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# ($= \text{mol.}\% \text{MgO}/(\text{MgO}+\text{FeO})$) [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$\mu$m Cs$^+$ 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 relic 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 ± 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 ~−3‰ to +2‰ with no correlation with Mg# [6-7], which are 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 ~99 and $\Delta^{17}\text{O}$ of ~4‰ are not observed in other chondrites and oxides that inherited oxygen isotope ratios of water ice are $^{16}$O-poor [e.g., 8]. In the former case, chondrules with Mg# of ~96 and $\Delta^{17}\text{O}$ of ~4‰, which are the lower end of the positive $\Delta^{17}\text{O}$-Mg# correlation, may correspond to the $^{16}$O-poor precursors. Such chondrules are rarely observed in Acfer 094 (ungr. C3.0) and CR chondrites [1,9]. The $^{16}$O-poor reductant may be carbon-rich organics, given the two observations; $^{16}$O-poor organics in primitive chondrites [10] and heating experiments that produced Fo90 olivine from Fo90 olivine and graphite or diamond [11]. Heating experiments produced chondrules with $\delta^{18}$O of +5.6‰ from type I chondrules with $\delta^{18}$O of +6.5‰ and 5 wt% graphite [12], indicating oxygen isotope mass fractionation between the chondrules and oxidized carbon (CO or CO2). Fig. 1b shows the deviation of $\delta^{18}$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 $^{16}$O-poor carbon-rich organics to $^{16}$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.