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**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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Abstract

Due to their common occurrences in various types of chondrites, igneous rims formed on pre-existing chondrules throughout chondrule-forming regions of the solar nebula. Although the peak temperatures are thought to reach similar values to those achieved during chondrule formation events, the heating duration in chondrule rim formation has not been well defined. We determined the two-dimensional chemical and oxygen isotopic distributions in an igneous rim of a chondrule within the Northwest Africa 3118 CV3_{oxA} chondrite with sub-micrometer resolution using secondary ion mass spectrometry and scanning electron microscopy. The igneous rim experienced aqueous alteration on the CV parent body. The aqueous alteration resulted in precipitation of the secondary FeO-rich olivine (Fa_{40–49}) and slightly disturbed the Fe-Mg distribution in the MgO-rich olivine phenocrysts (Fa_{11–22}) at about a 1 μm scale. However, no oxygen isotopic disturbances were observed at a scale greater than 100 nm. The MgO-rich olivine, a primary phase of igneous rim formation, has $\delta^{17}\text{O} = -6 \pm 3\text{‰}$ and $\delta^{18}\text{O} = -1 \pm 4\text{‰}$, and some grains contain extreme ^{16}O -rich areas ($\delta^{17}\text{O}$, $\delta^{18}\text{O} = \sim -30\text{‰}$) nearly 10 μm across. We detected oxygen isotopic migration of approximately 1 μm at the boundaries of the extreme ^{16}O -rich areas. Using oxygen self-diffusivity in olivine, the heating time of the igneous rim formation could have continued from several hours to several days at near liquidus temperatures ($\sim 2000\text{ K}$) in the solar nebula suggesting that the rim formed by a similar flash heating event that formed the chondrules.

1. INTRODUCTION

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

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

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

2. ANALYTICAL TECHNIQUES

2.1. Petrography and X-ray analysis

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

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

2.2. SIMS point analysis for O isotopic composition

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

2.3. Isotopography for O isotopic composition

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

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

3. RESULTS

3.1. Petrography and mineralogy

3.1.1. Chondrule

The Northwest Africa (NWA) 3118 is the oxidized Allende-like CV3_{oxA} carbonaceous chondrite (Russell et al., 2005). The chondrule studied is completely surrounded by a coarse-grained igneous rim (Fig. 2a). The chondrule has a rounded shape of 1.4 mm diameter with a type I (Mg# (= molar MgO/(MgO + FeO) %) $\geq$ 90)) porphyritic texture. The chondrule is mainly composed of olivine phenocrysts (Fa₂₋₅) and mesostasis in addition to opaque nodules. The major elemental compositions of these phases are listed in Tables S1–S3 (Figs. S1–S3).

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

3.1.2. Igneous rim

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

The coarse-grained rim is dominated by olivine. The olivine grains are enriched in FeO compared with those within the host chondrule. The FeO contents show a bimodal distribution with peaks at MgO-rich (Fa₁₁₋₂₂) and FeO-rich (Fa₄₀₋₄₉) compositions (Figs. S7, S8 and Table S4). Enstatite, pigeonite, feldspathic glasses, augite, nepheline, and Fe $\pm$ Ni sulfide (troilite, pyrrhotite or pentlandite) are identified in the rim (Fig. S6a). The major elemental compositions of these minerals are summarized in Tables S5–S8 (Figs. S9 and S10).

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

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

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

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

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

3.1.3 Matrix

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

3.2. Oxygen-isotope distribution of olivine

3.2.1 Chondrule and igneous rim

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

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

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

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

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

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

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

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

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

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

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

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

4. DISCUSSION

4.1. Aqueous alteration on the parent body

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

The oxygen self-diffusion coefficients (m^2/s) in olivine, D_{J_1980} (Jaoul et al., 1980), D_{Re} (Reddy et al., 1980), D_{J_1983} (Jaoul et al., 1983), $D_{O\&A}$ (Oishi and Ando, 1984; Ando et al., 1981), D_A (Andresson et al., 1989), $D_{G\&J}$ (Gérard and Jaoul, 1989), D_{Ry} (Ryerson et al., 1989) and D_{D_2002} (Dohmen et al., 2002) are expressed as follows:

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

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

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

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

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

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

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

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

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

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

4.2. Oxygen diffusion in MgO-rich olivine

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

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

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

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

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

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

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

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

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

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 oxygen fugacity in the experiment described above (see details in section 4.1).

The Fe-Mg interdiffusion coefficients in olivine at high temperatures (1673–1873 K) were also determined by Tachibana et al. (2013). The diffusion coefficient was obtained under a controlled $f\text{O}_2$ of $10^{-7.5}$ bars and Fo₉₅ composition:

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

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

$$D_T = \left(\frac{f\text{O}_{2\text{-IW}}}{f\text{O}_{2\text{-Ex}}}\right)^{\frac{1}{6}} \cdot 4.84 \times 10^{-5} \cdot \exp\left[\frac{-323000}{RT}\right] \quad (25)$$

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

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

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

5. Conclusions

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

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

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

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

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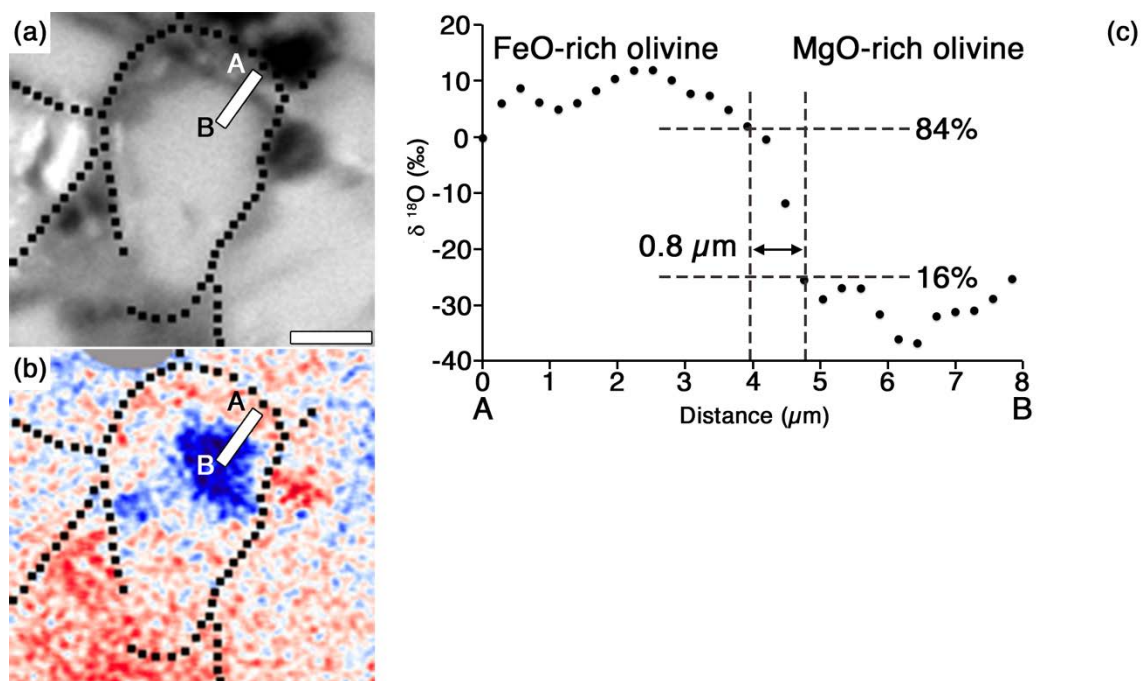
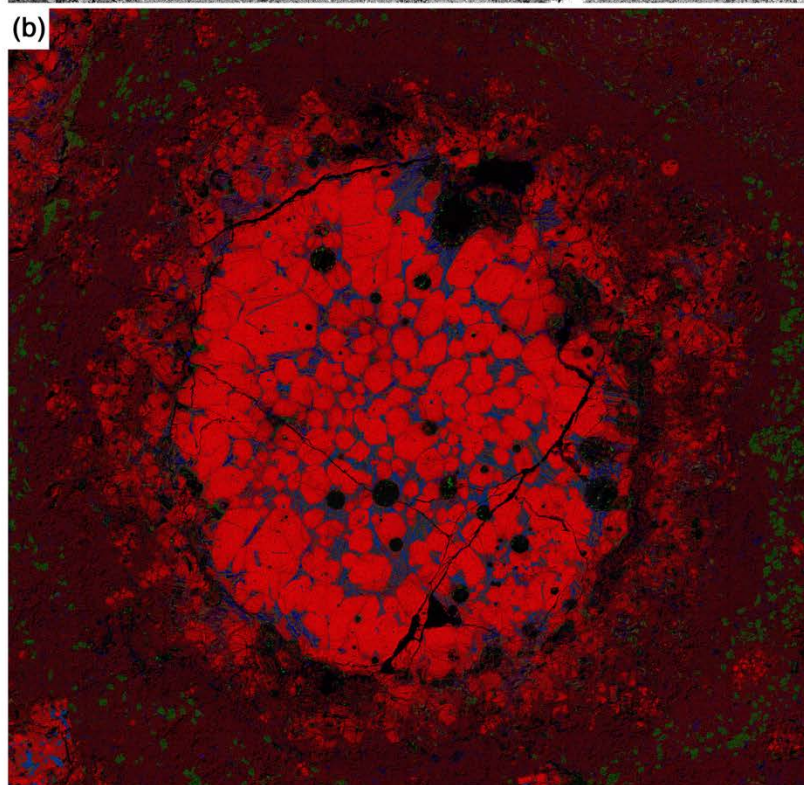
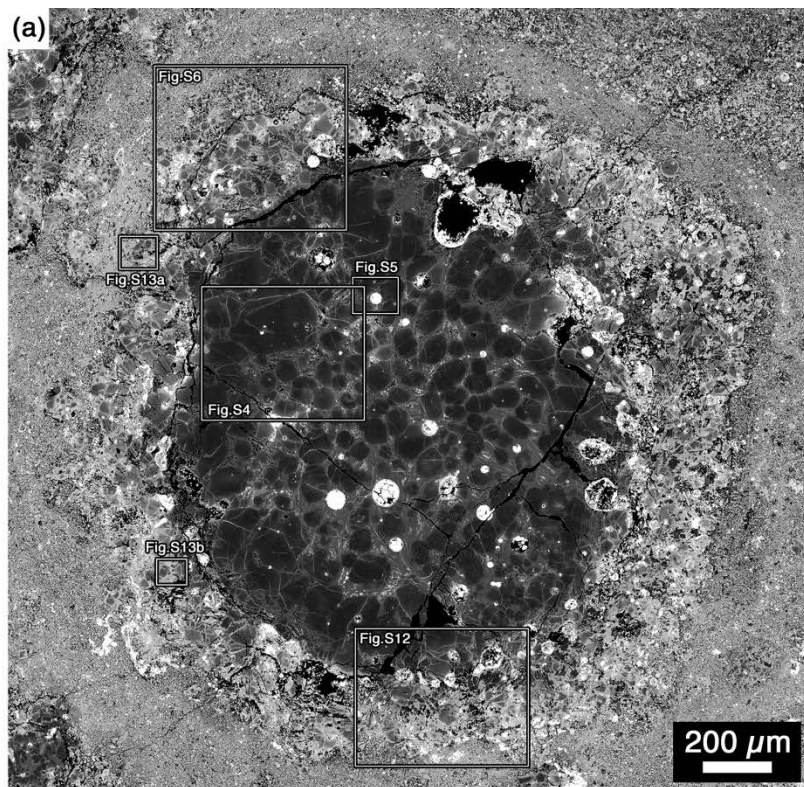


Fig. 1. (a) $^{24}\text{Mg}^{16}\text{O}^-$ isotopograph. (b) $\delta^{18}\text{O}$ isotopograph. (c) $\delta^{18}\text{O}$ profile along line A-B between the MgO-rich and FeO-rich olivine. The same isotopographs are used in Fig. 7c to determine the $\delta^{18}\text{O}$ profile within the MgO-rich olivine.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



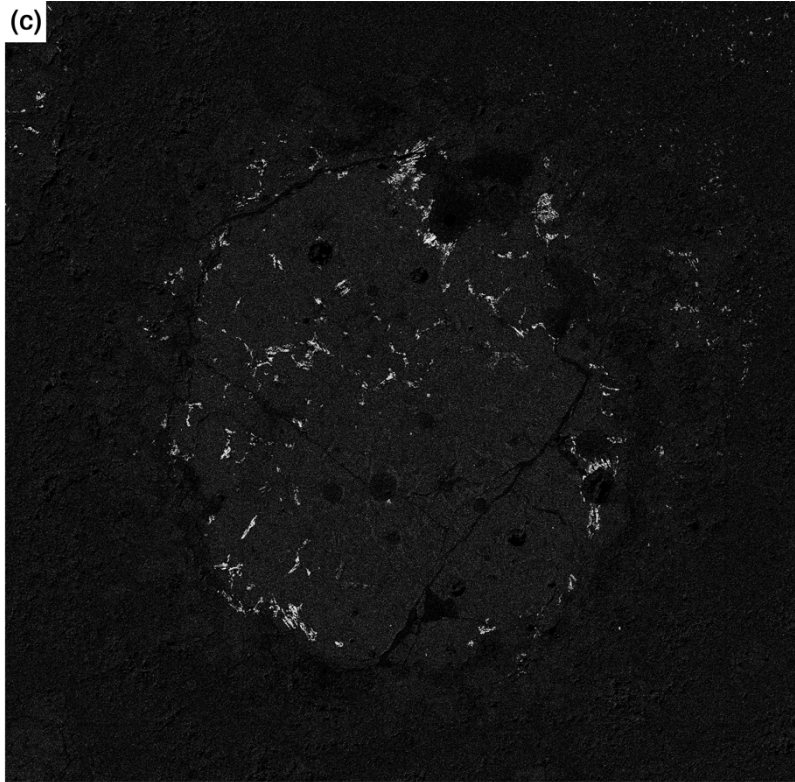


Fig. 2. (a) Back-scattered electron image of a type I chondrule surrounded by the igneous rim in the NWA3118 CV_{oxA} chondrite. The chondrule mainly consists of olivine, mesostasis, magnetite, sulfide and Fe-Ni metal. The areas outlined by squares are enlarged and shown in the figures indicated. (b) 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.

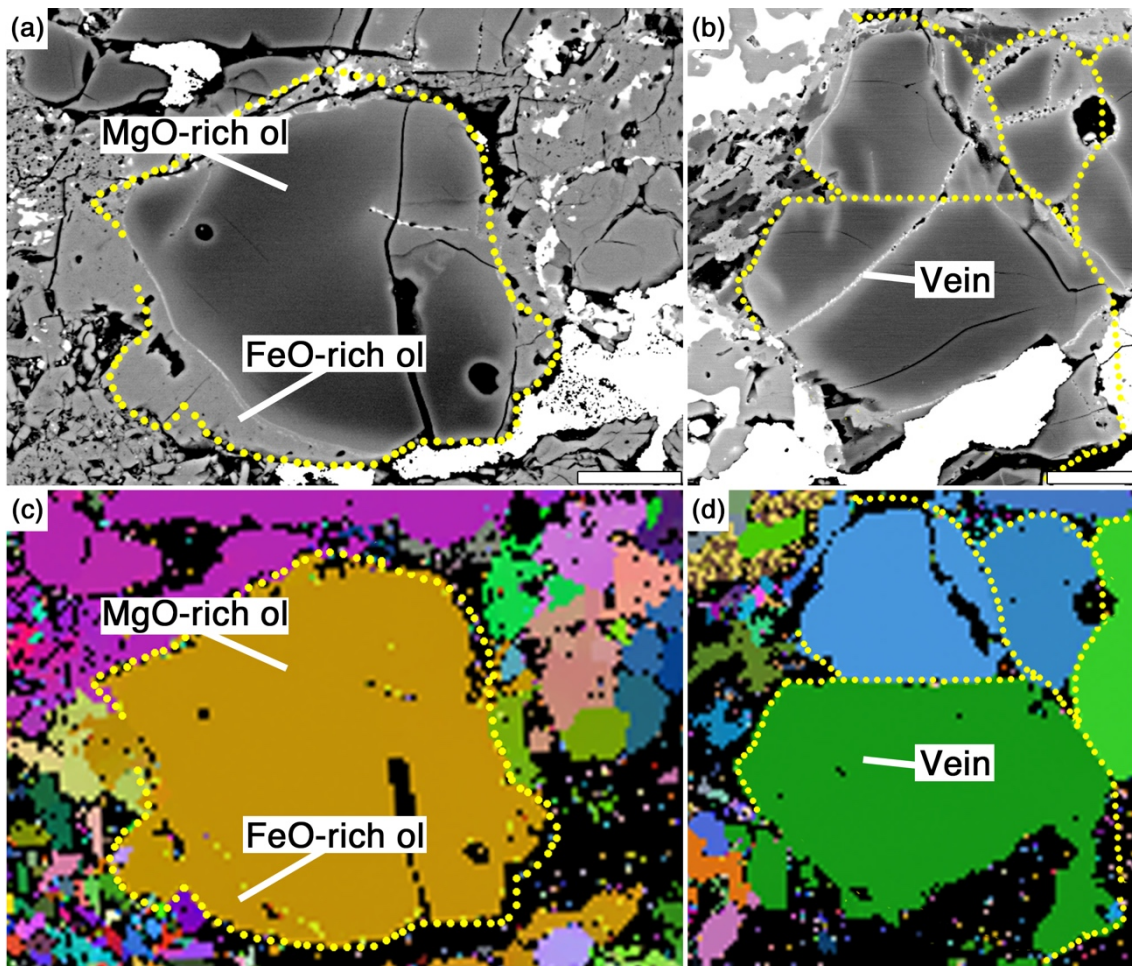
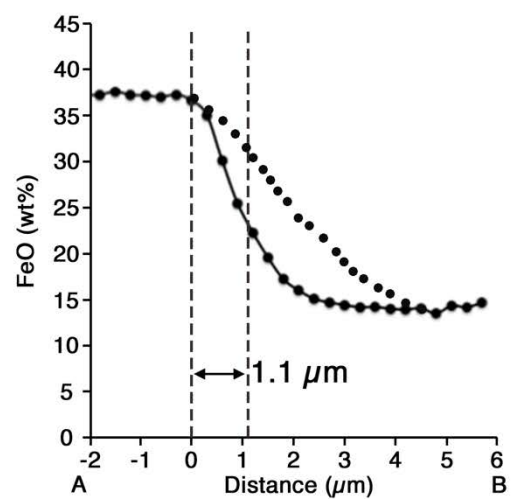
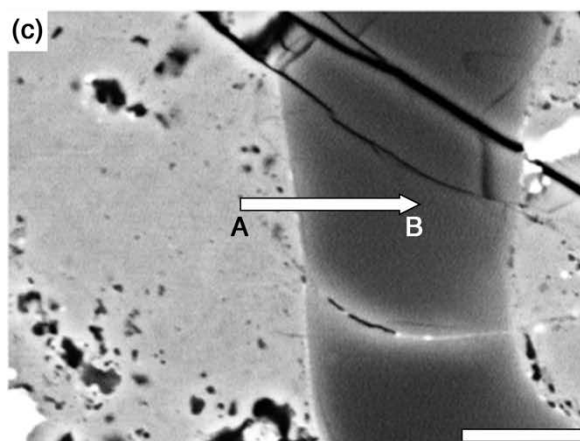
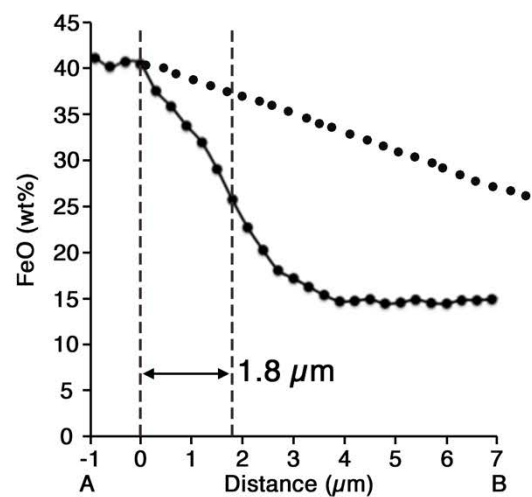
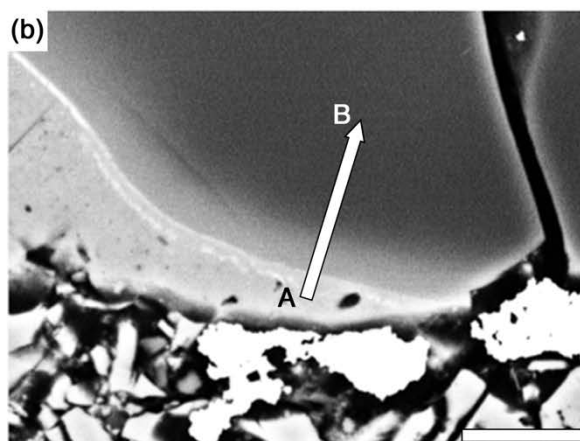
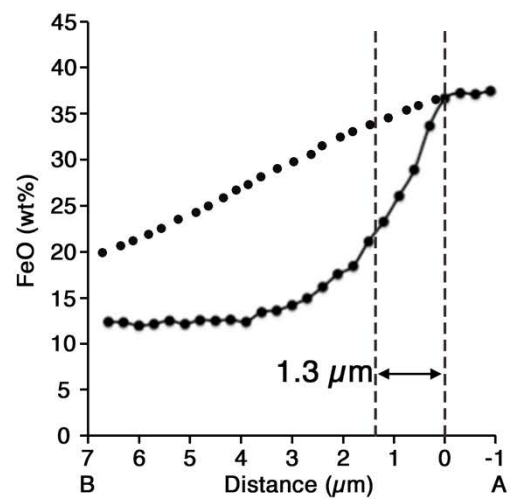
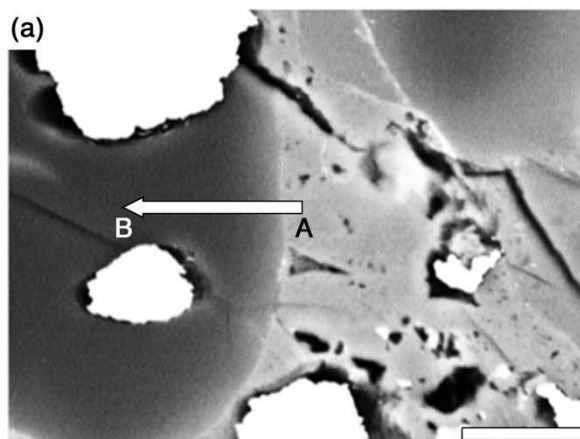
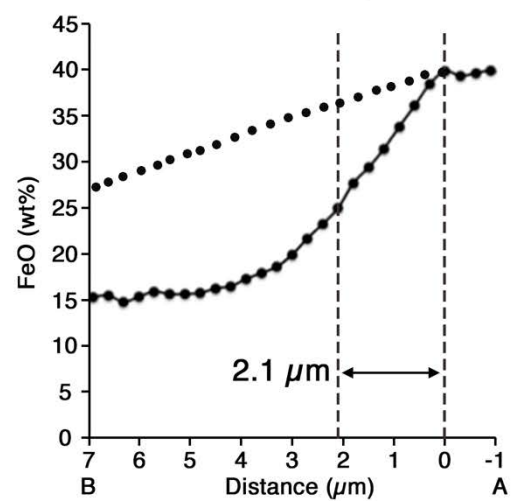
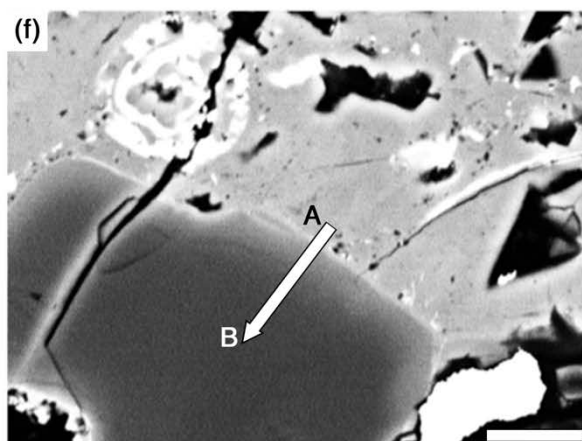
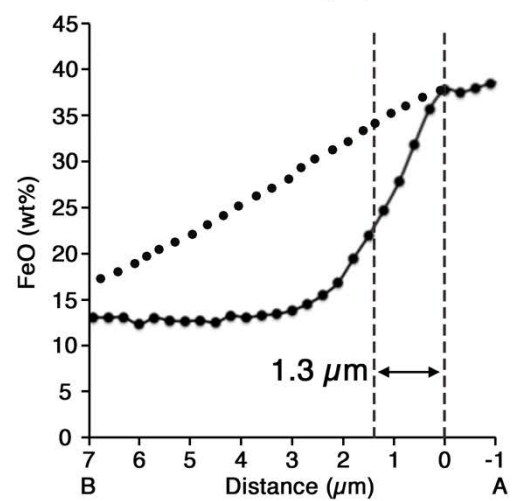
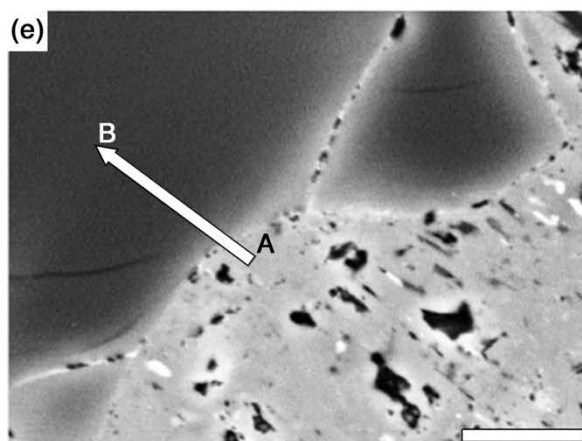
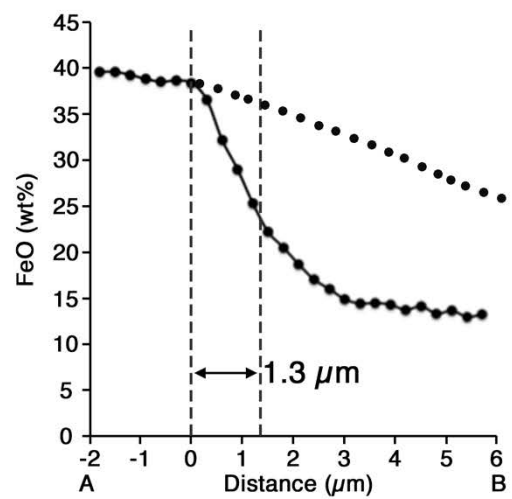
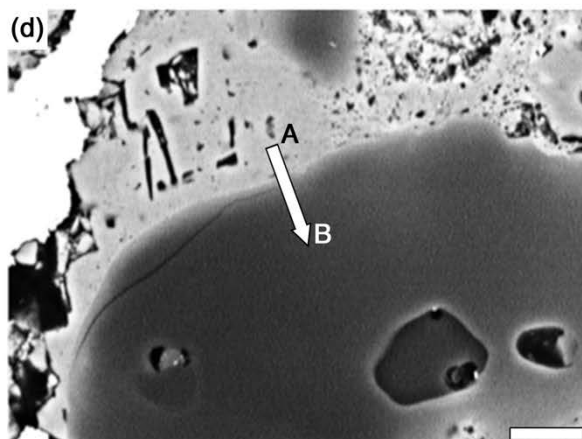


Fig. 3. Back-scattered electron images (a and b) and the EBSD images (c and d) of olivine grains in the igneous rim. Dotted lines represent grain boundaries. Scale bar is 10 μm . In (c), the EBSD images show that the FeO-rich olivine (FeO-rich ol) and MgO-rich olivine have the same crystallographic orientation.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





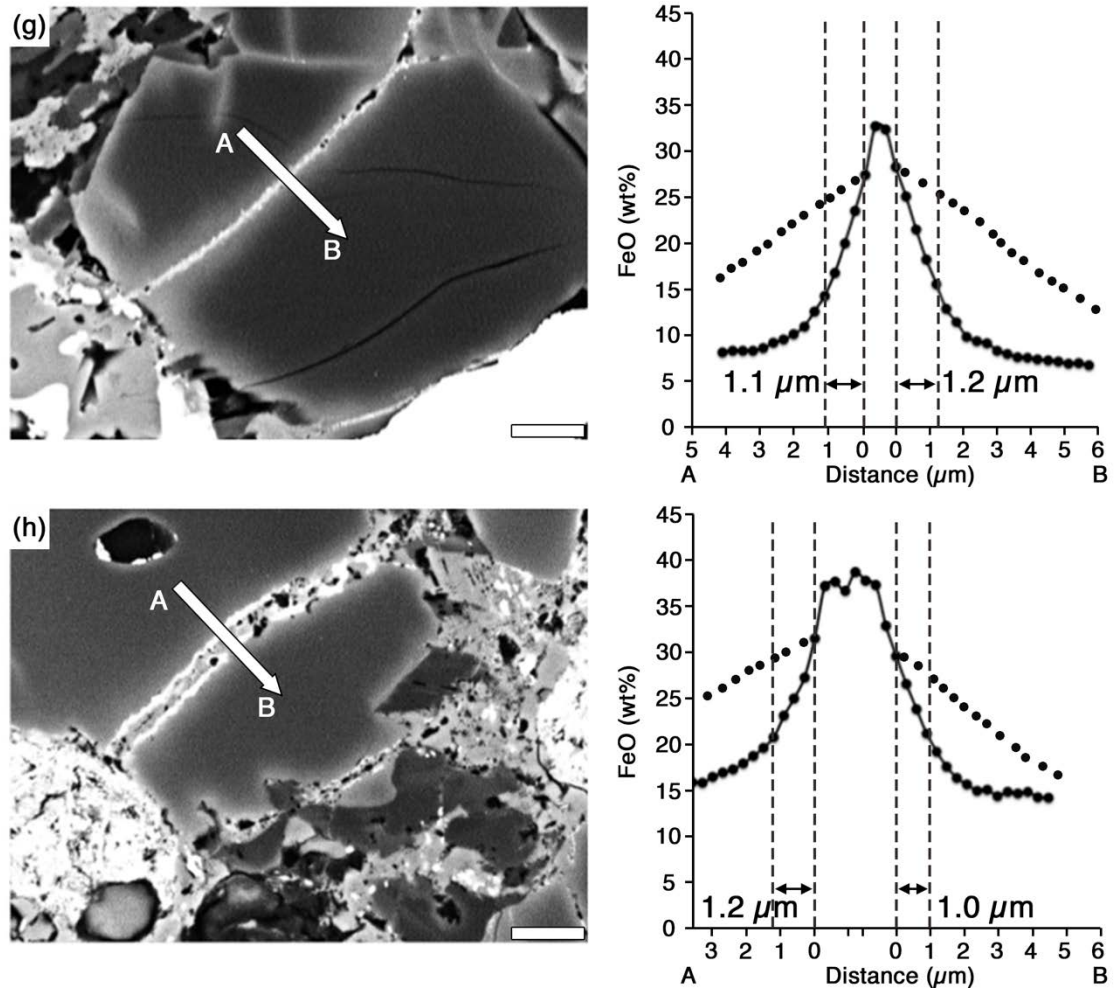


Fig. 4. (a–f) Compositional profiles along lines crossing the interfaces between FeO-rich olivine (A) and MgO-rich olivine (B). (g–h) Compositional profile across veins in MgO-rich olivine. Interdiffusion distances of the Fe-Mg zoning are shown in diagrams. Hypothetical interdiffusion profiles are shown 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 .

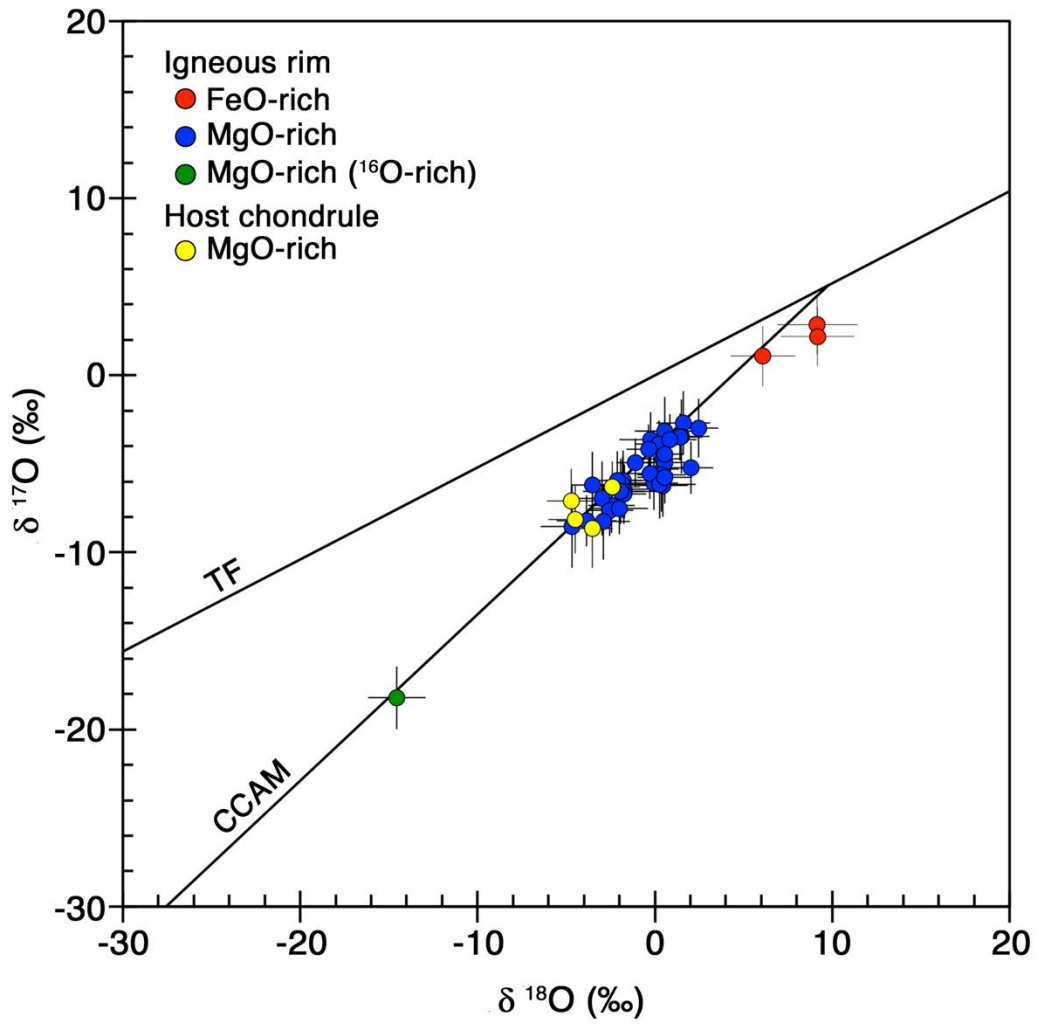


Fig. 5. Oxygen isotopic compositions of olivine in the igneous rim and the host chondrule. Error bar is 2σ TF: terrestrial fractionation line. CCAM: carbonaceous chondrite anhydrous mineral line.

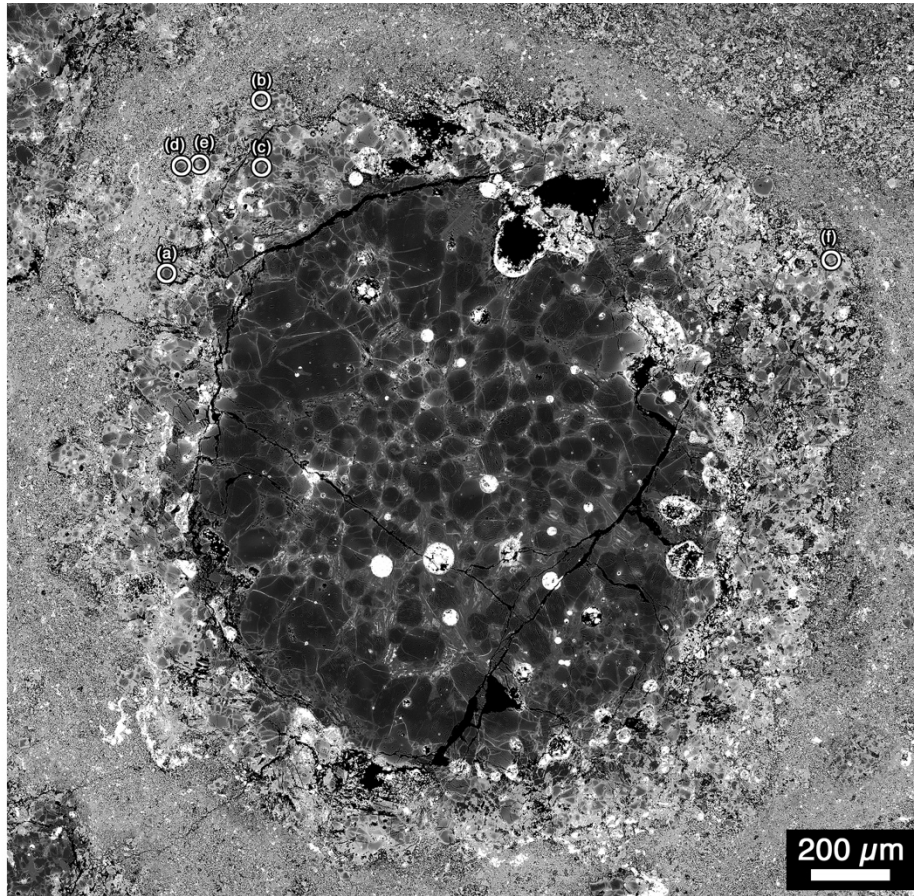


Fig. 6. Distribution of olivine grains containing the extreme ^{16}O -rich areas in the igneous rim. White circles show their positions. The magnified images at locations of circles labeled (a) to (f) show in Fig. 7 with the same labels.

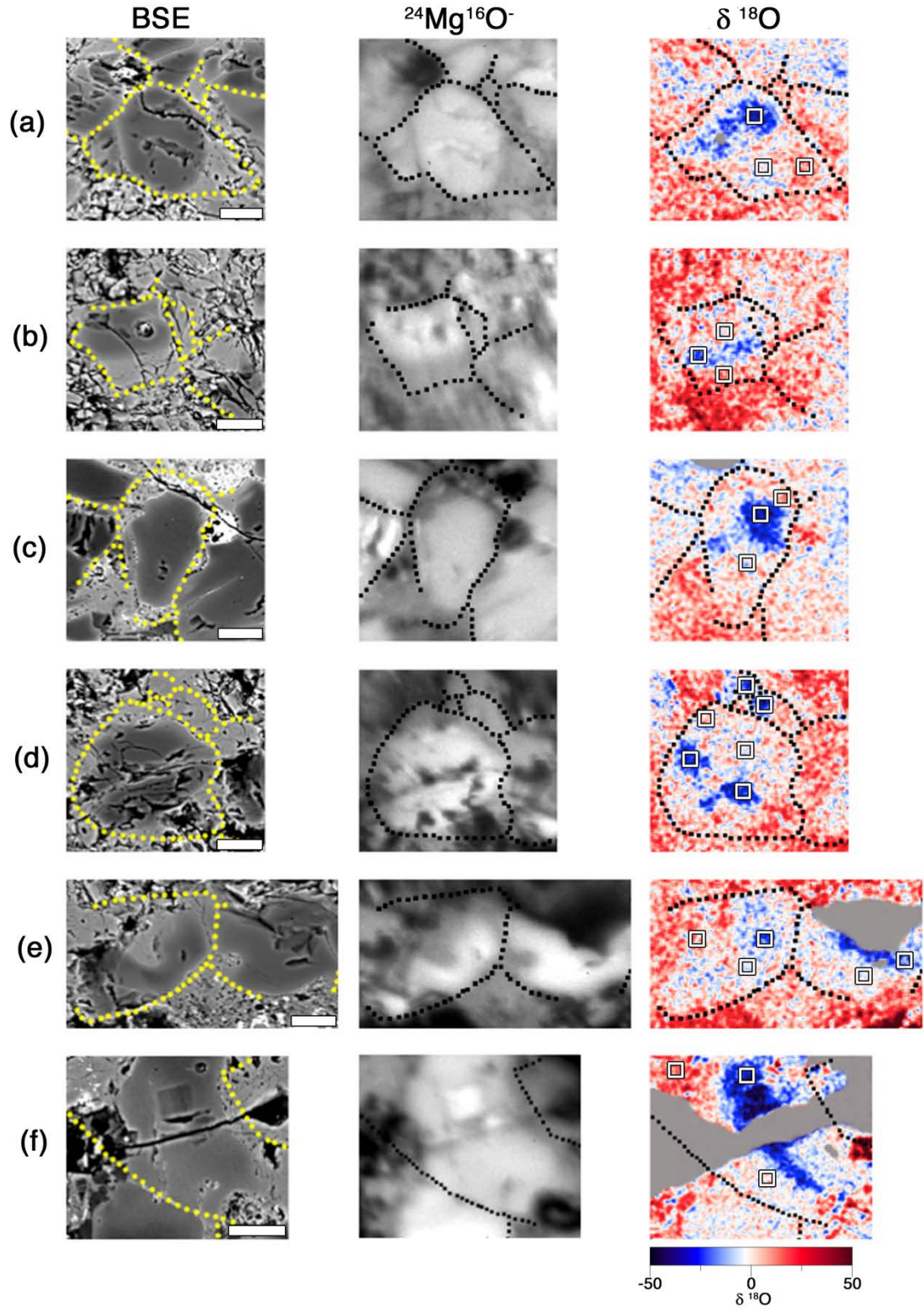


Fig. 7. BSE images, $^{24}\text{Mg}^{16}\text{O}^-$ and $\delta^{18}\text{O}$ isotopograph of olivine containing extreme ^{16}O -rich areas in the igneous rim. Dotted lines represent the grain boundaries determined by EBSD. Scale bar is 10 μm . Areas with artifacts are masked by the gray color. Squares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

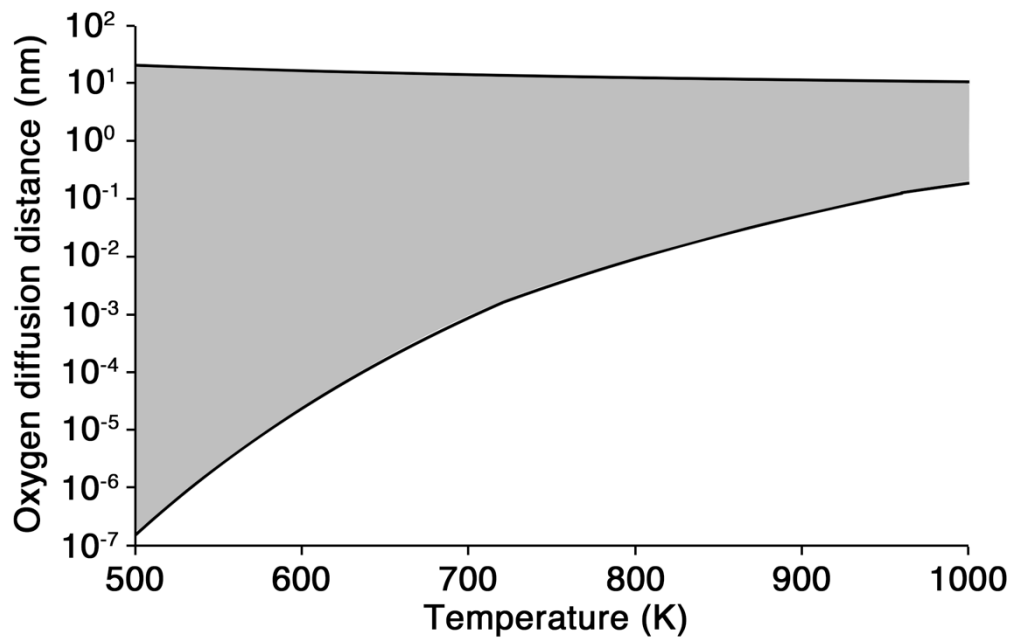


Fig. 8. Range of oxygen self-diffusion distance (nm) as a function of temperature (K) at Fe-Mg inter-diffusion distance of 1.3 μm .

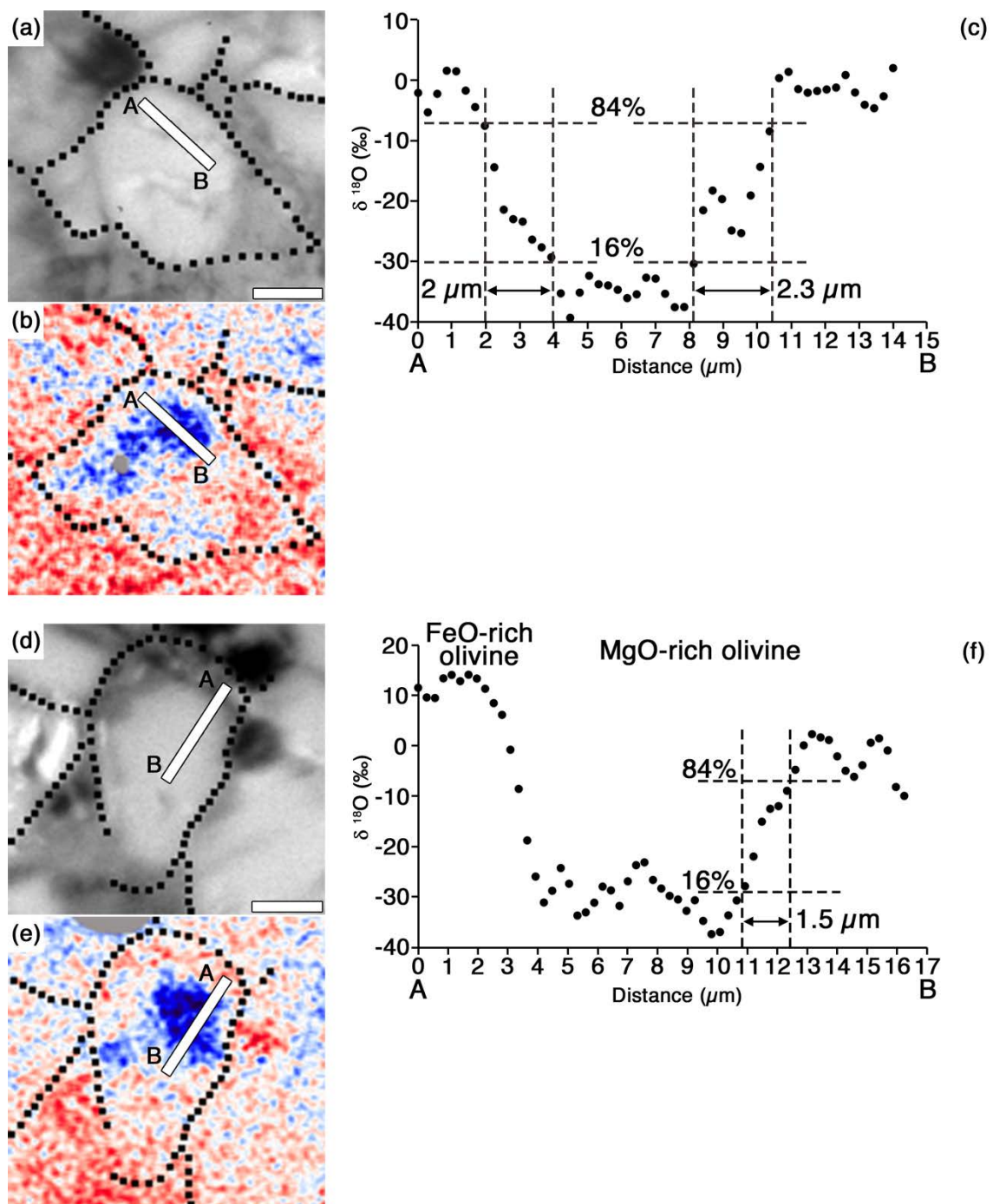


Fig. 9. (a and d) $^{24}\text{Mg}^{16}\text{O}^-$ isotopographs; (b and e) $\delta^{18}\text{O}$ isotopographs; (c and f) $\delta^{18}\text{O}$ profiles of lines A-B. Scale bar is 10 μm .

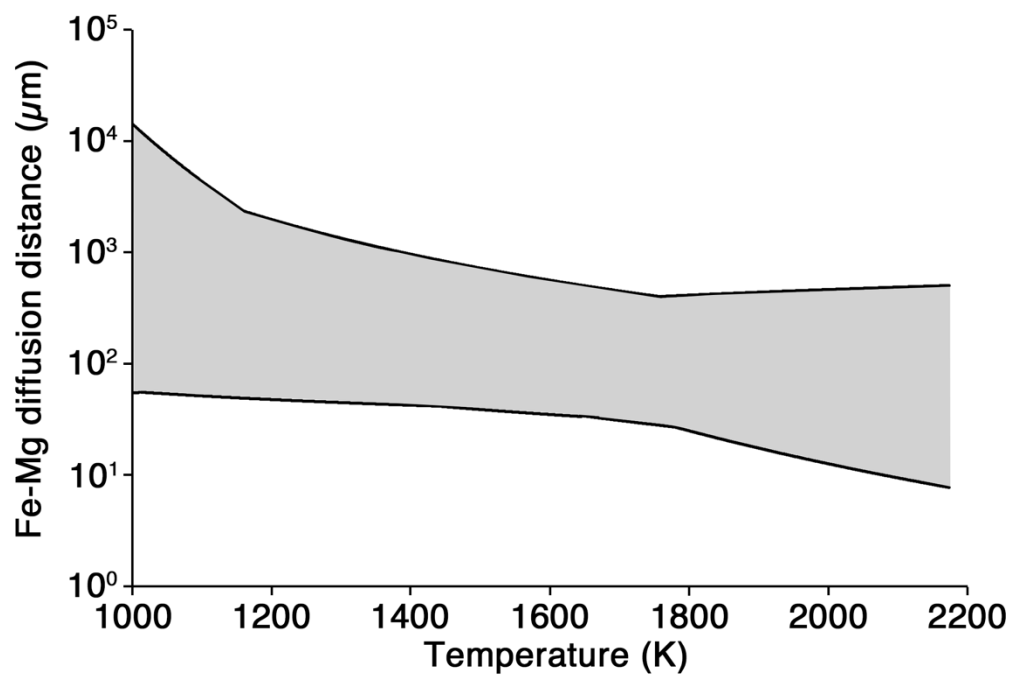


Fig. 10. Range of Fe-Mg interdiffusion distance (μm) as a function of temperature (K) at oxygen self-diffusion distance of 1 μm .

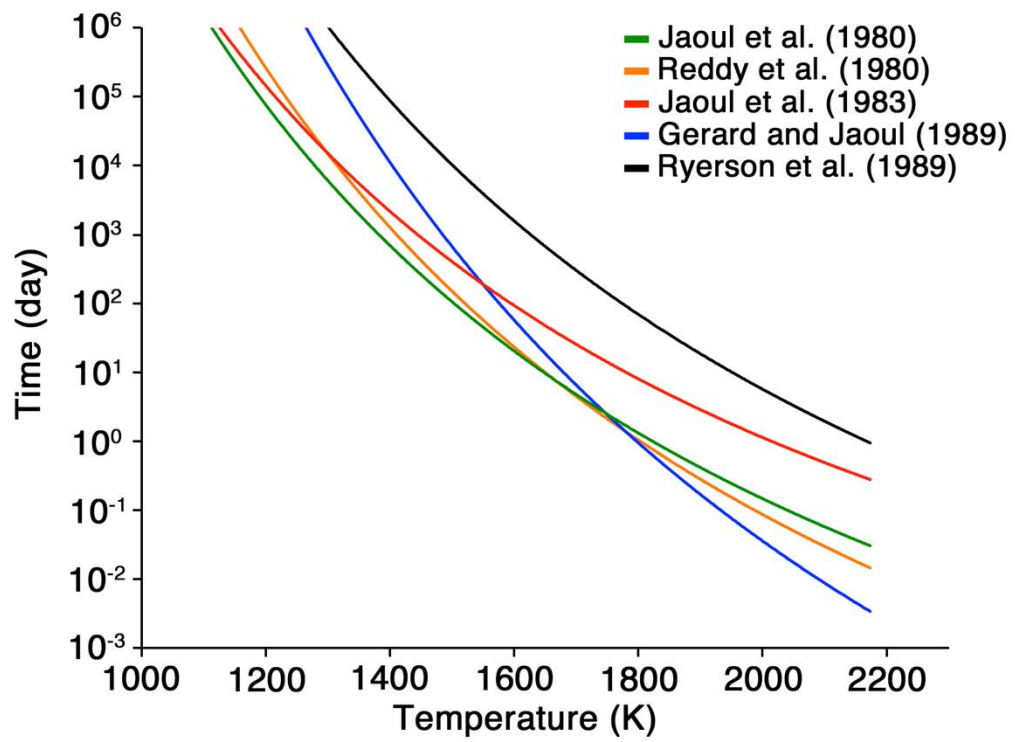


Fig. 11. Range of heating time (day) as a function of temperature (K) at oxygen self-diffusion distance of 1 μm .