

LONG FORMATION PERIOD OF SINGLE CAI: COMBINATION OF O AND MG ISOTOPE DISTRIBUTION. M. Yoshitake and H. Yurimoto, Department of Earth and Planetary Sciences, Tokyo Institute of Technology, Ookayama, 2-12-1, Meguro, Tokyo 152-8551, Japan (miwa@geo.titech.ac.jp)

Introduction: Coarse-grained calcium, aluminum-rich inclusions (CAIs) are known to have large O-isotopic heterogeneities. The O-isotopic heterogeneities are mainly established by analyzing of the Allende meteorite (oxidized-CV3) [1]. Melilite and anorthite are usually ^{16}O -poor with $\delta^{17,18}\text{O} \sim 10$ permil, whereas spinel and fassaite are ^{16}O -rich with $\delta^{17,18}\text{O} \sim -40$ permil. CAIs are also known to have live- ^{26}Al up to $(^{26}\text{Al}/^{27}\text{Al})_0 = 5 \times 10^{-5}$, called as canonical value [2]. The canonical value has been mainly determined by ^{16}O -poor minerals of Allende CAIs, *i.e.*, melilite and anorthite. However, it is still controversial whether the variations of initial $(^{26}\text{Al}/^{27}\text{Al})_0$ in a CAI correspond to chronology of crystallization of the minerals or not because oxidized-CV3 meteorites moderately altered aqueously/thermally and the Mg-isotope ratios had might be disturbed during the alteration.

On the other hand, Reduced-type CV3 meteorites are less altered and more preserved original texture than the oxidized-CV3 meteorites [3]. Therefore, CAIs of the reduced-CV3 are preferable to discuss relationships between variations of O-isotopes and $(^{26}\text{Al}/^{27}\text{Al})_0$ ratios. We have reported a type B2 CAI from the Vigarano reduced-CV3, which minerals have generally the same O isotopic characteristics observed in typical Allende CAIs except for anorthite [4]. The anorthite shows O-isotopic variations from $\delta^{17,18}\text{O} \sim -40$ to ~ 0 permil, which are comparable to the range observed among other minerals of the CAI. High precision isotope ratio micrographs (isotopographs [5]) revealed that the variations of anorthite grains were composed of two distinct O-isotope ratios [6]. Core of the anorthite is composed of ^{16}O -rich composition ($\delta^{17,18}\text{O} \sim -40$ permil) and thin ^{16}O -poor rims ($\sim 5\mu\text{m}$, $\delta^{17,18}\text{O} \sim 0$ permil) are directly attached on the core. The O-isotopic compositions abruptly changed at the boundary of ^{16}O -rich and ^{16}O -poor regions suggesting overgrowth of ^{16}O -poor anorthite from melt. The sharpness of the ^{16}O -rich/poor boundary is monitored as $3\mu\text{m}$ comparable to total aberration of the isotopograph ($\sim 2\mu\text{m}$).

We report here Mg-isotope distributions in melilite and anorthite in the type B2 CAI from the reduced-CV3 meteorite previously determined in O-isotope distributions.

Experimental procedure: The sample used in this study is a coarse-grained type B2 CAI (TTV1-01 [4]) found in a thick section from Vigarano CV3

chondrite. The petrological, mineralogical and O-isotopic studies of this CAI have been executed [4, 6].

Magnesium isotopic compositions of individual minerals have been measured by the TiTech CAMECA ims 1270 SIMS instrument. The primary ion beam was mass filtered positive ^{16}O ions accelerated to 23keV and the beam spot size was $\sim 5\mu\text{m}$ in diameter. Positive secondary ions of the ^{24}Mg , ^{25}Mg , ^{26}Mg , and ^{27}Al were analyzed by automatic peak jumping at a mass resolution power of ≥ 4000 , sufficient to completely eliminate hydride interferences at mass 25. Typical primary ion current was 30 - 60 pA in order to obtain the count rate of positive ^{27}Al or ^{24}Mg ion of $\sim 4 \times 10^5$ cps for the each analyses. In order to correct instrumental mass fractionation, Russian spinel (SPU), Miyake-jima anorthite, and synthetic åkermanite glass [7] were used as standard.

Results: Fig. 1 (a) shows ^{26}Al - ^{26}Mg systematics for anorthite of TTV1-01. Although O-isotopic compositions in each anorthite grain have two distinct values, *i.e.*, ^{16}O -rich in the center and ^{16}O -poor in the margin [6], no clear excesses of ^{26}Mg have been observed in the both distinct O-isotopic regions. The initial $(^{26}\text{Al}/^{27}\text{Al})_0$ ratio for all plots is calculated to be $-0.4 \pm 3.5 \times 10^{-6}$ ($\pm 2\sigma$) assuming zero intercept.

Fig. 1 (b) shows ^{26}Al - ^{26}Mg systematics for melilite. O-isotopic compositions of melilite are ^{16}O -poor [4]. The initial $(^{26}\text{Al}/^{27}\text{Al})_0$ are $4.5 \pm 1.1 \times 10^{-5}$ ($\pm 2\sigma$) assuming zero intercept, corresponding to the canonical value of the solar nebula [2].

Discussion: The ^{26}Mg - ^{26}Al systematics for melilite show that this CAI initially contained live- ^{26}Al of the canonical value. In order to reduce this live- ^{26}Al to the value observed in the anorthite, the anorthite crystallized at least 2.6 Myr (2σ) later if the isochron for anorthite has not been reset after solidification. Therefore, it is essential to determine the reset condition of the anorthite isochron by secondary disturbances after anorthite solidification.

After anorthite solidification, disturbance of isotope distribution can be thermally driven by solid-state diffusion of atoms. Because diffusion coefficients of atoms in anorthite have been determined [8, 9, 10] and O-isotope zoning-texture in anorthite has been precisely determined [6], we can evaluate diffusion distances of Mg-isotopes. Anorthite grains of TTV1-01 consist of ^{16}O -rich core surrounded by ^{16}O -poor

rim [5]. The sharpness of the boundary between ^{16}O -rich and ^{16}O -poor regions is about $3\text{ }\mu\text{m}$ comparable to spatial resolution of the isotopographs. Therefore, the diffusion distance of O-isotopes in the anorthite is defined as smaller than $1.5\text{ }\mu\text{m}$. Diffusion distances of Mg-isotopes is calculated as a function of temperature or time under this O-diffusion distance (Fig. 2). The curves in the diagram correspond to maximum diffusion distances of Mg-isotopes under the diffusion coefficients used. Temperature conditions higher than 1260°C cannot be accepted because anorthite crystals in the type B CAI composition would be melted [11]. Below the anorthite melting temperature, Mg-isotopes cannot diffuse across the anorthite grains because average size of the anorthite is $\sim 50\text{ }\mu\text{m}$ in width of lath shaped grain or in diameter of blocky grain. Therefore, even if the anorthite initially had live- ^{26}Al as the case of the melilite in the same CAI, it is difficult to erase the ^{26}Mg excesses by late stage heating processes after anorthite solidification.

The absence of clear ^{26}Mg -excesses (the initial $(^{26}\text{Al}/^{27}\text{Al})_0$ ratio of $-0.4 \pm 3.5 \times 10^{-6}$ (2σ)) in the anorthite strongly suggests that live- ^{26}Al had been disappeared in the CAI when the anorthite crystallized. Observed difference of the initial $(^{26}\text{Al}/^{27}\text{Al})_0$ ratio between melilite and anorthite in the CAI indicates that the anorthite crystallization was much later than the melilite crystallization, i.e., the CAI formation took place over a time of at least 2.6 Myr (2σ).

References: [1] Clayton R. N. (1993) *Ann. Rev. Earth Planet. Sci.* **21**, 115-149. [2] MacPherson et al. (1995) *Meteoritics*, **30**, 365-386. [3] Lee et al. (1996) *MAPS* **31**, 477-483. [4] Yoshitake M. et al. (2005) *GCA in press*. [5] Yurimoto H. et al. (2003) *ASS*, 203-204, 793-797. [6] Nagashima et al. (2004) *Chondrites and the Protoplanetary Disk conf.* 153-154. [7] Koike et al. (1993) *Proc. NIPR Symp. Antarct. Meteorites* **6**, 357-363. [8] Elphick S. C. et al. (1988) *Contrib. Mineral. Petrol.*, **100**, 490-495. [9] Ryerson F. G. and McKeegan K. D. (1994) *GCA* **58**, 3713-3734. [10] LaTourrette T. and Wasserburg G. J. (1998) *EPSL*, **158**, 91-108. [11] Stolper E. (1982) *GCA* **46**, 2159-2180.

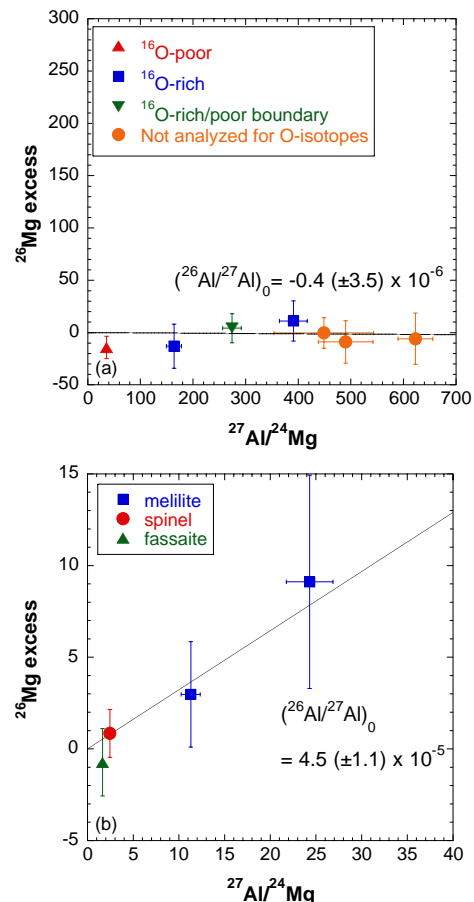


Fig. 1 Al-Mg evolution diagram for TTV1-01 type B2 CAI from the Vigarano meteorite. (a) anorthite, (b) melilite.

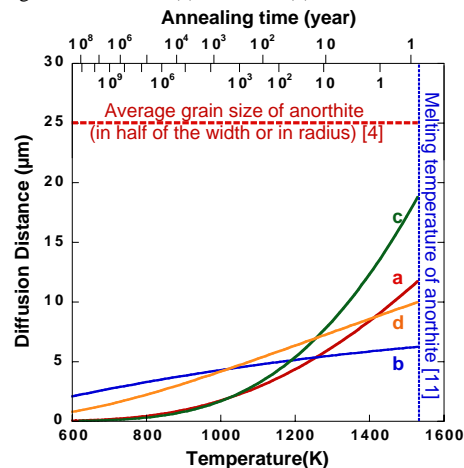


Fig. 2 Correlations between diffusion distances of Mg-isotopes and the annealing temperature/time. The diffusion distances were calculated under the diffusion distances of O-isotopes of $1.5\text{ }\mu\text{m}$ observed in anorthite grains of TTV1-01. Each line has been calculated by combinations between self-diffusion coefficients of O and of Mg in anorthite: a, [7] and //b axis [9]; b, [8] and //b axis [9]; c, [7] and //c axis [9]; d, [8] and //c axis [9], respectively. Upper and Lower scale of annealing time were calculated by self-diffusion coefficients of Oxygen of [9] and [8], respectively.