Distribution of ^{26}Al in the CR chondrite chondrule-forming region of the protoplanetary disk

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ABSTRACT

We report on the mineralogy, petrography, and *in situ* measured oxygen- and magnesium-isotope compositions of eight porphyritic chondrules (seven FeO-poor and one FeO-rich) from the Renazzo-like carbonaceous (CR) chondrites Graves Nunataks 95229, Grosvenor Mountains 03116, Pecora Escarpment 91082, and Queen Alexandra Range 99177, which experienced minor aqueous alteration and very mild thermal metamorphism. We find no evidence that these processes modified the oxygen- or Al-Mg isotope systematics of chondrules in these meteorites. Olivine, low-Ca pyroxene, and plagioclase within an individual chondrule have similar O-isotope compositions, suggesting crystallization from isotopically uniform melts. The only exceptions are relict grains in two of the chondrules; these grains are ^{16}O -enriched relative to phenocrysts of the host chondrules. Only the FeO-rich chondrule shows a resolvable excesses of ^{26}Mg , corresponding to an inferred initial $^{26}\text{Al}/^{27}\text{Al}$ ratio $[(^{26}\text{Al}/^{27}\text{Al})_0]$ of $(2.5 \pm 1.6) \times 10^{-6}$ ($\pm 2\text{SE}$). Combining these results with the previously reported Al-Mg isotope systematics of CR chondrules (Nagashima et al., 2014, *Geochem. J.* **48**, 561), 7 of 22 chondrules (32%) measured show resolvable excesses of ^{26}Mg ; the presence of excess ^{26}Mg does not correlate with the FeO content of chondrule silicates. In contrast, virtually all chondrules in weakly metamorphosed (petrologic type 3.0–3.1) unequilibrated ordinary chondrites (UOCs), Ornans-like carbonaceous (CO) chondrites, and the ungrouped carbonaceous chondrite Acfer 094 show resolvable excesses of ^{26}Mg . The inferred $(^{26}\text{Al}/^{27}\text{Al})_0$ in CR chondrules with resolvable excesses of ^{26}Mg range from $(1.0 \pm 0.4) \times 10^{-6}$ to $(6.3 \pm 0.9) \times 10^{-6}$, which is typically lower than $(^{26}\text{Al}/^{27}\text{Al})_0$ in the majority of chondrules from UOCs, COs, and Acfer 094. Based on the inferred $(^{26}\text{Al}/^{27}\text{Al})_0$, three populations of CR chondrules are recognized; the population characterized by low $(^{26}\text{Al}/^{27}\text{Al})_0$ ($< 3 \times 10^{-6}$) is dominant. There are no noticeable trends with major and minor element or O-isotope compositions between these populations. The weighted mean $(^{26}\text{Al}/^{27}\text{Al})_0$ of 22 CR chondrules measured is $(1.8 \pm 0.3) \times 10^{-6}$. An apparent agreement between the ^{26}Al - ^{26}Mg ages (using weighted mean value) and the revised (using $^{238}\text{U}/^{235}\text{U}$ ratio for bulk CR chondrites of 137.7789 ± 0.0085) ^{207}Pb - ^{206}Pb age of a set of chondrules from CR chondrites (Amelin et al., 2002, *Science* **297**, 1678) is consistent with the initial $^{26}\text{Al}/^{27}\text{Al}$ ratio in the CR chondrite chondrule-forming region at the canonical level ($\sim 5.2 \times 10^{-5}$), allowing the use of ^{26}Al - ^{26}Mg systematics as a chronometer for CR chondrules. To prove chronological significance of ^{26}Al for CR chondrules, measurements of Al-Mg and U-Pb isotope systematics on individual chondrules are required. The presence of several generations among CR chondrules indicates some chondrules that accreted into the CR chondrite parent asteroid avoided melting by later chondrule-forming events, suggesting chondrule-forming processes may have occurred on relatively limited spatial scales. Accretion of the CR chondrite parent body occurred at $> 4.0^{+0.5}_{-0.3}$ Ma after the formation of CAIs with the canonical $^{26}\text{Al}/^{27}\text{Al}$ ratio, although rapid accretion after formation of the major population of CR chondrules is not required by our data.

1. INTRODUCTION

^{26}Al is a short-lived radionuclide that decays to ^{26}Mg with a half-life of ~ 0.705 Myr (Norris et al., 1983). Its short half-life and presence in the early Solar System make ^{26}Al potentially one of the most precise chronometers of early Solar System processes, if it can be demonstrated that ^{26}Al was uniformly distributed in the protoplanetary disk (e.g., MacPherson et al., 1995; Jacobsen et al., 2008; Larsen et al., 2011; Wasserburg et al., 2012; Krot et al., 2012; Liu et al., 2012; Kita et al., 2013; Park et al., 2013). It is also possibly an important heat source for early-accreted planetesimals (e.g., Urey, 1955; Ghosh and McSween, 1998; Hevey and Sanders, 2006; Sahijpal et al., 2007; Elkins-Tanton et al., 2011; Schiller et al., 2015). Both local, solar energetic particle irradiation, and external, stellar origins for ^{26}Al have been proposed to explain the presence of ^{26}Al in the early Solar System at a high initial $^{26}\text{Al}/^{27}\text{Al}$ ratio, $(^{26}\text{Al}/^{27}\text{Al})_0$, relative to the Galactic chemical background (e.g., McKeegan et al., 2000; Chaussidon et al., 2006; Meyer, 2005; Arnould et al., 2006; Wasserburg et al., 2006; Duprat and Tatischeff, 2007; Larsen et al., 2011; Gounelle and Meynet, 2011). However, the maximum amount of ^{26}Al produced by irradiation during the life-time of the protoplanetary disk is insufficient to account for the amount of ^{26}Al that was present in the disk assuming its uniform distribution at the canonical level $[(^{26}\text{Al}/^{27}\text{Al})_0 \sim 5.2 \times 10^{-5}$ (Jacobsen et al., 2008)], which instead requires a recent stellar input (Duprat and Tatischeff, 2007). The stellar origin of ^{26}Al in the solar protoplanetary disk is generally accepted (e.g., Meyer, 2005; Arnould et al., 2006; Wasserburg et al., 2006; Duprat and Tatischeff, 2007; Gounelle and Meynet, 2011; Krot et al., 2012), but the abundance of ^{26}Al and the degree of its heterogeneity remain controversial (e.g., Larsen et al., 2011; Wasserburg et al., 2012; Kita et al., 2013; Bizzaro et al., 2014; Bizzaro, 2015; Bollard et al., 2015; Schiller et al., 2015).

Utilizing the ^{26}Al - ^{26}Mg chronometer depends upon ^{26}Al being homogeneously distributed in the protoplanetary disk during an object's formation (i.e., CAI [calcium-aluminum-rich inclusion], chondrule, or achondrite), and the possible non-radiogenic abundance variations of ^{26}Mg being accounted for. ^{26}Al appears to have been heterogeneously distributed during the epoch of CAI formation, and, therefore, cannot be used as a chronometer for the duration of this epoch (e.g., Krot et al., 2012; Liu et al., 2012; Makide et al., 2013; Park et al., 2013). In addition, some refractory grains and inclusions show large variations (typically deficits) in the initial $^{26}\text{Mg}/^{24}\text{Mg}$ ratio (e.g., Wasserburg et al., 1977; Liu et al., 2009; Wasserburg et al., 2012; Park et al., 2013). These observations provide clear evidence for Mg-isotope heterogeneity in the CAI-forming region. Despite these variations, the majority of CAIs from the Vigarano-like carbonaceous (CV) chondrites have the canonical $(^{26}\text{Al}/^{27}\text{Al})_0$ (e.g., MacPherson et al., 1995; Jacobsen et al., 2008; Larsen et al., 2011). While CV CAIs with the canonical $(^{26}\text{Al}/^{27}\text{Al})_0$ are generally considered to have formed over a short timescale (i.e., 0.004–0.02 Ma; Jacobsen et al., 2008; Larsen et al., 2011), uncertainties in the distribution of ^{26}Al make this interpretation ambiguous. Instead, variations in the $(^{26}\text{Al}/^{27}\text{Al})_0$ of CAIs may suggest that there are multiple CAI generations.

While the canonical $(^{26}\text{Al}/^{27}\text{Al})_0$ was reached in the CV CAI-formation region (e.g., Krot et al., 2012; Kita et al., 2013; Wadhwa et al., 2013), the distribution of ^{26}Al (homogeneous vs. heterogeneous) in the protoplanetary disk after the epoch of CAI formation remains controversial (e.g., Villeneuve et al., 2009; Larsen et al., 2011; Kleine et al., 2012; Krot et al., 2012; Kruijer et al., 2014). The U-corrected ^{207}Pb - ^{206}Pb and ^{182}Hf - ^{182}W ages applied to angrites have been found to be inconsistent with their inferred ^{26}Al - ^{26}Mg ages (Larsen et al., 2011; Kleine et al., 2012;

Schiller et al., 2015), suggesting either ^{26}Al and/or Mg-isotope heterogeneity in the disk, or resetting of the ^{26}Al - ^{26}Mg system in these meteorites. However, homogeneity of ^{182}Hf and ^{26}Al at least in the CV CAI and angrite formation region has been argued due to agreement of ^{182}Hf - ^{182}W and inferred ^{26}Al - ^{26}Mg ages using a revised initial $^{182}\text{Hf}/^{180}\text{Hf}$ ratio (Kruijer et al., 2014).

In contrast, evidence for heterogeneity of ^{26}Al in two chondrule-forming regions of the protoplanetary disk comes from correlated ^{26}Al - ^{26}Mg and U-corrected Pb-Pb studies of individual chondrules from Allende (CV3) and Northwest Africa (NWA) 5697 (L3.1). These coordinated analyses, where each isotopic system is studied on the same chondrule, show that the ^{26}Al - ^{26}Mg ages are younger than the U-corrected Pb-Pb ages by $\sim 1\text{--}3$ Ma (Bizzaro et al., 2014; Bollard et al., 2015; Bizzaro, 2015). If the Al-Mg and Pb-Pb systems were not disturbed, this age discrepancy argues for heterogeneity of ^{26}Al , at least in the regions of the protoplanetary disk where these chondrules formed.

Assuming a homogeneous distribution of magnesium isotopes in the disk, ingrowth of radiogenic ^{26}Mg in chondrules due to the decay of ^{26}Al is consistent with ^{26}Al homogeneity to $\pm 10\%$ of the canonical level (Villeneuve et al., 2009). We note, however, that due to the relatively low-precision measurements of $^{26}\text{Mg}/^{24}\text{Mg}$ ratio by SIMS ($> \pm 10$ ppm) compared to the total ingrowth of ^{26}Mg (~ 40 ppm) in solids having solar $^{27}\text{Al}/^{24}\text{Mg}$ ratio of 0.101, this conclusion is rather uncertain.

Using the ^{26}Al - ^{26}Mg system as a chronometer requires also that aluminum or magnesium did not migrate to, or from, their host minerals after crystallization. CAI minerals from metamorphosed CV chondrites often show disturbed isochrons (e.g., MacPherson et al., 1995 and references therein; Bullock et al., 2015). Mild thermal metamorphism probably disturbed the ^{26}Al - ^{26}Mg systematics of chondrules in some unequilibrated ordinary chondrites (UOCs; e.g., Kita et al., 2000; Huss et al., 2001). To confidently compare different chronometers, ‘pristine’ Solar System materials that formed after the epoch of CAIs and recorded $(^{26}\text{Al}/^{27}\text{Al})_0$ in their formation region (i.e., protoplanetary disk) are required. The Renazzo-like carbonaceous (CR) chondrites are ideal samples for this study because they generally experienced only small degrees of aqueous alteration and very mild thermal metamorphism (e.g., Weisberg et al., 1993; Zolensky et al., 1993; Busemann et al., 2007; Abreu and Brearley, 2010; Howard et al., 2015; Schrader et al., 2015), and therefore the ^{26}Al - ^{26}Mg systematics of their chondrules are likely undisturbed. Also, the CR chondrules appear to have formed relatively late; for instance, six handpicked chondrules from Acfer 059 (CR2) have an average ^{207}Pb - ^{206}Pb age of 4564.7 ± 0.6 Ma, which is 2.5 ± 0.8 (± 2 standard error of the mean [2SE]) Ma after the Efremovka (CV3) forsterite-bearing Type B CAI E60 having ^{207}Pb - ^{206}Pb age of 4567.2 ± 0.6 Ma (Amelin et al., 2002). This allows a comparison between relative (^{26}Al - ^{26}Mg) and absolute (^{207}Pb - ^{206}Pb) ages of CR chondrules, providing a test for ^{26}Al homogeneity in the disk region where CR chondrules formed. We note that the reported variations in $^{238}\text{U}/^{235}\text{U}$ ratio among CV CAIs (Brennecka et al., 2010) should be taken into consideration.

Here we present the major and minor element abundances and O-isotope compositions of individual minerals from, and the ^{26}Al - ^{26}Mg systematics, of eight chondrules (seven FeO-poor and one FeO-rich) from four CR chondrites. Major and minor element abundances and oxygen isotopic compositions are utilized to test if they are consistent with the typical CR chondrules, and to investigate potential relationships with $(^{26}\text{Al}/^{27}\text{Al})_0$. The three main hypotheses this study will address are: in the protoplanetary disk after the epoch of CAI formation, ^{26}Al was 1) homogeneously distributed at the canonical level, 2) homogeneously distributed at a lower level than the canonical value, and 3) heterogeneously distributed. To test these hypotheses, we

compare ^{26}Al - ^{26}Mg relative ages with U-corrected ^{207}Pb - ^{206}Pb absolute age of a set of chondrules from CR chondrites reported by Amelin et al. (2002).

2. EXPERIMENTAL

2.1. Mineralogy, petrology and chemical compositions

A total of eight chondrules from the CR chondrite thin sections Graves Nunataks (GRA) 95229,22, Grosvenor Mountains (GRO) 03116,15, Pecora Escarpment (PCA) 91082,15, and Queen Alexandra Range (QUE) 99177,6 were selected for detailed analysis (Table 1). Chondrule olivine, pyroxene, and plagioclase were analyzed for major and minor element abundances; representative analyses given in Tables 2–4. The major and minor element compositions of all olivine, pyroxene, and plagioclase are listed in Electronic Annex 1 (EA-1).

High-resolution backscattered electron (BSE) images of each chondrule studied were obtained with a JEOL Hyperprobe JXA 8500F electron probe microanalyzer (EPMA) at the University of Hawai'i at Mānoa (UH: EA-2). Additional images were taken on the JEOL JSM-5900LV scanning electron microscope (SEM) at UH. Apparent chondrule diameters were obtained by measuring major and minor axes of the chondrules from BSE images using Adobe Photoshop® (Table 1).

Major and minor element abundances were measured by the EPMA at UH; additional data from Schrader et al. (2015) is reproduced in EA-1. Polished and carbon coated thin sections were analyzed with a 1 μm beam, both individual points and line scans, with operating conditions of 15 keV and 15 nA, a ZAF correction method (a Phi-Rho-Z correction technique; Armstrong, 1988); peak and background counting times were varied per element to optimize detection limits. Only stoichiometric silicate analyses with totals between 98.0–102.0 wt.% were retained; standards and detection limits are listed in Tables 2–4 and EA-1.

2.2. Oxygen isotopic compositions

Olivine, low-Ca pyroxene, and plagioclase within chondrules were measured *in situ* for their O-isotope compositions with the UH Cameca ims 1280 ion microprobe, using the procedure described by Makide et al. (2009). A $10\times 10\ \mu\text{m}^2$ area was pre-sputtered for 90 sec by rastering with a $\sim 1\ \text{nA}$ Cs^+ primary ion beam before analysis. The raster area was reduced to $7\times 7\ \mu\text{m}^2$ in the center of pre-sputtered region for data collection. Secondary ions of $^{16}\text{O}^-$, $^{17}\text{O}^-$, and $^{18}\text{O}^-$ were measured in multicollection mode alternating between two cycles optimized for different isotopes. Each analysis consisted of 60 cycles alternating between a 4 sec cycle to measure $^{16}\text{O}^-$ on a Faraday cup (FC) and $^{17}\text{O}^-$ on an electron multiplier (EM) detector (to determine $^{17}\text{O}/^{16}\text{O}$) followed by a 10 sec cycle used to measure $^{16}\text{O}^-$ and $^{18}\text{O}^-$ on FCs (to determine $^{18}\text{O}/^{16}\text{O}$), for a total analysis time of ~ 9 min per measurement. Each ratio was calculated from counts summed over the entire analysis to minimize the additive positive bias associated with low counts in individual cycles (Ogliore et al., 2011); however, counts of $^{16}\text{O}^-$ (typically 5×10^8 cps (2×10^9 counts/cycle)) were not a concern for O-isotope analyses. Instrumental mass fractionation (IMF) was corrected using a standard analyzed before and after chondrule measurements each day (San Carlos olivine, Vitim Volcanic Field orthopyroxene (313 opx; Ionov et al., 1994, and Miyakejima anorthite). The $^{16}\text{OH}^-$ signal was monitored after each measurement, and $^{17}\text{O}^-$ was corrected for its contribution, which was typically $<0.1\%$. The variability of matrix effects in

olivines with a range of fayalite contents (Fa_{10–60}) can result in IMF variations of <2‰ in $\delta^{18}\text{O}$ between Fa_{10–40} and 3–4‰ between Fa_{10–60} (Jogo et al., 2012). Since IMF is expected to be mass-dependent it should not affect $\Delta^{17}\text{O}$. Therefore, $\Delta^{17}\text{O}$ is used to compare between FeO-poor and FeO-rich chondrule O-isotope compositions. Errors of individual analyses represent 2 σ analytical uncertainty including both the internal measurement precision and the external reproducibility for standard measurements during a given analytical session (Table 5). After O-isotope analyses, secondary ion mass spectrometry (SIMS) pits were imaged on the SEM to confirm their locations. Each 7×7 μm^2 pit is within the desired olivine grain (EA-2). Oxygen-isotope compositions are presented as standard delta-notation relative to Standard Mean Ocean Water, $\delta^{17}\text{O}_{\text{SMOW}}$ and $\delta^{18}\text{O}_{\text{SMOW}}$, and as deviation from the terrestrial fractionation line, $\Delta^{17}\text{O}$ ($=\delta^{17}\text{O}_{\text{SMOW}} - 0.52 \times \delta^{18}\text{O}_{\text{SMOW}}$).

2.3. Magnesium isotopic compositions

Olivine, low-Ca pyroxene, and plagioclase within chondrules were measured *in situ* for their Mg-isotope composition with the UH Cameca ims 1280 ion microprobe, using the procedure described by Makide et al. (2009).

For olivine and low-Ca pyroxene, we used ~4.5 nA O[−] primary ion beam with ~20 μm in diameter for pre-sputtering and collection of Al and Mg ions. Pre-sputtering was made for 180 seconds. Secondary ions of $^{24}\text{Mg}^+$, $^{25}\text{Mg}^+$, $^{26}\text{Mg}^+$, and $^{27}\text{Al}^+$ were measured simultaneously with four FCs. Data was collected over 30 cycles. The mass-resolving power was set to ~2300 using 500 μm exit slits. Although this does not allow a complete separation of $^{25}\text{Mg}^+$ from $^{24}\text{MgH}^+$ interference, a contribution of $^{24}\text{MgH}^+$ onto $^{25}\text{Mg}^+$ was estimated to be less than 1 ppm. The IMF was corrected using standards (San Carlos olivine and San Carlos pyroxene) analyzed before and after chondrule measurements each day.

For plagioclase, ~75–125 pA O[−] primary ion beam with ~5 μm in diameter was used for pre-sputtering and data collection (standard: Miyakejima anorthite). Pre-sputtering was made for 600 seconds. Secondary ions of $^{24}\text{Mg}^+$, $^{25}\text{Mg}^+$, and $^{26}\text{Mg}^+$ were measured on a monocollector EM sequentially by magnetic field jumping. $^{27}\text{Al}^+$ was measured by a multicollector FC together with $^{25}\text{Mg}^+$. Each measurement consisted of 120 cycles in total (~1.5 hours). The mass-resolving power was set to ~3800–4200, sufficient to separate interfering hydrides and doubly charged $^{48}\text{Ca}^{++}$.

In both types of measurements, automated centering of the secondary beam in the field aperture of the mass spectrometer and high-voltage offset control were applied before each measurement. Offset control and mass-peak centering (in the case of monocollection mode) were also applied at cycle 60. To minimize the positive bias associated with low counts in the denominator isotope, ratios were calculated from counts summed over the entire analysis (Ogliore et al., 2011).

For plagioclase, measured Mg-isotope ratios were corrected for mass fractionation (both instrumental and intrinsic) internally by a linear fractionation law normalized to the terrestrial Mg-isotope compositions [$^{25}\text{Mg}/^{24}\text{Mg} = 0.12663$, $^{26}\text{Mg}/^{24}\text{Mg} = 0.13932$ (Catanzaro et al., 1966)]. For olivine and pyroxene data, instrumental mass fractionation was corrected by standard-sample bracketing assuming all terrestrial standards used have the terrestrial Mg-isotope compositions, and then intrinsic mass fractionation was corrected with a linear law. Excesses of ^{26}Mg are defined relative to the terrestrial value. The choice of the mass fractionation law is not critical to the data set obtained in this study because the degree of intrinsic mass fractionation is very small,

and because the Al/Mg ratios of plagioclase that control the ^{26}Al - ^{26}Mg isochrons are sufficiently high that the uncertainty introduced by the choice of law is insignificant. The $^{27}\text{Al}^+ / ^{24}\text{Mg}^+$ ratios measured with the ion microprobe were corrected for Al/Mg relative sensitivity using the standards.

2.4. Uranium isotopic compositions

The U-isotope ($^{238}\text{U}/^{235}\text{U}$) composition from a total of four whole rock samples (10 analyses) taken from the CR chondrites Acfer 059, NWA 721, and NWA 801 were determined by multiple-collector inductively coupled plasma mass spectrometry (MC-ICPMS) at the University of Otago based on previously published methods (Stirling et al., 2005, 2006, 2007; Amelin et al., 2010). In brief, samples were spiked with a high purity ^{236}U - ^{233}U mixed tracer, and analysed without acid leaching, to prevent possible U-isotope fractionation induced by laboratory treatment. Approximately 0.1–0.4 g of each separated powdered sample was weighed into a PFA vial, then spiked with a weighed volume of ^{236}U - ^{233}U mixed tracer prior to sequential acid digestion with double-distilled, high-purity HF, HNO_3 and HCl on a hotplate held at $\sim 140^\circ\text{C}$. Following dissolution, uranium was chemically extracted from the sample matrix and purified using TRU and UTEVA ion exchange resins (Eichrom Technologies, U.S.A.). For each sample, uranium isotopic composition was measured using a Nu Plasma-HR MC-ICPMS (Nu Instruments Ltd, U.K.) coupled to a DSN-100 desolvating nebuliser by simultaneous detection of all U isotopes using an array of Faraday collectors operating with $10^{11} \Omega$ resistors. All $^{238}\text{U}/^{235}\text{U}$ ratios were corrected for both the contributions of ^{235}U and ^{238}U present in the ^{236}U - ^{233}U mixed spike and for instrumental mass fractionation, using the measured $^{236}\text{U}/^{233}\text{U}$ ratio, normalized to the true value (Stirling et al., 2005), and the exponential mass fractionation law (Hart and Zindler, 1989; Habfast, 1998). The $^{236}\text{U}/^{233}\text{U}$ ratio of the mixed spike tracer was calibrated against the U metal standard CRM 145, assuming CRM 145 has a $^{238}\text{U}/^{235}\text{U}$ value of 137.837 (Richter et al. 2010). The CRM 145 U standard, double-spiked in the same way as the samples, was run repeatedly throughout each analytical session to assess the performance of our analytical protocols. Variations in $^{238}\text{U}/^{235}\text{U}$ can be resolved at the ± 40 to 60 ppm level (2SD) using sample sizes equivalent to ~ 40 ng of ^{238}U and at the ± 100 to 200 ppm level (2SD) for ^{238}U load sizes corresponding to < 3 ng, dependent on the daily stability of the instrument (Stirling et al., 2007).

3. RESULTS

3.1. Mineralogy and petrology

Eight porphyritic chondrules from four thin sections with sufficiently sized plagioclase grains ($> 10 \times 10 \mu\text{m}^2$ to facilitate the $5 \times 5 \mu\text{m}^2$ analysis area) for SIMS analysis were studied (Table 1; GRA 95229 Ch29A and 29B, GRO 03116 Ch5 and Ch13, PCA 91082 Ch14 and Ch15, and QUE 99177 Ch13 and Ch14; preliminary data reported in Schrader et al., 2013a). Seven of these chondrules are FeO-poor ($\text{Fe}/(\text{Fe}+\text{Mg}) < 10\%$ atomic ratio), while GRO 03116 Ch5 is FeO-rich (i.e., $\text{Fe}/(\text{Fe}+\text{Mg}) > 10\%$ atomic ratio; e.g., Hewins, 1997; Jones et al., 2005). The FeO-poor chondrules consist of olivine, pyroxene (low-Ca and high-Ca), plagioclase, silica, glassy mesostasis, and Ni-poor Fe,Ni-metal (Figs. 1–3). The FeO-rich chondrule consists primarily of FeO-rich olivine, with minor amounts of FeO-poor olivine (i.e., relict grains), Na-rich

plagioclase, mesostasis, chromite, and Fe-sulfide (Figs. 1c, 2e,f). GRA 95229 Ch29A and Ch29B are a compound chondrule set, with Ch29B appearing to be deformed around Ch29A (Figs. 1a,b). GRA 95229 Ch29A and PCA 91082 Ch14 contain numerous spinel grains, while PCA 91082 Ch14 is compound with a smaller FeO-poor chondrule (Fig. 1e).

In the FeO-poor chondrules studied, olivine and low-Ca pyroxene are Mg-rich ($Fa_{0.6-3.4}$ and $Fs_{0.8-3.3}$, respectively), and plagioclase is Ca-rich ($An_{86.1-99.5}$). The FeO-rich chondrule contains olivine phenocrysts that range from $Fa_{10.8-48.1}$, with FeO-poor relict grains of $Fa_{0.4-0.5}$, and Na-rich plagioclase ($Ab_{63.7}An_{35.3}$; Tables 1–4; EA-1). The major and minor element compositions of chondrule silicates studied here are consistent with chondrules in the CR chondrites (e.g., Johnson and Prinz, 1991; Weisberg et al., 1993; Noguchi, 1995; Connolly et al., 2001; Krot et al., 2006; Connolly and Huss, 2010; Berlin et al., 2011; Schrader et al., 2013b, 2014a, 2015; Nagashima et al., 2014), and therefore will not be discussed in further detail.

3.2. Oxygen isotopic compositions

Within analytical uncertainties, chondrules are internally homogeneous with respect to the O-isotope compositions of their olivine, low-Ca pyroxene, and plagioclase; with the exception of chondrules that are considered to contain relict grains (i.e., PCA 91082 Ch14 and GRO 03116 Ch5) and GRA 95229 Ch29A (Fig. 4). PCA 91082 Ch14 is a spinel-bearing FeO-poor chondrule that contains olivine and plagioclase grains that are significantly more ^{16}O -enriched than the rest of the olivine, low-Ca pyroxene, and plagioclase in the chondrule (Fig. 4e). GRO 03116 Ch5 is an FeO-rich chondrule that contains two olivine grains that are more ^{16}O -enriched than its phenocrysts, and are considered to represent FeO-poor and FeO-rich relict grains ($Fa_{0.4}$ and $Fa_{10.8}$, respectively; Fig. 4c and Table 5); similar to those discussed in Schrader et al. (2013b). GRA 95229 Ch29A is an FeO-poor chondrule with pyroxene that is ^{16}O -enriched compared to olivine and plagioclase (Fig. 4a).

Oxygen isotopic compositions of FeO-poor chondrule olivines studied, including the relicts, range in $\Delta^{17}O$ from -23.1 to 0.1‰ (10 analyses, 4 chondrules), low-Ca pyroxenes range in $\Delta^{17}O$ from -5.2 to -2.1‰ (13 analyses, 6 chondrules), while plagioclases range from $\Delta^{17}O$ from -10.7 to -1.5‰ (14 analyses, 6 chondrules). If the two outliers in PCA 91082 Ch14 (i.e., olivine grain at $\Delta^{17}O = -23.1\text{‰}$ and plagioclase at $\Delta^{17}O = -10.7\text{‰}$) are excluded, the range for FeO-poor chondrule olivine changes to $\Delta^{17}O \sim -6.3$ to 0.1‰ and plagioclase changes to $\Delta^{17}O \sim -5.7$ to -1.5‰ . Oxygen-isotope compositions of the FeO-rich chondrule olivine phenocrysts range in $\Delta^{17}O$ from -1.0 to -0.7‰ (2 analyses), with FeO-poor and FeO-rich relict olivine grains at -5.1‰ and -3.4‰ , respectively (Fig. 5 and Table 5).

3.3. Magnesium isotopic compositions

One of the eight chondrules studied here, GRO 03116 Ch5 (FeO-rich), has a resolvable ^{26}Mg excess in plagioclase. The excess of ^{26}Mg corresponds to an inferred $(^{26}Al/^{27}Al)_0$ of $(2.5 \pm 1.6) \times 10^{-6}$ ($\pm 2SE$) (Figs. 6d,f,g and Table 6).

3.4. Uranium isotopic compositions

Uranium isotopic ratios of the ten CR chondrite analyses range from $^{238}U/^{235}U = 137.765 \pm 0.029$ to 137.818 ± 0.028 (2SD) (Table 7), within the $^{238}U/^{235}U$ range of bulk meteorites

studied by Goldmann et al. (2015). The $^{238}\text{U}/^{235}\text{U}$ whole rock values of the three studied CR chondrites are identical, within the magnitude of their associated uncertainties. The U-isotopic homogeneity between seven whole rock samples of Acfer 059, and among three different CR chondrites, suggests U-isotope homogeneity among the major components of the CR chondrites (e.g., chondrules and matrix; Schrader et al., 2011). The $^{238}\text{U}/^{235}\text{U}$ weighted average for the ten samples is 137.7789 ± 0.0085 (2SE) (Table 7). The $^{238}\text{U}/^{235}\text{U}$ compositions of the terrestrial reference materials BCR-2 and BHVO-2 (United States Geological Survey; USGS) were also analysed, giving rise to respective values of 137.803 ± 0.014 and 137.791 ± 0.011 (2SE), in very good agreement with previously published results (Weyer et al., 2008).

4. DISCUSSION

4.1. Mineralogy, petrology, and oxygen isotopic compositions

The chondrules studied here show no evidence for thermal metamorphism, and are therefore ideal for a study of their Al-Mg isotope systematics. The mineralogy and petrology of the chondrules, and the major and minor element and O-isotope compositions of chondrule olivine, pyroxene, and plagioclase are consistent with those in CR chondrites previously studied (Tables 1–5; e.g., Johnson and Prinz, 1991; Weisberg et al., 1993; Noguchi et al., 1995; Connolly et al., 2001; Krot and Keil, 2002; Krot et al., 2006; Connolly and Huss, 2010; Wasson and Rubin, 2009; 2010; Berlin et al., 2011; Schrader et al., 2013a, 2013b, 2014a, 2015; Nagashima et al., 2014; Tenner et al., 2013a, 2015a, 2015b). With few exceptions, the CR chondrites probably did not experience parent body temperatures above 240°C; with the least altered samples likely closer to 0°C (Zolensky et al., 1993; Busemann et al., 2007; Cody et al., 2008; Abreu, 2011; Alexander et al., 2013; Jilly-Rehak et al., 2015a; Schrader et al., 2015). However GRO 03116 was likely heated more extensively than the typical CR chondrite (Abreu, 2011; Alexander et al., 2013; Schrader et al., 2015). Yet, the degree of thermal metamorphism experienced by the CR chondrites in this study did not noticeably alter their silicate compositions. The Cr_2O_3 abundance in FeO-rich chondrule olivine suggests a petrologic subtype of approximately 3.00–3.05 for the CR chondrites in this study (disregarding their type 2 classifications due to aqueous alteration; Schrader et al., 2015). In thermally metamorphosed chondrites, due to different Fe-Mg diffusion rates in olivine and pyroxene (Ganguly and Tazzoli, 1994; Chakraborty, 1997), Fs has a lower Fe content than Fa (e.g., Wlotzka, 2005). The average Fa value of olivine and Fs value of low-Ca pyroxene in chondrules studied here show a 1:1 relationship (Fig. 7), indicating the chondrules were not thermally metamorphosed. Therefore, the chemical and isotopic data presented here are considered to be unaltered since their formation.

The chondrules studied here, except for relict grains identified in three chondrules, crystallized from uniform melts and are therefore appropriate for a study of their Al-Mg isotope systematics. With the exception of relict grains in FeO-poor chondrule GRA 95229 Ch29A, FeO-rich chondrule GRO 03116 Ch5, and Al-rich chondrule PCA 91082 Ch14, O-isotope compositions of olivine and low-Ca pyroxene phenocrysts and plagioclase within individual chondrules studied are similar (Fig. 5), suggesting that they crystallized from isotopically uniform melts, consistent with the O-isotope data for barred olivine chondrules in CR chondrites (Schrader et al., 2014a). The ^{16}O -enriched compositions of some olivine and plagioclase grains

in PCA 91082 Ch14 ($\Delta^{17}\text{O} = -23.1\text{‰}$ and -10.7‰) and the presence of abundant spinel grains in this chondrule are consistent with the presence of CAI and/or AOA (amoeboid olivine aggregate) material in the chondrule precursors; the spinel grains are probably relict (e.g., most CR CAIs have $\Delta^{17}\text{O} \sim -24\text{‰}$, Aléon et al., 2002; Makide et al., 2009). Several CR CAIs recycled during chondrule formation show large variations in $\Delta^{17}\text{O}$ (from -24‰ to -3‰ , Makide et al., 2009). In addition, the FeO-poor and FeO-rich relict grains in GRO 03116 Ch5 are ^{16}O -enriched relative to chondrule phenocrysts (Fig. 4c and Table 5), similar to relict grain-bearing FeO-rich chondrules discussed in Schrader et al. (2013b).

4.2. Inferred initial $^{26}\text{Al}/^{27}\text{Al}$ ratios

The Al-Mg isotope systematics of CR chondrules studied here are consistent with previous results. Here we summarize Al-Mg isotope systematics in 22 CR chondrules (19 FeO-poor and three FeO-rich) studied with the UH Cameca ims-1280: GRA 95229, GRO 03116, PCA 91082, and QUE 99177 (this study); and Acfer 311, Elephant Moraine (EET) 92042, EET 92174, El Djouf 001, and NWA 721 (Krot and Keil, 2002; Krot et al., 2006; Hutcheon et al., 2009; Nagashima et al., 2007, 2008, 2014). Only seven of these 22 chondrules ($\sim 32\%$) have resolvable ^{26}Mg excesses. These excesses correspond to the inferred $(^{26}\text{Al}/^{27}\text{Al})_0$ of (1.0 ± 0.4) , (2.3 ± 1.4) , (2.5 ± 1.6) , (2.9 ± 1.4) , (3.1 ± 1.3) , (4.6 ± 1.1) , and $(6.3 \pm 0.9) \times 10^{-6}$ ($\pm 2\text{SE}$), determined by weighted linear regressions (Figs. 6 and 8a and Table 6). The low percentage of chondrules in the CR chondrites with a resolvable ^{26}Mg excess is consistent with those reported by Kurahashi et al. (2008a) and Tenner et al. (2013a, 2015b), but distinct from those in weakly metamorphosed (petrologic type 3.0–3.1) unequilibrated ordinary chondrites (UOCs), Ornans-like carbonaceous (CO) chondrites, and Acfer 094 (ungrouped carbonaceous chondrite) (Kita and Ushikubo (2012) and references therein; Fig. 8b).

It has been hypothesized that aqueous alteration on the CR chondrite parent body may have modified chondrule ^{26}Al - ^{26}Mg systematics, resulting in low inferred $(^{26}\text{Al}/^{27}\text{Al})_0$ and corresponding young inferred chondrule ages (Kita and Ushikubo, 2012). However, neither of the two chondrules studied in QUE 99177, argued to be the least aqueously altered CR chondrite (Abreu and Brearley, 2010; Schrader et al., 2011, 2014b, 2015), has a resolvable ^{26}Mg excess (Table 6). In addition, only two of six chondrules in QUE 99177 studied by Tenner et al. (2013a) have resolvable ^{26}Mg excesses. Since Al-Mg isotope systematics in chondrules in QUE 99177 are consistent with those in chondrules in other CR chondrites (Nagashima et al., 2007, 2008, 2014; Tenner et al., 2013a, 2015b; this study), we suggest that aqueous alteration has not modified the ^{26}Al - ^{26}Mg systematics of chondrules within the CR chondrites measured.

4.2.1. Relationship between O-isotope composition, $\text{Fe}/(\text{Fe}+\text{Mg})$ ratio, and $(^{26}\text{Al}/^{27}\text{Al})_0$ of chondrule silicates

It has been suggested that the O-isotope composition and $\text{Fe}/(\text{Fe}+\text{Mg})$ atomic ratio (Fe#) of carbonaceous chondrite chondrule silicates correlate with the inferred $(^{26}\text{Al}/^{27}\text{Al})_0$ (e.g., Tenner et al., 2013a, 2015b). From six FeO-poor chondrules in QUE 99177 combined with chondrules in Acfer 094 (Ushikubo et al., 2012) and type 3.0 CO chondrites (Tenner et al., 2013b), it was inferred that ^{16}O -enriched ($\Delta^{17}\text{O} \sim -5\text{‰}$) and very Mg-rich (Fe# = 1) chondrules have $(^{26}\text{Al}/^{27}\text{Al})_0 > 0.3 \times 10^{-5}$, while ^{16}O -depleted ($\Delta^{17}\text{O} = -1.7$ to -2.8‰) and less Mg-rich (Fe# = 1.5 to 2.5) chondrules have $(^{26}\text{Al}/^{27}\text{Al})_0 < 0.3 \times 10^{-5}$ (Tenner et al., 2013a). Tenner et al. (2015b) also

reported this correlation in a study of 12 CR chondrules from QUE 99177 and Meteorite Hills 00426. Five chondrules with $\text{Fe}\# < 1$ and $\Delta^{17}\text{O}$ of -4.2 to -5.3‰ had resolvable ^{26}Mg excesses, with $(^{26}\text{Al}/^{27}\text{Al})_0$ of $(3.9 \pm 1.7) \times 10^{-6}$ to $(6.2 \pm 3.9) \times 10^{-6}$. The remaining seven chondrules had $\text{Fe}\#$ of 1.3 to 5.8 and $\Delta^{17}\text{O}$ of -0.9 to -4.9‰ , and no resolvable ^{26}Mg excesses (Tenner et al., 2015b). Confirmation of this observation is important as it would imply a temporal dependency of oxygen fugacity and O-isotope composition, potentially from an influx of $^{17,18}\text{O}$ -enriched water ice.

The 22 chondrules from the CR chondrites presented here (Nagashima et al., 2007, 2008, 2014; this study) do not show a relationship between O-isotope composition, $\text{Fe}\#$, and $(^{26}\text{Al}/^{27}\text{Al})_0$ (Figs. 5, 6, 9). In this study, we report on chondrules with a range in $\text{Fe}\#$ from 0.6 to 50.3 , and include the range in Fe content for each chondrule from both olivine and pyroxene (Fig. 9). Important differences from this study to that of Tenner et al. (2013a, 2015b), is that chondrules with $\text{Fe}\# < 1$ are found to have both resolvable and unresolvable ^{26}Mg excesses (Fig. 9b). Furthermore, there is clearly no relationship between $\text{Fe}\#$ and O-isotope composition with inferred $(^{26}\text{Al}/^{27}\text{Al})_0$ when FeO-rich chondrules are considered; two of three FeO-rich chondrules have resolvable ^{26}Mg excesses with inferred $(^{26}\text{Al}/^{27}\text{Al})_0$ values intermediate between those of FeO-poor chondrules (Fig. 8a). While there is a general relationship between $\text{Fe}\#$ and O-isotope composition of chondrule silicates (Fig. 5; Schrader et al., 2013b, 2014a; Tenner et al., 2013a; 2015a), there is no obvious relationship between $\text{Fe}\#$ and $(^{26}\text{Al}/^{27}\text{Al})_0$ observed here (Figs. 8a and 9).

4.2.2. Constraining the number of chondrule-forming events recorded by CR chondrules

Although the $(^{26}\text{Al}/^{27}\text{Al})_0$ values in chondrules in a single meteorite group vary (e.g., Fig. 8b), it is inferred that (1) most chondrules in UOCs (namely the LL3.0–3.1 chondrites), CO3.0, and Acfer 094, assuming a homogeneous distribution of ^{26}Al at the canonical level, formed within a narrow timescale of less than a million years in each group (Kita and Ushikubo, 2012); and (2) chondrules in each chondrite group did not all form in a single discrete chondrule formation event. This is consistent with distinct differences in the U-corrected ^{207}Pb - ^{206}Pb absolute ages of a limited number of individual chondrules measured so far from the type 3 ordinary, CV, and CR chondrites, including chondrules with measured $^{238}\text{U}/^{235}\text{U}$ ratio (Connelly et al., 2012; Bollard et al., 2014). However, it is unclear if chondrule formation occurred continuously over prolonged time or over discrete events per chondrite group. While asteroid accretion could not have occurred instantly after chondrule formation, it has been hypothesized that chondrule forming events may have eventually triggered the formation of chondritic parent bodies (Alexander et al., 2008). If the most significant chondrule-forming event for each chondrite group did lead to accretion, then chondrules from this event would be the youngest (i.e., last chondrules to form). To facilitate the gravitational collapse of gas and dust into a parent body, this ‘final’ chondrule formation event was conceivably substantial, and may be represented by the largest ‘group’ of chondrules. Here, we test these hypotheses by investigating whether or not chondrules within different chondrite groups, in particular the CR chondrites, formed during a single prolonged event or multiple events, and if the dominant formation event is also the youngest.

Chondrules from the CR chondrites show a similar distribution but lower mean value of inferred $(^{26}\text{Al}/^{27}\text{Al})_0$ values to those of other chondrite groups (Fig. 8). A major difference between CR chondrules and chondrules from other chondrule groups is that numerous CR chondrules have apparent negative $(^{26}\text{Al}/^{27}\text{Al})_0$ values, which are zero within uncertainty. Since negative initial ratios are impossible and these chondrules have values of zero within $+2\text{SE}$, we

consider these data valid (Fig. 8a and Table 6). We assume there is an average $(^{26}\text{Al}/^{27}\text{Al})_0$ value for chondrules from the CR chondrites, and consider both the positive and negative initial ratios as part of a statistical distribution around the true average $(^{26}\text{Al}/^{27}\text{Al})_0$.

To determine if there is a single, multiple, and/or dominant generation(s) of chondrules according to the $(^{26}\text{Al}/^{27}\text{Al})_0$ value for chondrules from the UOCs, CO chondrites, Acfer 094, and CR chondrites, we apply statistical tests (e.g., Fisher, 1932; Bevington and Robinson, 2003) to these datasets (graphically outlined in EA-3). First, we calculate a summed probability density distribution (Fig. 10) of all measured $(^{26}\text{Al}/^{27}\text{Al})_0$ values for UOC, CO, Acfer 094, and CR chondrites (Fig. 8b). This analysis shows that there are multiple peaks and shoulders of different widths for each chondrule type, suggesting the presence of multiple generations of chondrules. However, peaks in summed probability density distributions (Fig. 10) are dominated by analyses with relatively low uncertainties. To account for this, we proceed with a more detailed statistical analysis of the data. 1) For a given chondrite group, we calculate the mean of all measured $(^{26}\text{Al}/^{27}\text{Al})_0$ values (Fig. 8b) weighted by the measurement uncertainties, and the associated reduced chi-squared (χ^2) for this weighted mean. 2) From χ^2 we calculate a p -value, which is the probability of randomly obtaining a χ^2 at least as large as what we observed assuming that the chondrules were drawn from the same $(^{26}\text{Al}/^{27}\text{Al})_0$ reservoir. 3) We use the p -value to test our hypothesis, which for this case is: the chondrules were drawn from a single $(^{26}\text{Al}/^{27}\text{Al})_0$ reservoir. 4) If the p -value for the calculated χ^2 is greater than the usual significance level of 0.05, then we infer that the chondrules were drawn from a single generation. 5) If the p -value for one group is less than 0.05, we reject our hypothesis. 6) Then we determine the optimal two-generation grouping (by looking at all possible permutations of the measured $(^{26}\text{Al}/^{27}\text{Al})_0$ values divided into two groups) and determine the associated p -value (using Fisher's method to combine the two p -values for the two generations). 7) If this p -value is greater than 0.05, then we infer that the hypothesis that the chondrules were drawn from two generations, with ages given by the weighted-mean $(^{26}\text{Al}/^{27}\text{Al})_0$ values for the two groups, cannot be rejected. 8) If the p -value is still less than 0.05, then we calculate the optimal three-generation grouping, and so on.

If the measurements of chondrules in a chondrite group are very imprecise, we would not be able to reject the possibility that they were drawn from a single population. Similarly, improving measurement precision would result in more chondrule populations to be recognized in a given chondrite group, not fewer. For each dataset (i.e., UOCs, COs, Acfer 094, and CRs) we investigated the possibility of up to six distinct $(^{26}\text{Al}/^{27}\text{Al})_0$ chondrule populations (Figs. 11a–d) and calculated the p -values for each population (Figs. 11e–h). The optimum number of $(^{26}\text{Al}/^{27}\text{Al})_0$ populations for each dataset is the fewest groups that have a combined p -value greater than 0.05 (sub-dividing these groups further will result in larger p -values). However, we cannot rule out the possibility that there was a continuous production of chondrules, which would imply a significant number of groups.

The results of these statistical tests suggest that each chondrite group contains multiple chondrule populations based on $(^{26}\text{Al}/^{27}\text{Al})_0$ values, frequently with one dominant population (Figs. 11e,f,g,h); the p -value for a given chondrule population to be drawn from a single $(^{26}\text{Al}/^{27}\text{Al})_0$ is less than 0.05 in all cases. The UOCs, CO chondrites, and Acfer 094 (data from: Kita et al., 2000; Mostefaoui et al., 2002; Kurahashi et al., 2008b; Ushikubo et al., 2013) are all consistent with having two distinct $(^{26}\text{Al}/^{27}\text{Al})_0$ chondrule populations; with p -values of 0.069, 0.057, and 0.548, respectively (Figs. 11e–g). The $(^{26}\text{Al}/^{27}\text{Al})_0$ value of chondrules in different populations is sometimes similar, and their separation into different populations appear arbitrary (e.g., Fig. 11e); however it should be noted that while these populations are defined by their

common mean values, they are not strictly defined and overlaps of their end members into different generations is plausible. The CO chondrites have the lowest p -value for two generations and considering three generations gives a p -value of 0.447. However, the number of populations at which the p -value ‘jumps’ above the significance level of 0.05 is important and suggests two distinct $(^{26}\text{Al}/^{27}\text{Al})_0$ chondrule populations for these three chondrite groups (Figs. 11a–c). In contrast, the CR chondrites are consistent with having three distinct $(^{26}\text{Al}/^{27}\text{Al})_0$ chondrule populations; with a p -value of 0.100 (Fig. 11d). Two of the CR populations combined consist of eight chondrules, including six of the seven chondrules with resolvable $(^{26}\text{Al}/^{27}\text{Al})_0$ values. The third population consists of the 13 chondrules with unresolvable ratios and one chondrule with a low but resolved $(^{26}\text{Al}/^{27}\text{Al})_0$ (Fig. 11h). Therefore, neither the UOCs, CO chondrites, Acfer 094, or CR chondrites contain chondrules from a single $(^{26}\text{Al}/^{27}\text{Al})_0$ reservoir; suggesting either heterogeneous distribution of ^{26}Al or the presence of several populations of chondrules within a chondrite group which formed at different times in the disk with a uniform distribution of $(^{26}\text{Al}/^{27}\text{Al})_0$. This also implies that UOCs, CO chondrites, Acfer 094, and CR chondrites taken together did not form from a single $(^{26}\text{Al}/^{27}\text{Al})_0$ reservoir; consistent with previous work (e.g., Kita and Ushikubo, 2012). Furthermore, the dominant chondrule population is not always the youngest (Fig. 11e,f vs. 11g,h). However, we cannot rule out sampling bias and we recognize that we are pushing the Al-Mg chronometer to its limits (e.g., chondrules with lower $(^{26}\text{Al}/^{27}\text{Al})_0$ may be present, but we are not able to resolve them). Therefore, either 1) the final chondrule forming event triggered parent body accretion but it was not always the dominant event, or 2) the final chondrule forming event triggered parent body accretion and formed the most chondrules, however the current datasets for chondrule ages are not representative.

We infer that chondrules from the CR chondrites represent multiple generations of chondrules, but in general they have lower $(^{26}\text{Al}/^{27}\text{Al})_0$ ratios than chondrules from either the UOCs, CO chondrites, or Acfer 094 (Fig. 8b; Kita et al., 2000; Mostefaoui et al., 2002; Nagashima et al., 2007, 2008, 2014; Kurahashi et al., 2008b; Villeneuve et al. 2009; Kita and Ushikubo, 2012; Tenner et al., 2013a, 2015b; Ushikubo et al., 2013; this study). There are no systematic differences in observed chemical or stable isotope compositions between these chondrule generations. We infer that on average chondrules from the CR chondrites formed from a reservoir with low $(^{26}\text{Al}/^{27}\text{Al})_0$ ratios, although the CR chondrites contain chondrules that formed in reservoirs with higher $(^{26}\text{Al}/^{27}\text{Al})_0$ ratios than are resolvable via SIMS analyses (Nagashima et al., 2007, 2008, 2014; Kurahashi et al., 2008a; Tenner et al., 2013a, 2015b; this study).

4.3. Concordant ^{26}Al - ^{26}Mg relative and ^{207}Pb - ^{206}Pb absolute ages of CR chondrules

In this section, we use our data to test the distribution of ^{26}Al in the CR chondrule formation region in the protoplanetary disk by comparing their average ^{26}Al - ^{26}Mg ages and ^{207}Pb - ^{206}Pb ages for a set of six hand-picked CR chondrules reported by Amelin et al. (2002). We assume here that the chondrules reported by Amelin et al. (2002) are a representative population of chondrules from the CR chondrites. Despite the presence of multiple chondrule populations, we can attempt to compare the age ranges of chondrule formation determined with these two isotopic systems. Assuming a homogeneous distribution of ^{26}Al in the disk with the canonical $(^{26}\text{Al}/^{27}\text{Al})_0$ of $\sim 5.25 \times 10^{-5}$ (e.g., Jacobsen et al., 2008; Larsen et al., 2011), we can calculate a formation age after CV CAIs utilizing:

$$\Delta t \text{ (Ma)} = \ln[(^{26}\text{Al}/^{27}\text{Al})_{\text{CAI}}/(^{26}\text{Al}/^{27}\text{Al})_{\text{chondrule}}] \times t_{1/2}/\ln(2).$$

Where: Δt = time of chondrule formation after CV CAIs; $(^{26}\text{Al}/^{27}\text{Al})_{\text{CAI}}$ = the canonical $(^{26}\text{Al}/^{27}\text{Al})_0$; $(^{26}\text{Al}/^{27}\text{Al})_{\text{chondrule}} = (^{26}\text{Al}/^{27}\text{Al})_0$ for a particular chondrule; $t_{1/2}$ = half-life of ^{26}Al ~ 0.705 Ma (Norris et al., 1983). The single chondrule in group 1 (N=1, Fig. 11h) has a $(^{26}\text{Al}/^{27}\text{Al})_0$ of $(6.3 \pm 0.9) \times 10^{-6}$ (weighted mean $\pm 2\text{SE}$), corresponding to the inferred formation age of $2.2^{+0.1}_{-0.2}$ Ma (chondrule from Nagashima et al., 2014). The chondrules in group 2 (N=7) have $(^{26}\text{Al}/^{27}\text{Al})_0$ of $(3.2 \pm 0.6) \times 10^{-6}$ ($\chi^2 = 1.65$), corresponding to the inferred formation age of $2.9^{+0.2}_{-0.2}$ Ma. The $(^{26}\text{Al}/^{27}\text{Al})_0$ of the group 3 chondrules (N=14) is $(6.7 \pm 0.3) \times 10^{-7}$ ($\chi^2 = 1.38$), corresponding to the inferred formation age of $4.4^{+0.7}_{-0.4}$ Ma.

Using the $^{238}\text{U}/^{235}\text{U}$ ratio determined here for CR chondrites, and the isochron parameters reported by Amelin et al. (2002), the ^{207}Pb - ^{206}Pb age of the CR chondrules is 4563.64 ± 0.61 Ma. Comparing this age to the U-corrected ^{207}Pb - ^{206}Pb age of CV CAIs (4567.30 ± 0.16 Ma; Connelly et al., 2012), the U-corrected ^{207}Pb - ^{206}Pb age of CR chondrule formation is now 3.66 ± 0.63 Ma after CV CAI formation. Since the $^{238}\text{U}/^{235}\text{U}$ ratio used was not obtained for the same chondrules from Amelin et al. (2002) but rather bulk CR chondrites, we take a conservative approach and account for any potential U-isotope variation between chondrules, matrix, and bulk CR chondrites. This is done by including an additional uncertainty of 0.59 Ma suggested by Huyskens et al. (2016) when the U isotopic composition of chondrules can not be measured. The more conservative U-corrected ^{207}Pb - ^{206}Pb age of CR chondrule formation is 3.66 ± 0.86 Ma after CV CAI formation. The U-corrected average ^{207}Pb - ^{206}Pb age is consistent with the ^{26}Al - ^{26}Mg age of the group 3 chondrules from this study (i.e., $4.4^{+0.7}_{-0.4}$ vs. $3.7^{+0.9}_{-0.9}$ Ma after CV CAIs, respectively). The ^{207}Pb - ^{206}Pb age was determined from six hand-picked CR chondrules (Amelin et al., 2002) which were most likely sampled from both the group 2 and 3. If we combine group 2 and 3, which comprise 95% of the measured chondrules, the ^{26}Al - ^{26}Mg age of the CR chondrules is $3.7^{+0.3}_{-0.2}$. This agreement provides no evidence that initial abundance of ^{26}Al differed between the formation regions of the canonical CV CAIs and CR chondrules, and suggests a uniform distribution of ^{26}Al (within uncertainty of the reported ages) in the CR chondrule-forming region of the protoplanetary disk. These results provide support for a uniform distribution of ^{26}Al in the protoplanetary disk after CV CAIs.

Bulk isotope measurements of chondrules in the CR chondrites by MC-ICPMS and thermal ionization mass-spectrometry (TIMS) to define their ^{182}Hf - ^{182}W relative and U-corrected Pb-Pb absolute ages, will be able to further test the consistency of their relative and absolute ages.

While our results are consistent with a uniform distribution of ^{26}Al in the CR chondrite chondrule-forming region, heterogeneity of ^{26}Al in the ordinary and CV chondrite chondrule-forming regions of the protoplanetary disk was inferred by Bizzaro et al. (2014) and Bollard et al. (2015). These authors showed that correlated ^{26}Al - ^{26}Mg and U-corrected Pb-Pb studies of individual chondrules from Allende (CV3) and NWA 5697 (L3.1) show that the ^{26}Al - ^{26}Mg and U-corrected Pb-Pb chondrule ages differ up to ~ 1.5 Ma (Bizzaro et al., 2014; Bollard et al., 2015). However, the average Hf-W age for chondrules from Allende agree with Al-Mg ages of chondrules from CV chondrites (Budde et al., 2016). The disagreement between these results and our study may be the outcome of methodology (i.e., comparing means vs. coordinated analyses on individual chondrules) and/or the product of studying different chondrite groups. Until similar coordinated analyses are conducted on chondrules from CR chondrites this issue may not be

resolved, however we argue that comparing mean ages here is a valid first step. As chondrules in different chondrite groups likely accreted at different locations (e.g., Ghosh et al., 2006) and perhaps times (e.g., Villeneuve et al., 2009), it is unclear the heterogeneity inferred by Bizzarro et al. (2014) and Bollard et al. (2015) was localized or disk wide. It is also unclear if the ^{26}Al - ^{26}Mg system of chondrules in Allende were disturbed, as Allende was likely heated up to 590°C on its parent body (Busemann et al., 2007). Furthermore, discrepancies between the Al-Mg and U-Pb ages of individual chondrules may be the result of parent body thermal metamorphism disturbing the U-Pb system, and/or potential mixing of non-radiogenic Pb components (Kita et al., 2015).

Since our data are consistent with a uniform distribution of ^{26}Al in the protoplanetary disk, we proceed in our discussion with this assumption. The inferred uniform distribution of ^{26}Al in the protoplanetary disk and the observed variations in $(^{26}\text{Al}/^{27}\text{Al})_0$ in chondrules from each chondrite group studied suggest the presence of multiple chondrule generations within each of these groups, with one being dominant. However, the dominant chondrule generation is not always the last one inferred to have formed, although this may be sampling bias (e.g., Figs. 11e,f). Based on these observations, we infer that (i) accretion of a chondritic parent body did not occur immediately after chondrule formation (e.g., consistent with that suggested by Alexander et al. (2008) and Kita and Ushikubo (2012)), and (ii) some chondrules that accreted into an asteroid avoided melting by later chondrule-forming events (e.g., Figs. 11g,h), suggesting relatively small spatial scales of the chondrule-forming events. A similar conclusion was inferred based on the presence of CAIs unaffected by chondrule formation in all known chondrite groups (Krot et al., 2009).

The last chondrule to have formed in each chondrite group (i.e., the lowest $(^{26}\text{Al}/^{27}\text{Al})_0$), assuming a representative sampling, constrains the earliest age that parent body accretion could have started. This constrains accretion to $>2.8 \pm 0.9$ Ma, $>2.9^{+0.4}_{-0.2}$ Ma, and $>2.6^{+0.6}_{-0.4}$ Ma after CV CAIs, for the UOC (i.e., LL chondrite), CO chondrite, and Acfer 094 parent bodies respectively (Kita et al., 2000; Mostefaoui et al., 2002; Kurahashi et al., 2008b; Ushikubo et al., 2013). We note, however, that the late accretion of ordinary chondrite parent bodies would be difficult to reconcile with the high peak metamorphic temperatures reached (Huss et al., 2006) by these bodies, if ^{26}Al is assumed to be the major heat source. The last resolvable CR chondrule to have formed (Nagashima et al., 2014) places CR parent body accretion at $>4.0^{+0.5}_{-0.3}$ Ma after CV CAIs. However the abundance of chondrules with unresolvable ^{26}Mg excesses suggest that an age >4.5 Ma is more likely. These parent body accretion ages are within the uncertainties of with those estimated by Sugiura and Fujiya (2014), which were constrained with model calculations of the $^{26}\text{Al}/^{27}\text{Al}$ initial ratios required to reach peak metamorphic temperatures. In particular, Sugiura and Fujiya (2014) estimated parent body accretion ages of 2.1 ± 0.1 Ma for the ordinary chondrites, 2.7 ± 0.2 Ma for the CO chondrites, and 3.5 ± 0.5 Ma for the CR chondrites. Alternatively, the youngest chondrules may have been added later to a slowly accreting chondrite parent body, (e.g., during regolith gardening; e.g., Keller and Buseck, 1990), as accretion may have been a prolonged processes (e.g., Beyersdorf-Kuis et al., 2015; Johansen et al., 2015).

4.4. Consistency with parent body processes

The relatively young age of the CR chondrules suggests a low abundance of ^{26}Al in the CR parent body, which was insufficient to facilitate significant radioactive heating and subsequent parent body thermal alteration. The latter is supported by mineralogical observations indicating that (i) CR chondrites have experienced only weak parent body thermal metamorphism

(Weisberg et al., 1993; Abreu and Brearley, 2010; Schrader et al., 2011, 2014b, 2015), and (ii) the most thermally altered CR chondrites display evidence for impact generated thermal alteration (e.g., Abreu, 2011; Abreu and Bullock, 2014). The inferred young age of CR chondrule formation ($\sim 3.7^{+0.3}_{-0.2}$ [$\pm 2\text{SE}$] Ma after CV CAIs) and therefore young age of accretion for the CR chondrite parent body ($> 4.0^{+0.5}_{-0.3}$ Ma) is consistent with the young formation age of carbonates via aqueous alteration in CR chondrites (i.e., carbonate ages range from 2–15 Ma after CV CAIs; Jilly et al., 2013; Jilly-Rehak et al., 2015b), which may have been facilitated by impact heating.

5. CONCLUSIONS

From a comprehensive study of the major and minor element and O-isotope compositions of individual minerals from, and the ^{26}Al - ^{26}Mg systematics of chondrules in the CR chondrites we conclude that:

1. Only $\sim 32\%$ of chondrules in CR chondrites show resolvable excesses of ^{26}Mg , which is significantly lower than chondrules in UOCs, COs, and Acfer 094. The inferred initial $^{26}\text{Al}/^{27}\text{Al}$ ratios in CR chondrules are generally lower than those in other chondrite groups.
2. Based on the inferred $(^{26}\text{Al}/^{27}\text{Al})_0$, three generations of CR chondrules are recognized. The generation with low $(^{26}\text{Al}/^{27}\text{Al})_0$ ($< 3 \times 10^{-6}$) is dominant. There are no noticeable trends with major and minor element or O-isotope compositions between these populations. Furthermore, there is no obvious relationship between individual chondrule Fe# and $(^{26}\text{Al}/^{27}\text{Al})_0$ observed here.
3. The agreement between the average ^{26}Al - ^{26}Mg and the revised U-corrected ^{207}Pb - ^{206}Pb ages of chondrules from CR chondrites are consistent with ^{26}Al being uniformly distributed in the chondrule-forming region of the CR chondrites, and supports the use of the ^{26}Al - ^{26}Mg systematics as a chronometer for CR chondrules. Although, Al-Mg and Pb-Pb measurements have not been conducted on the same objects. Since the correlated studies of ^{26}Al - ^{26}Mg systematics and U-corrected ^{207}Pb - ^{206}Pb ages of individual chondrules from the UOC and CV chondrite groups indicate heterogeneity of ^{26}Al (e.g., Bizzaro et al., 2014; Bollard et al., 2015), similar measurements for individual CR chondrules are required to prove a uniform distribution of ^{26}Al in their formation region. We note, however, that validity of the U-corrected ^{207}Pb - ^{206}Pb ages of individual chondrules has been questioned (Kita et al., 2015).
4. The presence of multiple generations among CR chondrules indicate that some chondrules that accreted into the CR chondrite parent body avoided melting by the chondrule-forming event that produced the major population of chondrules, suggesting chondrule-forming events occurred over relatively limited spatial scales. Accretion of the CR chondrite parent body concluded no earlier than $> 4.0^{+0.5}_{-0.3}$ Ma after the formation of CV CAIs. Rapid accretion after chondrule formation is not required by our data.

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REFERENCES

- Abreu N. M. (2011) Petrographic evidence of shock metamorphism in CR2 chondrite GRO 03116 (abstract). *Meteorit. Planet. Sci.* **46**(Suppl.), A4–A4.
- Abreu N. M. and Brearley A. J. (2010) Early solar system processes recorded in the matrices of two highly pristine CR3 carbonaceous chondrites, MET 00426 and QUE 99177. *Geochim. Cosmochim. Acta* **74**, 1146–1171.
- Abreu N. M. and Bullock E. (2014) Opaque assemblages in CR2 Graves Nunataks (GRA) 06100 as indicators of shock-driven hydrothermal alteration in the CR chondrite parent body. *Meteorit. Planet. Sci.* **48**, 2406–2429.
- Alexander C. M. O'D., Grossman J. N., Ebel D. S., and Ciesla F. J. (2008) The formation conditions of chondrules and chondrites. *Science* **320**, 1617–1619.
- Alexander C. M. O'D., Howard K. T., Bowden R. and Fogel M. L. (2013) The classification of CM and CR chondrites using bulk H, C, and N abundances and isotopic compositions. *Geochim. Cosmochim. Acta* **123**, 244–260.
- Aléon J., Krot A. N. and McKeegan K. D. (2002) Calcium–aluminum-rich inclusions and amoeboid olivine aggregates from the CR carbonaceous chondrites. *Meteorit. Planet. Sci.* **37**, 1729–1755.
- Amelin Y., Krot A. N., Hutcheon I. D., and Ulyanov A. A. (2002) Lead isotopic ages of chondrules and calcium-aluminum-rich inclusions. *Science* **297**, 1678–1682.
- Amelin Y., Kaltenbach A., Iizuka T., Stirling C. H., Ireland T. R., Petaev M., and Jacobsen S. B. (2010) *Earth Planet. Sci. Lett.* **300**, 343–350.
- Armstrong J. T. (1988) Quantitative analysis of silicate and oxide minerals: Comparison of Monte Carlo, ZAF, and phi-rho-z procedures. In *Microbeam Analysis* (ed. Newbury D. E.). San Francisco Press, San Francisco. pp. 239–246.
- Arnould M., Goriely S., and Meynet G. (2006) The production of short-lived radionuclides by new non-rotating and rotating Wolf-Rayet model stars. *Astron. Astrophys.* **453**, 653–659.
- Berlin J., Jones R. H., and Brearley A. J. (2011) Fe-Mn systematics of type IIA chondrules in unequilibrated CO, CR, and ordinary chondrites. *Meteorit. Planet. Sci.* **46**, 513–533.
- Bevington P. R. and Robinson D. K. (2003) Data reduction and error analysis for the physical sciences. 3rd Edition. McGraw-Hill, New York, NY.
- Beyersdorf-Kuis U., Ott U., and Trierloff M. (2015) Early cosmic ray irradiation of chondrules and prolonged accretion of primitive meteorites. *Earth Planet. Sci. Lett.* **423**, 13–23.
- Bizzarro M. (2015) Calibrating the clock of creation – towards a high-resolution chronology of Solar System formation. *Goldschmidt Abstracts*, #310.
- Bizzarro M., Olsen M., Itoh S., Kawasaki N., Schiller M., Bonal L., Yurimoto H. (2014) Evidence for a reduced initial abundance of ²⁶Al in chondrule forming regions and implications for the accretion timescales of protoplanets. *Meteorit. Planet. Sci.* **49**(Suppl.), #A43–A43 (abstr.).
- Bollard J., Connelly J. N., and Bizzarro M. (2014) The absolute chronology of the early Solar System revisited. *Meteorit. Planet. Sci.* **49**(Suppl.), A49–A49 (abstr.).
- Bollard J., Kawasaki N., Sakamoto N., Larsen K., Wielandt D., Schiller M., Connelly J., Yurimoto H., and Bizzarro M. (2015) Early disk dynamics inferred from isotope systematics of individual chondrules. *Meteorit. Planet. Sci.* **50**(Suppl.), #5211 (abstr.).

- Brennecka G. A, Weyer S., Wadhwa M., Janney P. E., Zipfel J., Anbar A. D. (2010) $^{238}\text{U}/^{235}\text{U}$ variations in meteorites: Extant ^{247}Cm and implications for Pb-Pb dating. *Science* **327**, 449–451.
- Budde G., Kleine T., Kruijer T. S., Burkhardt C., and Metzler K. (2016) Tungsten isotopic constraints on the age and origin of chondrules. *Proc. Natl. Acad. Sci. U.S.A.* **113**, 2886–2891.
- Bullock E. S., Tenner T. J., Nakashima D., Kita N. T., MacPherson G. J., Ivanova M. A., Krot A. N., Petaev M. I., and Jacobsen S. B. (2015) High precision Al-Mg systematics of forsterite-bearing Type B CAIs. *46th Lunar and Planet. Sci. Conf.* CD-ROM (abstr. #1971).
- Busemann H., Alexander C.M.O'D., and Nittler L.R. (2007) Characterization of insoluble organic matter in primitive meteorites by microRaman spectroscopy, *Meteorit. Planet. Sci.* **42**, 1387–1416.
- Catanzaro E. J., Murphy T. J., Garner E. L., and Shields W. R. (1966) Absolute isotopic abundance ratios and atomic weight of magnesium. *J. Res. NBS. A Phys. Ch.* **70A**, 453–458.
- Chakraborty S. 1997. Rates and mechanisms of Fe-Mg interdiffusion in olivine at 980–1300°C. *Journal of Geophysical Research* **102**, 12,317–12,331.
- Chaussidon M., Robert F., and McKeegan K. D. (2006) Li and B isotopic variations in an Allende CAI: Evidence for the in situ decay of short-lived ^{10}Be and for the possible presence of the short-lived nuclide ^7Be in the early solar system. *Geochim. Cosmochim. Acta* **70**, 224–245.
- Clayton R. N., Onuma N., Grossman L., and Mayeda T. K. (1977) Distribution of the presolar component in Allende and other carbonaceous chondrites. *Earth Planet. Sci. Lett.* **34**, 209–224.
- Cody G. D., Alexander C. M. O'D., Yabuta H., Kilcoyne A. L. D., Araki T., Ade H., Dera P., Fogel M., Militzer B., and Mysen B. O. (2008) Organic thermometry for chondritic parent bodies. *Earth Planet. Sc. Lett.* **272**, 446–455.
- Connelly J. N., Bizzarro M., Krot A. N., Nordlund Å., Wielandt D., and Ivanova M. A. (2012) The absolute chronology and thermal processing of solids in the solar protoplanetary disk. *Science* **338**, 651–655.
- Connolly H. C. Jr. and Huss G. R. (2010) Compositional evolution of the protoplanetary disk: Oxygen isotopes of type-II chondrules from CR2 chondrites. *Geochim. Cosmochim. Acta* **74**, 2473–2483.
- Connolly H. C. Jr., Huss G. R., and Wasserburg G. J. (2001) On the formation of Fe-Ni metal in Renazzo-like carbonaceous chondrites. *Geochim. Cosmochim. Acta* **65**, 4567–4588.
- Duprat J. and Tatischeff V. (2007) Energetic constraints on in situ production of short-lived radionuclides in the early solar system. *Astrophys. J. Lett.* **671**, L69–L72.
- Elkins-Tanton L. T., Weiss B. P., and Zuber M. T. (2011) Chondrites as samples of differentiated planetesimals. *Earth Planet. Sci. Lett.* **305**, 1–10.
- Fisher R.A. (1932) Statistical methods for research workers, 4th Edition. (Olivier & Boyd, Edinburgh).
- Ganguly J. and Tazzoli V. (1994) Fe²⁺-Mg interdiffusion in orthopyroxene: Retrieval from the data on intracrystalline exchange reaction. *Am. Mineral.* **79**, 930–937.
- Ghosh A., McSween Jr. H.Y. (1998) A thermal model for the differentiation of asteroid 4 Vesta, based on radiogenic heating. *Icarus* **134**, 187–206.

- Ghosh A., Weidenschilling S. J., McSween H. C. Jr., and Rubin A. (2006) Asteroidal heating and thermal stratification of the asteroid belt. In *Meteorites and the Early Solar System II* (eds. D. S. Lauretta and H. Y. McSween). University of Arizona Press, Tucson, pp. 555–566.
- Goldmann A., Brennecka G., Noordmann J., Weyer S., and Wadhwa M. (2015) The uranium isotopic composition of the Earth and the Solar System. *Geochim. Cosmochim. Acta* **148**, 145–158.
- Gounelle M. and Meynet G. (2011) Short-lived radionuclides in the early solar system trace sequential star formation (abstract). In Workshop on Formation of the First Solids in the solar system, held November 7–9, 2011 in Kauai, Hawaii. LPI Contribution No. 1639. Houston, Texas: Lunar and Planetary Institute. p. 9008.
- Hart S. R. and Zindler A. (1989). Isotope fractionation laws: A test using calcium. *Int. J. Mass Spec. Ion Proc.* **89**, 287–301.
- Habfast K. (1998). Fractionation correction and multiple collectors in thermal ionization isotope ratio mass spectrometry. *Int. J. Mass Spec.* **176**, 133–48.
- Hevey P. and Sanders I. (2006) A model for planetesimal meltdown by ^{26}Al and its implications for meteorite parent bodies. *Meteorit. Planet. Sci.* **41**, 95–106.
- Howard K. T., Alexander C.M.O.'D, Schrader D. L., and Dyl K. A. (2015) Classification of hydrous meteorites (CR, CM and C2 ungrouped) by phyllosilicate fraction: PSD-XRD modal mineralogy and planetesimal environments. *Geochim. Cosmochim. Acta*. **149**, 206–222.
- Huss G. R., MacPherson G. J., Wasserburg G. J., Russell S. S., and Srinivasan G. (2001) ^{26}Al in CAIs and chondrules from unequilibrated ordinary chondrites. *Meteorit. Planet. Sci.* **36**, 975–997.
- Huss G. R., Rubin A. E. and Grossman J. N. (2006) Thermal metamorphism in chondrites. In *Meteorites and the Early Solar System II*, eds D. S. Lauretta & H. Y. McSween, Jr., University of Arizona Press, 567–586
- Ionov D. A., Harmon R. S., France-Lanord C., Greenwood P. B., and Ashchepkov I. V. (1994) Oxygen isotope composition of garnet and spinel peridotites in the continental mantle: Evidence from the Vitim xenolith suite, southern Siberia. *Geochim. Cosmochim. Acta* **58**, 1463–1470.
- Jacobsen B., Yin Q.-Z., Moynier F., Amelin Y., Krot A. N., Nagashima K., Hutcheon I. D., and Palme H. (2008) ^{26}Al - ^{26}Mg and ^{207}Pb - ^{206}Pb systematics of Allende CAIs: Canonical solar initial $^{26}\text{Al}/^{27}\text{Al}$ ratio reinstated. *Earth Planet. Sci. Lett.* **272**, 353–364.
- Jilly C. E., Huss G. R., and Nagashima K. (2013) Mn-Cr dating of secondary carbonates in CR chondrites. *44th Lunar and Planet. Sci. Conf.* CD-ROM. #2474 (abstr.).
- Jilly-Rehak C. E., Huss G. R., and Nagashima K. (2015a) Oxygen isotopes in secondary minerals in CR chondrites: Comparing components of different petrologic type. *46th Lunar and Planet. Sci. Conf.* CD-ROM. #1662 (abstr.).
- Jilly-Rehak C. E., Huss G. R., and Nagashima K. (2015b) ^{53}Mn - ^{53}Cr dating of secondary dolomite in a Renazzo (CR chondrite) dark inclusion. *78th Meeting of the Meteoritical Society*. #5136 (abstr.).
- Jogo K., Nagashima K., Hutcheon I. D., Krot A. N., and Nakamura T. (2012) Heavily metamorphosed clasts from the CV chondrite breccias Mokoia and Yamato-86009. *Meteorit. Planet. Sci.* **47**, 2251–2268.
- Johansen A., Mac Low M.-M., Lacerda P., and Bizzaro M. (2015) Growth of asteroids, planetary embryos, and Kuiper belt objects by chondrule accretion. *Science Advances* **17**, e1500109.

- Johnson C. A. and Prinz M. (1991) Chromite and olivine in type II chondrules in carbonaceous and ordinary chondrites: Implications for thermal histories and group differences. *Geochim. Cosmochim. Acta* **55**, 893–904.
- Keller L. P. and Buseck P. R. (1990) Aqueous alteration in the Kaba CV3 carbonaceous chondrite. *Geochim. Cosmochim. Acta* **54**, 2113–2120.
- Kita N. T. and Ushikubo T. (2012) Evolution of protoplanetary disk inferred from ^{26}Al chronology of individual chondrules. *Meteorit. Planet. Sci.* **47**, 1108–1119.
- Kita N. T., Nagahara H., Togashi S., and Morishita Y. (2000) A short duration of chondrule formation in the solar nebula: Evidence from ^{26}Al in Semarkona ferromagnesian chondrules. *Geochim. Cosmochim. Acta* **64**, 3913–3922.
- Kita N. T., Yin Q-Z., MacPherson G. J., Ushikubo T., Jacobsen B., Nagashima K., Kurahashi E., Krot A. N., and Jacobson S. B. (2013) ^{26}Al - ^{26}Mg isotope systematics of the first solids in the early solar system. *Meteorit. Planet. Sci.* **48**, 1383–1400.
- Kita N. T., Tenner T. J., Ushikubo T., Bouvier A., Wadhwa M., Bullock E. S., and MacPherson G. J. (2015) Why do U-Pb ages of chondrules and CAIs have more spread than their ^{26}Al ages? *78th Meeting of the Meteoritical Society*. #5360 (abstr.)
- Kleine T., Hans U., Irvin A. J., and Bourdon B. (2012) Chronology of the angrite parent body and implications for core formation in protoplanets. *Geochim. Cosmochim. Acta* **84**, 186–203.
- Krot A. N. and Keil K. (2002) Anorthite-rich chondrules in CR and CH carbonaceous chondrites: Genetic link between calcium-aluminum-rich inclusions and ferromagnesian chondrules. *Meteorit. Planet. Sci.* **37**, 91–111.
- Krot A. N., Libourel G., and Chaussidon M. (2006) Oxygen isotope compositions of chondrules in CR chondrites. *Geochim. Cosmochim. Acta* **70**, 767–779.
- Krot A. N., Makide K., Nagashima K., Huss G. R., Ogliore R. C., Ciesla F. J., Yang L., Hellebrand E., and Gaidos E. (2012) Heterogeneous distribution of ^{26}Al at the birth of the solar system: Evidence from refractory grains and inclusions. *Meteorit. Planet. Sci.* **47**, 1948–1979.
- Krot A. N., Nagashima K., van Kooten E. M. M. E., and Bizzarro M. (2015) Refractory inclusions recycled during formation of porphyritic chondrules from CH carbonaceous chondrites. *46th Lunar and Planetary Science Conference*, CD-ROM (abstr. #1596).
- Kruijer T. S., Kleine T., Fischer-Gödde M., Burkhardt C., and Wieler R. (2014) Nucleosynthetic W isotope anomalies and the Hf-W chronometry of Ca-Al-rich inclusions. *Earth Planet. Sci. Lett.* **403**, 317–327.
- Kurahashi E., Kita N. T., Nagahara H., and Morishita Y. (2008a) ^{26}Al - ^{26}Mg systematics and petrological study of chondrules in CR chondrites (abstract). *Geochim. Cosmochim. Acta* **72**, A504.
- Kurahashi E., Kita N. T., Nagahara H., and Morishita Y. (2008b) ^{26}Al - ^{26}Mg systematics of chondrules in a primitive CO chondrite. *Geochim. Cosmochim. Acta* **72**, 3865–3882.
- Larsen K. K., Trinquier A., Paton C., Schiller M., Wielandt D., Ivanova M. A., Connelly J. N., Nordlund Å., Krot A. N., and Bizzarro M. (2011) Evidence for magnesium isotope heterogeneity in the solar protoplanetary disk. *Astrophys. J.* **735**, L37.
- Liu M-C., Chaussidon M., Göpel C., and Lee T. (2012) A heterogeneous solar nebula as sampled by CM hibonite grains. *Earth Planet. Sci. Lett.* **327–328**, 75–83.
- MacPherson G. J., Davis A. M., and Zinner E. K. (1995) The distribution of aluminum-26 in the early solar system—A reappraisal. *Meteoritics* **30**, 365–386.

- MacPherson G. J., Bullock E. S., Janney P. E., Kita N. T., Ushikubo T., Davis A. M., Wadhwa M., and Krot A. N. (2010) Early solar system condensates with canonical, not supracanonical, initial $^{26}\text{Al}/^{27}\text{Al}$ ratios. *Astrophys. J. Lett.* **711**, L117–L121.
- Makide K., Nagashima K., Krot A. N., Huss G. R., Hutcheon I. D., and Bischoff A. (2009) Oxygen- and magnesium-isotope compositions of calcium–aluminum-rich inclusions from CR2 carbonaceous chondrites. *Geochim. Cosmochim. Acta* **73**, 5018–5050.
- McKeegan K. D., Chaussidon M., and Robert F. (2000) Incorporation of short-lived ^{10}Be in a calcium-aluminum-rich inclusion from the Allende meteorite. *Science* **289**, 1334–1337.
- Meyer B. S. (2005) Synthesis of short-lived radioactivities in a massive star. In Chondrites and the protoplanetary disk, edited by Krot A. N., Scott E. R. D., and Reipurth B. ASP Conference Series 341. San Francisco: Astronomical Society of the Pacific. pp. 515–526.
- Mostefaoui S., Kita N. T., Togashi S., Tachibana S., Nagahara H., and Morishita Y. (2002) The relative formation ages of ferromagnesian chondrules inferred from their initial aluminum-26/aluminum-27 ratios. *Meteorit. Planet. Sci.* **37**, 421–438.
- Nagashima K., Krot A. N., Chaussidon M. (2007) Aluminum-magnesium isotope systematics of chondrules from CR chondrites. *Meteorit. Planet. Sci.* **42**, Supplement, A115–A115.
- Nagashima K., Krot A. N., Huss G. R. (2008) ^{26}Al in chondrules from CR carbonaceous chondrites. 39th Lunar and Planet. Sci. Conf. CD-ROM. #2224 (abstr.).
- Nagashima K., Krot A. N., and Huss G. R. (2014) ^{26}Al in chondrules from CR2 chondrites. *Geochim. J.* **48**, 561–570.
- Noguchi T. (1995) Petrology and mineralogy of the PCA 91082 chondrite and its comparison with the Yamato-793495 (CR) chondrite. *Proc. NIPR Symp. Antarct. Meteorites* **8**, 33–62.
- Norris T. L., Gancarz A. J., Rokop D. J., and Thomas K. W. (1983) Half-life of ^{26}Al . *J. Geophys. Res.* **88**, B331–B333.
- Ogliore R. C., Huss G. R., Nagashima K. (2011) Ratio estimate in SIMS analysis. *Nucl. Instrum. Meth. B.* **269**, 1910–1918.
- Park C., Nagashima K., Hutcheon I. D., Wasserburg G. J., Papanastassiou D., Davis A. M., Huss G. R., and Krot A. N. (2013) Heterogeneity of Mg isotopes and variable $^{26}\text{Al}/^{27}\text{Al}$ ratio in FUN CAIs. 76th Annual Meteoritical Society Meeting, #5085(abstr.).
- Sahijpal S., Soni P., and Gupta G. (2007) Numerical simulations of the differentiation of accreting planetesimals with ^{26}Al and ^{60}Fe as the heat sources. *Meteorit. Planet. Sci.* **42**, 1529–1548.
- Schiller M., Connelly J. N., Glad A. C., Mikouchi T., and Bizzarro M. (2015) Early accretion of protoplanets inferred from a reduced inner solar system ^{26}Al inventory. *Earth Planet. Sci. Lett.* **420**, 45–54.
- Schrader D. L., Franchi I. A., Connolly H. C. Jr., Greenwood R. C., Lauretta D. S., and Gibson J. M. (2011) The formation and alteration of the Renazzo-like carbonaceous chondrites I: Implications of bulk-oxygen isotopic composition. *Geochim. Cosmochim. Acta* **75**, 308–325.
- Schrader D. L., Nagashima K., Krot A. N., Ogliore R. C., Yin Q-Z., and Amelin Y. (2013a) Testing the distribution of ^{26}Al in the protoplanetary disk using CR chondrules. 76th Meeting of the Meteoritical Society. #5141 (abstr.)
- Schrader D. L., Connolly H. C. Jr., Lauretta D. S., Nagashima K., Huss G. R., Davidson J. and Domanik K. J. (2013b) The formation and alteration of the Renazzo-like carbonaceous chondrites II: Linking O-isotope composition and oxidation state of chondrule olivine. *Geochim. Cosmochim. Acta* **101**, 302–327.

- Schrader D. L., Nagashima K., Krot A. N., Ogliore R. C., and Hellebrand E. (2014a) Variations in the O-isotope composition of gas during the formation of chondrules from the CR chondrites. *Geochim. Cosmochim. Acta*, **132**, 50–74.
- Schrader D. L., Davidson J., Greenwood R. C., Franchi I. A., and Gibson J. M. (2014b) A water-ice rich minor body from the early Solar System: The CR chondrite parent asteroid. *Earth Planet. Sci. Lett.* **407**, 48–60.
- Schrader D. L., Connolly H. C. Jr., Lauretta D. S., Zega T. J., Davidson J., and Domanik K. J. (2015). The formation and alteration of the Renazzo-like carbonaceous chondrites III: Towards understanding the genesis of ferromagnesian chondrules. *Meteorit. Planet. Sci.*, **50**, 15–50.
- Stirling, C.H., Halliday, A.N. and Porcelli, D., 2005. In search of live ^{247}Cm in the early solar system. *Geochim. Cosmochim. Acta* 69, 1059–1071.
- Stirling, C.H., Halliday, A.N., Potter, E.-K., Andersen, M.B. and Zanda, B., 2006. A low initial abundance of ^{247}Cm in the early solar system and implications for r-process nucleosynthesis. *Earth Planet. Sci. Lett.* 251, 386–397.
- Stirling, C.H., Andersen, M.B., Potter, E.-K. and Halliday, A.N., 2007. Low-temperature isotopic fractionation of uranium. *Earth Planet. Sci. Lett.* 264, 208–225.
- Sugiura N. and Fujiya W. (2014) Correlated accretion ages and $\epsilon^{54}\text{Cr}$ of meteorite parent bodies and the evolution of the solar nebula. *Meteorit. Planet. Sci.* **49**, 772–787.
- Tenner T. J., Ushikubo T., Nakashima D., Kita N. T., and Weisberg M. K. (2013a) ^{26}Al in chondrules from the CR3.0 chondrite Queen Alexandra Range 99177: A link with O isotopes. *44th Lunar and Planet. Sci. Conf.* CD-ROM. #2010 (abstr.).
- Tenner T. J., Ushikubo T., Kurahashi E., Kita N. T., and Nagahara H. (2013b) Oxygen isotope systematics of chondrule phenocrysts from the CO3.0 chondrite Yamato 81020: Evidence for two distinct oxygen reservoirs. *Geochim. Cosmochim. Acta* **102**, 226–245.
- Tenner T. J., Nakashima D., Ushikubo T., Kita N. T., and Weisberg M. K. (2015a) Oxygen isotope ratios of FeO-poor chondrules in CR3 chondrites: Influence of dust enrichment and H_2O during chondrule formation. *Geochim. Cosmochim. Acta* **148**, 228–250.
- Tenner T. J., Nakashima D., Ushikubo T., Weisberg M. K., and Kita N. T. (2015b) SIMS Al-Mg chronology of CR chondrite chondrules: Links with Mg# and O isotopes. *78th Meeting of the Meteoritical Society* #5325 (abstr.).
- Urey H.C. (1955) The cosmic abundances of potassium, uranium, and thorium and the heat balance of the Earth, the Moon, and Mars. *Proc. Natl. Acad. Sci.* **41**, 127–144.
- Ushikubo T., Kimura M., Kita N. T., and Valley J. W. (2012) Primordial oxygen isotope reservoirs of the solar nebula recorded in chondrules in Acfer 094 carbonaceous chondrite. *Geochim. Cosmochim. Acta* **90**, 242–264.
- Ushikubo T., Nakashima D., Kimura M., Tenner T. J., and Kita N. T. (2013) Contemporaneous formation of chondrules in distinct oxygen isotope reservoirs. *Geochim. Cosmochim. Acta* **109**, 280–295.
- Villeneuve J., Chaussidon M., and Libourel G. (2009) Homogeneous distribution of ^{26}Al in the solar system from the Mg isotopic composition of chondrules. *Science* **325**, 985–988.
- Wadhwa M., Bouvier A., and Janney P. E. (2013) Al-Mg systematics in a CAI from the NWA 6991 CV3 chondrite. *76th Meeting of the Meteoritical Society* #5253 (abstr.).
- Wasserburg, G., Lee, T., Papanastassiou, D. (1977) Correlated O and Mg isotopic anomalies in Allende inclusions: II. Magnesium. *Geophys. Res. Lett.* **4**, 299–302.

- Wasserburg G. J., Busso M., Gallino R., and Nollett K. M. (2006) Short-lived nuclei in the early solar system: Possible AGB sources. *Nuc. Phys. A* **777**, 5–69.
- Wasserburg G. J., Wimpenny J., and Yin Q-Z. (2012) Mg isotope heterogeneity, Al-Mg isochrones, and canonical $^{26}\text{Al}/^{27}\text{Al}$ in the early solar system. *Meteorit. Planet. Sci.* **47**, 1980–1997.
- Wasson J. T. and Rubin A. E. (2009) Composition of matrix in the CR chondrite LAP 02342. *Geochim. Cosmochim. Acta* **73**, 1436–1460.
- Wasson J. T. and Rubin A. E. (2010) Metal in CR chondrites. *Geochim. Cosmochim. Acta* **74**, 2212–2230.
- Weisberg M. K., Prinz M., Clayton R. N., and Mayeda T. K. (1993) The CR (Renazzo-like) carbonaceous chondrite group and its implications. *Geochim. Cosmochim. Acta* **57**, 1567–1586.
- Weyer S., Anbar A. D., Gerdes A., Gordon G. W., Algeo T. J., and Boyle E. A. (2008). Natural fractionation of $^{238}\text{U}/^{235}\text{U}$. *Geochim. Cosmochim. Acta* **72**, 345–359.
- Wlotzka F. (2005) Cr spinel and chromite as petrogenetic indicators in ordinary chondrites: Equilibration temperatures of petrologic types 3.7 to 6. *Meteorit. Planet. Sci.* **40**, 1673–1702.
- Zolensky M., Barrett R., and Browning L. (1993) Mineralogy and composition of matrix and chondrule rims in carbonaceous chondrites. *Geochim. Cosmochim. Acta* **57**, 3123–3148.

FIGURE CAPTIONS

Figure 1. BSE images of chondrules from this study. a) GRA 95229 Ch29A is an FeO-poor chondrule that contains spinel (sp), and is compound with Ch29B. b) GRA 95229 Ch29B is an FeO-poor chondrule that appears to be plastically deformed around Ch29A. c) GRO 03116 Ch5 is an FeO-rich chondrule that contains FeO-poor relict olivine (rel) and Na-rich plagioclase. d) GRO 03116 Ch13 is an FeO-poor chondrule. e) PCA 91082 Ch14 is an FeO-poor chondrule that is dominantly plagioclase (pl) but contains abundant spinel. The presence of spinel and the ^{16}O -enriched compositions of olivine (olv) and plagioclase (Fig. 4 and Table 5) suggest it may contain incompletely melted CAI fragments. f) PCA 91082 Ch15 is an FeO-poor chondrule. g) QUE 99177 Ch13 is an FeO-poor chondrule. h) QUE 99177 Ch14 is FeO-poor chondrule with plagioclase laths. Boxes correspond to magnified images in Figures 2 and 3. met = Fe,Ni metal, pyx = pyroxene, mx = matrix, CAI = calcium-aluminum rich inclusion, fc = fusion crust, and Ch = chondrule.

Figure 2. BSE images of selected regions of chondrules from a-d) GRA 95229, and e-h) GRO 03116. a-d) GRA 95229 Ch29A and Ch29B contain Fe,Ni-metal (met), olivine (olv), pyroxene (pyx), and plagioclase (pl). a,b) Ch29A also contains spinel (sp) and micron sized grains of silica (sil). e,f) GRO 03116 Ch5 consists predominantly of FeO-rich olivine, and minor amounts of chromite (chr), Na-rich plagioclase, and FeO-poor relict olivine (rel). g,h) GRO 03116 Ch13 is predominantly comprised of pyroxene (low- and high-Ca) and plagioclase, and minor amounts of Fe,Ni-metal, olivine, and silica. met = Fe,Ni metal, olv = olivine, rel = FeO-poor relict olivine, pyx = pyroxene, pl = plagioclase, sp = spinel, sil = silica, mx = matrix, sf = sulfide.

Figure 3. BSE images of selected regions of chondrules from a-d) PCA 91082, and e-h) QUE 99177. PCA 91082 Ch14 (a,b) and Ch15 (c,d) contain olivine (olv), pyroxene (high-Ca [Ca-pyx] and low-Ca [pyx]), plagioclase (pl), Fe,Ni-metal (met), and minor amounts of silica (sil); (b) also contains spinel (sp). QUE 99177 Ch13 (e,f) and Ch14 (g,h) contain olivine, pyroxene (low- and high-Ca), Fe,Ni-metal, plagioclase lasths, and minor amounts of silica. mx = matrix, sf = sulfide.

Figure 4. Three-isotope oxygen plots for chondrule olivine, low-Ca pyroxene, and plagioclase from: a) GRA 95229,22 Ch29A; b) GRA 95229,22 Ch29B; c) GRO 03116,15 Ch3; d) GRO 03116,15 Ch13; e) PCA 91082,15 Ch14; f) PCA 91082,15 Ch15; g) QUE 99177,6 Ch13; and h) QUE 99177,6 Ch14. With the exception of relict grains in c) GRO 03116 Ch5 and e) PCA 91082 Ch14 (shown in inset), the O-isotope composition of olivine, pyroxene, and plagioclase in each chondrule are homogeneous with respect to $\Delta^{17}\text{O}$. Terrestrial fractionation (TF) line, primitive chondrules minerals (PCM) line from Ushikubo et al. (2012), and carbonaceous chondrite anhydrous mineral (CCAM) line from Clayton et al. (1977) plotted for reference.

Figure 5. *In situ* Fe# ($\text{Fe}/(\text{Fe}+\text{Mg})\times 100$) vs. $\Delta^{17}\text{O}$ of chondrule silicates (olivine or pyroxene) for a) Fe-poor (Fe# 0–10) and Fe-rich chondrules (Fe# 10–60), and b) detail of FeO-poor chondrules. In general, FeO-poor chondrules are more ^{16}O -enriched than FeO-rich chondrules. Relict grains in FeO-rich chondrules have similar O-isotope compositions as FeO-poor chondrules. The O-isotope composition of the eight chondrules studied here [1] are consistent with chondrules from CR chondrites previously studied by [2] Schrader et al. (2013b) and [3] Schrader et al. (2014a). Error bars are 2σ .

Figure 6. Al–Mg isochron diagrams for the six chondrules with $^{27}\text{Al}/^{24}\text{Mg}$ ratios between 0 and 100; error bars on data points are 2σ . χ^2 = reduced χ^2 .

Figure 6 (continued). Al–Mg isochron diagrams for the two chondrules with $^{27}\text{Al}/^{24}\text{Mg}$ ratios between 0 and 250; error bars on data points are 2σ . g) GRO 03116,15 Ch5 is an FeO-rich chondrule with a resolvable ^{26}Mg excess corresponding to $(^{26}\text{Al}/^{27}\text{Al})_0$ of $(2.5 \pm 1.6) \times 10^{-6}$ ($\pm 2\text{SE}$).

Figure 7. Average Fa value of olivine vs. average Fs value of low-Ca pyroxene for individual chondrules, with 1σ error bars. Only chondrules with olivine and low-calcium pyroxene from this study are shown (Table 1). A linear regression through the data gives a slope 1 within 1σ uncertainties between Fs and Fa; $\text{Fs} = (1.11 \pm 0.14) \cdot \text{Fa} - (0.22 \pm 0.27)$ with a regression coefficient, $R^2 = 0.95$. The 1:1 relationship between Fa and Fs indicates that these chondrules were not thermally metamorphosed.

Figure 8. a) $(^{26}\text{Al}/^{27}\text{Al})_0$ for all chondrules from the CR chondrites of this study and Nagashima et al. (2007, 2008, 2014). b) $(^{26}\text{Al}/^{27}\text{Al})_0$ for the UOCs, CO chondrites, and Acfer 094; each group displays a range in $(^{26}\text{Al}/^{27}\text{Al})_0$. Chondrules from the CR chondrites generally have the lowest $(^{26}\text{Al}/^{27}\text{Al})_0$, and the majority of these chondrules have unresolvable ^{26}Mg excesses. Error bars are 2SE . Data from: [1] Nagashima et al. (2007, 2008, 2014), [2] this study, [3] Kita et al. (2000), [4] Mostefaoui et al. (2002), [5] Kurahashi et al. (2008b), and [6] Ushikubo et al. (2013).

Figure 9. a,b) Average Fe# (of olivine and/or low-Ca pyroxene) vs. $(^{26}\text{Al}/^{27}\text{Al})_0$ for individual chondrules with 2σ and 2SE uncertainties, respectively. b) Focus on FeO-poor chondrules. Chondrules with both olivine (black) and pyroxene (red) are plotted twice. There is no overall trend between Fe# vs. $(^{26}\text{Al}/^{27}\text{Al})_0$. Data from: [1] Nagashima et al. (2007, 2008, 2014) and [2] this study.

Figure 10. Probability density distributions of initial $(^{26}\text{Al}/^{27}\text{Al})_0$ for UOC, CO, Acfer 094, and CR (the same data as shown in Fig. 8). Estimated groupings as shown in Fig. 11 are indicated by dotted lines.

Figure 11. The p -value vs. proposed number of distinct $(^{26}\text{Al}/^{27}\text{Al})_0$ chondrule groups for a) UOCs, b) CO chondrites, c) Acfer 094, and d) CR chondrites (the group 1 p -value is 0 and therefore does not plot on this scale). Dashed line on (a–d) is the significant level (5%) used to test the hypothesis of minimum number of chondrule groups that were drawn from a single $(^{26}\text{Al}/^{27}\text{Al})_0$ reservoir. e–h) Minimum number of distinct $(^{26}\text{Al}/^{27}\text{Al})_0$ chondrule groups with corresponding p -values $> 5\%$. Error bars are 2SE . a,e) UOCs (Kita et al., 2000; Mostefaoui et al., 2002), b,f) CO chondrites (Kurahashi et al., 2008b), c,g) Acfer 094 (Ushikubo et al., 2013), and d,h) CR chondrites (Nagashima et al., 2007, 2008, 2014; this study). Weighted mean ($\pm 2\text{SE}$) $(^{26}\text{Al}/^{27}\text{Al})_0$ of all chondrules for each group: UOCs = $(0.73 \pm 0.08) \times 10^{-5}$; CO chondrites = $(0.60 \pm 0.03) \times 10^{-5}$; Acfer 094 = $(0.60 \pm 0.06) \times 10^{-5}$; CR chondrites = $(0.18 \pm 0.03) \times 10^{-5}$.

Table 1

Table 1. CR chondrule summary.																						
		Olivine											Ca-poor pyroxene									
Meteorite	Chondrule	Type	Major axis (mm)	Minor axis (mm)	(²⁶ Al/ ²⁷ Al) ₀	2SE	Fa min	Fa max	Fa avg	Fa std	# analyses	D ¹⁷ O olv weighted mean	2SE	# analyses	Fs min	Fs max	Fs avg	Fs std	# analyses	D ¹⁷ O pyx weighted mean ^a	2SE	# analyses
GRA 95229,22	Ch29A	FeO-poor	1.3	1.0	1.28E-06	1.70E-06	2.0	3.4	2.5	0.5	16	-0.4	0.7	3	2.1	2.1	2.1	0.0	1	-2.5	1.1	1
	Ch29B	FeO-poor	0.6	0.6	-4.24E-07	2.62E-06	2.1	2.2	2.2	0.0	5	-	-	-	2.0	3.0	2.5	0.3	10	-2.5	0.6	3
GRO 03116,15	Ch5	FeO-rich	0.7	0.6	2.47E-06	1.58E-06	0.4	48.1	26.7	12.5	22	(b)	(b)	4	-	-	-	-	-	-	-	-
	Ch13	FeO-poor	0.7	0.4	-8.60E-07	3.07E-06	-	-	-	-	-	-	-	-	1.4	2.8	1.7	0.3	17	-2.4	0.8	2
PCA 91082,15	Ch14	FeO-poor	1.4	1.1	-3.15E-07	3.40E-06	0.6	1.6	0.9	0.3	15	(c)	(c)	2	0.8	1.0	0.9	0.1	6	-5.2	0.8	2
	Ch15	FeO-poor	0.8	0.6	2.47E-06	3.29E-06	0.7	1.4	0.9	0.2	22	-5.2	0.7	3	-	-	-	-	-	-	-	-
QUE 99177,6	Ch13	FeO-poor	1.1	1.0	-1.96E-07	3.05E-06	1.7	2.0	1.9	0.1	7	-1.9	0.8	2	1.1	3.3	1.8	0.7	14	-2.8	0.8	2
	Ch14	FeO-poor	1.1	0.9	1.92E-06	2.76E-06	1.1	1.5	1.3	0.1	15	-	-	-	1.0	1.3	1.1	0.2	3	-3.7	0.6	3

(a) weighted mean D¹⁷O if more than 1 data point
(b) GRO 03116,15 Ch5 contains FeO-poor relict grains. D¹⁷O ranges from -0.7 to -5.1, both with 2s of 1.1.
(c) PCA 91082,15 Ch14 is likely a recycled CAI. Olivine D¹⁷O ranges from -23.1 to -6.1, both with 2s of 1.1.

Table 2

Table 2. Representative olivine analyses in ferromagnesian chondrules.

Meteorite	GRA 95229,22	GRA 95229,22	GRO 03116,15	GRO 03116,15	PCA 91082,15	PCA 91082,15	QUE 99177,6	QUE 99177,6
Chondrule	Ch29A	Ch29B	Ch5	Ch5	Ch14	Ch15	Ch13	Ch14
Grain	G1	G3	G7	G1	G1	G3	G7	G3
Point #	100	132	200	41	117	138	200	117
Grain location	olv-2	-	olv-16	olv-18	olv-7	olv2-13	olv-13	-
Ch. Type	FeO-poor	FeO-poor	FeO-rich	FeO-rich	FeO-poor	FeO-poor	FeO-poor	FeO-poor
<i>Chemical composition (wt. %)</i>								
SiO ₂	42.51	42.73	38.00	43.39	42.54	42.41	42.17	42.83
TiO ₂	bdl	0.06	bdl	0.00	bdl	bdl	0.05	bdl
Al ₂ O ₃	0.34	0.23	0.13	0.27	0.68	0.06	0.24	0.08
FeO	3.34	2.05	20.89	0.44	0.82	1.02	1.86	1.35
MnO	bdl	0.32	0.22	0.06	0.12	0.08	0.17	0.05
MgO	53.16	53.95	38.42	56.29	55.04	55.07	56.05	55.94
CaO	0.68	0.23	0.25	0.48	0.34	0.23	0.29	0.44
Na ₂ O	bdl	bdl	0.03	0.00	bdl	bdl	bdl	bdl
Cr ₂ O ₃	0.26	0.65	0.42	0.21	0.45	0.46	0.55	0.28
NiO	nd	nd	bdl	0.00	nd	nd	0.07	nd
Total	100.29	100.22	98.35	101.14	100.00	99.32	101.45	100.98
<i>Cation formula based on 4 oxygens</i>								
Si	1.008	1.009	1.000	1.007	1.000	1.005	0.986	1.000
Ti	bdl	0.001	bdl	bdl	bdl	bdl	0.001	bdl
Al	0.010	0.006	0.004	0.007	0.019	0.002	0.006	0.002
Fe	0.066	0.041	0.460	0.009	0.016	0.020	0.036	0.026
Mn	bdl	0.006	0.005	0.001	0.002	0.002	0.003	0.001
Mg	1.878	1.899	1.507	1.947	1.930	1.946	1.953	1.948
Ca	0.017	0.006	0.007	0.012	0.008	0.006	0.007	0.011
Na	bdl	bdl	0.002	bdl	bdl	bdl	bdl	bdl
Cr	0.005	0.012	0.009	0.004	0.008	0.009	0.010	0.005
Ni	nd	nd	bdl	bdl	nd	nd	0.001	nd
Total	2.984	2.981	2.993	2.987	2.984	2.988	3.005	2.995
Fe#	3.4	2.1	23.4	0.4	0.8	1.0	1.8	1.3
Mg#	96.6	97.9	76.6	99.6	99.2	99.0	98.2	98.7
Fe/Mg	0.04	0.02	0.31	0.00	0.01	0.01	0.02	0.01
Fe/Mn	-	6	93	7	7	12	11	25

Ch. = chondrule; G = silicate grain.

bdl = below detection limit.

K is bdl in all analyses here.

Standards used for silicate analyses (with detection limits in wt.%) were San Carlos olivine, augite, or anorthite for Si (0.01), sphene for Ti (0.02), plagioclase or anorthite for Al (0.01), San Carlos olivine or fayalite for Fe (0.02–0.04), garnet for Mn (0.03–0.04), San Carlos olivine for Mg (0.01), augite or sphene for Ca (0.01), albite for Na (0.01), orthoclase for K (0.01), chromite for Cr (0.02), and NiO for Ni (0.03).

Detection limits with ranges given varied between analytical sessions, but individual detection limits were applied per session.

Table 3. Representative pyroxene analyses in ferromagnesian chondrules.

Meteorite	GRA 95229,22	GRA 95229,22	GRO 03116,15	PCA 91082,15	QUE 99177,6	QUE 99177,6
Chondrule	Ch29A	Ch29B	Ch13	Ch14	Ch13	Ch14
Grain	G4	G4	G3	G2	G5	G2
Point #	118	136	113	108	106	203
Near SIMS	pyx-12	pyx-15	pyx-2	pyx-8	pyx-4	pyx-5
Ch. Type	FeO-poor	FeO-poor	FeO-poor	FeO-poor	FeO-poor	FeO-poor
<i>Chemical composition (wt.%)</i>						
SiO ₂	58.10	58.86	58.65	58.62	58.97	57.88
TiO ₂	0.33	0.23	0.14	0.47	0.21	0.20
Al ₂ O ₃	3.25	2.37	1.01	1.59	1.22	0.89
FeO	1.44	1.32	1.01	0.53	0.96	0.82
MnO	0.25	0.23	0.13	0.08	0.12	bdl
MgO	36.10	36.99	36.91	37.74	36.60	38.40
CaO	0.97	0.41	0.85	0.80	2.04	0.76
Cr ₂ O ₃	0.89	0.84	0.69	0.59	0.72	0.59
Total	101.34	101.25	99.40	100.43	100.84	99.55
<i>Cation formula based on 6 oxygens</i>						
Si	1.939	1.961	1.988	1.964	1.977	1.960
Ti	0.008	0.006	0.004	0.012	0.005	0.005
Al	0.128	0.093	0.040	0.063	0.048	0.035
Fe	0.040	0.037	0.029	0.015	0.027	0.023
Mn	0.007	0.006	0.004	0.002	0.003	bdl
Mg	1.796	1.837	1.865	1.885	1.830	1.939
Ca	0.035	0.015	0.031	0.029	0.073	0.028
Cr	0.024	0.022	0.018	0.016	0.019	0.016
Total	3.977	3.976	3.979	3.985	3.983	4.006
Fe#	2.1	2.0	1.5	0.8	1.4	1.2
Mg#	96.0	97.3	96.9	97.7	94.8	97.4
Ca#	1.9	0.8	1.6	1.5	3.8	1.4
Fe/Mg	0.02	0.02	0.02	0.01	0.01	0.01
Fe/Mn	6	6	7	6	8	-

Ch. = chondrule; G = silicate grain.

bdl = below detection limit.

Na, K, and Ni (when measured) are bdl in all analyses here.

Standards used for silicate analyses (with detection limits in wt.%) were San Carlos olivine, augite, or anorthite for Si (0.01), sphene for Ti (0.02), plagioclase or anorthite for Al (0.01), San Carlos olivine or fayalite for Fe (0.02–0.04), garnet for Mn (0.03–0.04), San Carlos olivine for Mg (0.01), augite or sphene for Ca (0.01), albite for Na (0.01), orthoclase for K (0.01), chromite for Cr (0.02), and NiO for Ni (0.03).

Detection limits with ranges given varied between analytical sessions, but individual detection limits were applied per session.

Table 4

Table 4. Representative plagioclase analyses in ferromagnesian chondrules.

Meteorite	GRA 95229,22	GRA 95229,22	GRO 03116,15	GRO 03116,15	PCA 91082,15	PCA 91082,15	QUE 99177,6	QUE 99177,6
Chondrule	Ch29A	Ch29B	Ch5	Ch13	Ch14	Ch15	Ch13	Ch14
Grain	G19	G5	G20	G7	G13	G9	G13	G8
Point #	201	203	1	202	7	31	202	3
Near SIMS	plag-3	plag-5	-	-	plag-7	-	plag-17	plag-19
Ch. Type	FeO-poor	FeO-poor	FeO-rich	FeO-poor	FeO-poor	FeO-poor	FeO-poor	FeO-poor
<i>Chemical composition (wt.%)</i>								
SiO ₂	45.59	45.09	58.05	45.65	44.93	48.68	44.91	49.20
TiO ₂	bdl	bdl	bdl	bdl	bdl	0.06	bdl	0.05
Al ₂ O ₃	33.87	33.91	25.05	32.09	34.25	32.43	32.98	32.20
FeO	0.35	1.05	1.16	0.41	bdl	0.08	0.40	0.07
MnO	bdl	0.05	nd	bdl	bdl	bdl	bdl	bdl
MgO	0.92	0.78	0.39	1.05	0.73	0.88	1.01	1.00
CaO	18.69	18.68	7.10	18.30	18.14	17.26	19.18	17.21
Na ₂ O	0.38	0.56	7.08	0.54	0.13	0.12	0.45	0.19
K ₂ O	0.03	bdl	0.18	bdl	bdl	bdl	bdl	bdl
Cr ₂ O ₃	bdl	bdl	nd	bdl	bdl	0.04	bdl	bdl
Total	99.83	100.12	98.99	98.05	98.18	99.55	98.94	99.92
<i>Cation formula based on 8 oxygens</i>								
Si	2.105	2.087	2.630	2.146	2.101	2.227	2.101	2.242
Ti	bdl	bdl	bdl	bdl	bdl	0.002	bdl	0.002
Al	1.843	1.849	1.338	1.778	1.887	1.749	1.818	1.729
Fe	0.013	0.041	0.044	0.016	bdl	0.003	0.016	0.002
Mn	bdl	0.002	nd	bdl	bdl	bdl	bdl	bdl
Mg	0.064	0.054	0.026	0.074	0.051	0.060	0.070	0.068
Ca	0.925	0.926	0.345	0.922	0.909	0.846	0.961	0.840
Na	0.034	0.051	0.622	0.049	0.012	0.010	0.041	0.016
K	0.002	0.000	0.010	0.000	0.000	0.000	0.000	0.000
Cr	bdl	bdl	nd	bdl	bdl	0.001	bdl	bdl
Total	4.986	5.009	5.015	4.984	4.960	4.900	5.007	4.899
Or	0.2	0.0	1.0	0.0	0.0	0.0	0.0	0.0
An	96.3	94.8	35.3	95.0	98.7	98.8	95.9	98.1
Ab	3.5	5.2	63.7	5.0	1.3	1.2	4.1	1.9

Ch. = chondrule; G = silicate grain.

bdl = below detection limit.

Ni (when measured) is bdl in all analyses here.

Standards used for silicate analyses (with detection limits in wt.%) were San Carlos olivine, augite, or anorthite for Si (0.01), sphene for Ti (0.02), plagioclase or anorthite for Al (0.01), San Carlos olivine or fayalite for Fe (0.02–0.04), garnet for Mn (0.03–0.04), San Carlos olivine for Mg (0.01), augite or sphene for Ca (0.01), albite for Na (0.01), orthoclase for K (0.01), chromite for Cr (0.02), and NiO for Ni (0.03).

Detection limits with ranges given varied between analytical sessions, but individual detection limits were applied per session.

Table 5

Table 5. O-isotope analysis of chondrules from the CR chondrites.

Meteorite	Ch.	Type	Grain	Spot #	Mineral Analyzed	EMPA #	Fe# near spot (olv/pyx)	d ¹⁸ O	2s	d ¹⁷ O	2s	D ¹⁷ O	2s
GRA 95229,22	Ch29A	FeO-poor	G1	olv-1	olv	104	3.1	2.1	1.0	0.6	1.0	-0.5	1.1
			G1	olv-2	olv	100	3.4	2.3	1.0	1.3	1.0	0.1	1.1
			G1	olv-3	olv	103	3.2	1.8	1.0	0.3	1.0	-0.7	1.1
			G4	pyx-12	pyx	118	2.1	-1.6	1.6	-3.3	0.7	-2.5	1.1
			-	plag-2	plag	39	-	0.3	1.1	-1.3	0.9	-1.5	1.1
			-	plag-3	plag	201	-	0.7	1.1	-2.4	0.9	-2.7	1.1
			-	plag-4	plag	202	-	1.4	1.1	-1.3	1.0	-2.1	1.1
GRA 95229,22	Ch29B	FeO-poor	G1	pyx-13	pyx	122	2.3	0.5	1.6	-2.6	0.6	-2.8	1.1
			G2	pyx-14	pyx	128	3.0	0.8	1.6	-2.1	0.7	-2.5	1.1
			G4	pyx-15	pyx	136	2.0	0.6	1.6	-1.8	0.7	-2.1	1.1
			G5	plag-5	plag+pyx+sp	203		1.4	1.0	-0.8	1.0	-1.5	1.1
GRO 03116,15	Ch5	FeO-rich	G7	olv-15	olv	201	16.1	1.1	1.0	-0.2	1.0	-0.7	1.1
			G7	olv-16	olv	200	23.4	1.1	1.0	-0.4	0.9	-1.0	1.1
			G14	olv-17	olv	66	10.8	-4.7	1.0	-5.8	1.0	-3.4	1.1
			G1	olv-18	olv	41	0.4	-6.3	1.1	-8.4	1.0	-5.1	1.1
GRO 03116,15	Ch13	FeO-poor	G2	pyx-1	pyx	107	1.6	1.6	1.6	-1.3	0.7	-2.2	1.1
			G3	pyx-2	pyx	113	1.5	1.8	1.6	-1.7	0.7	-2.6	1.1
PCA 91082,15	Ch14	FeO-poor	G1	olv-7	olv	117	0.8	-43.3	1.0	-45.6	1.0	-23.1	1.1
			G1	olv-8	olv/pyx/plag	202	0.9	-4.3	1.0	-8.5	1.0	-6.3	1.1
			G2	pyx-8	pyx	108	0.8	-3.4	1.6	-6.9	0.7	-5.1	1.1
			G2	pyx-9	pyx	104	0.9	-3.2	1.6	-6.9	0.7	-5.2	1.1
			-	plag-6	plag	-	-	-5.3	1.0	-8.5	1.0	-5.7	1.1
			-	plag-7	plag	7	-	-4.6	1.1	-8.0	1.0	-5.6	1.1
			-	plag-8	plag	201	-	-15.2	1.1	-18.6	1.0	-10.7	1.1
PCA 91082,15	Ch15	FeO-poor	G3	olv2-13	olv	138	1.0	1.4	1.5	-4.8	0.9	-5.5	1.2
			G5	olv2-14	olv	205	1.4	1.8	1.5	-4.6	0.8	-5.5	1.1
			G6	olv2-15	olv+Ca-pyx	206	0.8	1.2	1.5	-4.0	0.8	-4.7	1.1
			-	plag-11	plag	-	-	-0.8	1.2	-5.2	1.2	-4.8	1.3
			-	plag-12	plag	-	-	-1.3	1.0	-5.3	0.9	-4.6	1.1
			-	plag-16	plag	-	-	0.0	1.0	-5.2	0.9	-5.2	1.1
QUE 99177,6	Ch13	FeO-poor	G7	olv-13	olv	200	1.8	0.5	1.0	-2.1	1.0	-2.4	1.1
			G8	olv-14	olv+pyx	201	1.7	-0.9	1.0	-2.0	1.0	-1.5	1.1
			G3	pyx-3	pyx	100	1.4	0.4	1.6	-2.7	0.8	-3.0	1.1
			G4	pyx-4	pyx	106	1.4	0.0	1.6	-2.6	0.7	-2.6	1.1
			-	plag-17	plag	202	-	0.3	1.0	-2.7	1.0	-2.9	1.1
			-	plag-18	plag	20	-	0.8	1.1	-2.3	1.0	-2.8	1.1
QUE 99177,6	Ch14	FeO-poor	G2	pyx-5	pyx	203	1.2	-2.3	1.6	-4.9	0.8	-3.6	1.1
			G2	pyx-6	pyx	204	1.3	-2.3	1.6	-4.6	0.7	-3.4	1.1
			G1	pyx-7	pyx	205	1.0	-1.9	1.6	-4.9	0.7	-3.9	1.1
			-	plag-19	plag	3	-	-2.1	1.1	-4.6	1.0	-3.6	1.1
			-	plag-20	plag	2	-	-1.2	1.0	-4.7	0.9	-4.0	1.1

Ch. = Chondrule. Fe# = iron number, olivine or lo-Ca pyroxene. Fe# near spot is nearest EMPA data point to the center of the SIMS pit.

EMPA # and Fe# correspond to analyses within individual meteorites. This number corresponds to data in EA-1.

The external reproducibility of the San Carlos olivine standard was $\pm 0.89\text{‰}$, $\pm 0.96\text{‰}$, and $\pm 1.09\text{‰}$ for $\delta^{17}\text{O}$, $\delta^{18}\text{O}$, and $\Delta^{17}\text{O}$ (2σ), respectively, over a two day session.

The external reproducibility for the orthopyroxene standard was $\pm 0.57\text{‰}$, $\pm 1.58\text{‰}$, and $\pm 0.83\text{‰}$ for $\delta^{17}\text{O}$, $\delta^{18}\text{O}$, and $\Delta^{17}\text{O}$ respectively, over a single day session.

The external reproducibility for the pyroxene standard was $\pm 0.85\text{‰}$, $\pm 0.96\text{‰}$, $\pm 0.85\text{‰}$ for $\delta^{17}\text{O}$, $\delta^{18}\text{O}$, and $\Delta^{17}\text{O}$, respectively, over single day session.

Table 6

Table 6. Results of Al-Mg isotope analyses of chondrules from CR chondrites.

Meteorite	Chondrule (Type)	Phase	$^{27}\text{Al}/^{24}\text{Mg}$ (2s)	$\delta^{26}\text{Mg}^*$ (‰) (2s)	$(^{26}\text{Al}/^{27}\text{Al})_0 \times 10^{-6}$ (2SE)	Time after CV CAIs (Myr)
GRA 95229,22	Ch29A (FeO-poor)	plag	90.9 ± 2.2	1.0 ± 2.7	1.3 ± 1.7	>3.5
		plag	78.0 ± 1.9	1.9 ± 2.2		
		plag	88.4 ± 2.2	0.7 ± 2.3		
		plag	79.2 ± 1.9	0.5 ± 2.5		
		plag	80.9 ± 1.8	0.1 ± 1.8		
		olv	0.008 ± 0.001	-0.04 ± 0.13		
		olv	0.024 ± 0.002	0.04 ± 0.13		
GRA 95229,22	Ch29B (FeO-poor)	plag	70.3 ± 1.6	-1.0 ± 2.0	-0.4 ± 2.6	>3.0
		plag	78.4 ± 1.8	0.3 ± 2.9		
		plag	30.1 ± 2.2	0.0 ± 2.7		
		plag	65.7 ± 1.5	0.6 ± 2.4		
		olv	0.0024 ± 0.0002	0.00 ± 0.14		
GRO 03116,15	Ch5 (FeO-rich)	plag	173.2 ± 6.6	2.4 ± 3.4	2.5 ± 1.6	3.1 (-0.5/+1.1)
		plag	172.1 ± 6.5	2.8 ± 3.0		
		plag	122.7 ± 4.8	2.7 ± 2.9		
		olv	0.0012 ± 0.0001	-0.15 ± 0.13		
		olv	0.0013 ± 0.0001	-0.16 ± 0.12		
GRO 03116,15	Ch13 (FeO-poor)	plag	35.9 ± 0.8	0.2 ± 1.7	-0.9 ± 3.1	>2.9
		plag	40.3 ± 0.9	-0.4 ± 1.8		
		plag	41.3 ± 0.9	-1.3 ± 1.6		
		plag	35.0 ± 0.8	0.7 ± 1.6		
		lpx	0.0195 ± 0.0004	-0.01 ± 0.10		
		lpx	0.025 ± 0.001	-0.01 ± 0.08		
PCA 91082,15	Ch14 (FeO-poor)	plag	92.7 ± 3.9	1.6 ± 4.2	-0.3 ± 3.4	>2.8
		plag	85.1 ± 3.6	-3.1 ± 3.7		
		plag	199.3 ± 9.5	2.2 ± 7.7		
		olv	0.0017 ± 0.0001	0.02 ± 0.14		
		olv	0.0008 ± 0.0001	-0.04 ± 0.13		
PCA 91082,15	Ch15 (FeO-poor)	plag	38.5 ± 1.6	0.5 ± 2.9	2.5 ± 3.3	>2.8
		plag	47.0 ± 2.0	1.5 ± 2.5		
		plag	43.3 ± 0.9	0.2 ± 1.4		
		plag	41.0 ± 0.9	1.4 ± 4.6		
		plag	41.7 ± 0.9	1.5 ± 2.5		
		olv	0.0012 ± 0.0001	-0.05 ± 0.21		
		olv	0.0009 ± 0.0001	-0.03 ± 0.137		
QUE 99177,6	Ch13 (FeO-poor)	plag	42.2 ± 1.1	-1.3 ± 1.7	-0.2 ± 3.0	>2.9
		plag	41.0 ± 1.0	0.7 ± 1.5		
		plag	42.6 ± 1.1	-0.1 ± 1.5		
		olv	0.0054 ± 0.0004	-0.14 ± 0.13		
		low-Ca px	0.02 ± 0.0005	-0.03 ± 0.10		
QUE 99177,6	Ch14 (FeO-poor)	plag	43.2 ± 0.9	2.3 ± 2.8	1.9 ± 2.8	>3.0
		plag	45.3 ± 1.0	1.4 ± 1.8		
		plag	39.2 ± 1.0	0.9 ± 1.9		
		plag	35.0 ± 0.9	0.1 ± 1.4		
		plag	37.2 ± 1.0	-1.1 ± 1.5		
		olv	0.018 ± 0.001	-0.15 ± 0.13		
		olv	0.0015 ± 0.0001	-0.06 ± 0.15		

Ch = chondrule; olv = olivine; lo-Ca px = low calcium pyroxene; plag = plagioclase.

If no excess or unresolvable excess of ^{26}Mg , minimum ages are calculated from 2s uncertainties as upper limits.

Table 7

Table 7. Uranium isotopic composition of CR chondrites.

#	Meteorite	$^{238}\text{U}/^{235}\text{U}$	2SD
1	Acfer 059	137.787	0.044
2	Acfer 059	137.815	0.055
3	Acfer 059	137.765	0.029
4	Acfer 059	137.818	0.028
5	Acfer 059	137.780	0.016
6	Acfer 059	137.779	0.016
7	Acfer 059	137.779	0.016
8	NWA 721	137.777	0.021
9	NWA 801	137.777	0.030
10	NWA 801	137.774	0.016
	Acfer 059 wtd. average	137.783	0.013
	NWA 721 single analysis	137.777	0.021
	NWA 801 wtd. average	137.775	0.014
	CR chondrite wtd. average	137.7789	0.0085

wtd. = weighted; SD= stadard deviation.

Figure 1

Figure 2

Figure 3

Figure 4

Figure 6

Figure 5

Figure 6 continued

Figure 7

Figure 8

Figure 9

Figure 11