



Title	Heating duration of igneous rim formation on a chondrule in the Northwest Africa 3118 CV3(oxA) carbonaceous chondrite inferred from micro-scale migration of the oxygen isotopes
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Citation	Geochemistry, 79(4), UNSP 125524 https://doi.org/10.1016/j.chemer.2019.07.006
Issue Date	2019-12
Doc URL	http://hdl.handle.net/2115/82427
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Type	article (author version)
File Information	Geochemistry_v79_04_1255424_.pdf



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1 **Heating duration of igneous rim formation on a chondrule in the Northwest Africa 3118**
2 **CV3_{oxA} carbonaceous chondrite inferred from micro-scale migration of the oxygen**
3 **isotopes**

4

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14

15 **Abstract**

16 Due to their common occurrences in various types of chondrites, igneous rims formed on
17 pre-existing chondrules throughout chondrule-forming regions of the solar nebula. Although
18 the peak temperatures are thought to reach similar values to those achieved during chondrule
19 formation events, the heating duration in chondrule rim formation has not been well defined.
20 We determined the two-dimensional chemical and oxygen isotopic distributions in an igneous
21 rim of a chondrule within the Northwest Africa 3118 CV3_{oxA} chondrite with sub-micrometer
22 resolution using secondary ion mass spectrometry and scanning electron microscopy. The
23 igneous rim experienced aqueous alteration on the CV parent body. The aqueous alteration
24 resulted in precipitation of the secondary FeO-rich olivine (Fa₄₀₋₄₉) and slightly disturbed the
25 Fe-Mg distribution in the MgO-rich olivine phenocrysts (Fa₁₁₋₂₂) at about a 1 μm scale.
26 However, no oxygen isotopic disturbances were observed at a scale greater than 100 nm. The
27 MgO-rich olivine, a primary phase of igneous rim formation, has $\delta^{17}\text{O} = -6\pm 3\text{‰}$ and $\delta^{18}\text{O} =$
28 $-1\pm 4\text{‰}$, and some grains contain extreme ^{16}O -rich areas ($\delta^{17}\text{O}, \delta^{18}\text{O} = \sim -30\text{‰}$) nearly 10 μm
29 across. We detected oxygen isotopic migration of approximately 1 μm at the boundaries of the
30 extreme ^{16}O -rich areas. Using oxygen self-diffusivity in olivine, the heating time of the
31 igneous rim formation could have continued from several hours to several days at near
32 liquidus temperatures ($\sim 2000\text{ K}$) in the solar nebula suggesting that the rim formed by a
33 similar flash heating event that formed the chondrules.

34

35 1. INTRODUCTION

36 It is believed that chondrules formed during flash heating events in the solar nebula
37 (e.g., Gooding et al., 1980; Grossman and Wasson, 1982; Hewins, 1996). Chondrules are often
38 surrounded by rims that formed in the nebula after the host chondrules. Chondrule rims are
39 divided into two types: (1) fine-grained or matrix-like rims that are similar in chemical
40 composition and grain size to the host chondrite matrix (e.g., Ashworth, 1977; Allen et al.,
41 1980; King and King, 1981; Scott et al., 1984), and (2) coarse-grained or igneous rims that
42 show evidence of a high degree of melting (e.g., Rubin, 1984; Rubin and Wasson, 1987; Krot
43 and Wasson, 1995). Igneous rims surround ~50%, ~10% and <1% of chondrules in the CV3,
44 H-L-LL3, and CO3 chondrites, respectively (Rubin, 1984). The ubiquitous occurrence of
45 igneous rims across the chondrite groups suggests that igneous rim formation commonly
46 occurred in the solar nebula and was related to chondrule formation. However, the rim
47 formation process is not as well understood compared with chondrule formation. For example,
48 even the heating duration has not yet been evaluated.

49 Olivine phenocrysts within igneous rims sometimes contain very ¹⁶O-rich areas in
50 their interiors (e.g., Takeda et al., 2002; Nagashima et al., 2003, 2011, 2013, 2015). The
51 ¹⁶O-rich composition is clearly distinct from that of most olivine in the igneous rims and the
52 phenocrysts in the host chondrules, and is similar to that of amoeboid olivine aggregates. Such
53 ¹⁶O-rich olivine cannot form during chondrule formation because the oxygen isotopic
54 compositions of the minerals crystallized from chondrule melts are typically close to those of
55 the rocky planets (Yurimoto et al., 2008; Tenner et al., 2018). Therefore, these ¹⁶O-rich
56 olivines are the igneous rim feedstocks and heating process survivors (i.e. relict grain) from
57 the igneous rim formation, and the rim formation process could be traced using these olivines.

58 To constrain the rim formation process, we studied the two-dimensional
59 micro-distribution of the chemical compositions and oxygen isotopes in an igneous rim from
60 the Northwest Africa (NWA) 3118 CV_{oxA} chondrite intercorrelated with petrography.

61

62 2. ANALYTICAL TECHNIQUES

63 2.1. Petrography and X-ray analysis

64 A polished thin section sample of the NWA 3118 CV_{oxA} chondrite containing an
65 igneous rimmed porphyritic olivine chondrule was used in this study. The section was coated
66 with a thin film (~20 nm) of carbon for the petrographic observations, quantitative elemental
67 analysis, and X-ray elemental mapping using a field emission scanning electron microscope
68 (FE-SEM; JEOL JSM-7000F) equipped with an energy dispersive spectrometer (EDS;
69 Oxford X-Max 150) at the Hokkaido University. In addition, a crystallographic orientation
70 map was obtained using an electron backscattered diffraction (EBSD; Oxford HKL)

71 installed on the FE-SEM. A 15 keV electron beam probe was applied with currents of 0.3 nA
72 and 5 nA for the quantitative chemical analysis, and X-ray elemental mapping and EBSD
73 mapping, respectively. The step sizes and the dwell times for the mapping were typically set
74 at 0.5 μm and 10 μs , respectively.

75

76 **2.2. SIMS point analysis for O isotopic composition**

77 The section was coated with a Au thin film (70 nm) before the secondary ion
78 mass spectrometry (SIMS) analysis. The point analyses of oxygen isotopes were made using
79 a SIMS (Cameca ims-1280HR) at Hokkaido University. We applied a focused Cs^+ ion
80 primary beam (3–5 μm in diameter) with a total kinetic energy of 20 keV. A normal incident
81 electron gun was used for charge compensation on the primary beam irradiated area. The
82 secondary ions of $^{16}\text{O}^-$ sputtered by the primary beam were accelerated to 10 keV and
83 measured using a Faraday cup (FC), and the $^{17}\text{O}^-$ and $^{18}\text{O}^-$ were measured using an electron
84 multiplier (EM) in the peak jumping mode of the sector magnet. The mass resolution power
85 (MRP) of $M/\Delta M$ was set at ~ 5500 to ensure that the contribution of $^{16}\text{OH}^-$ to $^{17}\text{O}^-$ was
86 negligible. The $^{16}\text{O}^-$ signals were obtained by subtracting background counts of the FC from
87 the total FC counts of $^{16}\text{O}^-$. The EM was operated in the pulse counting mode with a dead
88 time of 48 ns. A typical count rate for $^{16}\text{O}^-$ was set as $\sim 8 \times 10^7$ cps. Each measurement was
89 conducted for 30 cycles of a counting sequence with $^{16}\text{O}^-$ for 1 s, $^{17}\text{O}^-$ for 2 s, and $^{18}\text{O}^-$ for 1
90 s with a waiting time of 2 s for every mass to stabilize the sector magnetic field. The
91 instrumental mass fractionation was corrected using an olivine crystal from San Carlos,
92 Arizona, USA (Fa_8 , $\delta^{17}\text{O}_{\text{SMOW}} = 2.7\text{‰}$, $\delta^{18}\text{O}_{\text{SMOW}} = 5.2\text{‰}$; Yurimoto et al., 2011).

93

94 **2.3. Isotopography for O isotopic composition**

95 The quantitative oxygen isotope distribution (isotopography) on the thin section
96 was performed using an isotope microscope system (Cameca ims-1270 + SCAPS) at the
97 Hokkaido University (Yurimoto et al., 2003). A static Cs^+ primary beam of 20 keV was
98 homogeneously irradiated on the sample area with an oval shape approximately 70 μm in
99 diameter with a beam current of $\sim 0.5\text{--}2$ nA. A normal incident electron gun was utilized to
100 compensate positive charging of the sputtered region due to the primary beam irradiation. An
101 accelerating voltage of 10 kV was applied and secondary ions of 10 keV ± 75 eV were
102 measured by a stacked CMOS-type active pixel sensor (SCAPS) using the energy band-pass
103 slit in the ims-1270. A contrast aperture of 150 μm in diameter was used. The exit slit was
104 narrowed enough to eliminate the contribution of interference ions to the isotope images. The
105 isotope image was projected on the SCAPS detector that has 600×576 pixels with a stacked
106 electrode for each pixel and directly detects the secondary ions (Takayanagi et al., 2003). Each

107 pixel can store the signal of up to 50,000 ions and the pixel size corresponds to 0.28 μm of the
108 isotope image. The stored charge in the capacitor of a pixel can be read out individually
109 through a readout transistor without destruction of the charge and an operating reset transistor
110 (a non-destructive readout mode). This function realizes the SCAPS detector as an
111 integral-type ion detector. The charge signal was acquired with the PXI system controlled by
112 LabVIEW software (Sakamoto and Yurimoto, 2006). The readout time for one image frame
113 was 5.25 s with a noise level of 55 μV corresponding to 1.8 ions in the non-destructive readout
114 mode (Yamamoto et al., 2010). The typical mass sequence for acquiring the secondary ion
115 images for one cycle was ‘reset’, $^{27}\text{Al}^{16}\text{O}^-$, $^{56}\text{Fe}^{16}\text{O}^-$, $^{24}\text{Mg}^{16}\text{O}^-$, $^{28}\text{Si}^-$, ‘reset’, $^{16}\text{O}^-$, ‘reset’, $^{18}\text{O}^-$
116 and $^{16}\text{O}^-$. The ‘reset’ images were taken during the resetting operation of the pixel capacitors
117 (a destructive readout mode) and used as the reset frames for the fixed pattern noise correction
118 (Matsumoto et al., 1993; Kunihiro et al., 2001; Nagashima et al., 2001; Takayanagi et al.,
119 2003). The typical measurement times for the secondary ion species during one cycle were 420
120 s for reset, 25 s for $^{27}\text{Al}^{16}\text{O}^-$, 50 s for $^{56}\text{Fe}^{16}\text{O}^-$, 50 s for $^{24}\text{Mg}^{16}\text{O}^-$, 10 s for $^{28}\text{Si}^-$, 420 s for reset,
121 5 s for $^{16}\text{O}^-$, 420 s for reset, 250 s for $^{18}\text{O}^-$, and 5 s for $^{16}\text{O}^-$. Each isotopograph was calibrated
122 in the non-destructive readout mode. The last part of the analytical cycle (‘reset’, $^{18}\text{O}^-$ and
123 $^{16}\text{O}^-$) was repeated several times in order to integrate the secondary ion signals to improve the
124 precisions of isotopograph. The acquired $^{16}\text{O}^-$ and $^{18}\text{O}^-$ isotopographs were averaged for each
125 isotope after an image shift correction (Sakamoto et al., 2007; Park et al., 2012). The amount
126 of image shift was 1 to 3 pixels corresponding to 0.28 to 0.84 μm on the sample surface. The
127 $\delta^{18}\text{O}$ isotopographs were obtained by calculating secondary ion ratios of the $^{18}\text{O}/^{16}\text{O}$ and
128 normalized to the SMOW scale using $\delta^{18}\text{O}$ values of minerals obtained by point analyses. The
129 typical precision ($2S_{\square}$) of the $\delta^{18}\text{O}$ isotopographs is 30‰ per pixel corresponding to 0.28 μm
130 \times 0.28 μm on the isotope image. An image processing method of moving average with 3×3
131 pixels was applied to the isotope ratio images to reduce the statistical fluctuation of the
132 secondary ions. Therefore, the precision of the isotope ratio is improved to $\sim 10\%$ per pixel for
133 the smoothed image. The $\delta^{18}\text{O}$ values of the olivine grains in the isotopography were
134 calculated from an 11×11 pixels area that corresponds to the beam spot size (approximately 3
135 $\times 3 \mu\text{m}^2$) for the spot analysis. The spatial resolution of the isotopography is evaluated by the
136 boundaries between the extremely ^{16}O -rich olivine and the FeO-rich olivine that is defined by
137 the width between 16% and 84% of the oxygen isotopic composition difference across the
138 boundary (Fig. 1). The spatial resolution of the $\delta^{18}\text{O}$ isotopograph is calculated to be 0.8 μm .

139

140 3. RESULTS

141 3.1. Petrography and mineralogy

142 3.1.1. Chondrule

143 The Northwest Africa (NWA) 3118 is the oxidized Allende-like CV_{3oxA}
144 carbonaceous chondrite (Russell et al., 2005). The chondrule studied is completely surrounded
145 by a coarse-grained igneous rim (Fig. 2a). The chondrule has a rounded shape of 1.4 mm
146 diameter with a type I (Mg# (= molar MgO/(MgO + FeO) %) \geq 90)) porphyritic texture. The
147 chondrule is mainly composed of olivine phenocrysts (Fa₂₋₅) and mesostasis in addition to
148 opaque nodules. The major elemental compositions of these phases are listed in Tables S1–S3
149 (Figs. S1–S3).

150 The olivine phenocrysts are euhedral and have a core-mantle structure. The grain
151 shapes and sizes are rounded and smaller (20–150 μ m) in the core, and an elliptic and larger
152 (200–300 μ m) in the mantle. The mesostasis of the chondrule is represented by dendritic
153 plagioclase and augite (Fig. S4). The plagioclase is often replaced by nepheline, a secondary
154 mineral formed during parent-body aqueous alteration and metamorphism. The secondary
155 nepheline is abundant on the periphery (Figs. 2b, c). Opaque phases occur as rounded to
156 sub-rounded nodules consisting of Fe-Ni metal, Fe sulfide, and metal-sulfide aggregates (Fig.
157 S5). The Fe-Ni metal and Fe sulfide are often replaced by magnetite.

158

159 3.1.2. Igneous rim

160 A coarse-grained igneous rim with thickness up to 400 μ m occurs continuously
161 around the chondrule (Fig. 2a). Mechanical cracks occur near the boundary between the
162 chondrule and the coarse-grained rim. The cracks are secondary in origin, formed after the
163 formation of the rim because the cracked olivine-grains show the same crystallographic
164 orientations across the cracks, indicating that they were originally single crystal grains (Figs.
165 S6a, b).

166 The coarse-grained rim is dominated by olivine. The olivine grains are enriched
167 in FeO compared with those within the host chondrule. The FeO contents show a bimodal
168 distribution with peaks at MgO-rich (Fa_{~11-22}) and FeO-rich (Fa_{~40-49}) compositions (Figs. S7,
169 S8 and Table S4). Enstatite, pigeonite, feldspathic glasses, augite, nepheline, and Fe \pm Ni
170 sulfide (troilite, pyrrhotite or pentlandite) are identified in the rim (Fig. S6a). The major
171 elemental compositions of these minerals are summarized in Tables S5–S8 (Figs. S9 and
172 S10).

173 Enstatite and pigeonite (Fs₁₋₄Wo₁₋₂ and Fs₁₋₄Wo₄₋₁₀) are subhedral and often
174 partially replaced by FeO-rich olivine (Fig. S11a). Augite (Fs₁₋₅Wo₂₇₋₄₂) overgrows on the
175 enstatite and pigeonite, (Fig. S11a) and occurs in the mesostasis of the rim. The rim contains
176 abundant spherical or irregular assemblages of Fe- and Fe-Ni sulfides (troilite, pyrrhotite,
177 and pentlandite), whereas Fe-Ni metal and magnetite are rare (Fig. S11b). They are also
178 replaced by FeO-rich olivine (Fig. S11c). Feldspathic glass and nepheline occur in the

179 mesostasis (Fig. S11a). These petrographic observations are similar to those reported in the
180 Allende meteorite by Brearley and Krot (2013), who showed that aqueous alteration took
181 place on the parent body.

182 The MgO-rich olivine grains are euhedral to subhedral. Their grain size gradually
183 changes radially from 100–150 μm near the chondrule-rim boundary to 10–50 μm at the
184 rim-matrix boundary (Figs. S6a, b). Some MgO-rich olivine grains are overgrown by
185 FeO-rich olivine with the same crystallographic orientation (Figs. 3a, c). The cracks of the
186 MgO-rich olivine grains are healed by FeO-rich olivine as veins (Figs. 3b, d).

187 The FeO-rich olivine appears anhedral and fills the spaces among the other
188 minerals in the rim (Fig. S11a). The compositions of FeO-rich olivine gradually change from
189 Fa_{42} at the chondrule-rim boundary to Fa_{48} at the rim-matrix boundary (Fig. S12). The
190 igneous rim is covered with FeO-rich olivine at the matrix-rim boundary (Fig. S13a). The
191 FeO-rich olivines show both smooth and porous surfaces on the polished thin section (Fig.
192 S13b).

193 The MgO-rich olivine grains in the igneous rim have Fe-Mg chemical zoning near
194 ~~many~~ crystal surfaces. Where there are grains with FeO-rich olivine overgrowths, the zoning
195 gradients occur perpendicular to the surfaces of the MgO-rich olivine (Figs. 4a–f). No
196 Mg-Fe zoning is observed ~~within the mantling~~ on the sides of the FeO-rich olivines.
197 Gradients in Mg-Fe composition also occur perpendicular to the veins in the MgO-rich
198 olivine grains (Figs. 4g–h).

199

200 3.1.3 Matrix

201 The matrix surrounding the igneous rim is predominantly composed of olivine with
202 Fa_{46-51} . The olivine composition is similar to the FeO-rich olivine in the igneous rim (Table.
203 S9). The olivine grains show a bimodal size distribution; the coarser-grains (5–10 μm) are
204 either needle-shaped or irregular; the finer-grains (< 1 μm) are needle-shaped (Fig. S14a).
205 Neither fayalite (> Fa_{90}) nor phyllosilicates are observed. Sulfides and Ca, Fe-rich silicates are
206 scattered throughout the matrix (Fig. S14b).

207

208 3.2. Oxygen-isotope distribution of olivine

209 3.2.1 Chondrule and igneous rim

210 On the oxygen three-isotope diagram ($\delta^{17}\text{O}$ vs. $\delta^{18}\text{O}$), the compositions of olivine
211 phenocrysts in the host chondrule measured by the point analyses plot on a \sim slope-1 line
212 (CCAM) ($\delta^{17}\text{O} = -7.6 \pm 2.1\%$, $\delta^{18}\text{O} = -3.8 \pm 2.1\%$, 2σ ; Fig. 5 and Table S10). The oxygen
213 isotopic compositions of olivines in the igneous rim that are also obtained by point analyses
214 are distributed heterogeneously along the \sim slope-1 line (Fig. 5, Table S11 and Fig. S15). The

215 oxygen isotopic compositions of the FeO-rich olivines are ^{16}O -poor ($\delta^{17}\text{O} = 2.0 \pm 1.8\%$, $\delta^{18}\text{O}$
216 $= 8.1 \pm 3.6\%$ on average). In contrast, the MgO-rich olivine grains are relatively ^{16}O -rich
217 ($\delta^{17}\text{O} = -5.6 \pm 3.2\%$, $\delta^{18}\text{O} = -0.7 \pm 3.6\%$ on average). The standard deviations for $\delta^{17}\text{O}$ and
218 $\delta^{18}\text{O}$ ($2\sigma = 1.8$ and 3.6% , respectively) suggest that the O isotope variation is mass-dependent
219 for the FeO-rich olivine, whereas the variation for MgO-rich olivine appears to be mass
220 independent with a slope-1 variation ($2\sigma = 3.2$ and 3.6% for $\delta^{17}\text{O}$ and $\delta^{18}\text{O}$). One distinct
221 composition ($\delta^{17}\text{O} = -18\%$ and $\delta^{18}\text{O} = -15\%$) was observed for a MgO-rich olivine grain in
222 the igneous rim. As discussed in the next section, this distinct composition could be a result of
223 a primary beam overlap across areas with two different oxygen isotopic compositions in the
224 MgO-rich olivine, i.e., $\delta^{17}\text{O} \sim -6\%$, $\delta^{18}\text{O} \sim -1\%$ and $\delta^{17,18}\text{O} \sim -30\%$.

225

226 **3.2.2. Heterogeneous oxygen isotope distribution within the MgO-rich olivine crystals**

227 Using isotopography, we discovered that MgO-rich olivine grains in the igneous rim
228 sometimes contain extreme ^{16}O -rich areas in their interiors. The MgO-rich olivine grains
229 containing the extreme ^{16}O -rich areas tend to be present near the boundary between the
230 igneous rim and adjacent matrix, and were not found near the rim-chondrule boundary (Fig. 6).

231 The MgO-rich olivine (about $27\ \mu\text{m}$ in size) in the center of Figure 7a has variable
232 intra-crystalline oxygen-isotope compositions. The oxygen isotopic compositions of MgO-rich
233 olivine in the area that appears dark blue in Figure 7a and the surrounding white to blue area
234 are $\delta^{18}\text{O} = -30.6 \pm 3.7\%$ and $-1.9 \pm 2.9\%$, respectively. The extreme ^{16}O -rich area is not
235 chemically distinct from the surrounding MgO-rich olivine. The oxygen isotopic composition
236 of the MgO-rich olivine surrounding the ^{16}O -rich ~~patch~~ area is consistent with most of the
237 measured point analyses (i.e., except for the one point analysis with $\delta^{18}\text{O} = -14.5\%$; see Fig. 5
238 and Table S11). The $\delta^{18}\text{O}$ for the FeO-rich olivine is $10.4 \pm 3.0\%$ consistent with those from
239 point analyses. The ^{16}O -rich analysis shown in Figure 5 ($\delta^{17}\text{O} = -18\%$, $\delta^{18}\text{O} = -15\%$; analysis
240 number 15 in Table S11) was collected from the boundary between the extreme ^{16}O -rich ~~patch~~
241 area and the surrounding olivine (Fig. S15).

242 The olivine crystal shown in Figure 7b has two different chemical compositions; an
243 upper MgO-rich area and a lower FeO-rich area. The MgO-rich area of the olivine includes an
244 extreme ^{16}O -rich area ($\delta^{18}\text{O} = -16.1 \pm 3.3\%$). The upper side of the extreme ^{16}O -rich area
245 directly contacts the MgO-rich olivine, whereas the lower side directly contacts the overgrown
246 FeO-rich olivine. The MgO-rich area of the olivine has Fe-Mg chemical zoning at the
247 boundary with the FeO-rich overgrowth. The $\delta^{18}\text{O}$ value in the upper area of the MgO-rich
248 olivine is $-2.3 \pm 3.4\%$. The two different oxygen isotope areas in the MgO-rich olivine are
249 indistinguishable in the chemical compositions. The $\delta^{18}\text{O}$ value of the FeO-rich overgrown
250 olivine is $11.1 \pm 3.3\%$.

251 The MgO-rich olivine (about 27 μm in size; $\delta^{18}\text{O} = -2.5 \pm 2.9\%$) in the center of
252 Figure 7c includes two extreme ^{16}O -rich areas. The larger area has $\delta^{18}\text{O}$ value of $-33.1 \pm 3.2\%$.
253 These two areas could be connected in three dimensions (above or below the thin section
254 surface) because they have the same crystallographic orientation. A part of the upper boundary
255 of the larger area is directly adjacent to the overgrown FeO-rich olivine (Fig. 1) with $\delta^{18}\text{O}$
256 value of $6.5 \pm 2.6\%$.

257 The MgO-rich olivine in the center of Figure 7d is about 26 μm in size and partially
258 overgrown by FeO-rich olivine. This olivine includes two extreme ^{16}O -rich areas with $\delta^{18}\text{O}$
259 values of $-30.2 \pm 3.4\%$ and $-28.3 \pm 4.9\%$, respectively. The oxygen isotopic compositions are
260 the same within the measurement error and their crystallographic orientations are also the
261 same. Therefore, these two areas could be one region connected above or below the thin
262 section surface. The surrounding MgO-rich olivine area has $\delta^{18}\text{O}$ value of $-1.4 \pm 3.1\%$ that is
263 typical for the MgO-rich olivine in this igneous rim. The FeO-rich olivine in Figure 7d also
264 has the typical oxygen isotopic composition of $\delta^{18}\text{O} = 5.9 \pm 3.2\%$. The olivine grains adjacent
265 to the upper right include two extreme ^{16}O -rich areas with $\delta^{18}\text{O}$ values of $-15.4 \pm 5.1\%$
266 (smaller area) and $-24.0 \pm 5.0\%$ (larger area). Although the olivine grains seem to be FeO-rich,
267 the extreme ^{16}O -rich areas are slightly enriched in MgO composition according to the BSE
268 image suggesting that the extreme ^{16}O -rich areas were originally enriched in MgO. Parts of the
269 FeO-rich olivine outside of the extreme ^{16}O -rich areas have the typical oxygen isotopic
270 composition of FeO-rich olivine in the igneous rim.

271 The MgO-rich olivine grain ($\delta^{18}\text{O} = -4.3 \pm 2.7\%$) on the left side of Figure 7e is
272 about 20 μm in size and includes an extreme ^{16}O -rich area with $\delta^{18}\text{O}$ values of $-19.8 \pm 3.4\%$.
273 The MgO-rich olivine is overgrown by FeO-rich olivine with $\delta^{18}\text{O}$ value of $6.2 \pm 4.0\%$. The
274 MgO-rich olivine grain (about 30 μm) on the right side ($\delta^{18}\text{O} = -1.7 \pm 2.7\%$) includes an
275 extreme ^{16}O -rich area with $\delta^{18}\text{O}$ value of $-16.9 \pm 3.0\%$.

276 The $\delta^{18}\text{O}$ image in Figure 7f was normalized to the SMOW scale using the average
277 $\delta^{18}\text{O}$ value of the MgO-rich olivine in the rim obtained by spot analyses ($\delta^{18}\text{O} = -0.7 \pm 3.6\%$,
278 Table S10) because spot analyses of oxygen isotopes were not applied in this image field. The
279 MgO-rich olivine grain (~ 35 μm across) is separated by a crack and includes an extreme $\delta^{18}\text{O}$
280 area across the crack ($\delta^{18}\text{O} = -31.5 \pm 4.7\%$). The FeO-rich olivine has an oxygen isotopic
281 composition of $\delta^{18}\text{O} = 12.0 \pm 2.0\%$.

282 In summary, most of the oxygen isotopic compositions of the FeO-rich and
283 MgO-rich olivines determined by SCAPS are $\sim 8\%$ and $\sim -2\%$ for $\delta^{18}\text{O}$, respectively, for all
284 the image fields and these values are consistent with those obtained by the SIMS spot analyses.
285 The MgO rich olivines often include extreme ^{16}O -rich areas with sizes typically less than 10
286 μm . The oxygen isotopic compositions of the extreme ^{16}O -rich areas are typically $\delta^{18}\text{O} \sim$

287 -30‰ (and $\delta^{17}\text{O} \sim -30\text{‰}$ inferred from combining the results of the point analysis), but tends
288 toward less ^{16}O -rich values for areas having narrow boundaries ($<3 \mu\text{m}$ across) with adjacent
289 ^{16}O -poor olivine. This tendency of oxygen isotopic variation can be interpreted by the spatial
290 resolution of isotopography of this study (Fig. 1), the oxygen self-diffusion discussed in
291 section 4.2, and supports that the extreme ^{16}O -rich areas in this igneous rim have $\delta^{17,18}\text{O} \sim \delta^{18}\text{O}$
292 values of $\sim -30\text{‰}$.

293

294 4. DISCUSSION

295 4.1. Aqueous alteration on the parent body

296 The occurrences of secondary minerals – FeO-rich olivine (Fa_{40-49}), nepheline,
297 magnetite, and Ca,Fe-rich silicates, and their petrographic textures observed in this study
298 suggest that NWA3118 experienced extensive aqueous/metasomatic alteration on the CV
299 chondrite parent body (e.g., Krot et al., 1995, 1998, 2004; Maruyama et al., 1999; Maruyama
300 and Yurimoto, 2003; Brearley, 2003). The FeO-rich secondary olivine exhibits textures such as
301 (1) overgrowths around individual MgO-rich olivine grains, (2) veins in the MgO-rich olivine
302 grains, (3) replacement of enstatite and pigeonite, and opaque nodules, and (4) lath-shaped
303 grains in the matrix, which are similar to those observed in Allende (e.g. Krot et al., 1995,
304 1998, 2004; Komatsu et al., 2015; Cuvillier et al., 2015) suggesting origin by aqueous
305 alteration origin on the CV parent body.

306 Chemical compositions are different between olivine grains in the igneous rim,
307 Fa_{11-22} , and in the host chondrule, Fa_{2-5} . The characteristics of Fa_{11-22} might be due to
308 metamorphic processes on the parent body (Huss et al., 2006; Tenner et al., 2015). If the
309 Fa_{11-22} are the result of metamorphic processes on the parent body, the MgO-rich olivine in
310 the host chondrule contacted to the Fa_{11-22} olivine also must have similar Fa composition.
311 However, the MgO-rich olivine of the host chondrule has clearly distinct chemical
312 composition, Fa_{2-5} and the chemical compositions of olivine are abruptly changed at the
313 chondrule-rim boundary (Figs. S6 and S8). If the Fa_{11-22} are the result of metamorphic
314 processes on the parent body, Fe-Mg interdiffusion should have occurred in the MgO-rich
315 olivine. We calculated Fe-Mg interdiffusion profiles assuming that Fa composition at the
316 center of olivine grain changed from Fa_2 to Fa_{15} on the parent body. The interdiffusion profiles
317 are inconsistent with compositional zoning in the igneous rim olivine (Fig. 4). Therefore, it is
318 unlikely that the Fa_{11-22} composition of the MgO-rich olivines in the igneous rim was achieved
319 by metamorphic processes on the parent body. The MgO-rich olivines with Fa_{11-22} in the
320 igneous rim are the original composition when the igneous rim formed.

321

322 The Fe-Mg chemical zoning between the MgO-rich olivine and adjacent FeO-rich
323 olivine in the igneous rim is an inter-diffusion profiles during the alteration on the parent body,
324 and can be used as a probe to determine timescales and temperature of the alteration (e.g.,
325 Weinbruch et al., 1994). We defined the Fe-Mg inter-diffusion distance from the interface
326 ($x_{\text{Fe-Mg}} = 0$) between the FeO-rich olivine (FeO content = C_1) and MgO-rich olivine (FeO
327 content in the core = C_2) as the distance $x_{\text{Fe-Mg}}$ with the FeO content of $(C_1 - C_2)/e + C_2$ (Fig. 4,
328 Tables S12). For example, in Figure 4a, the Fe-Mg inter-diffusion distance $x_{\text{Fe-Mg}}$ from the
329 FeO-rich olivine with 37 wt% FeO into the MgO-rich olivine with 21 wt% FeO is 1.3 μm .
330 Similarly, the Fe-Mg inter-diffusion distances shown in Figures 4b to 4f are 1.8, 1.1, 1.3, 1.3,
331 and 2.1 μm , respectively.

332 The diffusion distance from veins of the FeO-rich olivine was also analyzed on both
333 sides across veins in MgO-rich olivine (Figs. 4g, h). The interface between the vein and
334 MgO-rich olivine is defined as $x_{\text{Fe-Mg}} = 0$. Diffusion distances of 1.1 μm from 27 wt% FeO to 15
335 wt% FeO and 1.2 μm from 28 wt% FeO to 15 wt% FeO were obtained on both sides of the vein
336 for the grain shown in Figure 4g. Similarly, diffusion distances of 1.2 and 1.0 μm were
337 obtained across the vein from 31 to 21 wt% FeO and from 30 to 20 wt% FeO for the grain in
338 Figure 4h.

339 The interdiffusion distances observed in the olivine grains range from 1.0 μm to 2.1
340 μm . The longer diffusion distances shown in Figures 4b (1.8 μm) and 4f (2.1 μm) could reflect
341 the anisotropy of Fe-Mg diffusion in olivine that is faster along the c-axis (space group: Pbnm)
342 than along the a- or b-axis (Dohmen and Chakraborty, 2007). The Fe-Mg diffusion distance
343 recorded within the MgO-rich olivine in the igneous rim is an average of 1.3 μm regardless of
344 the grain location in the igneous rim (Fig. S8). Our observations of Fe-Mg diffusion agree with
345 those observed in the matrix olivine grains of the Allende CV_{oxA} chondrite (Cuvillier et al.,
346 2015).

347 The Fe-Mg interdiffusion profiles show that the concentration gradient is present
348 only on the MgO-rich olivine side, not within the overgrown FeO-rich olivine. Cuvillier et al.
349 (2015) proposed that the composition in the FeO-rich olivine is homogeneous because
350 diffusion in FeO-rich olivine is faster than in MgO-rich olivine and the compositional jump
351 derives from resistance that prevents equilibrium at the interface. They also mention that this
352 resistance is not known to occur in olivine grains. Alternatively, the Fe-Mg interdiffusion
353 could have mainly formed in the MgO-rich olivine with dissolved Fe ions in a
354 high-temperature aqueous fluid before ~~or during~~ precipitation of FeO-rich olivine. The
355 FeO-rich olivine may have precipitated interstitially on the MgO-rich olivine and the
356 intergranular spaces between minerals in the igneous rim after cooling. Fe-Mg diffusion

357 profiles in the Fe-rich olivines, thus, were not appeared due to the slow diffusivity at the low
358 temperature.

359 The diffusion distances for the Fe-Mg interdiffusion, $x_{\text{Fe-Mg}}$, and O self-diffusion,
360 x_{oxygen} , are approximated as:

$$361 \quad x_{\text{Fe-Mg}} = \sqrt{2D_{\text{Fe-Mg}} t} \quad (1)$$

$$362 \quad x_{\text{oxygen}} = \sqrt{2D_{\text{oxygen}} t} \quad (2)$$

363 where D is the diffusion coefficient for the corresponding species and t is the time. We ignore
364 the compositional dependence of $D_{\text{Fe-Mg}}$ for simplicity. If these diffusion processes occurred
365 simultaneously in the olivine of the igneous rim, the diffusion distances should be correlated as
366 follows;

$$367 \quad x_{\text{oxygen}} = x_{\text{Fe-Mg}} \sqrt{\frac{D_{\text{oxygen}}}{D_{\text{Fe-Mg}}}} \quad (3)$$

368 The average $x_{\text{Fe-Mg}}$ is 1.3 μm in this study and x_{oxygen} is then given as a function of
369 temperature independently from the diffusion time (duration of aqueous alteration). The
370 Fe-Mg inter-diffusion and oxygen self-diffusion coefficients in olivine have been measured in
371 several previous studies. Nakamura and Schmalzried (1984) determined the Fe-Mg
372 inter-diffusion coefficient ($D_{\text{N\&S}}$) at temperatures between 1323 and 1553 K and a composition
373 of Fo₈₇. Chacraborty (1997) also reported the Fe-Mg inter-diffusion coefficient (D_{C}) at
374 temperatures between 1253 and 1573 K and a composition of Fo₈₆. Their diffusion coefficients
375 were obtained under a controlled oxygen fugacity of $f\text{O}_2 = 10^{-12}$ bars. Dohmen et al. (2007)
376 determined the Fe-Mg interdiffusion coefficients of Fo₈₆ at temperatures from 973–1473 K.
377 The diffusion coefficients were obtained at $f\text{O}_2$ of 10^{-12} bars at temperatures between
378 1173–1473 K ($D_{\text{D_hightemp}}$), while those between 973–1123 K were obtained at $f\text{O}_2$ of
379 10^{-15} - 10^{-17} bars ($D_{\text{D_lowtemp}}$). The forsterite content of the MgO-rich olivine in the igneous rim
380 is adequate to choose these Fe-Mg interdiffusion coefficients in previous studies. However,
381 they need to be extrapolated to lower temperatures because they were determined at
382 temperatures higher than peak metamorphic temperatures inferred for Allende (e.g., Keil,
383 2000; Ito and Messenger, 2010) and were extrapolated down to 500 K in this study. The
384 Fe-Mg inter-diffusion coefficients (m^2/s), $D_{\text{N\&S}}$ (Nakamura and Schmaizried, 1984), D_{C}
385 (Chakraborty et al., 1997), $D_{\text{D_hightemp}}$ (Dohmen et al., 2007), and $D_{\text{D_lowtemp}}$ (Dohmen et al.,
386 2007) were expressed as follows:

$$387 \quad D_{\text{N\&S}} = 8.33 \times 10^{-10} \cdot \exp\left[\frac{-188000}{RT}\right] \quad (4)$$

388
$$D_C = 5.38 \times 10^{-9} \cdot \exp\left[\frac{-226000}{RT}\right] \quad (5)$$

389
$$D_{D_hightemp} = 6.66 \times 10^{-10} \cdot \exp\left[\frac{-201000}{RT}\right] \quad (6)$$

390
$$D_{D_lowtemp} = 1.54 \times 10^{-9} \cdot \exp\left[\frac{-223000}{RT}\right] \quad (7)$$

391 where T is the temperature (K) and R is the gas constant (8.314 J/mol·K).

392 The oxygen fugacity, fO_2 , for the CV chondrites was estimated to be close to the
 393 iron-wustite (IW) oxygen buffer during chondrule formation and close to the
 394 fayalite-magnetite-quartz (FMQ) buffer during aqueous alteration (Righter and Neff, 2007).
 395 We assumed that the diffusion occurred under the FMQ buffer during aqueous alteration that is
 396 expressed as follows (Myers and Eugster, 1983);

397
$$\log fO_2 = -\frac{24441.9}{T} + 8.290 \quad (8)$$

398 The fO_2 corrections for $D_{N\&S}$, D_C , and $D_{D_hightemp}$ to the FMQ were performed assuming that
 399 the Fe-Mg interdiffusion is controlled by the vacancy concentration determined by the Fe^{2+} -
 400 Fe^{3+} equilibrium in olivine (e.g., Chakraborty et al., 1997). The fO_2 corrected expressions for
 401 D_{Fe-Mg} are given as follows:

402
$$D_{N\&S} = \left(\frac{fO_{2_FMQ}}{fO_{2_Ex}}\right)^{\frac{1}{6}} \cdot 8.33 \times 10^{-10} \cdot \exp\left[\frac{-188000}{RT}\right] \quad (9)$$

403
$$D_C = \left(\frac{fO_{2_FMQ}}{fO_{2_Ex}}\right)^{\frac{1}{6}} \cdot 5.38 \times 10^{-9} \cdot \exp\left[\frac{-226000}{RT}\right] \quad (10)$$

404
$$D_{D_hightemp} = \left(\frac{fO_{2_FMQ}}{fO_{2_Ex}}\right)^{\frac{1}{6}} \cdot 6.66 \times 10^{-10} \cdot \exp\left[\frac{-201000}{RT}\right] \quad (11)$$

405 where fO_{2_FMQ} is the oxygen fugacity at the FMQ buffer and fO_{2_Ex} is that in the
 406 experiments. Dohmen et al. (2007) reported that Fe-Mg interdiffusion at low temperatures
 407 (below 1173 K) are independent of fO_2 , most likely due to a change in the diffusion
 408 mechanism where the vacancy concentration is controlled extrinsically by the impurity ions.
 409 Thus, the fO_2 correction was not made for $D_{D_lowtemp}$.

410 The oxygen self-diffusion coefficients (m²/s) in olivine, D_{J_1980} (Jaoul et al., 1980),
 411 D_{Re} (Reddy et al., 1980), D_{J_1983} (Jaoul et al., 1983), $D_{O\&A}$ (Oishi and Ando, 1984; Ando et al.,
 412 1981), D_A (Andresson et al., 1989), $D_{G\&J}$ (Gérard and Jaoul, 1989), D_{Ry} (Ryerson et al., 1989)
 413 and D_{D_2002} (Dohmen et al., 2002) are expressed as follows:

414
$$D_{J_1980} = 1.46 \times 10^{-8} \cdot \exp\left[\frac{-328000}{RT}\right] \quad (12)$$

415
$$D_{\text{Re}} = 3.50 \times 10^{-7} \cdot \exp\left[\frac{-372000}{RT}\right] \quad (13)$$

416
$$D_{\text{L}_{1983}} = 2.30 \times 10^{-10} \cdot \exp\left[\frac{-293000}{RT}\right] \quad (14)$$

417
$$D_{\text{O\&A}} = 2.85 \times 10^{-6} \cdot \exp\left[\frac{-416000}{RT}\right] \quad (15)$$

418
$$D_{\text{A}} = 6.86 \times 10^{-10} \cdot \exp\left[\frac{-302000}{RT}\right] \quad (16)$$

419
$$D_{\text{G\&J}} = 6.7 \times 10^{-6} \cdot (fO_2_{\text{FMQ}})^{0.34} \cdot \exp\left[\frac{-38000}{T}\right] \quad (17)$$

420
$$D_{\text{Ry}} = 2.6 \times 10^{-10} \cdot (fO_2_{\text{FMQ}})^{0.21} \cdot \exp\left[\frac{-226000}{RT}\right] \quad (18)$$

421
$$D_{\text{D}_{2002}} = 4.57 \times 10^{-9} \cdot \exp\left[\frac{-338000}{RT}\right] \quad (19)$$

422 The diffusivities given by Eqs. (12–16) are independent of fO_2 because they were
 423 determined for pure forsterite, and the oxygen fugacity dependence was not determined for
 424 $D_{\text{D}_{2002}}$. We applied all the combinations of $D_{\text{Fe-Mg}}$ given by Eqs. (7) and (9–11) and D_{oxygen}
 425 given by Eqs. (12–19) to Eq. (3) to calculate the possible range of x_{oxygen} for a given $x_{\text{Fe-Mg}}$ of
 426 1.3 μm .

427 Figure 8 shows the change of x_{oxygen} with temperature. The x_{oxygen} is widespread over
 428 the range from 10^{-7} nm to 0.1 μm at temperatures below 1000 K. Because the parent body of
 429 the CV_{oxA} chondrite was not heated above 1000 K (e.g., Keil, 2000; Ito and Messenger, 2010),
 430 x_{oxygen} should be smaller than 0.1 μm . Therefore, the oxygen isotopic compositions in the
 431 olivine of NWA3118 should not have been disturbed on the parent body except for a very thin
 432 layer ($< 0.1 \mu\text{m}$) of the grain boundary. Because the thickness of the possible oxygen diffusion
 433 layer is much smaller than the spatial resolution of the oxygen isotopography (Fig. 1), the thin
 434 layer was not observed in this study.

435

436 4.2. Oxygen diffusion in MgO-rich olivine

437 Figure 9 shows the $\delta^{18}\text{O}$ profiles at the edge of the extreme ^{16}O -rich area in a
 438 MgO-rich olivine single crystal within the igneous rim. The sharpness of the boundary defined
 439 by 2σ , a width between 16% and 84% of the compositional difference, is 1.5–2.3 μm . This
 440 value is larger than the spatial resolution of the isotopography (0.8 μm , Fig. 1). This indicates
 441 that oxygen isotope self-diffusion ($\sim 1 \mu\text{m}$) occurred in the MgO-rich olivine across the
 442 boundary between the different oxygen isotopic compositions prior to the parent body
 443 accretion. The oxygen isotope self-diffusion distance of $\sim 1 \mu\text{m}$ may also explain the

444 observation that the oxygen isotopic compositions in the extreme ^{16}O -rich areas tend toward
 445 ^{16}O -depletion when the area is smaller than $3\ \mu\text{m}$ across the short side (Fig. 7).

446 Because the MgO-rich olivine of the igneous rim was crystallized from a melt
 447 and the extreme ^{16}O -rich areas in the MgO-rich olivine should have been relict crystals in
 448 the melt, oxygen isotope exchanges in the ^{16}O -rich areas should have principally occurred
 449 during the igneous rim formation through oxygen self-diffusion in the MgO-rich olivine. We
 450 note that forsterite dissolution occurred during the igneous-rim formation from the irregular
 451 shapes of extreme ^{16}O -rich areas and forsterite dissolution in chondrule-like melts is a very
 452 fast process (Soulié et al. 2017). This suggests that forsterite dissolution might not have a
 453 significant effect for the oxygen isotope self-diffusion.

454 Because the $f\text{O}_2$ during chondrule formation was estimated to be close to the
 455 iron-wustite (IW) oxygen buffer (Righeter and Neff, 2007), we assumed that the diffusion
 456 occurred under the IW buffer (IW~0) as follows (Myers and Eugster, 1983);

$$457 \quad \log f\text{O}_2 = -\frac{26834.7}{T} + 6.471 \quad (20)$$

458 We note that chondrule formed under the oxygen fugacity was one or two log units below
 459 IW (IW-1–IW-2) (Grossman et al., 2012). This difference in $f\text{O}_2$ has no influence on our
 460 calculation, even if $f\text{O}_2$ is 2 orders of magnitude (IW-2) lower than our assumption (IW~0).
 461 Because our data were normalized to an $f\text{O}_2$ using an exponent of 1/6 for $f\text{O}_2$ dependence of
 462 Fe-Mg inter-diffusion in olivine (Dohmen et al., 2007). That is, only an exponent of 1/6
 463 difference affects a diffusion coefficient. Furthermore, the Fe-Mg diffusion distance is also
 464 proportional to the square root of a diffusion coefficient, indicating the effect of oxygen
 465 fugacity difference is not significantly large. The $D_{\text{Fe-Mg}}$ corrected to the IW $f\text{O}_2$ are
 466 expressed as follows:

$$467 \quad D_{N\&S} = \left(\frac{f\text{O}_{2_IW}}{f\text{O}_{2_Ex}}\right)^{\frac{1}{6}} \cdot 8.33 \times 10^{-10} \cdot \exp\left[\frac{-188000}{RT}\right] \quad (21)$$

$$468 \quad D_C = \left(\frac{f\text{O}_{2_IW}}{f\text{O}_{2_Ex}}\right)^{\frac{1}{6}} \cdot 5.38 \times 10^{-9} \cdot \exp\left[\frac{-226000}{RT}\right] \quad (22)$$

$$469 \quad D_{D_hightemp} = \left(\frac{f\text{O}_{2_IW}}{f\text{O}_{2_Ex}}\right)^{\frac{1}{6}} \cdot 6.66 \times 10^{-10} \cdot \exp\left[\frac{-201000}{RT}\right] \quad (23)$$

470 The $f\text{O}_{2_IW}$ in the equations is the oxygen fugacity at the IW buffer and the $f\text{O}_{2_Ex}$ is the
 471 oxygen fugacity in the experiment described above (see details in section 4.1).

472 The Fe-Mg interdiffusion coefficients in olivine at high temperatures (1673–1873
 473 K) were also determined by Tachibana et al. (2013). The diffusion coefficient was obtained
 474 under a controlled $f\text{O}_2$ of $10^{-7.5}$ bars and Fo_{95} composition:

475
$$D_T = 4.84 \times 10^{-5} \cdot \exp\left[\frac{-323000}{RT}\right] \quad (24)$$

476 The diffusion coefficient (m^2/s) after the $f\text{O}_2$ correction to IW is given by:

477
$$D_T = \left(\frac{f\text{O}_2\text{-IW}}{f\text{O}_2\text{-Ex}}\right)^{\frac{1}{6}} \cdot 4.84 \times 10^{-5} \cdot \exp\left[\frac{-323000}{RT}\right] \quad (25)$$

478 We calculated the $x_{\text{Fe-Mg}}$ in olivine for $x_{\text{oxygen}} = 1 \mu\text{m}$ using Eq. (3). The $x_{\text{Fe-Mg}}$
 479 decreases with increasing temperature (Fig. 10) and is estimated to be larger than $10 \mu\text{m}$ for
 480 all the combinations of $D_{\text{Fe-Mg}}$ and D_{oxygen} up to 2000 K that is an inferred peak temperature
 481 of chondrule formation (e.g., Hewins and Connolly, 1996; Hewins et al., 2005).

482 Because the size of the extreme ^{16}O -rich areas in this study were less than $10 \mu\text{m}$
 483 across (Fig. 7), the calculated $x_{\text{Fe-Mg}}$ is large enough to homogenize the Fe-Mg distribution
 484 between the ^{16}O -rich areas and the surrounding overgrown MgO-rich olivine. This is
 485 consistent with the observations of the homogeneous chemical compositions of the
 486 MgO-rich olivines with different oxygen isotopic compositions.

487 The heating duration for $x_{\text{oxygen}} = 1 \mu\text{m}$ within the MgO-rich olivine in the
 488 igneous rim was calculated to be from several hours to several days at near liquidus
 489 temperature (2000 K) and longer than one year at near solidus temperature (1400 K) (Fig.
 490 11). The estimated heating duration was similar to those for chondrule formation at near
 491 liquidus temperature that was molten for up to several tens of hours (e.g., Hewins and
 492 Connolly, 1996; Hewins et al., 2005) suggesting that the igneous rim was also formed during
 493 transient heating events.

494

495 5. Conclusions

496 The igneous rim surrounding a type I chondrule in the Northwest Africa 3118 CV_{oxA}
 497 chondrite mainly consists of two types of olivine with MgO-rich (Fa₁₁₋₂₂) and FeO-rich
 498 (Fa₄₀₋₄₉) compositions. The igneous rim contains secondary minerals that were formed by
 499 aqueous alteration on the CV chondrite parent body, including ~~ferroan~~ FeO-rich olivine,
 500 nepheline, and Fe-Ni sulfides.

501 The oxygen isotopic composition of the FeO-rich olivine ($\delta^{17}\text{O} = 2.0 \pm 1.8\%$, $\delta^{18}\text{O}$
 502 $= 8.1 \pm 3.6\%$, 2SD), the mass-dependent isotope fractionation, and the porous texture filling
 503 between the MgO-rich olivine grains indicate that the FeO-rich olivine precipitated from an
 504 ^{16}O -poor aqueous fluid on the CV chondrite parent body. The MgO-rich olivine shows Fe-Mg
 505 chemical zoning at the interface with the FeO-rich olivine, indicating that Fe-Mg
 506 inter-diffusion occurred during the aqueous alteration on the parent body. The inter-diffusion
 507 distance indicates that the oxygen isotopic composition of the MgO-rich olivine ($\delta^{17}\text{O} = -5.6 \pm$

508 3.2‰, $\delta^{18}\text{O} = -0.7 \pm 3.6\text{‰}$) was not disturbed on the parent body and should preserve the rim
509 formation events in the solar nebula.

510 The MgO-rich olivine includes extreme ^{16}O -rich areas ($\delta^{17,18}\text{O} \sim -30\text{‰}$). These
511 areas are anhedral and less than 10 μm across, indicating that they represent relict grains that
512 survived the igneous rim formation event. The oxygen diffusion distance of about 1 μm in the
513 extreme ^{16}O -rich area constrains that the heating duration of the igneous rim formation was
514 from several hours to several days at about 2000 K, suggesting flash heating events similar to
515 chondrule formation formed the igneous rim. The MgO-rich olivine grains with $\delta^{18}\text{O}$ values of
516 $\sim -1\text{‰}$ overgrew the extreme ^{16}O -rich relict olivine grains during igneous rim formation.

517

518 **Acknowledgments**

519 This article is dedicated to Klaus Keil in honor of his outstanding contributions to
520 cosmochemistry. Constructive comments by Timothy J. Fagan, Kazuhide Nagashima and the
521 Associate Editor Alexander N. Krot significantly improved the quality of the manuscript. This
522 study is partly supported by the Monbu-kagaku-sho grants.

523

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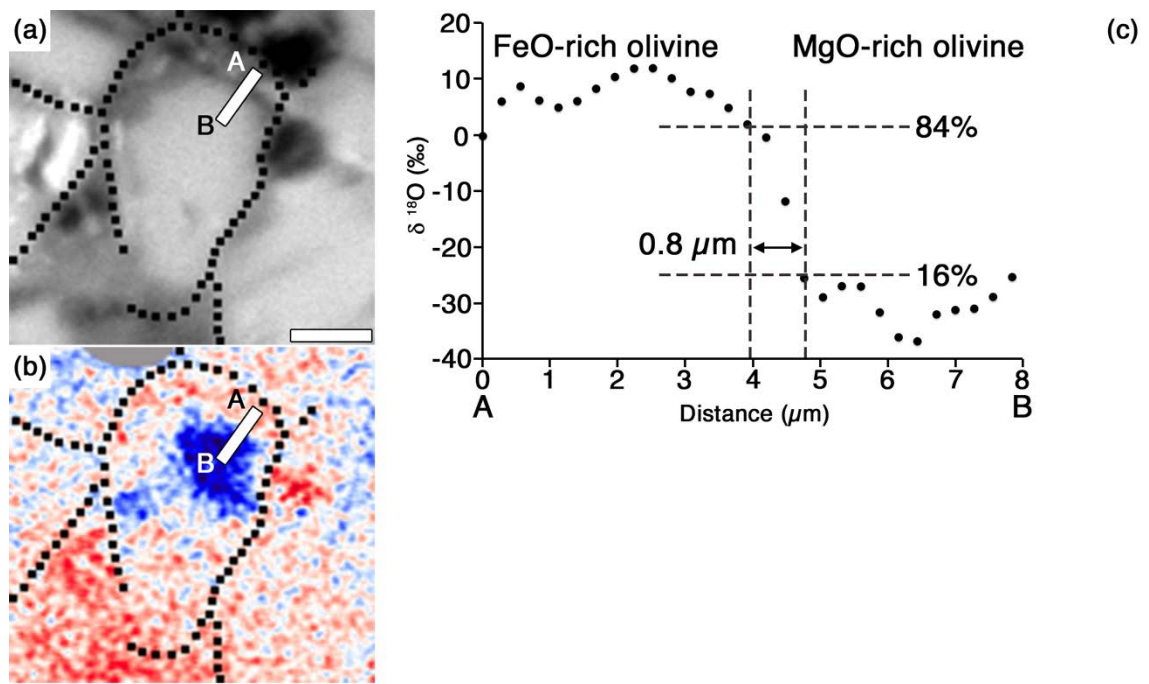
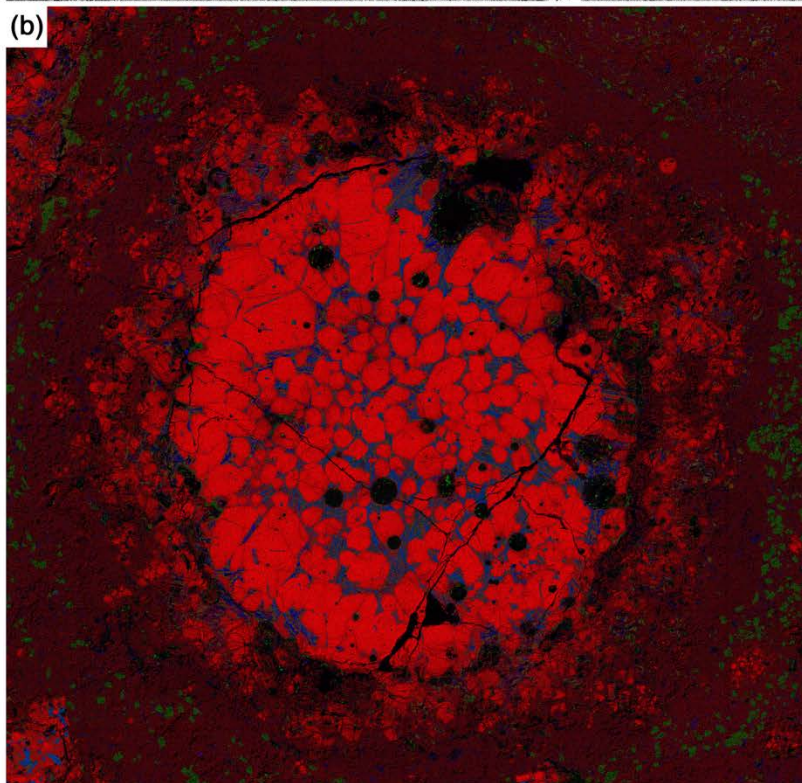
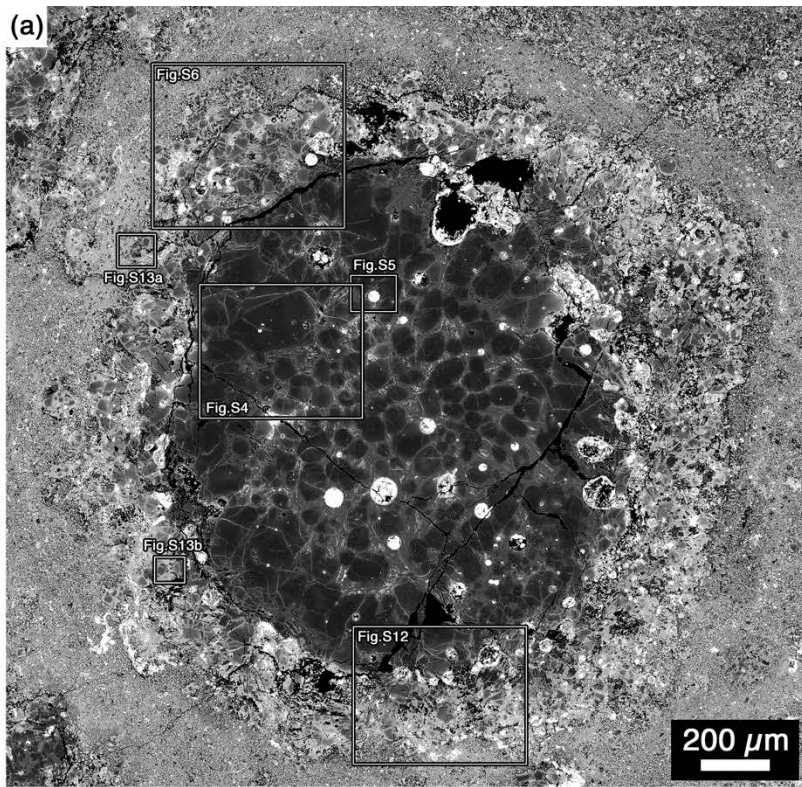


Fig. 1. (a) $^{24}\text{Mg}^{16}\text{O}^-$ isotopograph. (b) $\delta^{18}\text{O}$ isotopograph. (c) $\delta^{18}\text{O}$ profile along line A-B between the MgO-rich and FeO-rich olivine. The same isotopographs are used in Fig. 7c to determine the $\delta^{18}\text{O}$ profile within the MgO-rich olivine. Scale bar is 10 μm .



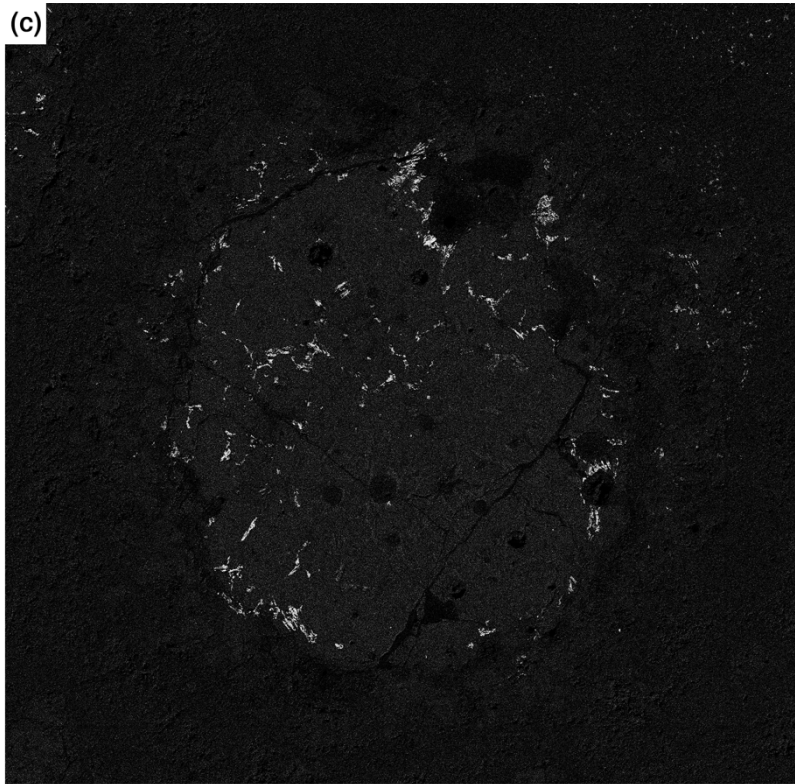


Fig. 2. (a) Back-scattered electron image of a type I chondrule surrounded by the igneous rim in the NWA3118 CV_{oxA} chondrite. The chondrule mainly consists of olivine, mesostasis, magnetite, sulfide and Fe-Ni metal. The areas outlined by squares are enlarged and shown in the figures indicated. (b) Combined X-ray elemental map of K α lines of Mg (red), Ca (green), and Al (blue). Blue-colored areas are mainly composed of plagioclase and nepheline. (c) X-ray elemental map of Na K α . White-colored areas are nepheline-rich.

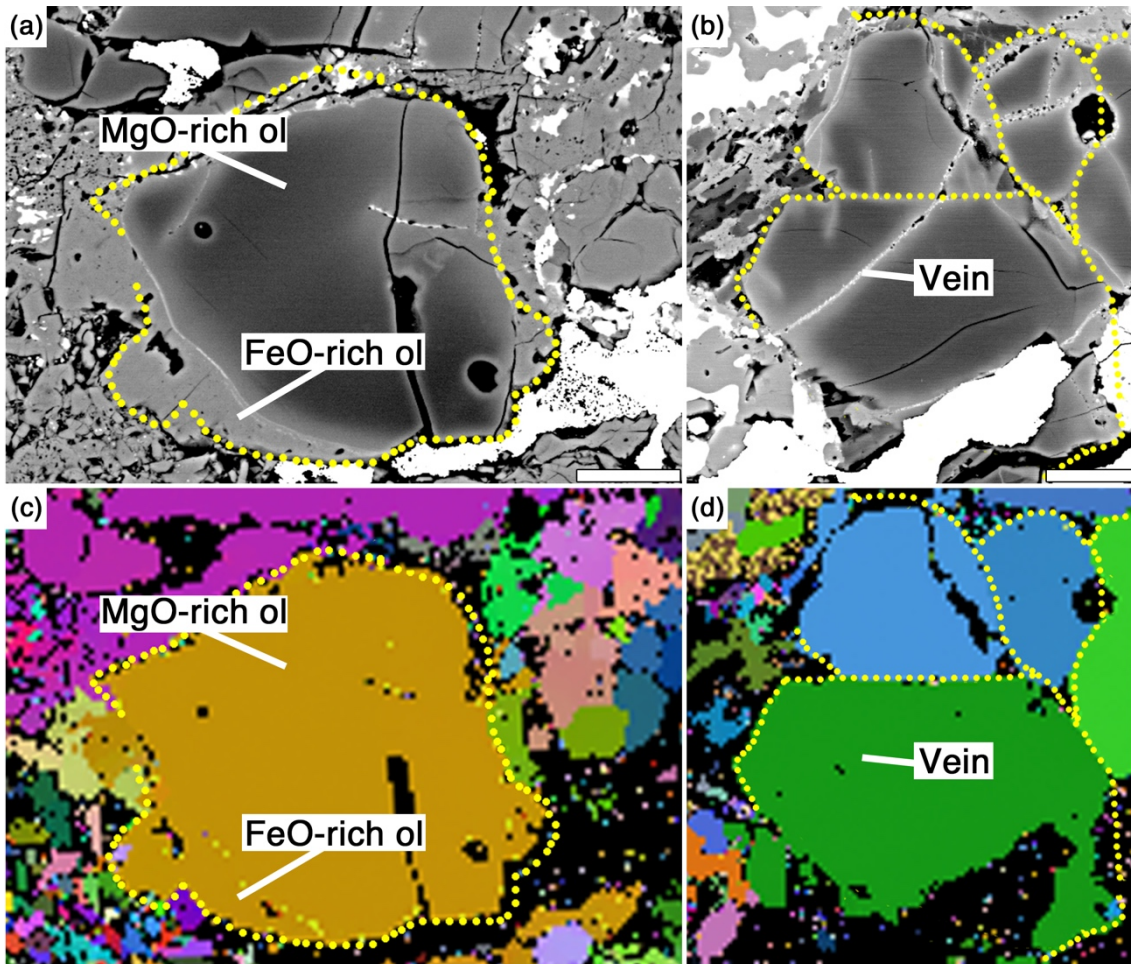
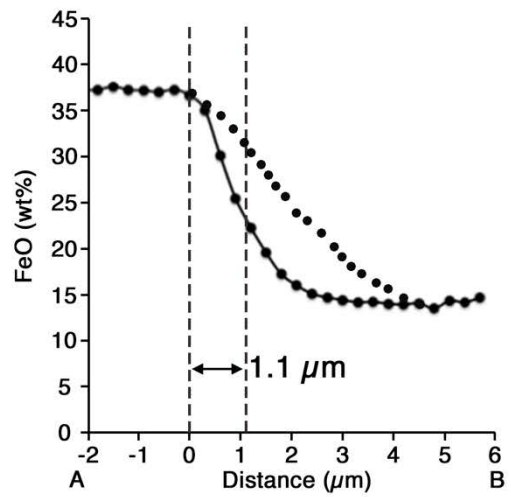
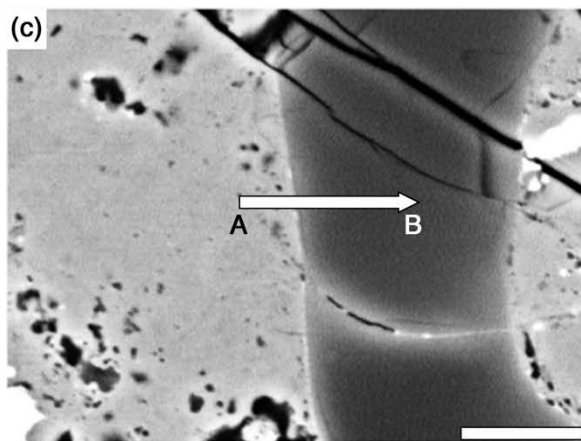
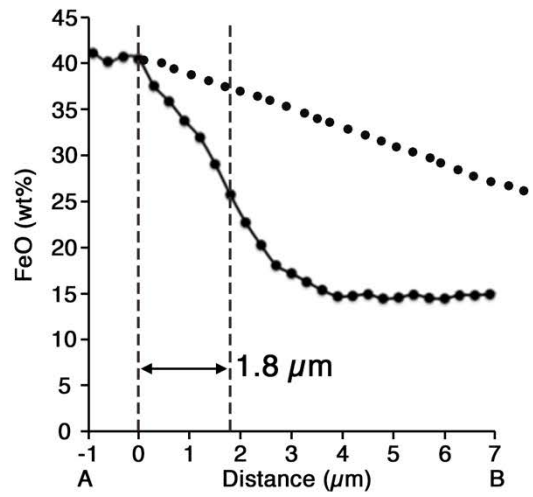
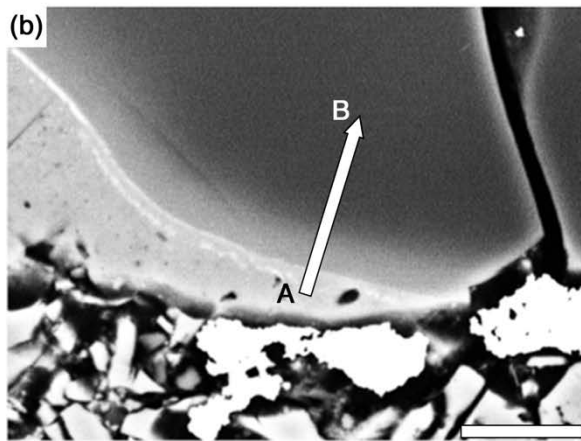
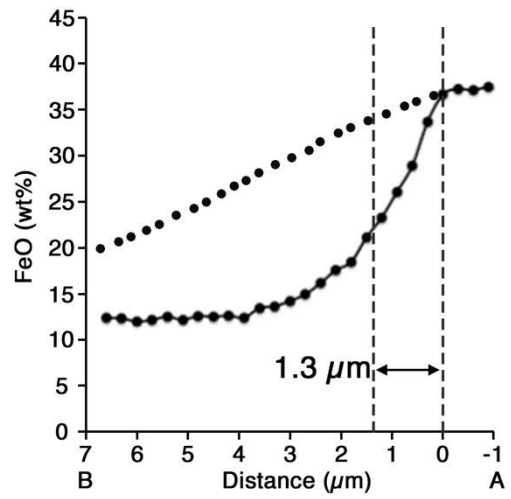
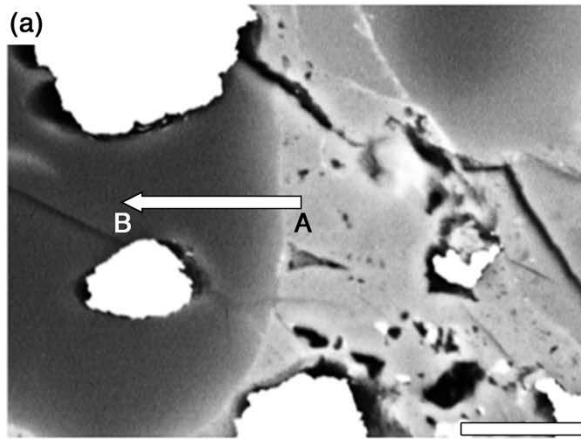
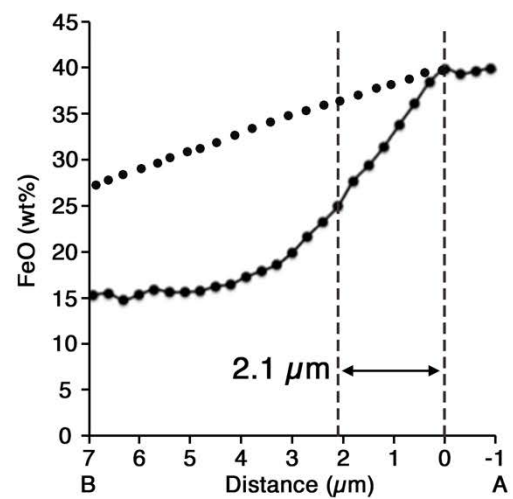
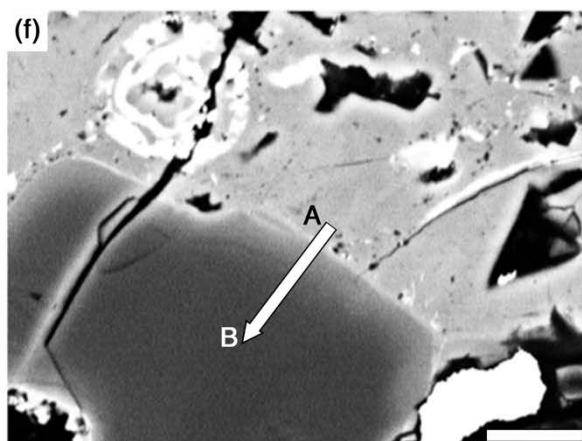
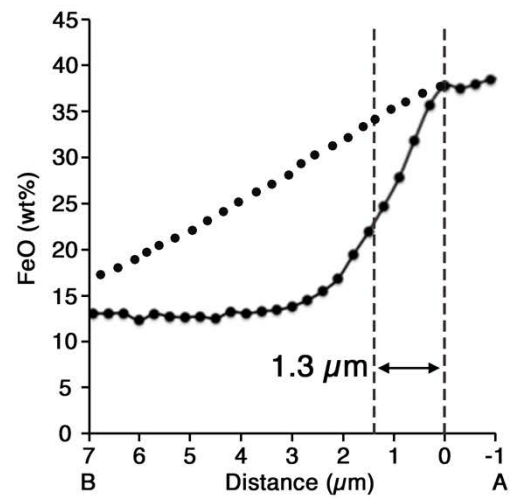
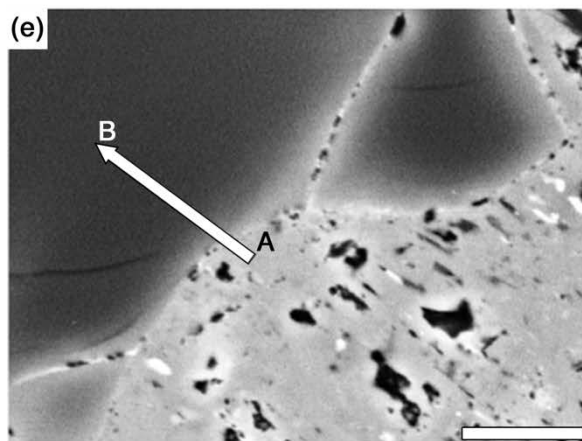
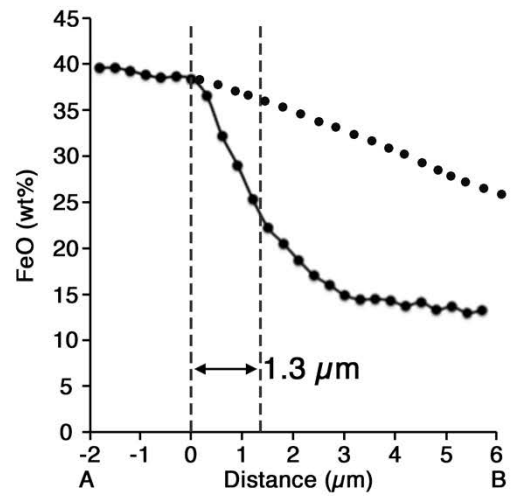
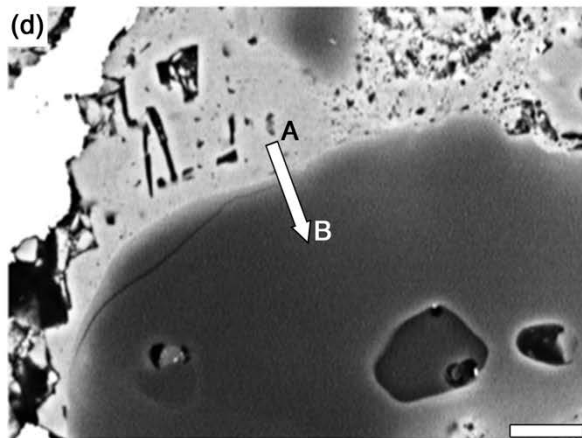


Fig. 3. Back-scattered electron images (a and b) and the EBSD images (c and d) of olivine grains in the igneous rim. Dotted lines represent grain boundaries. Scale bar is 10 μm . In (c), the EBSD images show that the FeO-rich olivine (FeO-rich ol) and MgO-rich olivine have the same crystallographic orientation. In (d), the MgO-rich olivine grains separated by veins have the same crystallographic orientation.





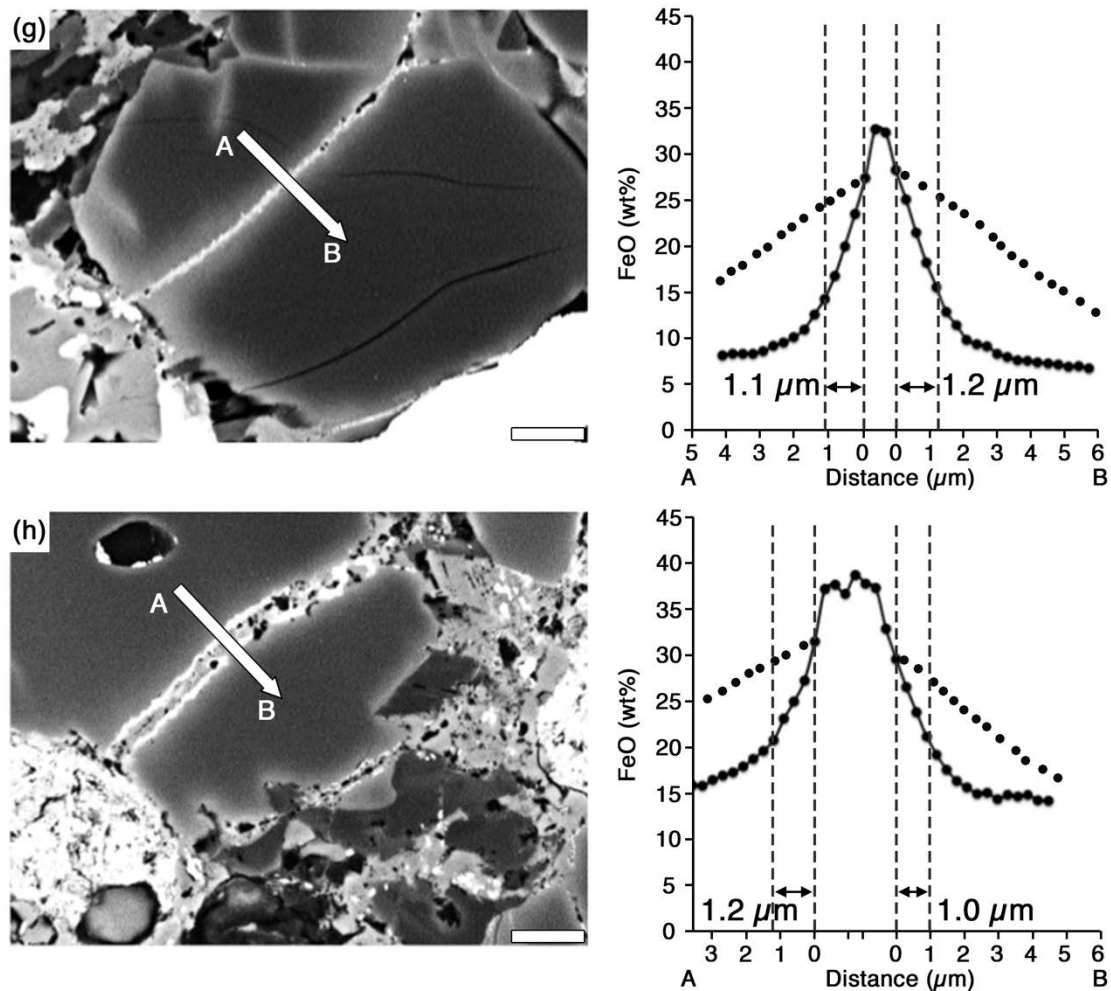


Fig. 4. (a–f) Compositional profiles along lines crossing the interfaces between FeO-rich olivine (A) and MgO-rich olivine (B). (g–h) Compositional profile across veins in MgO-rich olivine. Interdiffusion distances of the Fe-Mg zoning are shown in diagrams. Hypothetical interdiffusion profiles are shown by dotted lines in each diagram. The hypothetical diffusion is calculated using a spherical body model of constant surface concentration (Crank, 1975). The boundary conditions are as follows: initial concentration is Fa_2 and the final concentration at the center of spherule is Fa_{15} . Measured surface concentration and measured radius for each olivine are used for calculation. Scale bar is 5 μm .

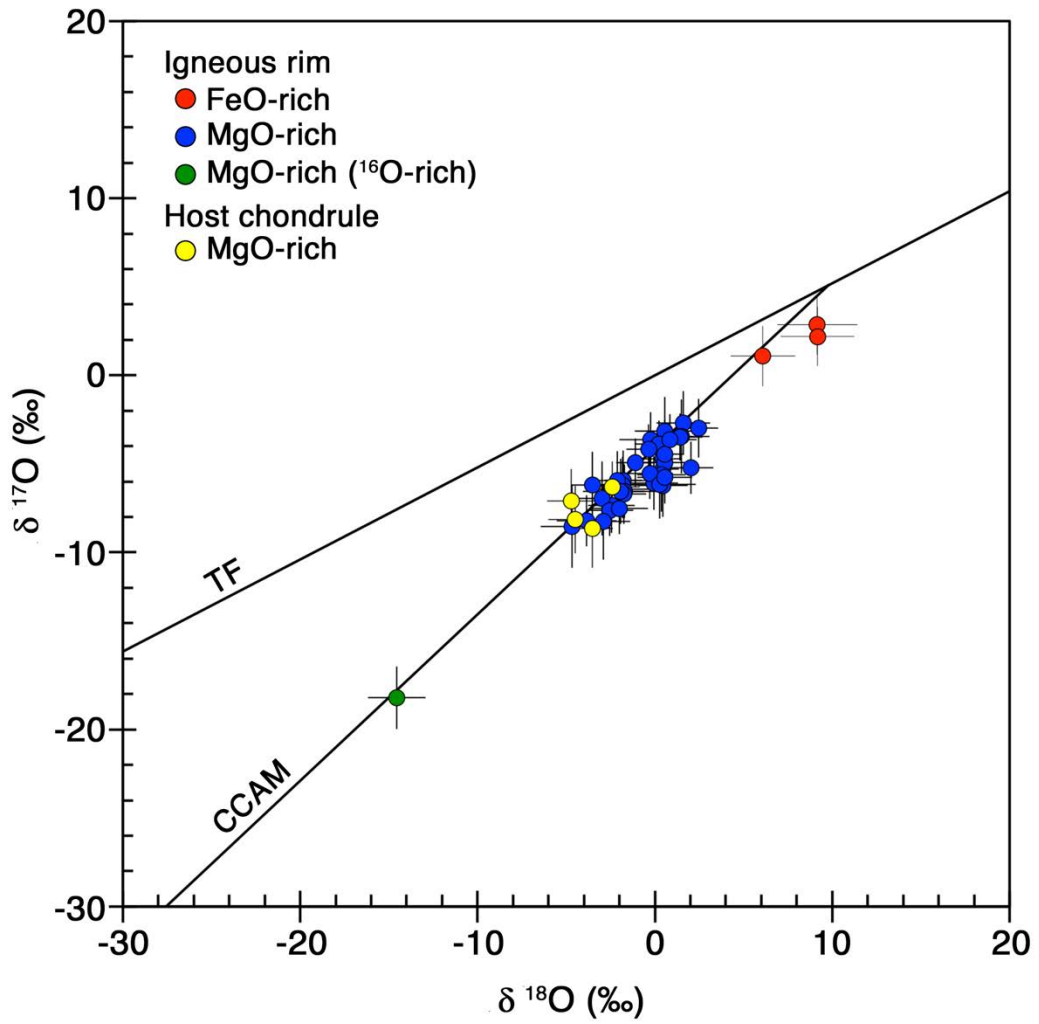


Fig. 5. Oxygen isotopic compositions of olivine in the igneous rim and the host chondrule. Error bar is 2σ TF: terrestrial fractionation line. CCAM: carbonaceous chondrite anhydrous mineral line.

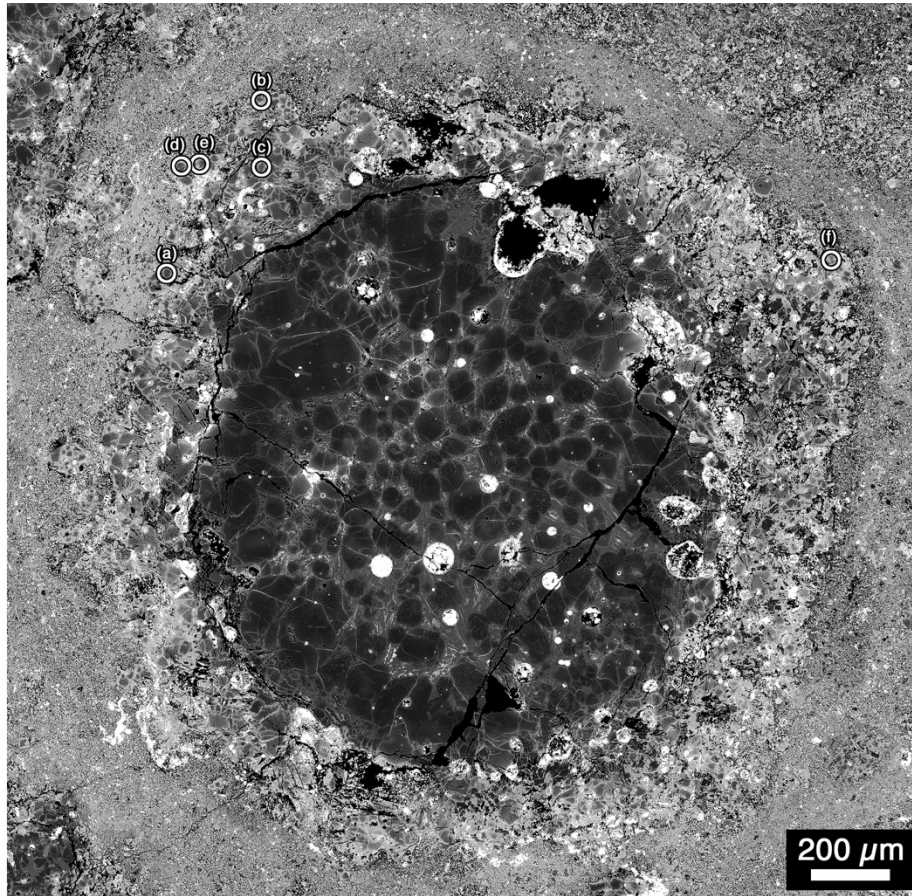


Fig. 6. Distribution of olivine grains containing the extreme ^{16}O -rich areas in the igneous rim. White circles show their positions. The magnified images at locations of circles labeled (a) to (f) show in Fig. 7 with the same labels.

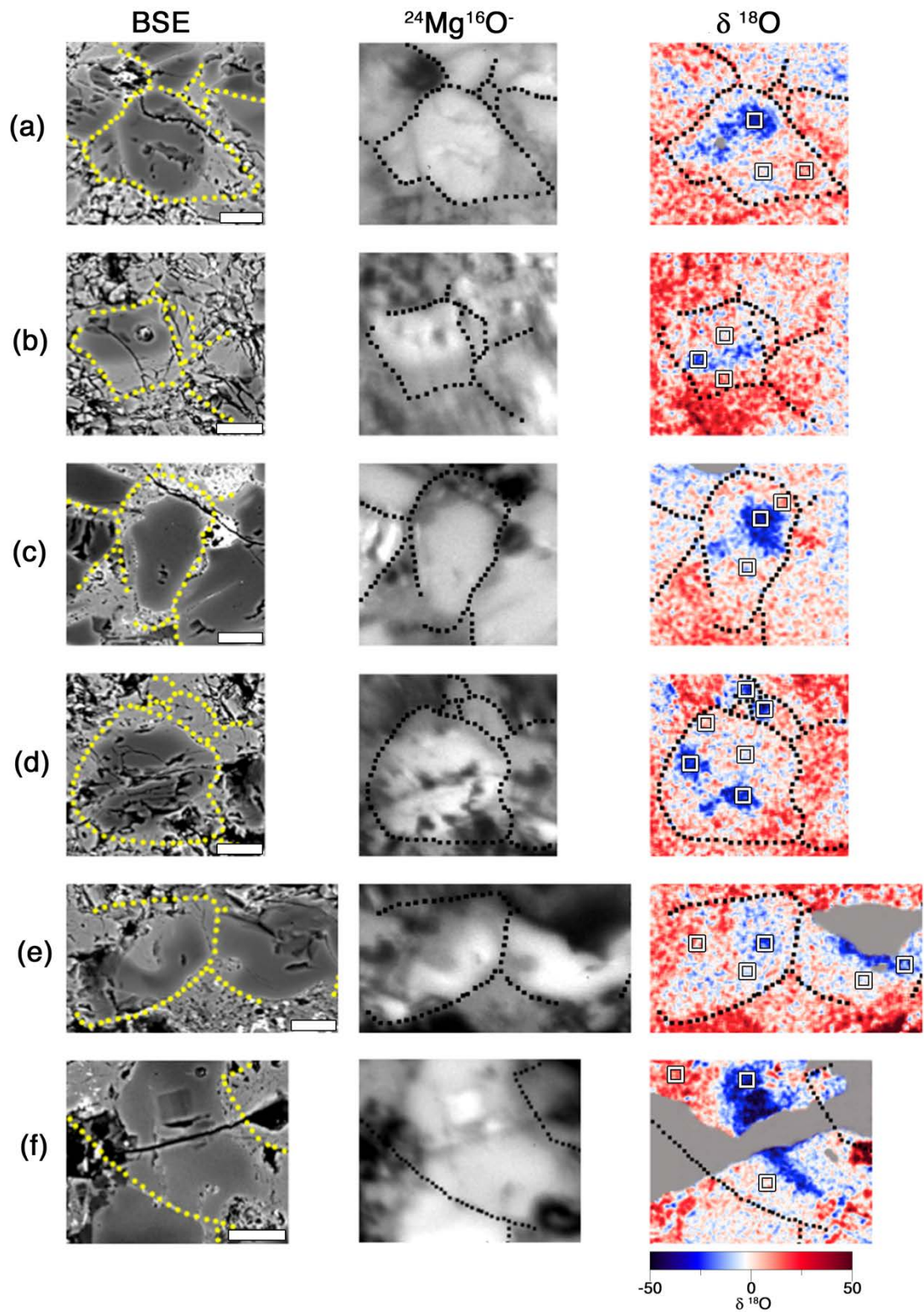


Fig. 7. BSE images, $^{24}\text{Mg}^{16}\text{O}^-$ and $\delta^{18}\text{O}$ isotopograph of olivine containing extreme ^{16}O -rich areas in the igneous rim. Dotted lines represent the grain boundaries determined by EBSD. Scale bar is 10 μm . Areas with artifacts are masked by the gray color. Squares in the $\delta^{18}\text{O}$ isotopograph correspond to areas where $\delta^{18}\text{O}$ values were calculated (see details in the text).

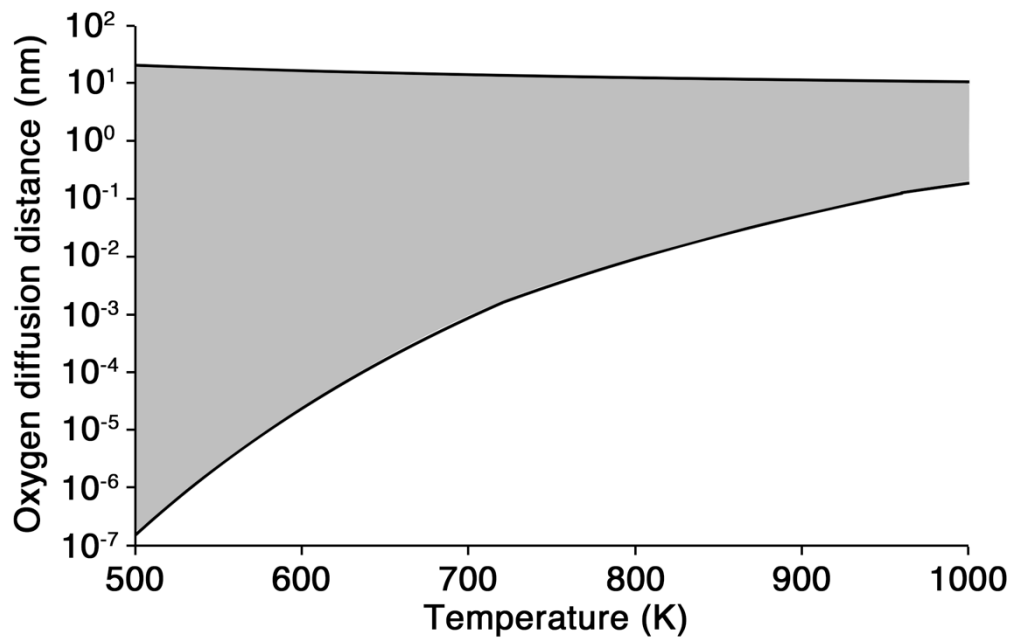


Fig. 8. Range of oxygen self-diffusion distance (nm) as a function of temperature (K) at Fe-Mg inter-diffusion distance of $1.3 \mu\text{m}$.

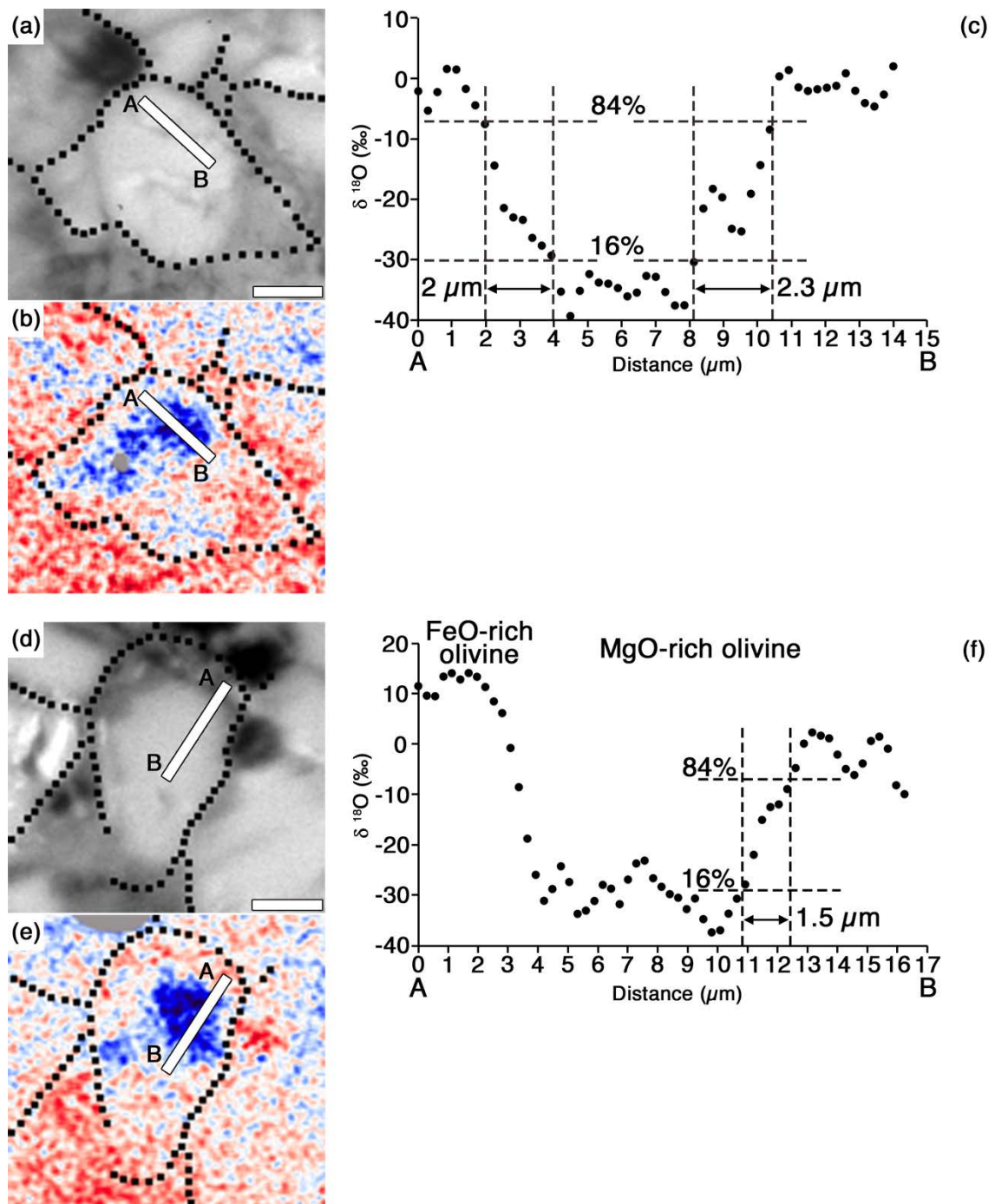


Fig. 9. (a and d) $^{24}\text{Mg}^{16}\text{O}^-$ isotopographs; (b and e) $\delta^{18}\text{O}$ isotopographs; (c and f) $\delta^{18}\text{O}$ profiles of lines A-B. Scale bar is 10 μm .

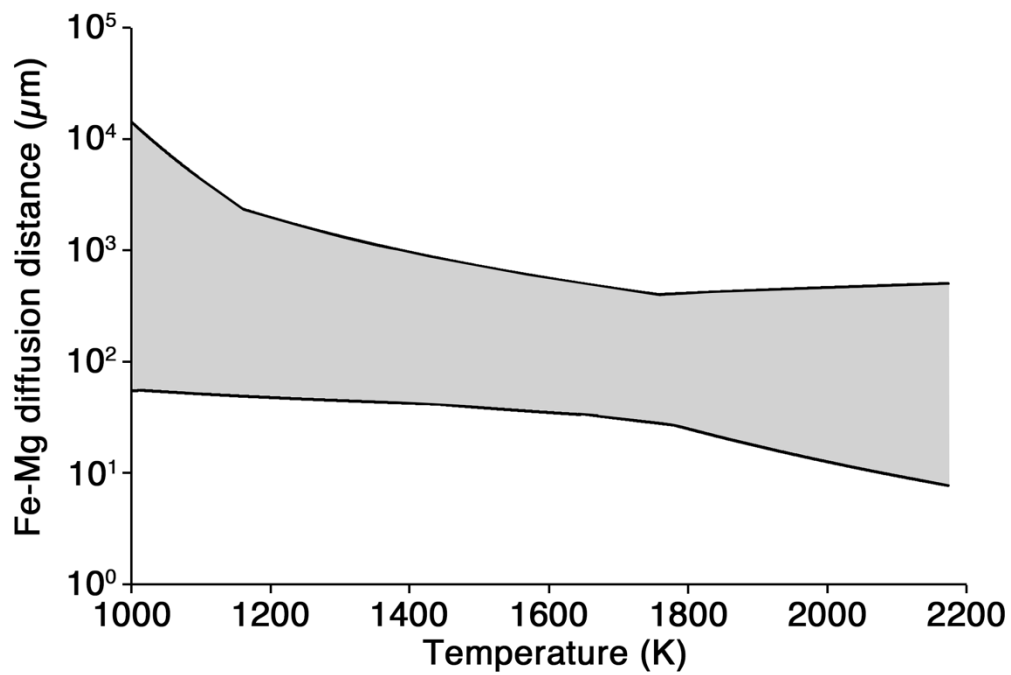


Fig. 10. Range of Fe-Mg interdiffusion distance (μm) as a function of temperature (K) at oxygen self-diffusion distance of 1 μm .

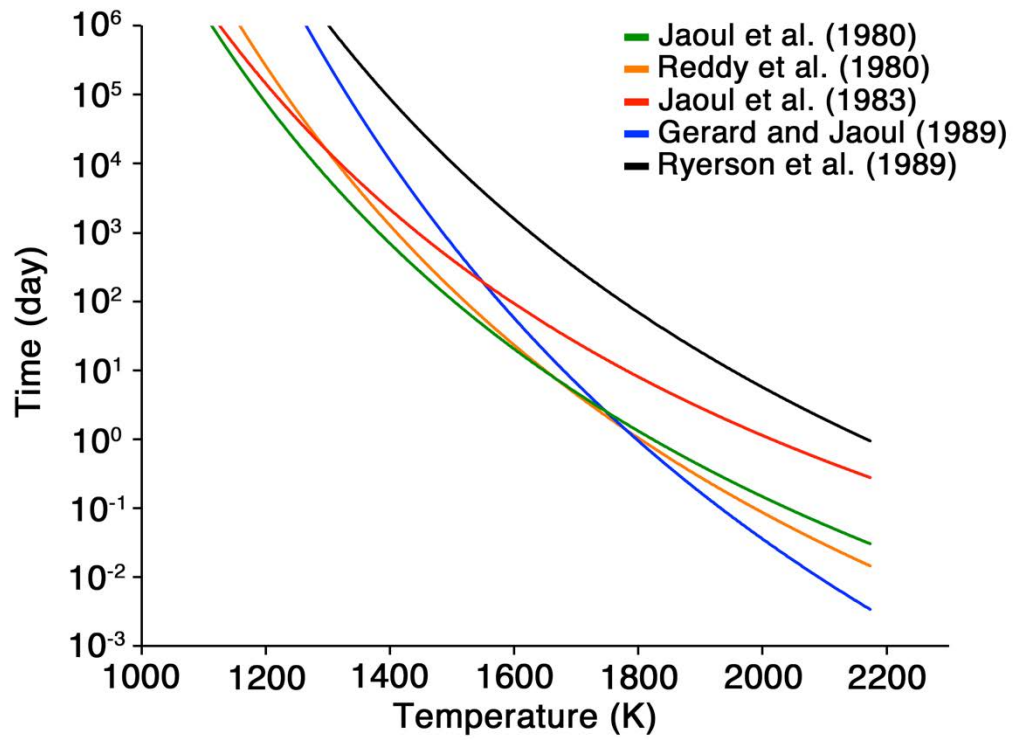


Fig. 11. Range of heating time (day) as a function of temperature (K) at oxygen self-diffusion distance of 1 μm .