

CONSTRAINTS ON THE ORIGIN OF MAGNESIAN CHONDRULES AND ON THE GASEOUS RESERVOIRS IN THE EARLY SOLAR SYSTEM: AN OXYGEN ISOTOPIC STUDY. G. Libourel^{1,2}, M. Chaussidon¹, A.N. Krot³, ¹CRPG-CNRS 15, Rue Notre-Dame des Pauvres, BP20, 54501 Vandoeuvre les Nancy, France, ²ENSG-INPL, BP40, 54501 Vandoeuvre les Nancy, France (libou@crpg.cnrs-nancy.fr), ³Hawai'i Institute of Geophysics and Planetology, School of Ocean and Earth Science and Technology, University of Hawai'i at Manoa, Honolulu, HI 96822, USA.

Introduction: The origin of chondrules, which constitute one of the major high-temperature components of primitive meteorites (chondrites), remains enigmatic [1,2]. Although the mineralogy, textures and chemical compositions of chondrules suggest that they formed by incomplete melting of solid precursors during localized, brief, repetitive heating events, possibly by shock waves, in the protoplanetary disk, the nature of chondrule precursor materials and open- vs. closed-system behavior of chondrule melts are still controversial [3,4].

It has been recently suggested that gas-melt interactions played an important role during the formation of magnesium-rich (Fa and Fs <5 mol%), porphyritic (type I) chondrules [5-7] and that coarse-grained aggregates of magnesian olivine and Fe, Ni-metal (with or without spinel) were among the major solid precursors of type I chondrules [8]. These olivine aggregates (Fig. 1a) have granoblastic textures which might imply that they are fragments of thermally processed planetesimals.

Based on these observations, Libourel and Krot [8] proposed that i) typical type I chondrules are complex objects composed of an inherited Fe, Ni-metal bearing dunite-like aggregates that progressively react with an igneous component (glass, low-Ca pyroxene, high-Ca pyroxene, and \pm silica phase) formed during the chondrule melting episode and equilibrated with the nebular gas, and (ii) most of the chemical variability of type I chondrules could have resulted from variations in the extent of chemical exchange between the chondrule melt and nebular gas during chondrule-forming event(s).

If gas-melt exchange played an important role during chondrule formation, it may have also left specific traces in the oxygen isotopic compositions of chondrule minerals (if chondrule precursors and nebular gas had different oxygen isotopic compositions). Therefore, the two major predictions which can be made for oxygen isotopes from this model of chondrule formation are that (i) isotopic disequilibrium is expected in an individual chondrule between "relict" olivines and phases derived from chondrule melts (pyroxene and glassy mesostasis) and (ii) a common isotopic component should be detectable in all chondrules if they incorporated SiO from the same nebular gas. In order to test these predictions, we studied the oxygen isotopic systematics of the major silicate phases (olivine, pyroxene, mesostasis, and silica phase) of type I chondrules from CV and CR chondrites at $\approx 20 \mu\text{m}$ scale ($\approx \text{ng}$ scale) using *in situ* ion microprobe measurements.

Analytical techniques: The O isotopic compositions were measured on the CRPG-CNRS ims 1270 ion microprobe in multi-collection mode. Analytical procedure for O isotope measurements is described in [9, 10]. Here, we report *in situ* ion microprobe analyses of the oxygen isotopic composition of the major silicate phases (olivine, low-Ca pyroxene, silica and mesostasis) of 37 magnesian type I porphyritic chondrules from CV (Vigarano USNM 477-2, Vigarano UH5, Mokoia and Efremovka) and CR (EET 92042, EET 92147, EET 87770, El Djouf 001, MAC 87320 and GRA 95229) carbonaceous chondrites. This represents a total of 225 individual mineral analyses.

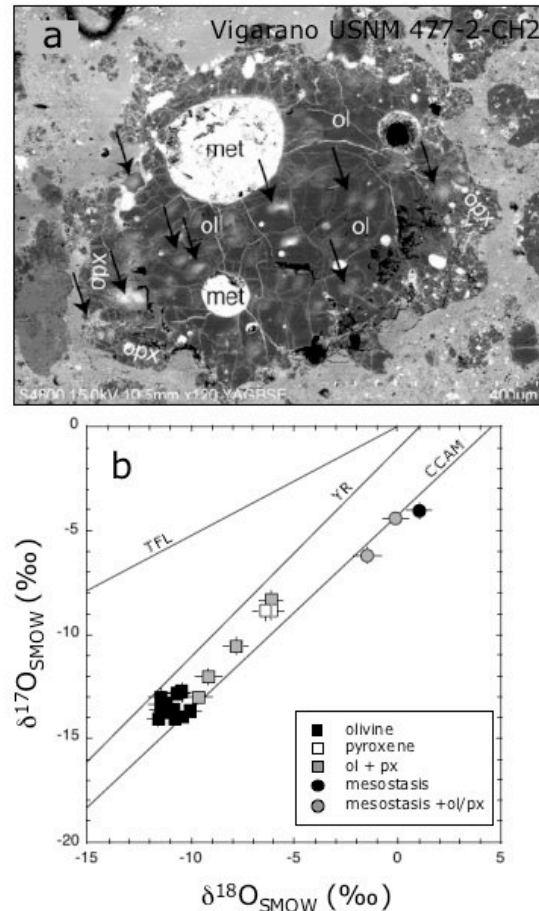


Fig. 1. a) Backscattered electron image of chondrule-Ch2 from Vigarano containing coarse-grained aggregates composed of Fe,Ni-metal (met) and magnesian olivines surrounded by the igneous shell composed of low-Ca pyroxene and glassy mesostasis. Some olivine-olivine grain boundaries in the aggregate lack any layer of glassy mesostasis and show triple junctions with granoblastic equilibrium textures [8]. b) Twenty *in situ* ion microprobe spot analyses were performed in this chondrule (arrows in Fig.1a). Notice the homogeneity of the olivine oxygen isotopic composition and the ^{16}O -depleted composition of the low-Ca pyroxenes relative to olivines.

Results: In spite of significant variations of the modal proportions of major mineral phases in CR and CV chondrules, we show that the same isotopic characteristics are observed: (i) olivines are isotopically homogeneous at the ‰ level within a chondrule although they may vary significantly from one chondrule to another, (ii) low-Ca pyroxenes are also isotopically homogeneous but systematically ^{16}O -depleted relative to olivines of the same chondrule and (iii) all chondrule minerals

analyzed show ^{16}O -enrichments relative to the terrestrial mass

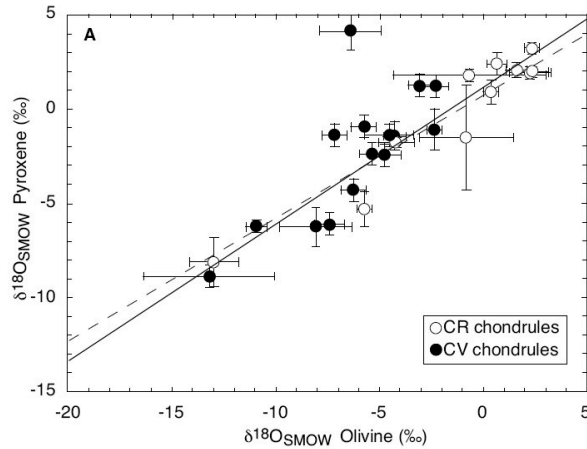


Fig. 2. Relationships between the average oxygen isotopic compositions ($\delta^{18}\text{O}$) of olivine and pyroxene for chondrules where these two phases could be analysed. Solid and open symbols are for chondrules from CV chondrites and CR chondrites, respectively. Note the strong positive correlations shown by all chondrules (with the exception of two) between pyroxene and olivine for either $\delta^{18}\text{O}$ or $\delta^{17}\text{O}$ values. The solid lines are the regression lines calculated from the chondrule data (excluding the two chondrules which don't plot on the correlations): slope of 0.768 ± 0.120 and intercept of $1.15 \pm 0.52\text{‰}$ for $\delta^{18}\text{O}$ values, slope of 0.774 ± 0.098 and intercept of $0.30 \pm 0.55\text{‰}$ for $\delta^{17}\text{O}$ values. The dashed lines show the "theoretical" correlation lines of slope 0.66 predicted by the model of gas-melt interaction. The observed and predicted correlation lines agree within uncertainties.

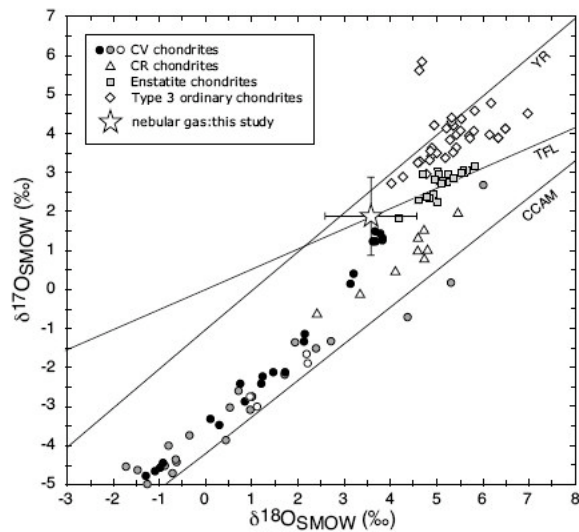


Fig. 3. Three oxygen isotope diagram for bulk chondrules from various chondrite class (because of the choice of the scale some ^{16}O -rich chondrules from CV chondrites are not shown). Data for CV chondrites are from [11], for solid dots, from [12] for grey dots and from [13] for open dots. Data for E chondrites are from [14] for CR chondrites from [15] and for type 3 ordinary chondrites from [16]. Note that the gaseous reservoir ($\delta^{18}\text{O}_{\text{gas}} \sim 3.6 \pm 1\text{‰}$ and $\delta^{17}\text{O}_{\text{gas}} \sim 1.8 \pm 1\text{‰}$) inferred from the correlation between olivine and pyroxene oxygen isotopic compositions plots at the convergence of the trends defined by bulk chondrules from carbonaceous, enstatite and ordinary chondrites. This composition is also very close to that of the reservoir required to explain the slope 1 correlation in unaltered CAI minerals (Young and Russel line).

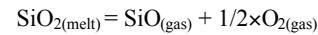
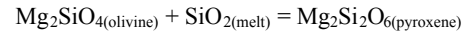
fractionation line which are decreasing from olivine (\pm spinel) to low-Ca pyroxene and to silica and mesostasis. The observation that, in most of the type I chondrules studied, the coexisting olivine and pyroxene crystals and glassy mesostasis have different oxygen isotopic compositions (Fig. 1b) implies that the olivine and pyroxene grains are not co-magmatic and that the glassy mesostasis is not the parent liquid of the olivine.

The $\delta^{18}\text{O}$ (Fig. 2) and $\delta^{17}\text{O}$ values of pyroxene and olivine appear to be strongly correlated for all the studied CR and CV chondrules according to :

$$\delta^{18}\text{O}_{\text{(pyroxene)}} = 0.72 \left({}^{+0.13}_{-0.16} \right) \times \delta^{18}\text{O}_{\text{(olivine)}} + 1.21 \left({}^{+0.79}_{-0.54} \right)$$

$$\delta^{17}\text{O}_{\text{(pyroxene)}} = 0.81 \left({}^{+0.13}_{-0.15} \right) \times \delta^{17}\text{O}_{\text{(olivine)}} + 0.60 \left({}^{+0.95}_{-1.00} \right)$$

These correlations are consistent with mass balance of oxygen isotopes calculated in the case of chondrule formation according to the following gas-melt-mineral reactions:



These equations imply that 2/3 of the oxygen incorporated by low-Ca pyroxene during its crystallization in the melt is derived from the dissolution of precursor olivine and 1/3 from the addition of chemical components such as SiO from the nebular gas. The composition of the nebular gaseous reservoir in which type I CR and CV chondrules formed is determined from the present data set to be $\delta^{18}\text{O}_{\text{(gas)}} \approx 3.6 \pm 1\text{‰}$ and $\delta^{17}\text{O}_{\text{(gas)}} \approx 1.8 \pm 1\text{‰}$. This isotopic composition (Fig. 3) is in excellent agreement with that of the gaseous reservoir previously proposed from the oxygen isotope systematics of bulk chondrules from various class (i.e., CV, CR, UOC, EH) of chondrites.

Conclusions: The oxygen isotopic systematic strongly supports the idea that most of the olivines in type I chondrules are relict grains and that gas-melt interaction played a major role in the chondrule crystallization history during which the O-isotopic composition of the melt was continuously modified by exchange with the surrounding nebular gas. Detailed geochemical and chronological studies are now needed for disentangling this complex history and for shedding light on both the nature of the relict olivine precursors, i.e., recycling of previous generation of chondrules or of differentiated planetesimals, and on the existence of a nebular gaseous reservoir, its persistence and its evolution over the first million years of the solar protoplanetary disk.

References: [1] R.H. Hewins (1997) *Rev. Earth Planet. Sci.* 25, 61. [2] B. Zanda (2004) *EPSL* 224, 1. [3] S.J. Desch, H. C. Connolly, Jr. (2002) *MAPS* 37, 183. [4] R.H. Hewins et al. *In: Chondrites and the Protoplanetary Disk, ASP Conferences Series* 341, 286. [5] L. Tissandier et al., (2002) *MAPS* 37, 1377. [6] A.N. Krot et al. (2004) *MAPS* 39, 1931. [7] G. Libourel et al. (2006) *EPSL* 251, 232. [8] G. Libourel et A.N. Krot *EPSL* 254, 1. [9] A. Gurenko and M. Chaussidon (2002) *EPSL* 202,63-79. [10] A.N. Krot et al. (2006) *GCA* 70, 767. [11] Clayton et al. (1983) *In: Chondrules and Their Origins. Lunar Planetary Institute, Houston*, 37. [12] R.H. Jones et al. (2004) *GCA* 68, 3423. [13] Rubin et al. (1983) *EPSL* 96, 247. [14] R.N. Clayton, T.K. Mayeda *GCA* 60, 1999. [15] Weisberg et al. (1983) *GCA* 57, 1567. [16] Clayton et al. (1991) *GCA* 55, 2317.