

### Oxygen isotope systematics of porphyritic chondrules and their fragments in CH and CB chondrites.

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**Introduction:** Chondrules in carbonaceous chondrites including non-porphyritic chondrules in CH and CB chondrites show systematic trends that the  $\Delta^{17}\text{O}$  ( $= \delta^{17}\text{O} - 0.52 \times \delta^{18}\text{O}$ ) values increase with decreasing Mg# (= mol.% MgO/[MgO+FeO]) [e.g., 1-2], which are explained by an addition of  $^{16}\text{O}$ -poor water ice as an oxidant to the  $^{16}\text{O}$ -rich anhydrous precursors [e.g., 3]. In contrast, porphyritic chondrules in CH and CB chondrites show no systematic difference in the  $\Delta^{17}\text{O}$  values between type I and II chondrules ( $-4.7\%$  to  $+4.1\%$  vs.  $-2.1\%$  to  $+2.7\%$ ; [2,4]), though the detailed relationship with Mg# is unclear. Here we report new SIMS oxygen isotope analyses of olivine and pyroxene fragments in the Asuka-881020 CH chondrite ( $15\mu\text{m}$  Cs<sup>+</sup> primary beam, WiscSIMS). We further discuss the  $\Delta^{17}\text{O}$ -Mg# systematics of the porphyritic chondrules in CH and CB chondrites.

**Results and discussion:** Thirteen of 20 fragments are FeO-poor with Mg# of 90.7 – 99.3 and others are FeO-rich with Mg# of 50.5 – 80.0. The oxygen isotope ratios plot along with the primitive chondrule mineral (PCM) line [1] with  $\Delta^{17}\text{O}$  values from  $-5.0\%$  to  $+3.2\%$ . These data are compared to the average  $\Delta^{17}\text{O}$  values of the individual porphyritic chondrules (excluding relict grains; [2,4]) in  $\Delta^{17}\text{O}$  versus Mg# plot in Fig. 1a. The  $\Delta^{17}\text{O}$  ranges of FeO-poor and -rich fragments overlap those of the type I and II chondrules in CH and CB chondrites, respectively, suggesting that the olivine and pyroxene fragments are fragments of the porphyritic chondrules. Skeletal olivine-pyroxene chondrules may not be a source of the fragments, as they have a narrow range of  $\Delta^{17}\text{O}$  values of  $-2.6 \pm 1.4\%$  [5].

The  $\Delta^{17}\text{O}$  values of the type I chondrules and FeO-poor fragments increase with increasing Mg#, while those of the type II chondrules and FeO-rich fragments show no correlation with Mg# (Fig. 1a). Similarly, type II chondrules in CR and Tagish Lake-type chondrites have  $\Delta^{17}\text{O}$  variations from  $\sim -3\%$  to  $+2\%$  with no correlation with Mg# [6-7], which are explained by an addition of  $^{16}\text{O}$ -poor water ice to the  $^{16}\text{O}$ -rich precursors [e.g., 3]. However, the positive  $\Delta^{17}\text{O}$ -Mg# correlation for chondrules and fragments with Mg# > 96 cannot be explained by this process, but may be explained by addition of  $^{16}\text{O}$ -poor reductant to the  $^{16}\text{O}$ -rich precursors or addition of  $^{16}\text{O}$ -rich oxidant (water ice) to the  $^{16}\text{O}$ -poor precursors. The latter case is less likely, as chondrules with Mg# of  $\sim 99$  and  $\Delta^{17}\text{O}$  of  $\sim +4\%$  are not observed in other chondrites and oxides that inherited oxygen isotope ratios of water ice are  $^{16}\text{O}$ -poor [e.g., 8]. In the former case, chondrules with Mg# of  $\sim 96$  and  $\Delta^{17}\text{O}$  of  $\sim -4\%$ , which are the lower end of the positive  $\Delta^{17}\text{O}$ -Mg# correlation, may correspond to the  $^{16}\text{O}$ -rich precursors. Such chondrules are rarely observed in Acfer 094 (ungr. C3.0) and CR chondrites [1,9]. The  $^{16}\text{O}$ -poor reductant may be carbon-rich organics, given the two observations;  $^{16}\text{O}$ -poor organics in primitive chondrites [10] and heating experiments that produced Fo<sub>99</sub> olivine from Fo<sub>94</sub> olivine and graphite or diamond [11]. Heating experiments produced chondrules with  $\delta^{18}\text{O}$  of  $+5.6\%$  from type I chondrules with  $\delta^{18}\text{O}$  of  $+6.5\%$  and 5 wt% graphite [12], indicating oxygen isotope mass fractionation between the chondrules and oxidized carbon (CO or CO<sub>2</sub>). Fig. 1b shows the deviation of  $\delta^{18}\text{O}$  values from the PCM lines ( $\Delta^{18}\text{O}_{\text{PCM}}$ ) for the type I chondrules and FeO-poor fragments along with the  $\Delta^{17}\text{O}$  values. The chondrules and fragments with high  $\Delta^{17}\text{O}$  values have negative  $\Delta^{18}\text{O}_{\text{PCM}}$  values. This suggests the oxygen isotope mass fractionation during the reduction and supports the view of addition of  $^{16}\text{O}$ -poor carbon-rich organics to  $^{16}\text{O}$ -rich precursors.

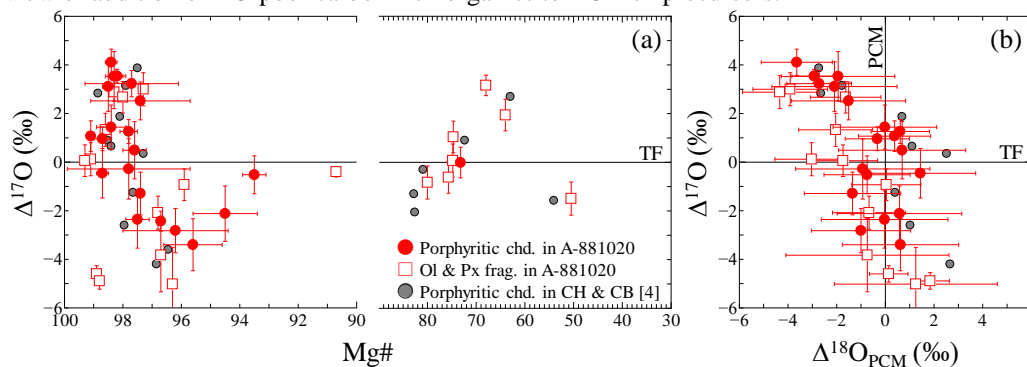


Fig. 1: (a)  $\Delta^{17}\text{O}$  vs. Mg# of porphyritic chondrules and olivine and pyroxene fragments in CH and CB chondrites and (b)  $\Delta^{17}\text{O}$  vs.  $\Delta^{18}\text{O}_{\text{PCM}}$  of type I porphyritic chondrules and FeO-poor fragments in CH and CB chondrites.

**References:** [1] Ushikubo T. et al. (2012) *GCA* 90:242-264. [2] Nakashima D. et al. (2010) *M&PS* 47:A148. [3] Tenner T.J. et al. (2015) *GCA* 148: 228-250. [4] Krot A.N. et al. (2010) *GCA* 74:2190-2211. [5] Krot A.N. and Nagashima K. (2017) *GJ* 51:45-68. [6] Connolly Jr. H.C. and Huss G.R. (2010) *GCA* 74:2473-2483. [7] Yamanobe M. et al. (2018) *Polar Sci.* 15:29-38. [8] Sakamoto N. et al. (2007) *Science* 317:231-233. [9] Schrader D.L. et al. (2013) *GCA* 101:302-327. [10] Hashizume K. et al. (2011) *Nature Geosci.* 4:165-168. [11] Connolly Jr. H.C. et al. (1994) *Nature* 371:136-139. [12] Ash R.D. et al. (1998) *M&PS* 33:A11.