

The chromium isotopic composition of Almahata Sitta

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Abstract—Nine ureilitic fragments of the anomalous polymict ureilite Almahata Sitta have been analyzed for their Cr isotopic compositions. All the samples, including both nonmagnetic and magnetic portions, show essentially the same $\epsilon^{54}\text{Cr}$ deficit (-0.77 ± 0.10), relative to the terrestrial Cr standard. This contrasts with the variable positive ^{54}Cr anomalies observed for carbonaceous chondrites, but agrees with the values measured for eucrites, diogenites, and mesosiderites (Trinquier et al. 2007). This implies that, contrary to previous suggestions based on O isotopes, ureilites were not derived from any known carbonaceous chondrite parent body. Instead, the Almahata Sitta parent body may have accreted in a nebular region/environment similar to that of the howardite, eucrite, and diogenite (HED) parent body. In addition, the lack of variation in $\epsilon^{54}\text{Cr}$ combined with variable O isotopic compositions in the meteorite fragments suggests that whatever process(es) caused the O isotopic heterogeneity of the solar system was probably not responsible for heterogeneity in $\epsilon^{54}\text{Cr}$. The samples show resolvable variations in $\epsilon^{53}\text{Cr}$ (0.15–0.41) that are correlated with Mn/Cr ratios, suggesting that live ^{53}Mn was present at the time of formation of Almahata Sitta. The isochron yields an initial $^{53}\text{Mn}/^{55}\text{Mn}$ value of $3.1 (\pm 1.1) \times 10^{-6}$, corresponding to an age of $4563.6 \pm (2.2)$ Ma when related to U-Pb and Mn-Cr data for the D'Orbigny angrite. This age is consistent with the Mn-Cr and Al-Mg ages of two other polymict ureilites (Goodrich et al. 2010). Magmatic activity on the ureilites' parent body seems to have postdated the formation of refractory inclusions by approximately 4–5 Ma.

INTRODUCTION

Ureilites are a unique group of achondrites. They are composed mostly of olivine and pigeonite with igneous textures and, in this sense, resemble differentiated terrestrial ultramafic rocks. However, ureilites also have features that make them appear to be undifferentiated and “primitive” (Goodrich et al. 2001). In a 3-oxygen-isotope plot ($\delta^{17}\text{O}$ versus $\delta^{18}\text{O}$), ureilites occupy the upper one-third of the slope-1 carbonaceous

chondrite anhydrous minerals (CCAM) line (Clayton and Mayeda 1988). Thus, a carbonaceous chondrite precursor material has been suggested for these meteorites (Kita et al. 2004). However, ureilites exhibit considerable variation along the CCAM line, making it difficult to pair them with any particular carbonaceous chondrite. Recent studies have shown significant differences in $^{54}\text{Cr}/^{52}\text{Cr}$ ratios among different groups of chondrites, differentiated meteorites, and terrestrial rocks (Shukolyukov and Lugmair 2006a; Trinquier

et al. 2007; Qin et al. 2010). Compared to terrestrial rocks and enstatite chondrites, carbonaceous chondrites show variable excesses in $^{54}\text{Cr}/^{52}\text{Cr}$ from 0.4 to 1.6 ϵ (relative deviation of the Cr isotope ratio from a terrestrial standard times 10,000) and ordinary chondrites show constant deficits of -0.4ϵ . Thus, Cr isotopic compositions may resolve the question of whether ureilites originated from a carbonaceous chondrite parent body that underwent incomplete melting. Earlier work suggested that ureilites may have different $^{54}\text{Cr}/^{52}\text{Cr}$ ratios (approximately -0.9ϵ) from carbonaceous chondrites (Shukolyukov and Lugmair 2006b; Ueda et al. 2006).

Almahata Sitta consists of hundreds of centimeter-sized fragments collected along a linear path coincident with the projected Earth-impacting orbit of the asteroid 2008 TC₃ in the Nubian Desert, Sudan (Jenniskens et al. 2009). The asteroid exploded in the stratosphere. Initial investigation on piece #7 of the meteorite identified it as a polymict ureilite (Jenniskens et al. 2009). An additional 10 pieces were analyzed for O isotopic composition and the results demonstrated that Almahata Sitta is a ureilite (Rumble et al. 2010). Those are the samples we concentrate on here. These ureilite samples display O isotopic heterogeneity that spans the entire range previously observed for both monomict and polymict ureilites (Rumble et al. 2010). This makes Almahata Sitta an ideal test for correlations between $\Delta^{17}\text{O}$ and $\epsilon^{54}\text{Cr}$, like those displayed by carbonaceous chondrites.

Aside from fingerprinting the provenance of ureilites, the Cr isotopic system is further explored to determine whether ^{53}Mn - ^{53}Cr chronology can provide additional constraints on the timing of differentiation in Almahata Sitta's parent body. The ultramafic mineralogy of ureilites, and the associated incompatible lithophile element depletions, limits the precision with which their ages can be determined using the U-Th-Pb and Sm-Nd systems (Goodrich et al. 1991; Goodrich and Lugmair 1995; Torigoye-Kita et al. 1995a, 1995b). The ^{53}Mn - ^{53}Cr short-lived chronometer has proven to be a very useful tool to date a wide variety of solar system objects because both elements are present at relatively high abundances in a variety of meteorites. Results with ^{53}Mn - ^{53}Cr for the feldspathic clasts in polymict ureilites Dar al Gani (DaG) 165 and DaG 139 indicate that they formed 0.70 ± 0.18 Ma after the angrite D'Orbigny (Goodrich et al. 2010). This age is consistent with a ^{26}Al - ^{26}Mg age of 0.5 ($-0.3/+0.5$) Ma after D'Orbigny for another polymict ureilite DaG 139 (Goodrich et al. 2010). Using the recently reported Pb-Pb age of 4563.8 ± 0.4 Ma for D'Orbigny that calculates the age based on the measured, nonterrestrial, U isotopic composition of D'Orbigny (Brennecka et al.

2010), the corresponding absolute ages recalculated here are 4563.10 ± 0.22 and 4563.3 ($+0.4/-0.5$) Ma, respectively. These ages are also in accord with a less precise U-Pb age of 4563 ± 21 Ma for another ureilite (Torigoye-Kita et al. 1995b).

SAMPLES AND METHODS

Chromium isotopic compositions of nine fragments of Almahata Sitta were measured on aliquots of samples that have also been analyzed for their O isotopic compositions by Rumble et al. (2010). These samples span the entire range of O isotopic compositions previously observed for ureilites (Rumble et al. 2010). Fusion crust and desert weathering products were removed from each sample, which were then crushed to a grain size of approximately 0.5 mm, and washed in diluted HCl (Rumble et al. 2010). For each sample, the materials that adhered to a hand magnet were separated from the less magnetic components of the sample, and this fraction is designated as the "magnetic fraction" (Rumble et al. 2010). The less magnetic component is designated as the "nonmagnetic fraction." A total of eight nonmagnetic fractions and five magnetic fractions were analyzed separately. Approximately 3–8 mg of each fraction were dissolved in a mixture of HNO_3 and HF (in 1:2 ratio) in a Teflon beaker at approximately 150 °C for 24 h. All materials went into solution except for some black residues. Raman analyses showed that these residues are mostly graphitic carbon with a trace amount of Al-Mg-bearing spinel. None of these phases have significant amounts of Cr. No chromite grains, which usually survive open-beaker HNO_3 -HF digestion, were found in the residues. Thus, not incorporating these materials into the bulk samples will not affect the accuracy of the Cr isotopic analyses of the bulk ureilite samples.

Chemical separation of Cr followed a two-step cation exchange column procedure that was used in our previous study (Qin et al. 2010). The Cr isotopic composition was analyzed on a Triton multicollector thermal ionization mass spectrometer at DTM. Each sample and the terrestrial Cr standard, NIST 3112a, were loaded onto three to four single, degassed Re filaments, each load containing approximately 1–2 μg of Cr. The Cr isotope ratios were measured in static mode, and a total of 420 ratios were collected during each run. Each filament was run twice. Only the average of the Cr isotope ratios from all the runs is reported. A detailed description of the analytical methods can be found in Qin et al. (2010).

An aliquot of each sample after sample digestion was saved for the determination of its Mn/Cr ratio. The Mn/Cr ratios were determined using a Nu plasma

multicollector inductively coupled plasma–mass spectrometry at DTM. The samples were diluted in 0.5% HNO₃ to make the Cr concentrations close to 0.1 ppb. This was done to minimize the matrix effects. The intensities of ⁵⁵Mn, ⁵⁴Fe + ⁵⁴Cr + ⁴⁰Ar¹⁴N, and ⁵³Cr were measured on Faraday cups H6, H2, and L2 with the zoom lens adjusted to provide a mass separation of 0.25 atomic mass between adjacent cups. The samples were analyzed in dry plasma mode to minimize the interferences from hydrides. The beam was centered on mass 54 at the beginning of each analysis. The interferences on ⁵⁵Mn and ⁵³Cr were corrected by subtracting the measured intensities on these peaks in a blank solution. The typical ⁵³Cr signal is approximately 15 mV. The background/signal ratio for ⁵³Cr is approximately 5%. The ⁵⁵Mn signal in the samples is an order of magnitude higher than ⁵³Cr and the background/signal ratio is only approximately 0.5%. Gravimetrically prepared Mn, Cr mixed standards with Mn/Cr ratios of 0.5, 1, 2, and constant Cr concentration of 0.1 ppb were used to construct the working curve. The blank solution and three standards were measured at the beginning, middle, and end of the analytical section to monitor the drift of instrumental mass fractionation, sensitivity, and background with time. We found that the Mn/Cr ratio reproducibility (2 sigma SD) for all standard measurements was approximately 0.5%.

RESULTS

The Cr isotopic data of Almahata Sitta are given in Table 1 and shown in Fig. 1. With one exception, the samples, including both nonmagnetic fractions and magnetic fractions, have similar deficits (from -0.85 ± 0.20 to -0.67 ± 0.12) in $\epsilon^{54}\text{Cr}$ with a mean at approximately -0.77 ± 0.10 , relative to the terrestrial laboratory Cr standard NIST SRM 3112a. The one exception is the magnetic fraction of sample #4 (not plotted on Fig. 1). For the exceptional sample, we obtained a significantly lower value of -1.86 ± 0.20 . This lower value might be an analytical artifact. A higher than usual ⁵⁶Fe/⁵²Cr intensity ratio ($>10^{-4}$) was found for this sample compared to all the others. Although the ⁵⁴Fe interference is usually correctable for even higher Fe/Cr ratios (Qin et al. 2010), we have found that on occasion lower $\epsilon^{54}\text{Cr}$ values were obtained associated with high Fe/Cr ratio, especially in runs where the ionization temperature was higher than usual (>1400 °C). In these situations, the accuracy in $\epsilon^{53}\text{Cr}$ seemed not to have been affected. Other than this measurement, no resolvable difference in $\epsilon^{54}\text{Cr}$ was observed between nonmagnetic and magnetic fractions. The Cr isotopic compositions of meteorites can be

Table 1. Cr isotopic compositions of Almahata Sitta fragments.

Sample	$\epsilon^{53}\text{Cr}$	$\epsilon^{54}\text{Cr}$	⁵⁵ Mn/ ⁵² Cr	$\Delta^{17}\text{O}^a$
Almahata Sitta #4				
	0.22 ± 0.06	-0.72 ± 0.11	0.729	-1.018
Magnetic	0.41 ± 0.08	-1.86 ± 0.20^b	1.41	–
Almahata Sitta #15				
	0.26 ± 0.05	-0.76 ± 0.14	0.591	-0.952
Almahata Sitta #27				
	0.18 ± 0.04	-0.81 ± 0.18	0.572	-1.527
Magnetic	0.20 ± 0.09	-0.80 ± 0.19	0.811	-1.546
Almahata Sitta #36				
	0.31 ± 0.04	-0.69 ± 0.07	0.847	-1.539
Magnetic	0.33 ± 0.07	-0.80 ± 0.20	1.23	-1.524
Almahata Sitta #44				
	0.15 ± 0.03	-0.80 ± 0.10	0.561	-1.045
Almahata Sitta #47				
	0.18 ± 0.05	-0.78 ± 0.08	0.548	-0.940
Magnetic	0.21 ± 0.07	-0.75 ± 0.26	0.725	–
Almahata Sitta #49				
	0.19 ± 0.05	-0.79 ± 0.13	0.591	-0.992
Almahata Sitta #51				
Magnetic	0.23 ± 0.06	-0.85 ± 0.20	1.06	–
Almahata Sitta #54				
	0.34 ± 0.04	-0.67 ± 0.12	1.20	-2.256
Average		-0.77 ± 0.10^c		

Unless otherwise indicated, the Cr isotopic data are for the nonmagnetic portions of the samples. All uncertainties quoted for individual analyses are two standard errors, and are the larger of the external errors of the standard and sample measurements in the analysis sequence.

^aThe O isotopic data are from Rumble et al. (2010).

^bThis negative value is likely caused by analytical artifacts.

^cThe uncertainty quoted for the ‘‘Average’’ is the 2 sigma SD of all the analyses, excluding the magnetic fraction of fragment #4.

modified by exposure to cosmic ray irradiation (Shima and Honda 1966; Birck and Allègre 1985). The noble gas measurements of Almahata Sitta revealed that the exposure age of Almahata Sitta is approximately 15 Ma (Ott et al. 2010; Welten et al. 2010). Such a low exposure age is unlikely to have affected the Cr isotopic compositions of our samples.

In contrast to the almost uniform values of $\epsilon^{54}\text{Cr}$, resolvable variations in $\epsilon^{53}\text{Cr}$ values were found between different Almahata Sitta fragments. Like the $\epsilon^{54}\text{Cr}$ results, no systematic differences in $\epsilon^{53}\text{Cr}$ were found between nonmagnetic and magnetic fractions of the same sample, except sample #4. Little is known about the differences in the mineralogies of the nonmagnetic and magnetic fractions of the samples, except that the latter contains most of the metal. The

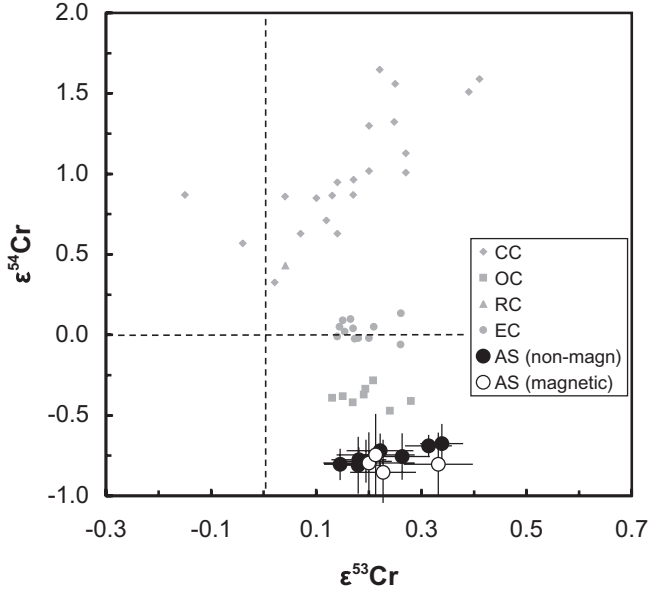


Fig. 1. The Cr isotopic compositions of Almahata Sitta fragments compared to chondrites. The data for chondrites are from Qin et al. (2010), Shukolyukov and Lugmair (2006a), and Trinquier et al. (2007, 2008).

$\epsilon^{53}\text{Cr}$ values in the samples range from 0.15 ± 0.03 to 0.41 ± 0.08 , corresponding to a total variation of 0.26 ppm, which is resolvable with our precision. Variations in $\epsilon^{53}\text{Cr}$ values correlate reasonably well with variations in Mn/Cr ratios (0.548–1.41) (Fig. 2), suggesting that there was live ^{53}Mn at the time of the last Cr isotopic equilibrium. Using Isoplot (Ludwig 2003) and a model 2 fit, we obtained an initial $^{53}\text{Mn}/^{55}\text{Mn}$ value of $(3.1 \pm 1.1) \times 10^{-6}$, and an initial $^{53}\text{Cr}/^{52}\text{Cr}$ of 0.02 ± 0.09 . Using the angrite D’Orbigny as a time reference with $^{53}\text{Mn}/^{55}\text{Mn} = (3.23 \pm 0.04) \times 10^{-6}$ (Glavin et al. 2004) at a U-Pb age of 4563.8 ± 0.4 Ma (Brennecka et al. 2010), we calculated the absolute age of Almahata Sitta to be 4563.6 ± 2.2 Ma. If we use a model 1 fit, the initial $^{53}\text{Mn}/^{55}\text{Mn}$ and $^{53}\text{Cr}/^{52}\text{Cr}$ values of $(2.8 \pm 1.0) \times 10^{-6}$ and 0.04 ± 0.09 were obtained, respectively. These are indistinguishable from the values from the model 2 fit. Thus, hereafter only the results from the model 2 fit are discussed.

DISCUSSION

The deficits in $\epsilon^{54}\text{Cr}$ in the Almahata Sitta fragments are consistent with measurements of deficits (-0.9 ± 0.2) for the ureilites Kenna and Lewis Cliff 85440 (Shukolyukov and Lugmair 2006b). The $\epsilon^{54}\text{Cr}$ deficits are distinct from the positive anomalies (0.4–1.6) observed for CV chondrites and all other known types

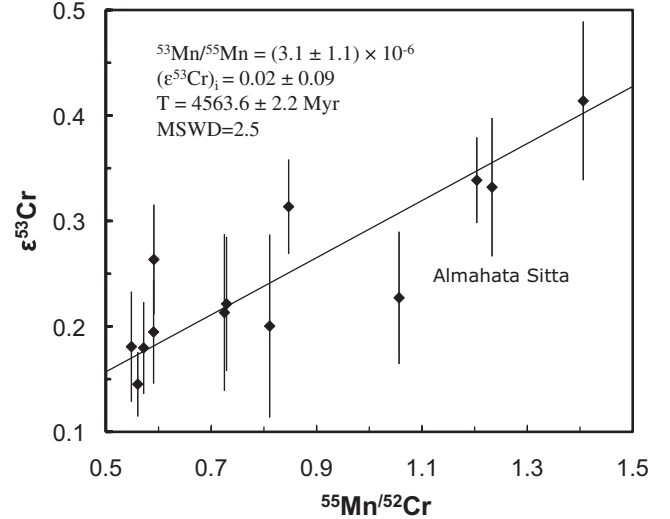


Fig. 2. The ^{53}Mn - ^{53}Cr systematics of the Almahata Sitta fragments.

of carbonaceous chondrites. Thus, ureilites seem unlikely to be derived from any carbonaceous chondrite parent body, at least one that is similar to any of those that have been measured for their Cr isotopic compositions. The deficits in $\epsilon^{54}\text{Cr}$ found for Almahata Sitta are also resolvable more negative than the uniform deficits (approximately -0.4) found for ordinary chondrites.

The only sample that has a distinctly different $\epsilon^{54}\text{Cr}$ value is the magnetic fraction from sample #4, which we interpret as an analytical artifact. The deficits in $\epsilon^{54}\text{Cr}$ for all other Almahata Sitta samples are consistent with the values observed for eucrites, diogenites, and mesosiderites (Trinquier et al. 2007). Because the variations in $\epsilon^{54}\text{Cr}$ cannot be caused by in situ radioactive decay of any known nuclides or known chemical fractionation processes, they must reflect the isotopic heterogeneity of the solar nebula. The close match between the $\epsilon^{54}\text{Cr}$ compositions of the ureilite and howardite, eucrite, and diogenite (HED) meteorites suggests that their parent bodies accreted in the same or similar nebular regions. We note that another group of basaltic achondrites, the angrites, is characterized by less negative $\epsilon^{54}\text{Cr}$ anomalies (Trinquier et al. 2007), so there is not a unique Cr isotopic composition for all differentiated or partially differentiated parent bodies.

As mentioned earlier, a potential correlation between $\Delta^{17}\text{O}$ and $\epsilon^{54}\text{Cr}$ has been suggested for carbonaceous chondrites by previous studies (Trinquier et al. 2007; Yin et al. 2009). This correlation was used to question the photodissociation explanation for the nonmass dependent fractionation of O isotopes in meteoritic materials (Yin et al. 2009). The total variation in $\Delta^{17}\text{O}$ among Almahata Sitta samples is

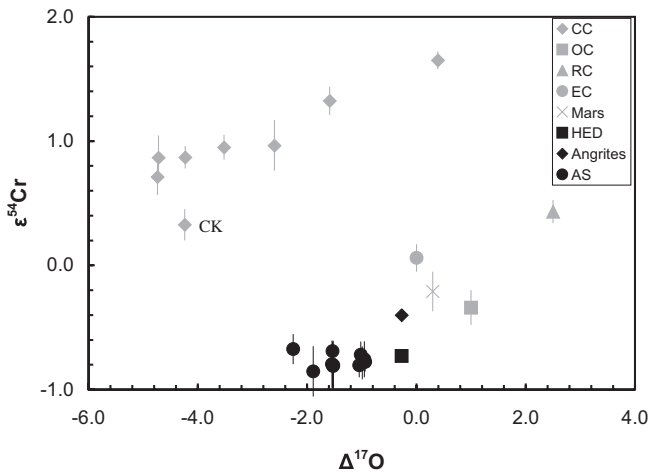


Fig. 3. $\epsilon^{54}\text{Cr}$ versus $\Delta^{17}\text{O}$ in the Almahata Sitta fragments and in other inner solar system materials. The O isotope data were taken from the literature (Clayton et al. 1984, 1991; Weisberg et al. 1991; Clayton and Mayeda 1999). The O isotopic compositions for CR, CK5, O, E chondrites, and differentiated meteorites are the mean values of the respective groups. The $\epsilon^{54}\text{Cr}$ data are from this study and Qin et al. (2010), except for the basaltic achondrites (Trinquier et al. 2007).

about 20% of the variation among carbonaceous chondrites (Fig. 3). The total variation in $\epsilon^{54}\text{Cr}$ among all carbonaceous chondrites is approximately 1.5 ϵ . If the Cr isotopic compositions of ureilites are correlated with those of O in the same manner as carbonaceous chondrites, the expected variation in $\epsilon^{54}\text{Cr}$ among Almahata Sitta samples is approximately 0.3 ϵ . Although this value is only slightly larger than the total distribution of approximately 0.2 ϵ observed for this sample, we should be able to resolve this amount of variation. For instance, we have clearly resolved a difference of -0.2ϵ between Nakhla and a terrestrial standard and a difference of -0.4ϵ between ordinary chondrites and a terrestrial standard (Qin et al. 2010). The fact that all Almahata Sitta samples have essentially identical $\epsilon^{54}\text{Cr}$ values within error, regardless of their O isotope compositions, seems to support our earlier conclusion that variations in $\Delta^{17}\text{O}$ and $\epsilon^{54}\text{Cr}$ observed in meteorites were not necessarily caused by the same process (Fig. 3), although we cannot completely rule out the possibility that the $\epsilon^{54}\text{Cr}$ values were initially heterogeneous among different pieces and were homogenized by later events. The potential correlation between $\Delta^{17}\text{O}$ and $\epsilon^{54}\text{Cr}$ for carbonaceous chondrites may simply reflect a mixing between two reservoirs with distinct O and Cr isotopic compositions.

The ^{53}Mn - ^{53}Cr results show that there was live ^{53}Mn in the Almahata Sitta parent body at the time of the final closure of the Mn-Cr system. The Mn-Cr

isochron yields an initial $^{53}\text{Mn}/^{55}\text{Mn}$ value of $(3.1 \pm 1.1) \times 10^{-6}$, which, calculated relative to the Mn-Cr (Glavin et al. 2004) and U-Pb ages (Brennecka et al. 2010) for the angrite D'Orbigny translates to an absolute age for Almahata Sitta of 4563.6 ± 2.2 Ma. The initial $^{53}\text{Mn}/^{55}\text{Mn}$ value determined here for Almahata Sitta is indistinguishable within uncertainty of the $^{53}\text{Mn}/^{55}\text{Mn}$ value of $(2.84 \pm 0.10) \times 10^{-6}$ determined for the feldspathic clasts of polymict ureilite DaG-165, and is also concordant with the Al-Mg age of feldspathic clasts of polymict ureilite DaG-319 of 0.5 ($-0.3/+0.5$) Ma after D'Orbigny (Goodrich et al. 2010). Thus, all three polymict ureilites record an event at approximately 4564 Ma, approximately 4–5 Ma after the formation of refractory inclusions in CV chondrites (Amelin et al. 2002; Connelly et al. 2008). This is consistent with all three ureilites originating from the same parent body. Although it is difficult to directly relate the feldspathic clasts to the main group ureilites petrologically and chemically, they appear to be indigenous products of differentiation that are in some way complementary to the olivine-pigeonite residues represented by the majority of monomict ureilites (Kita et al. 2003; Cohen et al. 2004; Goodrich et al. 2004). Thus, the age recorded in the three ureilites likely reflects the time of igneous differentiation of the parent body of ureilites. Hence, a ureilite melting event occurred almost simultaneously with magmatic activity in the HED parent body, as suggested by the Mn-Cr and Al-Mg ages (4563–4565 Ma) of eucrites and diogenites (Lugmair and Shukolyukov 1998; Bizzarro et al. 2005; Trinquier et al. 2008), and in the angrite parent body (Glavin et al. 2004; Amelin 2008). Ureilites demonstrate, once again, that extensive melting of small planetesimals occurred in the first few Ma of the young solar system (Kita et al. 2003; Bizzarro et al. 2005; Qin et al. 2008). The early melting is consistent with decay of short-lived nuclides, such as ^{26}Al and possibly ^{60}Fe , as the major heat source, when these short-lived nuclides were abundant enough to induce melting.

The Cr isotopic composition of Almahata Sitta suggests that ^{53}Cr isotopic equilibrium was reached at the time of magmatic activity in its parent body, which is in contrast to the heterogeneous O isotopic compositions. One explanation of the discrepant behavior of Cr- and O-isotope systematics is that melting and melt segregation in the ureilite parent body was incomplete and rapid. Melt segregation might have been accelerated by smelting reactions between Fe-bearing olivine and carbon-rich materials, precluding appreciable oxygen isotope exchange with their wall rock (Goodrich et al. 2007). If this is the case, the Cr isotope exchange must have occurred much faster than

O isotopes. At the present stage, it is not fully understood yet.

CONCLUSIONS

The $^{54}\text{Cr}/^{52}\text{Cr}$ ratio of Almahata Sitta is identical to that of eucrites, diogenites, and mesosiderites, suggesting that the parent body of Almahata Sitta accreted in the same or a similar nebular region to the HED parent body. The $^{54}\text{Cr}/^{52}\text{Cr}$ value is clearly resolved from those of known carbonaceous chondrites or any other known type of chondrite. This refutes the suggested genetic link between ureilites and known carbonaceous chondrites. In addition, the heterogeneous $\Delta^{17}\text{O}$ values of the Almahata Sitta samples are not accompanied by variations in $^{54}\text{Cr}/^{52}\text{Cr}$, in contrast to the covariations observed for carbonaceous chondrites. This is supportive of our earlier suggestions that the variations in $\Delta^{17}\text{O}$ and $^{54}\text{Cr}/^{52}\text{Cr}$ in the solar system materials were not governed by the same process (Qin et al. 2010).

The ^{53}Mn - ^{53}Cr systematics show that there was live ^{53}Mn at the time of the final closure of the Mn-Cr system in Almahata Sitta, with an initial $^{53}\text{Mn}/^{55}\text{Mn}$ value of $3.1 (\pm 1.1) \times 10^{-6}$. The Mn-Cr isochron translates to an age of 4563.6 (± 2.2) Ma, which is consistent with Mn-Cr and Al-Mg ages obtained for two other polymict ureilites, and probably reflects the age of magmatic activity in the ureilite parent body. Ureilite igneous differentiation is similar to or slightly younger than the Mn-Cr and Al-Mg ages of eucrites, diogenites, and mesosiderites (Lugmair and Shukolyukov 1998; Bizzarro et al. 2005; Trinquier et al. 2008). Large-scale magmatic activity took place on a number of asteroidal bodies in the first few Ma following formation of the solar system powered by the decay of short-lived radionuclides, especially ^{26}Al . But melting and separation of melt from solid residues was probably rather incomplete and fast in the ureilite parent body (Goodrich et al. 2007).

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