

¹⁶O-POOR GROSSITE IN REFRACTORY INCLUSIONS FROM PRISTINE CO3.0 CHONDRITES. J.-D.

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Introduction: Thanks to its ubiquity in all phases in the early Solar System and its three stable isotopes, O is a very powerful tool to study the processes and conditions in the very early stages of Solar System evolution. An important challenge when studying refractory inclusions and other chondritic components is the disentangling of primary signatures and the effects of secondary processes, notably experienced on the meteorites parent body. To avoid this complication, it is important to study the least altered samples possible.

Calcium-, Aluminium-rich Inclusions (CAIs) containing the very first phases predicted to condense from a gas of Solar composition [1,2], such as corundum, hibonite, or grossite, have been mostly found in the aqueously altered CM chondrites but have also been discovered more recently in the least altered CO3.0 chondrites [e.g. 3]. The focus of this study is grossite bearing inclusions in the least altered CO chondrites, the paired Dominion Range (DOM) 08004 and DOM 08006 [e.g. 4], and Allan Hills (ALH) A77307.

While rarer than most CAI forming phases, grossite bearing inclusions have been found in several groups of carbonaceous chondrites [5], some of which are known to have undergone significant thermal and/or aqueous secondary processes. Studying such inclusions in the least altered samples available removes the complications linked to these secondary processes.

Samples and Methods: Grossite bearing inclusions were studied in one polished thin section each of DOM 08004 and DOM 08006, and in two thin sections of ALHA 77307. Each of the four studied thin sections was mapped with EDX and BSE using a Leo 1530 field emission electron microscope at the Max Planck Institute for Chemistry (MPIC) with an acceleration voltage of 15 kV and a magnification of typically 200 \times . Major element compositions were measured with a JEOL JXA 8200 electron microprobe at the University of Mainz using an acceleration voltage of 20 kV and a 20 nA beam with a probe size of 2 μ m.

Oxygen isotopic ratios were measured in-situ using the NanoSIMS 50 at MPIC. The primary current of the Cs $^+$ beam was 50 pA and the instrument was tuned to achieve a mass resolving power (MRP, Cameca NanoSIMS definition) of over 8000, sufficient to resolve 16 OH from 17 O. 16 O was measured using a Faraday cup while 17 O and 18 O were measured using electron multipliers. 30 Si was also monitored to ascertain that no drift to another phase occurred during analysis. A 5 μ m raster was used for pre-sputtering and analysis was done

with a 3 μ m raster. The pre-sputter and analysis times were 5 min and 6 min, respectively. All results were standardised to mineral standards of similar mineralogy to correct for matrix effects except for grossite that was standardised to Madagascar hibonite. 2σ errors quoted include external standard deviation of standards and internal precision for each analysis.

Results: 165 CAIs were characterised in DOM 08004, 6 of which contain grossite. In DOM 08006, 425 CAIs were found, including 24 grossite bearing inclusions. A total of 267 CAIs were found in the two sections of ALHA 77307 in this study, only 2 of which are grossite bearing inclusions.

So far, two inclusions found in DOM 08004 have been analysed for O isotopes in grossite: DOM18 and DOM139. DOM18 (Fig. 1) is $\sim 85 \times 75 \mu$ m in size, composed of a core of grossite with perovskite and surrounded by a rim of melilite and diopside in places. DOM139 (Fig. 2) is a subrounded object composed of grossite, perovskite, spinel and melilite and is $\sim 90 \times 85 \mu$ m in dimensions.

Grossite in these two inclusions show essentially stoichiometric compositions with averages and standard deviations of 76.47 ± 0.90 wt.% Al₂O₃ and 21.55 ± 0.14 wt.% CaO. MgO concentrations are below detection limits and FeO (≤ 0.62 wt.%) and TiO₂ (≤ 0.28 wt.%) concentrations are low.

Unlike other CAI forming phases such as hibonite, spinel, or diopside measured in inclusions from the same section, and even within the very same inclusions, that have 16 O-rich compositions typical of the least altered CAIs ($\Delta^{17}\text{O} = \sim -24 \text{ ‰}$), grossite in both DOM18 and DOM139 is 16 O-poor (Fig. 3), with values ranging from $\Delta^{17}\text{O} = -3.4 \pm 2.2 \text{ ‰}$ to $\Delta^{17}\text{O} = 2.5 \pm 2.2 \text{ ‰}$. These compositions plot within error of the CCAM [6], Y&R [7] and PCM [8] lines. While the use of Madagascar hibonite as a standard for grossite O isotopic compositions could introduce some disturbances by matrix effect, it is highly unlikely to be sufficient to explain the discrepancy of composition between the highly 16 O-rich minerals and the 16 O-poor grossite. Moreover, such matrix effects typically introduce mass fractionation effects, causing deviations from the mixing lines that are not observed here.

Similar results were previously reported from a CAI from Murchison (CM) [9], although this inclusion showed mass fractionation effects, and another found in the CO3 Yamato-81020 [10]. These results are in

contrast with other O isotopic analyses in CH [11] and an ungrouped [12] chondrites.

The results presented here suggest that alteration on the CO parent body is unlikely to be the origin of the ^{16}O -rich signature of grossite as DOM 08004 is amongst the least altered chondrites. This is supported by the composition of spinel in DOM139. It has very low concentration of FeO (≤ 0.43 wt.%) which tends to be higher in inclusions that have undergone alteration on the parent body (e.g. [13]). Thus, this O isotopic signature is most likely to have been acquired before the accretion of these inclusions to their parent body.

The O isotopic composition of spinel in DOM139 is indistinguishable from the main CAI composition with $\Delta^{17}\text{O} = -21.6 \pm 3.7\text{‰}$. This is noteworthy as spinel is predicted to condense after grossite [2] and has similar O isotopic composition to the ones found in phases predicted to form before grossite, such as hibonite found in other inclusions of this section. It however seems unlikely that the O isotopic composition of the environment in which these CAIs formed changed between the formation of hibonite and grossite before reverting to its original composition between grossite and spinel formation.

It appears that the ^{16}O -poor signature in grossite is much more widely spread than thought before, and the analysis of more grossite-bearing inclusions is required to better constrain the O isotopic reservoirs that played a role in the formation of grossite. Using additional isotopic systems such as Al-Mg or Ca-Ti could also be very useful in an attempt to define the timing and processes involved in this formation.

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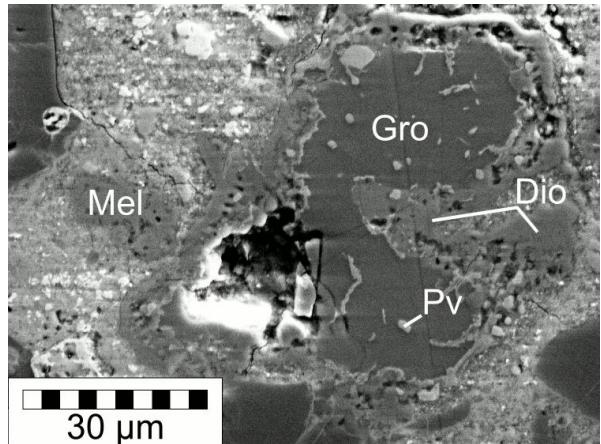


Figure 1. BSE image of DOM18. Dio: diopside, Gro: grossite, Mel: melilite, Pv: perovskite.

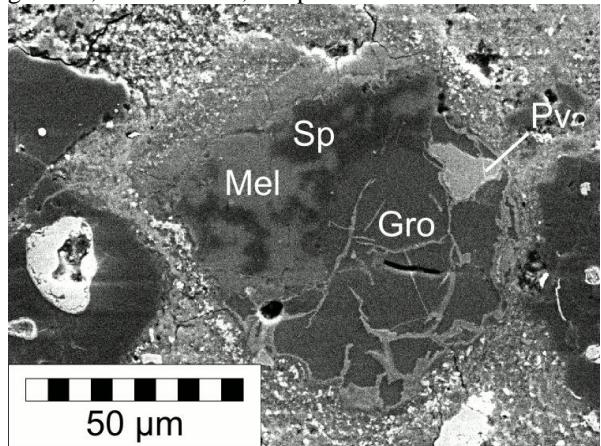


Figure 2: BSE image of DOM139. Gro: grossite, Mel: melilite, Pv: perovskite, Sp: spinel.

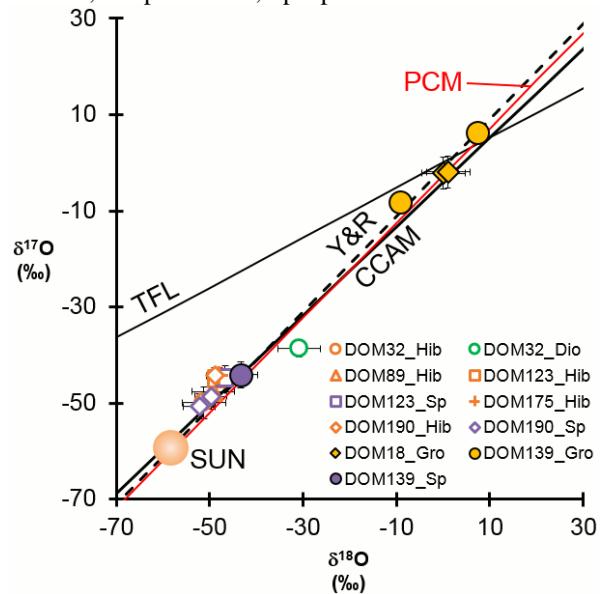


Figure 3: O isotopic composition of minerals from hibonite and grossite bearing CAIs from DOM 08004. Composition of the Sun from [14].