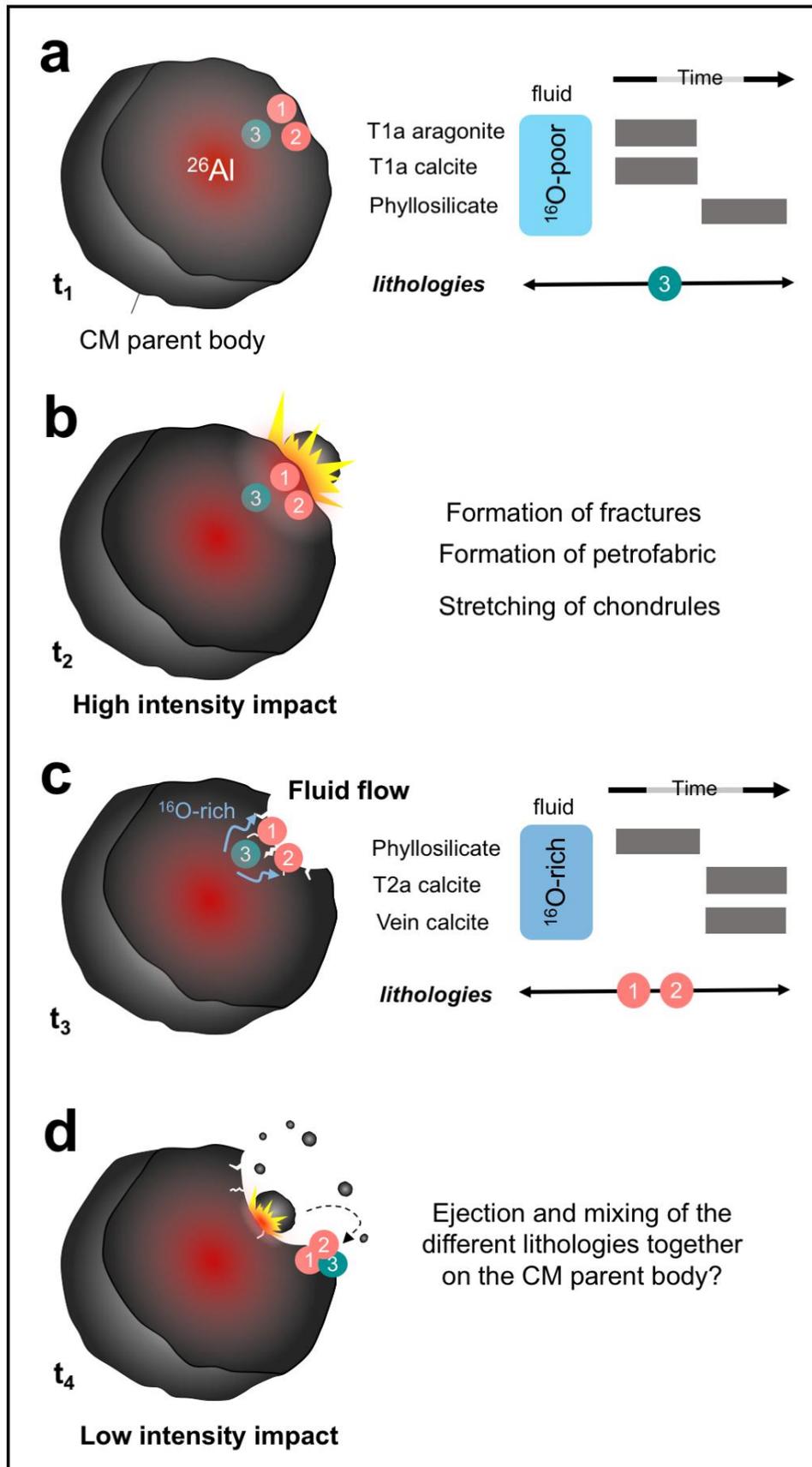
Figure 12



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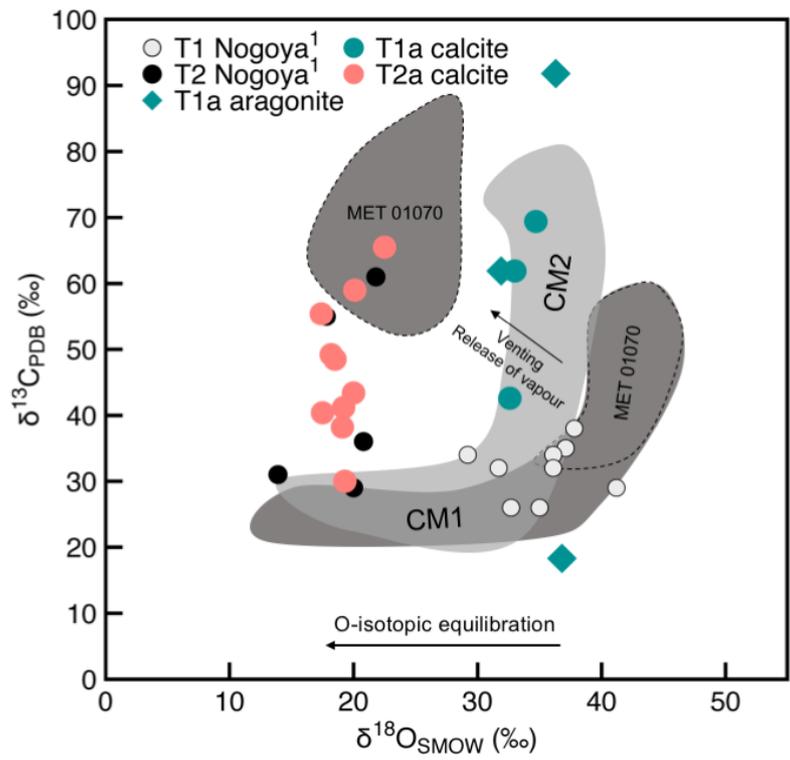
Figure 13



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Figure 14

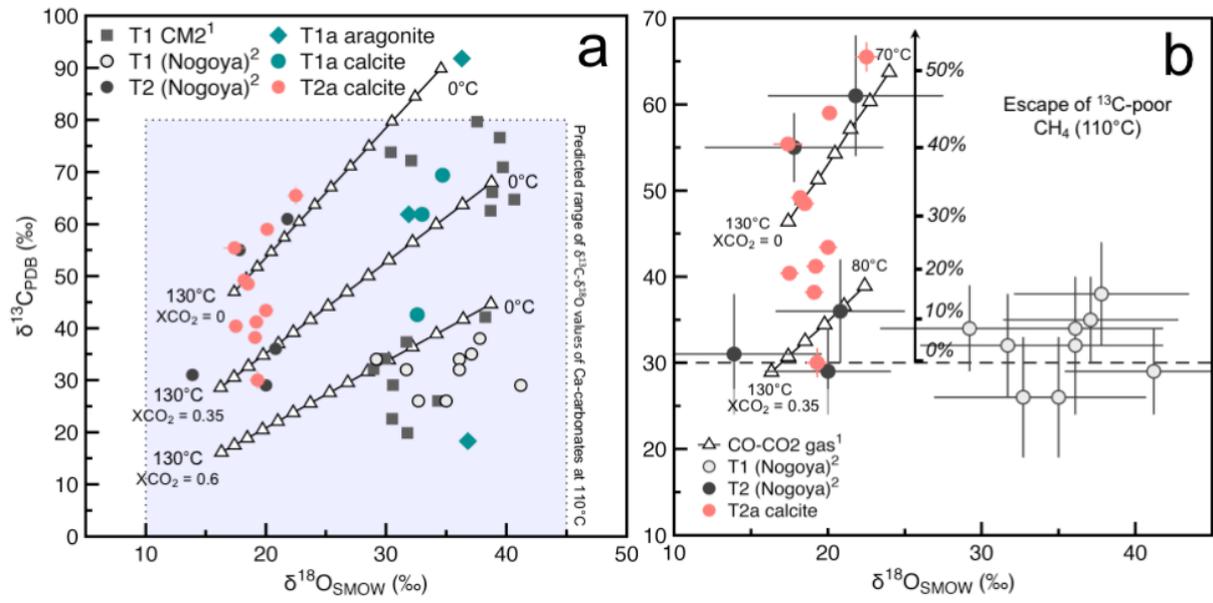


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**Figure 15**



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Figure 16

<b>Lithologies</b>	<b>L1 (n = 510)</b>	<b>L2 (n = 501)</b>	<b>L3 (n = 251)</b>
Surface counting (mm <sup>2</sup> )	6	4	2
Matrix (vol. %)	60.15	62.3	68
Chondrule (vol. %)	30	34.7	23.9
Sulphide (vol. %)	8.6	1.6	5.2
Ca-carbonate (vol. %)	1.2	1.4	2.4
Metallic Fe-Ni (vol. %)	~0.05	Not detected	~0.5
Chondrule mesostasis	Phyllosilicate	Phyllosilicate	Phyllosilicate
Mafic silicate into chondrule	Partially altered	Partially altered	Unaltered
Type of Ca-carbonate	T2a and veins	T2a	T1a

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**Table 1.** Modal abundance of mixture of phases and components (in vol. %) determined by point counting and petrographic observations of section #4 of Boriskino.

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	Mean	StDev	Min	Max
	(n = 61)			
Primary axis length ( $\mu\text{m}$ )	249	109	82	531
Secondary axis length ( $\mu\text{m}$ )	187	79	71	447
Tertiary axis length ( $\mu\text{m}$ )	132	48	61	300
Volume ( $\text{mm}^3$ )	0.03	0.04	0.002	0.2
Aspect ratio <sup>1</sup>	1.87	0.46	1.11	3.34
Elongation <sup>2</sup>	1.34	0.26	1.01	2.07
Flatness <sup>3</sup>	1.4	0.29	1.06	2.46

<sup>1</sup>Primary axis length/tertiary axis length  
<sup>2</sup>Primary axis length/secondary axis length  
<sup>3</sup>Secondary axis length/tertiary axis length

**Table 2.** Bulk best-fit ellipsoid measurements of dark-toned objects (interpreted as type I chondrules) in the CM Boriskino.

#	Type	Polymorph	Lith.	$\delta^{18}\text{O}$ (‰)	2 $\sigma$	$\delta^{17}\text{O}$ (‰)	2 $\sigma$	$\Delta^{17}\text{O}$ (‰)	2 $\sigma$	$\delta^{13}\text{C}$ (‰)	2 $\sigma$
CC10	1a	Aragonite	(3)?	36.8	0.8	18.3	0.5	-0.8	0.8	18.3	0.6
CC4	1a	Aragonite	(3)	36.3	0.7	17.8	0.5	-1.1	0.7	91.8	0.8
CC6	1a	Aragonite	(3)	33.4	1.2	16.4	0.7	-1.0	1.1	-	-
CC5	1a	Aragonite	(3)	31.9	0.6	15.6	0.6	-0.9	0.7	61.9	1.4
CC15-1	1a	Calcite	(3)	35.4	0.7	16.6	0.6	-1.8	0.8	25.6	1.5
CC15-2	1a	Calcite	(3)	29.8	0.7	15.2	0.5	-0.4	0.7	59.6	0.5
CC3	1a	Calcite	(3)	33	0.7	16.6	0.6	-0.6	0.8	61.9	1.2
CC7-1	1a	Calcite	(3)	34.7	0.7	16.7	0.6	-1.3	0.7	71.6	0.9
CC7-2	1a	Calcite	(3)	-	-	-	-	-	-	67.2	0.9
<b>Mean</b>				<b>33.9</b>		<b>16.7</b>		<b>-1.0</b>		<b>55.8</b>	
<i>StDev</i>				<i>2.4</i>		<i>1</i>		<i>0.4</i>		<i>25.7</i>	
CC6-3	2a	Calcite	(1)	22.5	0.7	8.2	0.6	-3.5	0.8	56.8	0.9
CC6-2	2a	Calcite	(1)	-	-	-	-	-	-	68.9	0.6
CC6-1	2a	Calcite	(1)	-	-	-	-	-	-	70.7	1.3
CC11	2a	Calcite	(2)	20.1	0.5	8.8	0.6	-1.7	0.7	59	0.4
CC7	2a	Calcite	(1)	20	0.6	7.3	0.5	-3.1	0.7	43.4	0.5
CC9	2a	Calcite	(1)	19.3	0.7	7.6	0.6	-2.4	0.8	30	1.7
CC12	2a	Calcite	(2)	19.2	0.7	6.8	0.7	-3.2	0.8	-	-
CC5	2a	Calcite	(1)	19.2	0.6	7.4	0.5	-2.6	0.7	41.2	0.6
CC3	2a	Calcite	(1)	19.1	0.6	6.7	0.5	-3.3	0.7	38.2	0.6
CC10	2a	Calcite	(2)	18.6	0.7	7.2	0.6	-2.4	0.8	-	-
CC1	2a	Calcite	(1)	18.5	0.6	6.2	0.5	-3.4	0.7	48.5	0.5
CC14	2a	Calcite	(2)	18.2	0.6	7.2	0.6	-2.2	0.7	49.2	0.6
CC8-2	2a	Calcite	(1)	17.6	0.5	8.6	0.5	-0.6	0.7	-	-
CC8-1	2a	Calcite	(1)	17.2	0.8	7.1	0.7	-1.8	0.9	55.4	0.5
CC16	2a	Calcite	(3)	17.5	0.6	6.6	0.6	-2.5	0.7	40.4	0.5
<b>Mean</b>				<b>19.0</b>		<b>7.4</b>		<b>-2.5</b>		<b>50.1</b>	
<i>StDev</i>				<i>1.4</i>		<i>0.8</i>		<i>0.8</i>		<i>12.4</i>	
CC2	Vein	Calcite	(1)	16.9	0.6	6.7	0.5	-2.0	0.7	-	-

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1286 **Table 3.** Mineralogies, petrographic types, and oxygen and carbon isotopic compositions of  
 1287 Boriskino Ca-carbonates.