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Initial $^{26}\text{Al}/^{27}\text{Al}$ in carbonaceous-chondrite chondrules: Too little ^{26}Al to melt asteroids

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Abstract—We report ^{26}Mg excesses correlated with Al/Mg ratios in five chondrules from the primitive CO3.0 chondrite Yamato 81020 that yield a mean initial $^{26}\text{Al}/^{27}\text{Al}$ ratio of only $(3.8 \pm 0.7) \times 10^{-6}$, about half that of ordinary chondrite (OC) chondrules. Even if asteroids formed immediately after chondrule formation, this ratio and the mean Al content of CO chondrites is only capable of raising the temperature of a well-insulated CO asteroid to 940 K, which is more than 560 K too low to produce differentiation. The same ratio combined with the higher Al content of CV chondrites results in a CV asteroid temperature of 1100 K. We calculate that the mean initial $^{26}\text{Al}/^{27}\text{Al}$ ratio of about 7.4×10^{-6} found in LL chondrules is only able to produce small amounts of melting, too little to produce differentiation. These results cast serious doubt on the viability of ^{26}Al as the heat source responsible for asteroid differentiation. Inclusion of ^{60}Fe raises temperatures about 160 K, but this increment is not enough to cause differentiation, even of an LL-chondrite asteroid. Copyright © 2004 Elsevier Ltd

1. INTRODUCTION

The refractory calcium-aluminum-rich inclusions (CAIs) common in most carbonaceous chondrites (CC) generally show excesses of ^{26}Mg indicative of extinct ^{26}Al ($t_{1/2} = 720$ ka); observed initial $^{26}\text{Al}/^{27}\text{Al}$ ratios [$(^{26}\text{Al}/^{27}\text{Al})_0$] are typically $(4\text{--}5) \times 10^{-5}$ (e.g., MacPherson et al., 1995). Recent studies of chondrules in ordinary chondrites (OC) have also shown small excesses in ^{26}Mg that correlate with the Al/Mg ratio. Kita et al. (2000) reported isochrons for OC chondrules with a mean initial $^{26}\text{Al}/^{27}\text{Al}$ ratio of 7.4×10^{-6} ; other less-precise studies by Hutcheon and Hutchison (1989), Russell et al. (1996), McKeegan et al. (2000), Huss et al. (2001), and Mostefaoui et al. (2002) yielded similar ratios of $(5\text{--}10) \times 10^{-6}$, approximately 5 to 10 times lower than those in CAIs.

Although several studies of Mg isotopes in (mostly Al-rich) chondrules from CC have been published, most $(^{26}\text{Al}/^{27}\text{Al})_0$ values were below the detection limits. Resolvable excesses in ^{26}Mg were derived for only one or two chondrules from each study (Sheng et al., 1991; Galy et al., 2000; Hutcheon et al., 2000; Srinivasan et al., 2000; Yurimoto and Wasson, 2002; Hsu et al., 2003); the mean of inferred $(^{26}\text{Al}/^{27}\text{Al})_0$ ratios is similar to the lowest ratios reported for OC.

Largely because there is no other accepted heat source, ^{26}Al is widely assumed to have been responsible for melting asteroids to form differentiated meteorites (e.g., Urey, 1955; Srinivasan et al., 1999; Huss et al., 2001; Zinner and Göpel, 2002). Several teams have modeled the interior thermal evolution of asteroids with ^{26}Al as a heat source (e.g., Miyamoto et al., 1981; Miyamoto, 1991; Grimm and McSween, 1993; Ghosh and McSween, 1998; LaTourrette and Wasserburg, 1998; Merk et al., 2002). Such models typically assume an initial abundance of ^{26}Al , and it is recognized that this assumption critically depends on the time at which planetesimals were accreting relative to the time at which CAIs formed (e.g., LaTourrette

and Wasserburg, 1998). One way to try to constrain this time estimate is to examine the isotopic record preserved in pristine chondrules. Chondrules are clearly nebular products, and under the assumption that the chondrule formation epoch(s) preceded the accretion of appreciably sized asteroids, the minimum value of $^{26}\text{Al}/^{27}\text{Al}$ observed in chondrules represents the best estimate of the maximum $^{26}\text{Al}/^{27}\text{Al}$ present when asteroids formed. To obtain such a constraint it is, of course, necessary that the isotopic record reflects nebular processes and not isotopic closure achieved at lower temperatures over a protracted period during cooling in an asteroidal setting. For these reasons, it is important to examine the ^{26}Al record in pristine CC chondrules. Because parent-body thermal metamorphism and aqueous alteration play significant roles in the preservation of Al-Mg systematics, we studied one of the most primitive and unaltered carbonaceous chondrites, CO3.0 Yamato 81020 (Kojima et al., 1995).

2. EXPERIMENTAL PROCEDURES

2.1. Petrographic Techniques

We studied chondrules in section 51–1 of Yamato 81020 (Y-81020) prepared by the National Institute of Polar Research (NIPR) in Tokyo. A careful search of the section revealed only a handful of chondrules containing plagioclase coarse enough ($\geq 15 \mu\text{m}$ across) to analyze with our ion microprobe; these chondrules all have high FeO contents. Spots having high Al/Mg ratios were located by X-ray mapping using a JEOL JXA-8200 electron microprobe analyzer. Mineral compositions were determined using a wavelength-dispersive system. To minimize volatile-element loss in plagioclase, the 15 keV electron beam was defocused to a diameter of $2 \mu\text{m}$, and the intensity was held to 5 nA. In some regions elemental maps were created.

Backscattered-electron (BSE) images of the chondrules were made with a LEO-1430VP scanning electron microscope.

2.2. Ion-Microprobe Techniques

Mg isotope measurements were performed by secondary ion mass spectrometry (SIMS) with the Cameca IMS-1270 instrument at the University of California, Los Angeles (UCLA). A primary ion beam of 23 keV $^{16}\text{O}_2^+$ was focused to a diameter of 5 to 10 μm with an ion intensity of 30 to 90 pA. An optical square gate (field aperture)

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corresponding to a $30 \times 30 \mu\text{m}$ image field on the sample surface reduced contributions to the Mg ion signal from adjacent Mg-rich silicates. Positive secondary ions $^{24}\text{Mg}^+$, $^{25}\text{Mg}^+$, $^{26}\text{Mg}^+$, and $^{27}\text{Al}^+$ were analyzed in a peak-jumping mode at a mass resolution of ~ 4000 , sufficient to eliminate molecular ion interferences. A typical run consisted of 100 blocks of four cycles each (1, 6, 4 s integration times for Mg ions and 2 s for Al ions). The Mg^+ ion intensities were determined by pulse-counting with an electron multiplier, and the Al^+ ions were collected on a Faraday cup. The $^{24}\text{Mg}^+$ secondary ion intensity was between 2×10^4 and 8×10^4 counts per s.

Appropriate corrections were made for detector deadtime. The detection efficiency of the electron multiplier was calibrated against the Faraday cup at a count rate of $\sim 10^6$ per s. A relative sensitivity factor (≈ 1.2) for converting measured $^{27}\text{Al}^+ / ^{24}\text{Mg}^+$ ion ratios to atomic values was determined by analyses of a synthetic glass standard. This glass has an anorthite composition with a small amount of Mg added (Armstrong et al., 1982); it contains 0.15 wt.% MgO, similar to the MgO contents in our plagioclase samples, 0.1 to 0.2 wt.%.

The Mg-isotopic ratios of olivine were measured with a multi-collection system using Faraday cups. The primary ion beam was $50 \mu\text{m}$ in diameter with 25 nA intensity. The $^{24}\text{Mg}^+$ intensity was between 1×10^8 and 3×10^8 counts per s equivalent.

Measured Mg-isotope ratios were corrected for instrumental mass fractionation by a linear fractionation law normalized to the terrestrial ratios: $^{25}\text{Mg}/^{24}\text{Mg} = 0.12663$, $^{26}\text{Mg}/^{24}\text{Mg} = 0.13932$ (Catanzaro et al., 1966). Excesses of radiogenic ^{26}Mg ($\delta^{26}\text{Mg}^* \neq 0$) are defined relative to the terrestrial values.

3. PRISTINE NATURE OF THE YAMATO 81020 CO3.0 CHONDRITE

Y-81020 was chosen for this study because it is the least-equilibrated CO3 chondrite and one of the carbonaceous chondrites that best preserves the nebular record (Kojima et al., 1995; Wasson et al., 2001; Yurimoto and Wasson, 2002; Chizmadia et al., 2002; Wasson and Rubin, 2003; Itoh and Yurimoto, 2003). Chondrules in Y-81020 are essentially unmodified by thermal metamorphism or aqueous alteration. Chizmadia et al. (2002) showed that, even at very low degrees of metamorphism, thin fayalitic rims surround forsterite grains in amoeboid olivine inclusions (AOI) in most CO3 chondrites. Although every forsterite grain in CO3.2 Kainsaz AOI is rimmed by fayalite, no forsterite grain in Y-81020 AOI has a fayalitic rim.

Grossman and Rubin (1999) showed that in type-3 chondrites, sulfur leaves the matrix and enters sulfide grains with increasing petrologic subtype. Semarkona (LL3.0), the least-metamorphosed ordinary chondrite (OC), has a sulfur-rich matrix; matrices in more-equilibrated type-3 OC contain significantly less sulfur. Grossman and Rubin (1999) found that Y-81020 has a sulfur-rich matrix analogous to that of Semarkona; most other CO3 chondrites have matrices that contain much less sulfur.

Many porphyritic chondrules in Y-81020 contain clear isotropic glass. Moderate thermal metamorphism would have caused the glass to devitrify and recrystallize. More than 90% of the type-II (high-FeO) porphyritic chondrules in Y-81020 contain low-FeO relict grains with thin ($\sim 5 \mu\text{m}$) overgrowths of fayalitic olivine that have steep compositional gradients (Wasson and Rubin, 2003). Thermal metamorphism would have caused substantial cation diffusion that would have flattened the compositional gradients and equilibrated the olivine grains.

Refractory inclusions in CO3 chondrites exhibit systematic mineralogical changes with increasing metamorphic grade. The

least-equilibrated CO3 chondrites contain melilite-rich inclusions with perovskite and FeO-poor spinel; metamorphosed CO chondrites contain refractory inclusions with feldspathoids, pyroxene, FeO-rich spinel and ilmenite, but little melilite and perovskite (Kojima et al., 1995; Russell et al., 1998). The refractory inclusions in Y-81020 show no evidence of metamorphism; they consist mainly of melilite and FeO-poor spinel, and many contain perovskite (Kojima et al., 1995). We also note that, with rare exceptions, CAIs and AOIs show $\Delta^{17}\text{O}$ ($=\delta^{17}\text{O} - 0.52 \cdot \delta^{18}\text{O}$) < -20 [ppt] in all mineral phases in Y-81020 (Wasson et al., 2001; Itoh et al., 2002). This offers additional evidence that the isotopic record in Y-81020 is well preserved and is relatively free from disturbances due to parent-body metamorphism or aqueous alteration.

Careful study of BSE images of two Y-81020 thin sections shows that it is not a breccia; there are no apparent clasts $> 800 \mu\text{m}$ in the rock. In a BSE survey of 400 type-I porphyritic chondrules in Y-81020, none of them contained high-FeO olivine grains as would be expected if these chondrules were incorporated after having been thermally metamorphosed elsewhere. Because type-I chondrules outnumber type-II chondrules, the apparent absence of metamorphosed type-I chondrules implies that very few (or none) of the type-II chondrules are expected to have been metamorphosed. These properties confirm that Y-81020 and its chondrules experienced only negligible thermal metamorphism.

4. RESULTS

4.1. Petrographic Observations

An X-ray survey of a $6 \times 9 \text{ mm}$ area revealed five high-FeO type-II porphyritic olivine chondrules (Fig. 1) that contain adequately coarse crystalline plagioclase (Ab57–87). Petrographic details are in the Appendix. The elemental compositions of the plagioclase grains analyzed by electron microprobe are presented in Table 1.

4.2. Magnesium Isotopes

A total of 28 measurements were made on eight different grains of plagioclase in the five chondrules (three to eight measurements per chondrule; Table 2). Most spots had a small portion of the beam overlapping olivine or Ca-rich pyroxene; the resulting $^{27}\text{Al}/^{24}\text{Mg}$ ratios were between 100 and 200. To estimate the initial Mg-isotopic composition of these chondrules, we also analyzed olivine in four of the chondrules. $\delta^{26}\text{Mg}^*$ values of the olivine are the same as those in the terrestrial standard ($\equiv 0$) to within 0.1 ‰. The olivine grains in chondrule G7g are smaller than the primary ion beam; the analyzed spot was mainly located on two olivine grains ($> 90\%$) with a minor portion overlapping on glass and plagioclase.

The $^{26}\text{Al}/^{26}\text{Mg}$ isochron diagrams are shown in Figure 2. Initial $^{26}\text{Al}/^{27}\text{Al}$ ratios were inferred for each chondrule by the slopes of the correlation line fitted by weighted-least squares (Williamson, 1968; Fig. 2). Even though no olivine analysis was made for C6w, it seems safe to assume that the intercept of the correlation diagram is the same as those of the other chondrules, i.e., it is probably within 0.1[pctpt] of the terrestrial

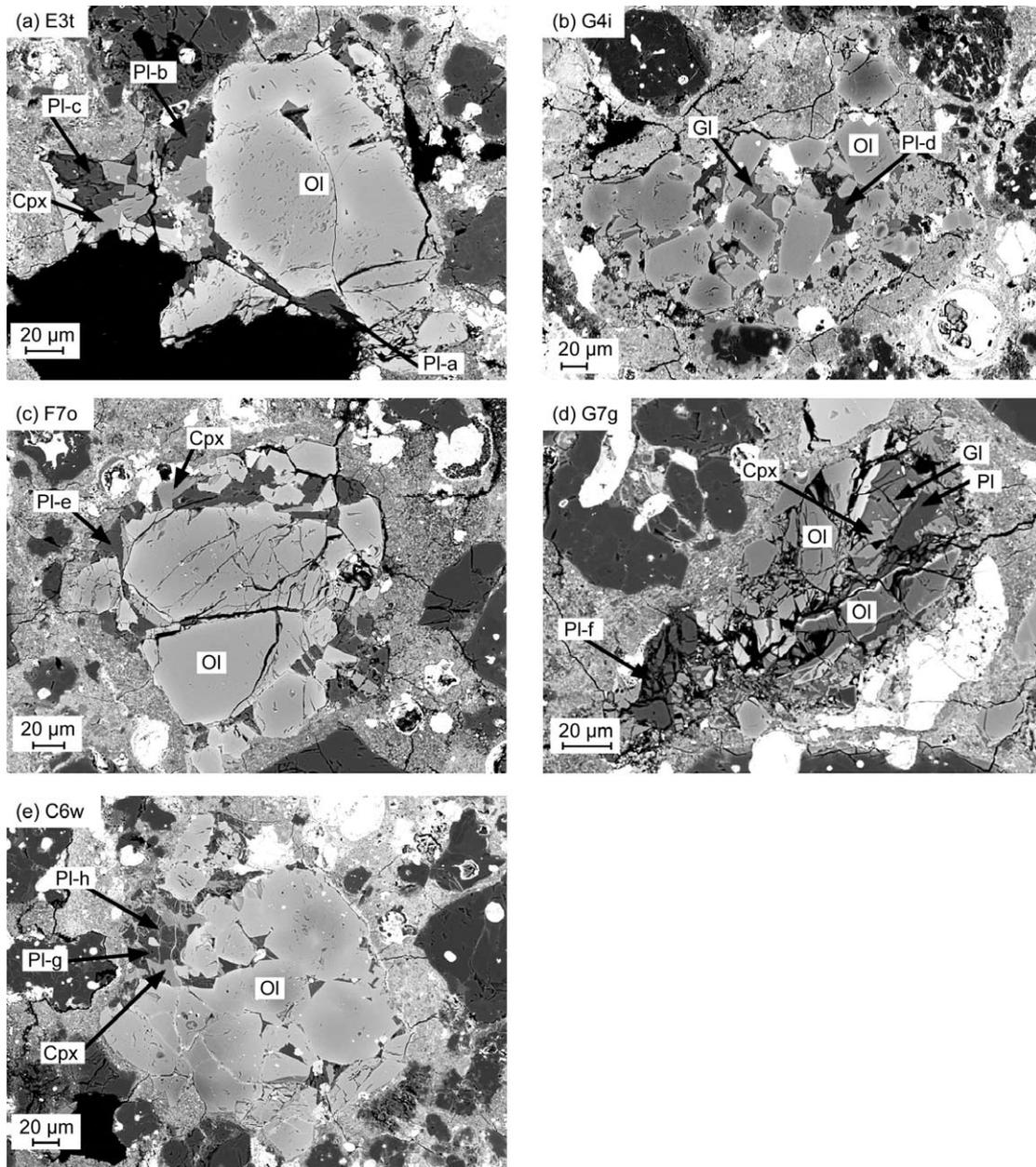


Figure 1. Backscattered electron images of five high-FeO type-II porphyritic olivine chondrules that contain coarse crystalline plagioclase (Ab57–87). Arrows show locations of plagioclase grains analyzed by SIMS. (a) E3t (b) G4i (c) F7o (d) G7g (e) C6w. The abbreviations Ol, Pl, Gl, Cpx denote olivine, plagioclase, glass, and Ca-rich pyroxene, respectively.

standard. All five chondrules in which the plagioclase was coarse enough to analyze contain resolved ^{26}Mg excesses that are well-correlated with Al/Mg. The estimated initial $^{26}\text{Al}/^{27}\text{Al}$ ratios with 2σ uncertainties of chondrules E3t, G4i, F7o, G7g, and C6w are $(3.7 \pm 1.1) \times 10^{-6}$, $(4.0 \pm 1.9) \times 10^{-6}$, $(4.2 \pm 1.3) \times 10^{-6}$, $(2.4 \pm 1.7) \times 10^{-6}$, and $(6.5 \pm 3.2) \times 10^{-6}$, respectively. It is noteworthy that we found excess ^{26}Mg correlated with the Al/Mg ratio in all chondrules analyzed. In four of five chondrules the positive trend is recognizable in the plagioclase alone. Although statistics are still limited, it is clear that ^{26}Al was alive at the time CO carbonaceous-chondrite chondrules formed.

5. DISCUSSION

5.1. Studies of ^{26}Al in Chondrules

Past reports of ^{26}Mg excesses correlated with Al/Mg ratios in chondrules have largely been confined to OC. All five ferromagnesian chondrules in LL3.0 Semarkona analyzed by Kita et al. (2000) showed clear ^{26}Mg excesses. Mostefaoui et al. (2002) also reported ^{26}Mg excesses from several chondrules from LL3.1 Bishunpur, but isochrons were less well-constrained and some chondrules did not yield resolvable excesses in ^{26}Mg . Similar results were obtained from one Al-rich chon-

Table 1. Composition (wt. %) of plagioclase in chondrules from Yamato 81020.

Chondrule	SiO ₂	Cr ₂ O ₃	Na ₂ O	MnO	K ₂ O	Al ₂ O ₃	TiO ₂	MgO	FeO	CaO	Total	Composition
E3t-a	64.9	<0.03	9.7	0.05	0.3	21.0	<0.03	0.1	1.0	3.3	100.3	Ab83 Or1.6
E3t-b	65.5	<0.03	10.3	0.06	0.2	20.5	0.04	0.1	1.0	2.5	100.2	Ab87 Or1.2
E3t-c	63.5	<0.03	9.6	<0.03	0.1	21.7	<0.03	0.1	1.2	4.1	100.4	Ab80 Or0.5
G4i-d	57.9	0.07	6.9	<0.03	0.2	25.2	0.09	0.2	1.1	8.5	100.2	Ab59 Or1.0
F7o-e	58.2	<0.03	6.8	<0.03	0.1	26.0	0.05	0.2	1.2	9.0	101.5	Ab57 Or0.7
G7g-f	60.2	0.06	8.3	<0.03	0.3	23.0	0.05	0.1	1.0	6.0	99.1	Ab70 Or1.8
D6w-g	61.9	<0.03	9.1	<0.03	0.1	22.3	0.06	0.1	1.0	5.7	100.3	Ab74 Or0.7
D6w-h	57.5	0.06	8.9	0.06	0.1	20.8	0.08	0.2	2.6	7.1	97.3	Ab69 Or0.5

drule in Semarkona by Hutcheon and Hutchison (1989), and ferromagnesian chondrules in Semarkona and Bishunpur by McKeegan et al. (2000). Studies of L/LL3.4 Chainpur revealed ²⁶Mg excesses in only one of seven Al-rich chondrules (Russell et al., 1996; Huss et al., 2001); ²⁶Mg excesses were not resolved in one Al-rich chondrule in Semarkona (Huss et al., 2001).

The only extensive study of CC chondrules is by Sheng et al. (1991); they observed ²⁶Mg excesses in two of thirteen Al-rich chondrules, mostly called POI (plagioclase-olivine inclusions),

Table 2. Mg-isotopic compositions of plagioclase and olivine in chondrules of Yamato 81020.

Chondrule/Phase	²⁷ Al/ ²⁴ Mg ± 2σ	δ ²⁶ Mg* ± 2σ ^M (‰)
E3t-pl-a	110 ± 3	-0.5 ± 4.1
	86 ± 6	1.5 ± 2.7
E3t-pl-b	187 ± 12	4.9 ± 4.3
	175 ± 5	4.9 ± 4.5
	186 ± 9	7.7 ± 3.5
	200 ± 10	4.1 ± 3.6
E3t-pl-c	162 ± 23	1.6 ± 3.6
	184 ± 13	6.7 ± 3.8
E3t-ol	0	-0.2 ± 0.1
	0	-0.1 ± 0.1
G4i-pl-d	79 ± 9	4.1 ± 4.6
	157 ± 8	6.6 ± 5.0
	149 ± 5	2.8 ± 6.9
	98 ± 9	4.5 ± 4.5
	122 ± 12	2.5 ± 2.8
	114 ± 17	2.5 ± 3.0
G4i-ol	5 ± 3	0.2 ± 3.0
	0	0.1 ± 0.1
F7o-pl-e	0	-0.1 ± 0.1
	103 ± 3	-0.2 ± 6.0
	115 ± 11	5.6 ± 2.4
	149 ± 6	4.5 ± 2.0
F7o-ol	135 ± 15	2.2 ± 2.3
	0	-0.1 ± 0.1
G7g-pl-f	0	-0.1 ± 0.1
	170 ± 30	3.2 ± 3.8
	198 ± 10	1.6 ± 4.8
G7g-ol	178 ± 34	4.1 ± 3.2
	0	0.0 ± 0.1
C6w-pl-g	0	0.2 ± 0.1
	120 ± 15	1.8 ± 11.5
	130 ± 16	6.2 ± 5.7
	116 ± 12	2.1 ± 6.9
	114 ± 5	6.9 ± 4.7
C6w-pl-h	91 ± 26	5.4 ± 6.3
	44 ± 7	3.0 ± 3.6
	36 ± 2	0.2 ± 4.1

pl = plagioclase; ol = olivine.

from CV3 Allende. Galy et al. (2000) also detected ²⁶Mg excesses in one Al-rich chondrule from Allende. Srinivasan et al. (2000) observed ²⁶Mg excesses in one of four Al-rich chondrules from CV3 Axtell. Hsu et al. (2003) found ²⁶Mg excesses in one POI and two ferromagnesian chondrules from CK3-an Ningqiang. Recent abstracts (Hutcheon and Jones, 1995; Hutcheon et al., 2000 and references therein; Marhas et al., 2000) reported resolved Al-Mg isochrons in seven of twenty-eight anorthite-rich CC chondrules (2 of 2 from Acfer 094, 1 of 11 from CR, 4 of 10 from CV, 0 of 1 from CH, and 0 of 4 from CO).

Although a lack of phases with very high Al/Mg ratios combined with high analytical uncertainties may have been the dominant reasons that ²⁶Mg excesses were not observed in most samples studied by these teams, with the possible exception of Acfer 094 these chondrites are more altered than Y-81020. It is therefore possible that diffusion occurring during parent-body thermal metamorphism and aqueous alteration caused changes in key phases in these chondrites (Huss et al., 2001).

There also is some uncertainty in interpreting the data on ²⁶Mg excesses in Al-rich and anorthite-rich chondrules. The precursor materials of these chondrules seem to have included anomalously high fractions of debris from refractory inclusions; some of the relict Al-rich phases may have preserved ²⁶Mg excesses that predated chondrule formation. In fact, Hutcheon et al. (2000) stated that some of the anorthite may be relict in their set of anorthite-rich chondrules. The entire anorthite grain need not be a relict, but if there were regions that preserved excess “fossil” ²⁶Mg*, this would lead to δ²⁶Mg* values that are higher than they would be if Mg isotopes were homogenized during chondrule formation. We therefore strongly suspect that this is the explanation of the “high” (²⁶Al/²⁷Al)₀ ratio of 1.2 × 10⁻⁵ in the two anorthite-rich chondrules from Acfer 094 studied by these authors and 1.9 × 10⁻⁵ in the Al-rich chondrule from Allende by Galy et al. (2000). Even if these Al-rich chondrules did not contain fossil ²⁶Mg*, their (²⁶Al/²⁷Al)₀ are probably not representative of ²⁶Al/²⁷Al ratios at the time of common ferromagnesian chondrule formation (Tachibana et al., 2003).

Since, on average, the Al/Mg ratios in this study are higher than in previous studies, the uncertainties on inferred (²⁶Al/²⁷Al)₀ are smaller than previous studies. Furthermore, our data are from chondrules in pristine CO3.0 Y-81020 chondrite. These factors combine to make our data more precise than others available for carbonaceous-chondrite chondrules.

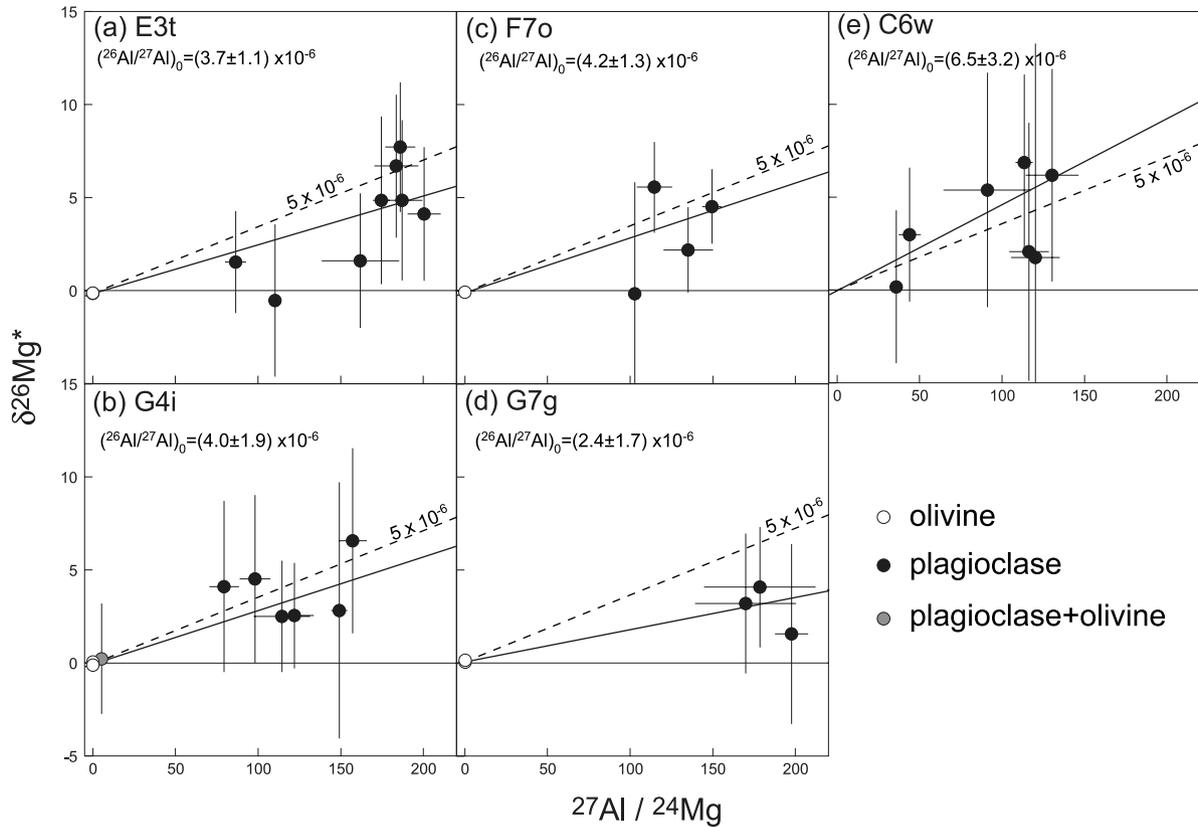


Figure 2. Al-Mg isochron diagrams for five chondrules from Yamato 81020 (CO3.0). Open and filled circles show olivine and plagioclase, respectively. Gray filled circles show plagioclase analyses that significantly overlap adjacent olivine grains. Correlation lines fitted by weighted-least squares (Williamson, 1968) are shown as solid lines. $\delta^{26}\text{Mg}^*$ correlates with the Al/Mg ratio in all chondrules. It is clear that ^{26}Al was alive at the time CO carbonaceous-chondrite chondrules formed. Reference lines corresponding to initial $^{26}\text{Al}/^{27}\text{Al}$ isotope ratios of 5×10^{-6} are also shown as dotted lines. (a) E3t (b) G4i (c) F7o (d) G7g (e) C6w.

5.2. Nebular vs. Parent-Body Formation of Crystalline Plagioclase in Y-81020 Chondrules

Most type-II chondrules in Y-81020 consist of large ferroan olivine phenocrysts surrounded by feldspathic glass that also encloses small grains of Ca pyroxene, small ferroan olivine grains, and tiny grains of chromite. Crystalline plagioclase is uncommon. The five type-II chondrules in this study are thus unusual in also possessing relatively coarse ($\sim 10\text{--}40\text{-}\mu\text{m}$) grains of crystalline plagioclase. The plagioclase in these chondrules is appreciably more sodic (Ab57–87) than in chondrules in other CO3 chondrites (Ab \sim 31 in CO3.0 ALHA 77307; Ab \sim 29 in CO3.5 Lancé; Ab16–36 in CO3.6 ALHA 77003; Kurat, 1975; Ikeda, 1982; Miura and Tomisaka, 1984).

Because two of the chondrules (G4i, Fig. 1b; G7g, Fig. 1d) in this study contain glass in addition to plagioclase, it is clear that the plagioclase was not produced by thermal metamorphism in an asteroid since isotropic igneous glass would not have survived metamorphism. In both chondrules, at least one plagioclase grain is subhedral and has a crystal face at the interface with the glass. One other chondrule (E3t, Fig. 1a) includes tiny patches that appear to be glass. The fourth chondrule (C6w, Fig. 1e) shows patchy zoning in olivine on a scale of 10 to 20 μm . These observations indicate that the plagioclase

crystallized from the chondrule melt. We conclude that the plagioclase in the Y-81020 chondrules was not produced by parent-body thermal alteration. We suggest that the coarse plagioclase could have formed in chondrule melting events involving small melt fractions and temperatures near the solidus.

5.3. Comparison of Initial $^{26}\text{Al}/^{27}\text{Al}$ Ratios in CC and OC Ferromagnesian Chondrules

As discussed above, past studies of ^{26}Al anomalies in chondrules in carbonaceous chondrites have, with rare exceptions, been of Al-rich or anorthite-rich chondrules. Although most of these studies yielded $^{26}\text{Al}/^{27}\text{Al}$ ratios that are the same as our mean within experimental error, the abstract by Hutcheon et al. (2000) reported that two anorthite-rich chondrules in Acfer 094 had $^{26}\text{Al}/^{27}\text{Al}$ ratios of 1.2×10^{-5} , ~ 3 times higher. As discussed above, some of the anorthite in this set of chondrules appears to be relicts from CAI progenitors, and may thus contain fossil $^{26}\text{Mg}^*$. For this reason we do not use these data in our comparison of $^{26}\text{Al}/^{27}\text{Al}$ ratios in CC and OC chondrules.

For discussion purposes, we assume that the CO carbonaceous-chondrite ferromagnesian chondrules measured by us are

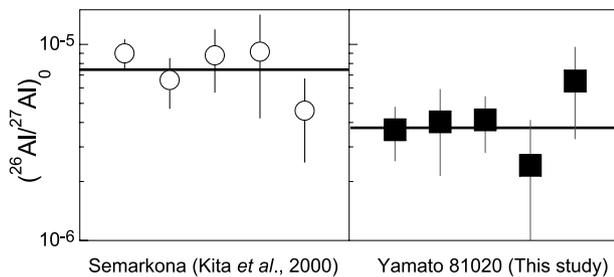


Figure 3. Comparison of initial $^{26}\text{Al}/^{27}\text{Al}$ ratios in OC Semarkona chondrules (Kita et al., 2000) with those observed in CC Yamato 81020 chondrules (this study). Ranges overlap but the means are different. Mean values (7.4×10^{-6} in OC, 3.8×10^{-6} in CC) are shown as solid lines.

broadly representative of unaltered CC. Kita et al. (2000) observed initial $^{26}\text{Al}/^{27}\text{Al}$ ratios in five Semarkona OC ferromagnesian chondrules within a range of $\sim(5\text{--}9) \times 10^{-6}$ and concluded that members of the set formed within 1 Ma of each other. In fact, the errors are large enough to permit all of the chondrules to have formed during a much shorter period, perhaps <0.2 Ma.

We determined a distribution in initial $^{26}\text{Al}/^{27}\text{Al}$ ratio in CC chondrules within a range of $\sim(2\text{--}7) \times 10^{-6}$. Our CC data are compared with the OC data of Kita et al. (2000) in Figure 3; it is evident that the highest values we observe are similar to the lowest values from OC chondrules. It appears that CC chondrule values are systematically lower than those in OC chondrules, however, the uncertainties are not small and there is considerable overlap in the initial $^{26}\text{Al}/^{27}\text{Al}$ ratios in the two data sets.

The initial $^{26}\text{Al}/^{27}\text{Al}$ ratio estimated by a weighted-mean of five CC chondrules is $(3.8 \pm 0.7) \times 10^{-6}$. Addition of the $(^{26}\text{Al}/^{27}\text{Al})_0$ of $(4.6 \pm 3.0) \times 10^{-6}$, for the Y-81020 chondrule studied by Yurimoto and Wasson (2002), to the set gives the same weighted-mean ratio. In contrast, the weighted-mean initial $^{26}\text{Al}/^{27}\text{Al}$ ratio of OC chondrules in Kita et al. (2000) is $(7.4 \pm 1.0) \times 10^{-6}$. Within the uncertainties of our data, all of the CC chondrules could have had the same initial $^{26}\text{Al}/^{27}\text{Al}$ ratio. If we assume that these two sets are representative of the respective chondrule populations, the initial $^{26}\text{Al}/^{27}\text{Al}$ ratio of CC chondrules is nearly a factor of two lower than those of OC chondrules.

The refractory inclusions in CC and OC show similar initial $^{26}\text{Al}/^{27}\text{Al}$ ratios (Russell et al., 1996; Huss et al., 2001) as well as O-isotope compositions (McKeegan et al., 1998). This is interpreted to indicate that the rare refractory inclusions in OC and the more common refractory inclusions in CC were formed at the same time from the same Al- and O-isotopic reservoirs. In contrast, differences between OC chondrules and CC chondrules in initial $^{26}\text{Al}/^{27}\text{Al}$ ratios and O-isotope compositions (e.g., Clayton et al., 1991) imply formation environments that were spatially and/or temporally distinct. This picture contrasts with the statement by Huss et al. (2001) that chondrules of all types from both CC and OC formed from an isotopic reservoir that was approximately homogeneous with respect to ^{26}Al .

5.4. Chronological Interpretations

If we assume that Al isotopes in the solar nebula were well mixed (with the exception of a negligible fraction of surviving presolar grains and isotopically unusual CAIs; Sahijpal and Goswami, 1998) and that the apparent differences in initial $^{26}\text{Al}/^{27}\text{Al}$ ratios are representative of the respective populations, the difference in initial $^{26}\text{Al}/^{27}\text{Al}$ ratios between refractory inclusions [$(^{26}\text{Al}/^{27}\text{Al})_0 \sim 4.5 \times 10^{-5}$] and CC chondrules implies a time difference of 2.6 ± 0.2 Ma. This value is consistent with the interval in absolute ages given by Pb-Pb isochrons between refractory inclusions in CV Efremovka and chondrules from CR Acfer 059, 2.5 ± 1.2 Ma (Amelin et al., 2002).

If Al isotopes were initially uniform in the inner solar system, the difference in initial $^{27}\text{Al}/^{27}\text{Al}$ ratios between OC and CC chondrules indicates that CC chondrules formed significantly (ca. 0.7 Ma) later than OC chondrules. This chronological interpretation implies that CC chondrules obtained their lower mean $\Delta^{17}\text{O}$ values after OC chondrules formed with higher $\Delta^{17}\text{O}$ values. This sequence contrasts with past suggestions that the temporal O-isotopic evolution of solar-nebular solids is from lower to higher $\Delta^{17}\text{O}$ values (Clayton and Mayeda, 1984; Rubin et al., 1990; Wasson, 2000). One or the other of these chondrule-formation sequences must be incorrect. It is possible that Al isotopes were not well-mixed and that the initial $^{26}\text{Al}/^{27}\text{Al}$ ratios in chondrules belonging to different classes (i.e., OC or CC) do not yield an accurate nebular chronology.

5.5. The ^{26}Al Heating of CO Asteroids

About 11% of meteorite falls are derived from asteroids that experienced large-scale melting that fractionated metal from silicates (Grady, 2000). Many of these differentiated meteorites are quite ancient, consistent with the idea that short-lived radioactivity was the primary heat source. However, the results presented here cast doubt on the viability of ^{26}Al as a heat source to melt the parent asteroids of differentiated meteorites.

There are several benchmark temperatures relevant to the melting of asteroids made up of anhydrous chondrites. The first melt is the Fe-FeS eutectic at ~ 1260 K (Brandes and Brook, 1992). Albitic plagioclase is the first silicate to melt at ~ 1390 K (Deer et al., 1965). At least 20% of the solids must be melted to allow silicates and metal to segregate in small (radius ~ 100 km) anhydrous asteroids (Taylor, 1992); it appears that this requires a temperature of 1500 to 1600 K.

We assume that the Al isotopes were mixed at each chondrite formation location. Because chondrules formed in the solar nebula and contain more of the bulk Al than refractory inclusions and matrix, the initial $^{26}\text{Al}/^{27}\text{Al}$ ratios in chondrules provide the best estimate of the amount of ^{26}Al available to heat asteroids. Ratios would be lower if appreciable time elapsed between chondrule formation and the accretion of kilometer-size asteroids.

We calculated temperature profiles for spherical asteroids (Carslaw and Jaeger, 1959) heated by the energy released by ^{26}Al decay using parameters similar to those of LaTourrette and Wasserburg (1998): a thermal conductivity of $2.1 \text{ W m}^{-1} \text{ K}^{-1}$, a heat capacity of $837 \text{ J kg}^{-1} \text{ K}^{-1}$, a heat production of 3.16

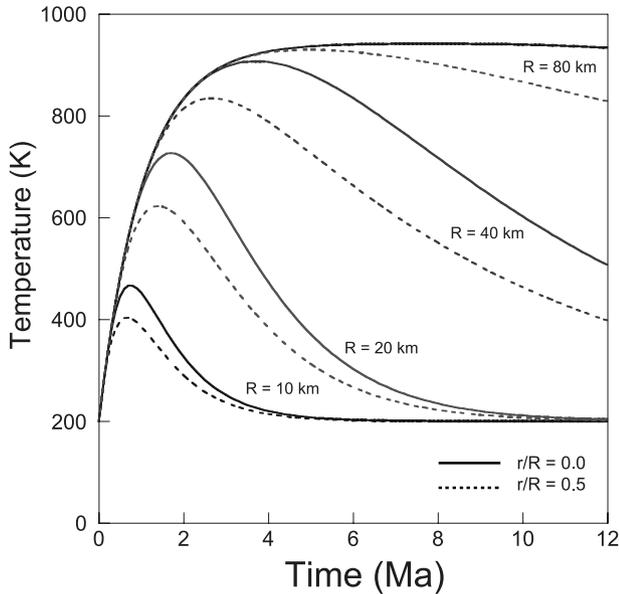


Figure 4. Time-temperature profiles for an asteroidal parent body heated only by the radioactive decay of ^{26}Al ; see text for details. Curves are shown for the core ($r/R = 0$) and the middle ($r/R = 0.5$) of asteroids with radii of 10 km, 20 km, 40 km, and 80 km. The initial and surface temperatures were set at 200 K. Maximum temperatures are reached at asteroidal radii around 40 km.

MeV per ^{26}Al atom (Schramm et al., 1970), and a parent-body density of 3370 kg m^{-3} (Yomogida and Matsui, 1983). We used our CO initial $^{26}\text{Al}/^{27}\text{Al}$ ratio of 3.8×10^{-6} , an Al abundance in CO chondrites of 1.43 wt.% (Wasson and Kallemeyn, 1988), and surface and initial asteroid bulk temperatures of 200 K. These choices are intended to be conservative; if anything, they err on the side of yielding higher temperatures. We assumed instant accretion to obtain the maximum temperature.

The calculated relationships between temperature and asteroid size, depth, and time after parent-body formation are shown in Figure 4. Because the mean life of ^{26}Al is so short, all asteroids with radii >40 km reach essentially identical maximum interior temperatures. The calculations show that, independent of the size of the parent body, the maximum temperature reached in a CO asteroid is 940 K (Fig. 4). This is far lower than 1260 K, the temperature necessary to form the Fe-FeS eutectic, the first melt to form in a metal-bearing chondritic asteroid, and 560 K too low to differentiate the body and segregate the silicates into a mantle and crust and the metal into a core. It is also not possible to melt CV-chondrite asteroids with the initial $^{26}\text{Al}/^{27}\text{Al}$ ratio observed in CC chondrules. Srinivasan et al. (2000) detected ^{26}Mg excess in one of four Al-rich chondrules from CV Axtell; the initial $^{26}\text{Al}/^{27}\text{Al}$ ratio was $(3.4 \pm 1.5) \times 10^{-6}$. Sheng et al. (1991) also reported ^{26}Mg excesses in two Al-rich chondrules from the Allende CV meteorites; these yield ratios of 6.1×10^{-6} and 2.6×10^{-6} (the latter were also obtained from Al-rich chondrules). Hutcheon et al. (2000) reported a value of $(5.6 \pm 3.3) \times 10^{-6}$ in CV Efremovka in an anorthite-rich chondrule. These data measured on Al-rich chondrules are consistent with our more precise ratios even though some fossil ^{26}Mg may be present in relict anorthite. We also note that, if Al isotopes were well mixed, it

is the lowest chondrule initial ratios that determine the maximum asteroid temperatures that can be produced.

We therefore calculated maximum temperatures in CV-chondrite asteroids using the CO initial $^{26}\text{Al}/^{27}\text{Al}$ ratio of 3.8×10^{-6} but the 20% higher Al content of 1.75 wt.% observed in CV chondrites (Wasson and Kallemeyn, 1988). The higher Al content raises the maximum temperature to 1100 K. Thus, even the higher ^{26}Al content of CV chondrites would not have produced melting, despite the fact that CV chondrites are the main carriers of the refractory inclusions with high $(^{26}\text{Al}/^{27}\text{Al})_0$ that have been measured to date. High contents of refractory inclusions by themselves do not result in melting because of the large time interval between CAI formation and parent-body accretion.

Similar calculations for the ordinary chondrites show that, although minor melting can be produced, the temperatures are too low to differentiate a small asteroid. The OC had higher initial $^{26}\text{Al}/^{27}\text{Al}$ ratios (7.4×10^{-6} ; Kita et al., 2000) but Al contents (1.19 wt.% in LL chondrites; Wasson and Kallemeyn, 1988) are appreciably lower than in CO and CV chondrites. Calculations in which no allowance is made for the latent heat of melting yield OC-asteroid temperatures of 1370 K, which is 100 K higher than the Fe-FeS eutectic value, but too low to melt plagioclase and well below the minimum temperature of 1500 K necessary to differentiate a small asteroid.

Latent heat is a major sink of energy; Wasson (1996) estimated that to melt chondrite minerals requires about the same amount of energy as is required to heat the solid phases by 400 K. We thus calculate that the production of a 20% melt fraction in an OC asteroid at 1500 K requires $\sim 20\%$ more heat than needed to raise the temperature from 200 to 1370 K. It thus seems clear that the measured $(^{26}\text{Al}/^{27}\text{Al})_0$ in OC chondrules would not have produced differentiation in OC asteroids even though, as noted below, the parent of the IVA iron meteorites seems to have been an LL-chondrite body.

At 1260 K (the Fe-FeS eutectic) and at 1400 K the S contents of the melt are 31 and 28 wt.%, respectively (Brandes and Brook, 1992). As a point of information we calculated the melt fractions at these temperatures in LL chondrites given that the total S content is 2.3 wt.% (Wasson and Kallemeyn, 1988). By this estimate, the total FeS-metal melt fraction at the eutectic is ~ 7.4 wt.% and that at 1400 K is ~ 8.1 wt.%.

The $(^{26}\text{Al}/^{27}\text{Al})_0$ value that we find for CO chondrules is similar to the initial $^{26}\text{Al}/^{27}\text{Al}$ ratio assumed by Ghosh and McSween (1998) for Vesta (presumed to be the HED parent body) immediately following accretion. Our calculations show that this ratio is not sufficient to cause melting of an asteroid of CO or CV composition. In contrast, the Ghosh-McSween model produces melting and differentiation. One reason their model achieves higher temperatures is that these authors assume that incipient melting at the Fe-FeS eutectic leads to instantaneous separation of a core, thereby removing 33% of the body's mass (which is no longer heated by ^{26}Al). Core segregation results in a large increase in the Al content in the residuum, and thereby increases the heating rate of the mantle and produces silicate partial melting. However, it is not clear if core separation can be rapidly achieved under these conditions. Other uncertainties (e.g., regarding the bulk composition of the HED parent body) lead to considerable ambiguity as to whether this level of ^{26}Al can melt any silicate, even under the most

optimistic assumptions. Alternative models based on ^{26}Al heating at similar levels do not yield temperatures sufficient for silicate differentiation (e.g., Miyamoto, 1991).

Our data and calculations show that ^{26}Al is grossly inadequate for melting carbonaceous-chondrite asteroids, and our calculations cast serious doubt on whether ^{26}Al alone could have melted OC asteroids. Because asteroids are small, they are not heated appreciably by accretion. Instantaneous collapse of a particle cloud to form a 100-km asteroid having a density of 3370 kg m^{-3} releases a gravitational energy amount of 5700 J kg^{-1} , sufficient to raise the temperature only $\sim 7 \text{ K}$. The temperature reached by our model, although too low to melt, could provoke some thermal metamorphism, obviously absent from the meteorite studied. Possibilities to explain this are that the meteorite was located in near-surface regions, or that accretion did not follow chondrule formation instantly (as we assumed). If accretion lagged the time interval between chondrule formation significantly, that time interval would have allowed ^{26}Al to decay, thus decreasing its heating ability in the parent body. It is also possible that parent body was significantly smaller than 100 km.

If one separates OC and CC material at a benchmark $\Delta^{17}\text{O} = 0[\text{ppt}]$, the most common differentiated stony (HED) and iron (IIIAB) meteorites are “carbonaceous.” Thus, these “CC” materials have in some cases experienced melting even though our data suggest that ^{26}Al was only able to provide at most ~ 50 to 80% of the required energy. There are several reports (Srinivasan et al., 1999; Nyquist et al., 2003) of excess ^{26}Mg , corresponding to $(^{26}\text{Al}/^{27}\text{Al})_0 \sim 1 \times 10^{-6}$, in eucrites, although Wadhwa et al. (2003) reported that the Juvinas eucrite shows no excesses, i.e., $(^{26}\text{Al}/^{27}\text{Al})_0 < 1 \times 10^{-7}$. We note that the presence of excess ^{26}Mg in HED meteorites demonstrates that the eucrites formed while ^{26}Al was still alive, but does not prove that ^{26}Al was the dominant heat source causing differentiation of the parent body. If the initial $^{26}\text{Al}/^{27}\text{Al}$ ratios and Al concentrations were no higher than in CC and OC chondrites, our model would require an additional heat source.

One way to salvage the possibility of ^{26}Al as a heat source would be if the differentiated meteorites were formed from nebular materials (i.e., chondrites) that contained $\sim 2\times$ higher $(^{26}\text{Al}/^{27}\text{Al})_0$ than CC or OC. It is possible that some of these chondritic materials could be in our meteorite collections. Some fraction of the asteroidal parents of these chondritic materials may have been located in near-surface regions that were not melted by internal heat sources, and some may have formed bodies too small ($\leq 3 \text{ km}$) to have retained enough ^{26}Al -deposited heat to melt. If these hypothetical chondrites exist, they would probably have been stored in asteroidal orbits similar to those in which the differentiated asteroids are located, and thus could sometimes fall and be present in meteorite collections. It will be necessary to demonstrate low initial ratios of $^{26}\text{Al}/^{27}\text{Al}$ for most other classes of chondrites before one can be sure that such chondrites did not exist. Although this confirmatory research needs to be done, the fact that the two most common classes (or “clans”) of chondrites have low $^{26}\text{Al}/^{27}\text{Al}$ ratios makes it unlikely that ^{26}Al was a major planetary heat source.

Another short-lived nuclide, ^{60}Fe , may also have contributed heat. Kita et al. (2000) reported an upper-limit initial $^{60}\text{Fe}/^{56}\text{Fe}$ ratio [$(^{60}\text{Fe}/^{56}\text{Fe})_0$] in LL3.0 Semarkona chondrules of $\sim 1.4 \times$

10^{-7} . Recently, Tachibana and Huss (2003) presented detailed evidence of live ^{60}Fe in five troilite grains in LL3.1 Bishunpur and LL3.1 Krymka. The mean $(^{60}\text{Fe}/^{56}\text{Fe})_0$ value is 1.4×10^{-7} ; the weighted mean is 1.2×10^{-7} . We calculated the heat contribution of ^{60}Fe by assuming $(^{60}\text{Fe}/^{56}\text{Fe})_0$ of OC and CC chondrules as 1.4×10^{-7} with the materials properties as mentioned above, and a heat production of 2.78 MeV (Lederer and Shirley, 1978) per ^{60}Fe atom. We used Fe abundances in CO, CV, and LL chondrites of 24.8, 23.5, and 18.5 wt.%, respectively (Wasson and Kallemeyn, 1988). The temperature increase due to ^{60}Fe for CO, CV, and LL was calculated to be 180, 170, and 140 K, respectively. If we neglect latent heat, the combination of the ^{26}Al and ^{60}Fe sources yields CO, CV, and LL maximum temperatures of 1120, 1370, and 1510 K, respectively. The result is that the CO body still fails to reach the Fe-FeS eutectic and the CV body exceeds the eutectic temperature but, after allowance for latent heat, would not melt plagioclase. It is beyond the scope of the present paper to carry out a calculation that includes a detailed assessment of latent heat effects, but we estimate that melting 15% of the solids would necessitate subtracting 60 K from the LL sum temperature of 1510 yielding 1450 K, almost certainly too low to allow differentiation, even in large ($R > 100 \text{ km}$) asteroids that are able to retain most of the heat.

In a recent abstract, Mostefaoui et al. (2003) reported evidence of $(^{60}\text{Fe}/^{56}\text{Fe})_0$ ratios much higher than those published by Tachibana and Huss (2003). Their estimated $(^{60}\text{Fe}/^{56}\text{Fe})_0$ value is $(7.5 \pm 2.6) \times 10^{-7}$. Our estimates of the temperature increase due to such high amounts of ^{60}Fe for CO, CV, and LL asteroids was 980, 900, and 730 K, respectively. If this high $(^{60}\text{Fe}/^{56}\text{Fe})_0$ value is confirmed to be representative of chondrules, then a combination of the ^{26}Al and ^{60}Fe sources could indeed melt asteroids.

5.6. Melting Differentiated Asteroids

Given that short-lived radioactivity appears insufficient to drive differentiation (the Mostefaoui et al. (2003) ^{60}Fe result notwithstanding), it is reasonable to ask what heat source melted the differentiated asteroids. A plausible heat source might be localized (rather than global) heating by impacts occurring on porous asteroids (Wasson and Kallemeyn, 2002; Housen and Holsapple, 2003). In the past, impact sources were considered ineffective, largely because much of the heat generated during cratering of low-porosity planets (e.g., the Earth) and large satellites goes into high-velocity ejecta that would be lost from asteroids.

In contrast, impact heat is retained much more efficiently on high-porosity target materials (Melosh, 1989; Wasson and Kallemeyn, 2002; Housen and Holsapple, 2003). Because there is no nebular mechanism to produce compact planetesimals, the first asteroids must have had high ($\geq 50\%$) porosities. Such objects offer two advantages for achieving high internal temperatures: (a) these bodies are easily compressed and thus, (because $P \cdot \Delta V$ (pressure times the change in volume) is higher than for compact targets) a larger fraction of the kinetic energy is converted to heat, and (b) the heat is deposited at deeper levels in the body (craters would have depths much greater than their diameters). Appreciable melting occurs at typical asteroidal impact velocities of 5 km s^{-1} (Wasson and Kallemeyn,

2002). Orbital simulations (W. Bottke, personal communication, 2001) show that 20% of impact velocities are $\geq 7 \text{ km s}^{-1}$, and thus deposit $\geq 2x$ more heat than impacts at typical velocities. Of course, magmas are self-selected to form in the impacts that deposit the most heat. And, compared to their porous siblings, melted meteoroids are more likely to survive space erosion and atmospheric disruption and fall as meteorites.

We note that one large group of differentiated meteorites, the group-IVA iron meteorites, has O-isotopic compositions (in their silicates) that fall within the LL chondrite field (Clayton et al., 1983). If these irons did form from LL chondrites, the chondrule Al-Mg data of Kita et al. (2000), and our calculations suggest that ^{26}Al alone was unable to produce the IVA magma. Perhaps the bulk of the heat was deposited in a large impact.

6. CONCLUSIONS

Our study of Mg isotopes in high-FeO ferromagnesian chondrules in CO3.0 Yamato 81020 revealed the presence of excess (radiogenic) ^{26}Mg that correlates with the Al/Mg ratio. Thus, ^{26}Al was alive in the CC chondrule-forming region; initial $^{26}\text{Al}/^{27}\text{Al}$ ratios ranged from 2.4×10^{-6} to 6.5×10^{-6} with a mean ratio, $(3.8 \pm 0.7) \times 10^{-6}$, that is ~ 2 times lower than the mean observed in OC chondrules (Kita et al., 2000). Calculations show that ^{26}Al in CO chondrite asteroids can raise the temperature only to $\approx 940 \text{ K}$, too low to produce melting. The same initial $^{26}\text{Al}/^{27}\text{Al}$ ratio in CV chondrites also produces no melt. The higher ($^{26}\text{Al}/^{27}\text{Al}$)₀ ratios in OC chondrules produce minor amounts of melting that are too low to cause differentiation of the asteroid. The lower initial $^{26}\text{Al}/^{27}\text{Al}$ ratio in CC relative to OC chondrules may have chronological significance or it may reflect nebular heterogeneity.

Recent studies have revealed an $^{60}\text{Fe}/^{56}\text{Fe}$ ratio of $\sim 1.4 \times 10^{-7}$ in LL-chondrite FeS (Tachibana and Huss, 2003). The addition of ^{60}Fe in this amount does not produce enough melt to cause asteroidal differentiation, even in ordinary chondrites.

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APPENDIX: PETROGRAPHIC DESCRIPTION OF THE FIVE CHONDRULES

Chondrule E3t (Fig. 1a) is a ~160 × 200 μm-size fragment of a type-II porphyritic-olivine chondrule. It contains one large (140 μm) subhedral olivine phenocryst (Fa35–47), a few 15 to 30 μm grains of sodic plagioclase (Ab80–87), numerous ~5 μm olivine grains, relatively coarse (15–40 μm) Ca-rich pyroxene grains, and accessory blebs of metallic Fe-Ni. The small olivine grains occur mainly within the Ca-rich pyroxene. Plucking during polishing has produced a >100-μm hole at the edge of the chondrule fragment.

Chondrule G4i (Fig. 1b) is a ~200 × 320 μm type-II porphyritic-olivine chondrule, that (in the plane of the section) contains seven 50 to 90 μm euhedral to subhedral olivine phenocrysts (Fa21–32), many smaller (5–40 μm) olivine grains, small Ca-rich pyroxene grains at the edge of coarse olivine, irregular coarse (10–30 μm) grains of plagioclase (Ab57–64), irregular 10 to 25-μm-size metal grains, and 5 to 20-μm patches of glassy mesostasis. Numerous small olivine grains resembling cauliflower are concentrated in the outer region of the chondrule.

Chondrule F7o (Fig. 1c) is a ~200 × 220 μm-size type-II porphyritic-olivine chondrule that contains two large (80–120 μm) subhedral olivine phenocrysts (Fa35–37), several smaller (10–30 μm) olivine grains, 10 to 20 μm Ca-rich pyroxene phenocrysts, coarse (10–40 μm)

plagioclase (Ab54–76) grains surrounding the large olivine grains, and rare small metal blebs.

Chondrule G7g (Fig. 1d) is a $\sim 80 \times 180 \mu\text{m}$ -size type-II chondrule fragment that contains several 5 to 20- μm -size high-FeO (Fa21–58) olivine phenocrysts and a large, apparently relict, elongated FeO-poor (Fa15) olivine grain rimmed by fayalitic olivine. The chondrule also contains olivine grain fragments, ~ 10 - μm -size irregularly shaped Ca-rich pyroxene grains, two 20- μm -size grains of plagioclase (Ab68–78), 15×20 - μm -size patches of glassy mesostasis, and rare small grains of metallic Fe-Ni.

Chondrule C6w (Fig. 1e) is a $\sim 200 \times 250 \mu\text{m}$ -size type-II porphyritic-olivine chondrule that contains several ~ 40 to $80 \mu\text{m}$ subhedral olivine phenocrysts (Fa21–43), a few 20 to 30 μm Ca-rich pyroxene grains, interstitial grains of plagioclase (Ab69–89) and small irregular metal grains. The angular shapes of some of the plagioclase grains are controlled by the surrounding prismatic faces of the olivine phenocrysts. The patchy zoning in the olivine shows the presence of several relict grains with compositions near Fa20. Gradients are strong in some cases (e.g., $\sim 40 \mu\text{m}$ NW of the “Ol” label) on a scale of $\sim 10 \mu\text{m}$.