



Hf–W thermochronometry: II. Accretion and thermal history of the acapulcoite–lodranite parent body

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ARTICLE INFO

Article history:

Received 11 November 2008

Received in revised form 8 April 2009

Accepted 9 April 2009

Available online 3 June 2009

Editor: R.W. Carlson

Keywords:

chronology

isochron

acapulcoite

lodranite

hafnium

tungsten

accretion

metamorphism

partial melting

cooling

thermal modeling

closure temperature

parent body

ABSTRACT

Acapulcoites and lodranites are highly metamorphosed to partially molten meteorites with mineral and bulk compositions similar to those of ordinary chondrites. These properties place the acapulcoites and lodranites between the unmelted chondrites and the differentiated meteorites and as such acapulcoites–lodranites are of special interest for understanding the initial stages of asteroid differentiation as well as the role of ²⁶Al heating in the thermal history of asteroids. To constrain the accretion timescale and thermal history of the acapulcoite–lodranite parent body, and to compare these results to the thermal histories of other meteorite parent bodies, the Hf–W system was applied to several acapulcoites and lodranites. Acapulcoites Dhofar 125 and NWA 2775 and lodranite NWA 2627 have indistinguishable Hf–W ages of $\Delta t_{\text{CAI}} = 5.2 \pm 0.9$ Ma and $\Delta t_{\text{CAI}} = 5.7 \pm 1.0$ Ma, corresponding to absolute ages of 4563.1 ± 0.8 Ma and 4562.6 ± 0.9 Ma. Closure temperatures for the Hf–W system for acapulcoites and lodranites, estimated from numerical simulations of W diffusion in high-Ca pyroxene, are 975 ± 50 °C and 1025 ± 50 °C, respectively. Owing to these high closure temperatures, the Hf–W ages provide information on the earliest high-temperature evolution, and combined with thermal modeling indicate that the acapulcoite–lodranite parent body accreted ~ 1.5 – 2 Ma after CAI formation, was internally heated by ²⁶Al decay, and reached its thermal peak ~ 3 Ma after CAI formation. Cooling rates for acapulcoites decreased from ~ 120 °C/Ma just below the thermal peak to ~ 50 °C/Ma at ~ 600 °C. Over the same temperature interval the cooling rate for lodranites decreased from ~ 100 °C/Ma to ~ 40 °C/Ma. These thermal histories may reflect cooling in the uppermost ~ 10 km of a parent body with a radius of ~ 35 – 100 km. Acapulcoites and lodranites evolved with a ¹⁸⁰Hf/¹⁸⁴W ratio of ~ 0.64 , which is indistinguishable from that of H chondrites but significantly lower than ¹⁸⁰Hf/¹⁸⁴W ~ 1.23 for carbonaceous chondrites. The low ¹⁸⁰Hf/¹⁸⁴W ratios of acapulcoites–lodranites were established before ~ 2 Ma and, hence, prior to partial melting in the parent body at ~ 3 Ma. Thus, they must reflect Hf–W fractionation of the precursor material by processes in the solar nebula. Combined with Hf–W ages of $\Delta t_{\text{CAI}} < 1$ Ma for differentiation of the parent bodies of magmatic iron meteorites and an Hf–W age of $\Delta t_{\text{CAI}} \sim 2.5$ Ma for the accretion of the H chondrite parent body, the Hf–W results for acapulcoites and lodranites reveal an inverse correlation between accretion age of asteroids and peak temperature in their interiors. The different thermal histories of most meteorite parent bodies, therefore, primarily reflect variations in their initial ²⁶Al abundance, which is determined by their accretion time.

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1. Introduction

Determining the accretion timescale and early thermal history of meteorite parent bodies is key for understanding the formation of planetesimals and for constraining the parameters that controlled their subsequent evolution. Such information can be obtained by applying short- and long-lived chronometers to meteorites (or

preferably a suite of meteorites derived from a single parent body) but also requires knowledge of the closure temperature (T_c) for diffusive exchange of parent and daughter elements among the different minerals in a meteorite (Dodson, 1973; Ganguly and Tirone, 2001). Most chronometers such as the U–Pb, Al–Mg or Ar–Ar systems closed at temperatures much lower than the peak temperatures reached in most meteorite parent bodies and as such provide constraints on the mid- to low-temperature evolution of meteorite parent bodies (Göpel et al., 1994; LaTourrette and Wasserburg, 1998; Trierloff et al., 2003). As a consequence, their applicability for determining the high-temperature thermal history is limited. However, such information is of special

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interest because it can provide constraints on the accretion timescale and the energy source(s) responsible for parent body heating. In contrast to the aforementioned chronometers, the Hf–W system ($t_{1/2} \sim 9$ Ma) has a relatively high closure temperature, which for H5–6 chondrites is almost as high as their peak metamorphic temperatures (Kleine et al., 2008). Thus, the Hf–W system appears ideally suited for determining the accretion timescale and high-temperature evolution of meteorite parent bodies.

Recent advances in the chronology of meteorites reveal that the accretion of the parent bodies of most if not all differentiated meteorites occurred within the first Ma after formation of Ca–Al-rich inclusions (CAIs) (Bizzarro et al., 2005; Kleine et al., 2005; Markowski et al., 2006; Schärsten et al., 2006; Burkhardt et al., 2008). In contrast, accretion of most chondrite parent bodies occurred later, more than ~ 2 Ma after CAI formation (Kita et al., 2000; Kunihiro et al., 2004). These results indicate that the early evolution of asteroids was determined by their initial ^{26}Al content, which was sufficiently abundant in the early-formed planetesimals to cause global melting and differentiation but had decayed to levels insufficient to trigger melting in the late-formed chondrite parent bodies (Bizzarro et al., 2005; Kleine et al., 2005; Schärsten et al., 2006).

Acapulcoites and lodranites have mineral and bulk compositions similar to those of ordinary chondrites but they have non-chondritic textures that result from recrystallization during high-temperature metamorphism and from partial melting. The acapulcoites show evidence for partial melting at the FeS–FeNi cotectic and in some cases silicate melts may have formed (Palme et al., 1981; Zipfel et al., 1995). The lodranites were heated to higher temperatures than the acapulcoites and are partial melting residues depleted in troilite and plagioclase (McCoy et al., 1997a). Thus, acapulcoites and lodranites have properties that place them between the unmelted chondrites and the differentiated meteorites (Palme et al., 1981; Zipfel et al., 1995; McCoy et al., 1996, 1997a) and as such are of special interest for understanding the differentiation of asteroids as well as the role of ^{26}Al heating in the thermal history of asteroids. Several lines of evidence indicate that acapulcoites and lodranites, despite their different thermal histories, were derived from a common parent body: the occurrence of meteorites that are transitional between acapulcoites and lodranites (McCoy et al., 1997a); similar O isotope compositions (Clayton and Mayeda, 1996); similar mineral compositions (McCoy et al., 1997b); and indistinguishable cosmic ray exposure ages (Eugster and Lorenzetti, 2005).

To constrain the accretion timescale and high-temperature thermal history of the acapulcoite–lodranite parent body, we applied the Hf–W chronometer to several acapulcoites and lodranites. The closure temperatures of the Hf–W system in these meteorites were estimated numerically using the model of Van Orman et al. (2001, 2006). The Hf–W ages and T_c estimates are then used in conjunction with previously published ages for acapulcoites and lodranites and thermal modeling to constrain the accretion timescale and thermal history of the acapulcoite–lodranite parent body. Finally, these results are compared to the thermal history of other meteorite parent bodies.

2. Samples and analytical techniques

2.1. Samples

Three acapulcoites (NWA 2656, NWA 2775, Dhofar 125) and 2 lodranites (NWA 2627, NWA 4663) were selected for determining internal Hf–W isochrons. In addition, Hf–W data for a ~ 700 mg whole-rock of the lodranite Gibson and for a ~ 40 mg metal separate from Acapulco were obtained. Detailed descriptions of the acapulcoites and lodranites investigated for this study can be found elsewhere (Greshake et al., 2001; Connolly et al., 2007; Irving et al., 2007). NWA 2775 and 2656 have an average grain size within the typical range of acapulcoites (150–230 μm), whereas Dhofar 125 is somewhat

finer grained (~ 100 μm). The high-Ca pyroxenes in NWA 2656 have an average grain size of 115 ± 12 μm , and a modal content of 13 vol.%. The high-Ca pyroxene in NWA 2775 is larger (190 ± 12 μm) but less abundant (10 vol.%). The average grain sizes in NWA 2627 and 4663 are larger than in acapulcoites and within the range of lodranites. Note that NWA 2627 has initially been classified as an acapulcoite but its average grain size indicates that it is a lodranite (Irving et al., 2007). The high-Ca pyroxenes in NWA 2627 and NWA 4663 have average grain sizes of 430 ± 25 μm (modal content of 8 vol.%) and 373 ± 18 μm (modal content of 7 vol.%). In addition there are tiny high-Ca pyroxene grains that occur as inclusions in olivine and probably represent trapped liquid within the crystallizing olivine. All acapulcoites and lodranites analyzed here show minimal to moderate degrees of alteration (W1 and W2, or W3 in the case of NWA 2627), resulting in partial oxidation of metals.

2.2. Analytical techniques

Pieces of meteorite (with a total weight of ~ 3 –5 g for each sample) were cleaned with abrasive paper and with a succession of 0.05 M HNO_3 , de-ionized H_2O and ethanol in an ultrasonic bath to remove any contamination introduced during cutting from larger samples. Whole-rock powders were obtained by crushing a 0.5–1 g piece of each sample in an agate mortar. The remaining material (~ 2 –4 g) was gently crushed in an agate mortar and separated into < 40 μm and 40–150 μm fractions using nylon sieves. During crushing metal grains were removed using a hand-magnet and silicate dust attached to or silicate grains intergrown with the metal grains were removed by repeated crushing of the magnetic fraction under ethanol. Between 15 and 100 mg of pure metal separates were obtained for most samples with the exception of NWA 2656, for which no pure metal grains could be obtained. Although all visible metal grains were removed, the 40–150 μm fractions were still slightly magnetic, most likely reflecting the presence of tiny metal inclusions in the silicate and oxide grains. The 40–150 μm fractions were further separated using a hand-magnet into a “weakly-magnetic” and “non-magnetic” fraction. These were then further separated in several weakly- and non-magnetic fractions that were labeled WM- n and NM- n , $n = 1, 2, 3, \dots$, where $n = 1$ always denotes the least magnetic fraction among the weakly- and non-magnetic fractions for each sample. All WM and NM fractions were inspected under the binocular microscope and they consist mainly of olivine and pyroxene but most fractions also contain some feldspar. All the WM and NM fractions were cleaned with ethanol in an ultrasonic bath and powdered in an agate mortar. Remaining metal grains were then removed from these powders using a hand-magnet and the resulting fractions weighted between 150 and 300 mg.

The metal separates were dissolved in 15 ml Savillex[®] vials at ~ 120 $^\circ\text{C}$ on a hotplate using a 6 M HCl–0.06 M HF mixture. In some cases, a few drops of concentrated HNO_3 were added. The NM fractions were dissolved in 60 ml Savillex[®] vials at ~ 180 $^\circ\text{C}$ on a hotplate using HF– HNO_3 (7:3). After digestion, the samples were dried and re-dissolved in HNO_3 – H_2O_2 to destroy organic compounds. Then the samples were completely dissolved in 6 M HCl–0.06 M HF and depending on the W contents a 1–10% aliquot was spiked with a mixed ^{180}Hf – ^{183}W tracer that was calibrated against pure Hf and W metals (Kleine et al., 2004).

The methods for the separation of Hf and W from the sample matrix were identical to those outlined in Kleine et al. (2008). Total procedural blanks ranged from ~ 100 to ~ 300 pg for the W isotope composition measurements and ~ 10 to 30 pg W and ~ 1 to 10 pg Hf for the isotope dilution measurements. For most samples, blanks are negligible but for some of the most W-depleted fractions blank corrections on the measured $^{182}\text{W}/^{184}\text{W}$ ratios were significant and ranged from < 1 to $\sim 4 \epsilon^{182}\text{W}$ (Table 1).

All isotope measurements were performed using a Nu Plasma MC-ICP-MS at ETH Zurich, equipped with a Cetac Aridus desolvating

Table 1
Hf–W data for acapulcoites and lodranites.

Sample	Hf (ppb)	W (ppb)	$^{180}\text{Hf}/^{184}\text{W} \pm 2\sigma$	$\epsilon^{182}\text{W} \pm 2\sigma$ meas.	$\epsilon^{182}\text{W} \pm 2\sigma$ corr.	$\epsilon^{183}\text{W} \pm 2\sigma$
<i>Dhofar 125 (acapulcoite)</i>						
WR-1	190.3	109.9	2.042 ± 0.020	-1.79 ± 0.28		-0.01 ± 0.26
WR-2	189.0	118.6	1.881 ± 0.019	-1.54 ± 0.18		0.01 ± 0.15
M	0.4	901.1	0.0005 ± 0.0001	-2.97 ± 0.21		0.11 ± 0.17
Mean (2 SE)				-3.33 ± 0.24		-0.20 ± 0.18
Fines-1	133.1	373.9	0.415 ± 0.004	-3.15 ± 0.36		-0.04 ± 0.31
Fines-2	131.8	375.1	0.420 ± 0.004	-2.90 ± 0.21		-0.08 ± 0.11
WM-1	171.8	91.3	2.22 ± 0.02	-2.77 ± 0.17		0.26 ± 0.13
WM-2	173.1	105.9	1.92 ± 0.02	-1.25 ± 0.19		-0.17 ± 0.21
NM-1	257.9	14.3	21.3 ± 0.4	-1.28 ± 0.18	13.5 ± 1.2	-0.18 ± 0.13
NM-2	237.4	25.9	10.8 ± 0.2	13.2 ± 0.8	4.1 ± 1.3	-0.30 ± 0.56
NM-3	270.9	30.0	10.6 ± 0.2	4.01 ± 0.78	4.1 ± 1.3	-0.11 ± 0.48
NM-4	174.2	29.4	6.99 ± 0.07	4.78 ± 0.52	4.9 ± 1.2	-0.14 ± 0.36
<i>NWA 2775 (acapulcoite)</i>						
WR	179.8	174.4	1.21 ± 0.01	-2.30 ± 0.21		-0.47 ± 0.17
Mean (2 SE)				-2.22 ± 0.32		-0.44 ± 0.21
M	6.2	1380.6	0.0053 ± 0.0001	-2.26 ± 0.09		-0.46 ± 0.04
WM-1	121.7	329.0	0.436 ± 0.004	-3.15 ± 0.37		-0.26 ± 0.18
Mean (2 SE)				-2.29 ± 0.23		-0.25 ± 0.13
WM-2	122.8	416.6	0.348 ± 0.003	-2.72 ± 0.14		-0.04 ± 0.12
Mean (2 SE)				-2.92 ± 0.18		-0.09 ± 0.11
Fines	159.7	213.4	0.883 ± 0.009	-2.64 ± 0.37		-0.13 ± 0.13
Mean (2 SE)				-2.64 ± 0.19		-0.15 ± 0.13
NM-1	189.9	3.8	58.2 ± 5.7	-2.74 ± 0.19		-0.04 ± 0.13
NM-2	165.4	5.1	38.1 ± 2.9	-2.75 ± 0.14		-0.17 ± 0.11
NM-3	180.2	27.0	7.63 ± 0.15	-2.71 ± 0.07		-0.12 ± 0.08
Mean (2 SE)				-2.28 ± 0.18		-0.20 ± 0.15
NM-1	189.9	3.8	58.2 ± 5.7	-2.14 ± 0.18		-0.28 ± 0.14
NM-2	165.4	5.1	38.1 ± 2.9	-1.94 ± 0.26		-0.27 ± 0.19
NM-3	180.2	27.0	7.63 ± 0.15	-2.12 ± 0.20		-0.25 ± 0.05
Mean (2 SE)				-2.41 ± 0.16	39.7 ± 2.9	2.21 ± 0.69
Mean (2 SE)				-2.29 ± 0.17	26.8 ± 2.3	1.76 ± 1.10
Mean (2 SE)				-2.35 ± 0.11	2.5 ± 1.2	-0.25 ± 0.55
Mean (2 SE)				-2.23 ± 0.18		
Mean (2 SE)				-1.06 ± 0.17		
Mean (2 SE)				-0.93 ± 0.16		
Mean (2 SE)				-1.00 ± 0.13		
Mean (2 SE)				-0.43 ± 0.44		
Mean (2 SE)				-0.47 ± 0.26		
Mean (2 SE)				-0.01 ± 1.10		
<i>Acapulco</i>						
M	0.747	1002.6	0.000879 ± 0.000005	-3.03 ± 0.12		0.25 ± 0.08
<i>NWA 2627 (lodranite)</i>						
WR	95.7	168.6	0.670 ± 0.007	-2.42 ± 0.16		-0.09 ± 0.23
M-1	18.3	1163.4	0.0185 ± 0.0002	-2.85 ± 0.27		0.22 ± 0.27
M-2	36.4	841.7	0.051 ± 0.001	-2.89 ± 0.30		-0.08 ± 0.27
Mean (2 SE)				-3.01 ± 0.22		0.14 ± 0.27
WM	66.8	58.5	1.346 ± 0.013	-2.95 ± 0.28		0.03 ± 0.28
Fines	125.2	38.4	3.85 ± 0.04	-2.15 ± 0.37	-2.2 ± 0.6	1.74 ± 0.23
NM-1	126.5	8.8	17.0 ± 0.6	-0.42 ± 0.41	-0.4 ± 0.6	-0.30 ± 0.36
NM-2	92.7	12.7	8.59 ± 0.22	7.77 ± 0.39	9.0 ± 0.9	1.74 ± 0.31
Mean (2 SE)				3.24 ± 0.35	3.6 ± 0.6	0.31 ± 0.23
<i>NWA 4663 (lodranite)</i>						
M	7.6	1162.8	0.0077 ± 0.0001	-3.12 ± 0.28		0.04 ± 0.15
Mean (2 SE)				-2.65 ± 0.14		0.05 ± 0.09
WM-2	308.9	224.7	1.62 ± 0.02	-2.74 ± 0.13		0.02 ± 0.12
WM-3	135.7	664.0	0.241 ± 0.002	-2.72 ± 0.16		0.09 ± 0.11
NM-1	651.1	28.5	26.9 ± 0.4	-0.56 ± 0.21		0.21 ± 0.19
NM-2	344.9	23.7	17.2 ± 0.4	-2.72 ± 0.14		-0.04 ± 0.12
Mean (2 SE)				3.24 ± 0.30	3.39 ± 0.31	0.91 ± 0.53
Mean (2 SE)				4.09 ± 0.33	4.38 ± 0.36	0.72 ± 0.15
<i>Gibson (lodranite)</i>						
WR	154.8	153.5	1.19 ± 0.01	-1.78 ± 0.24		-0.03 ± 0.15

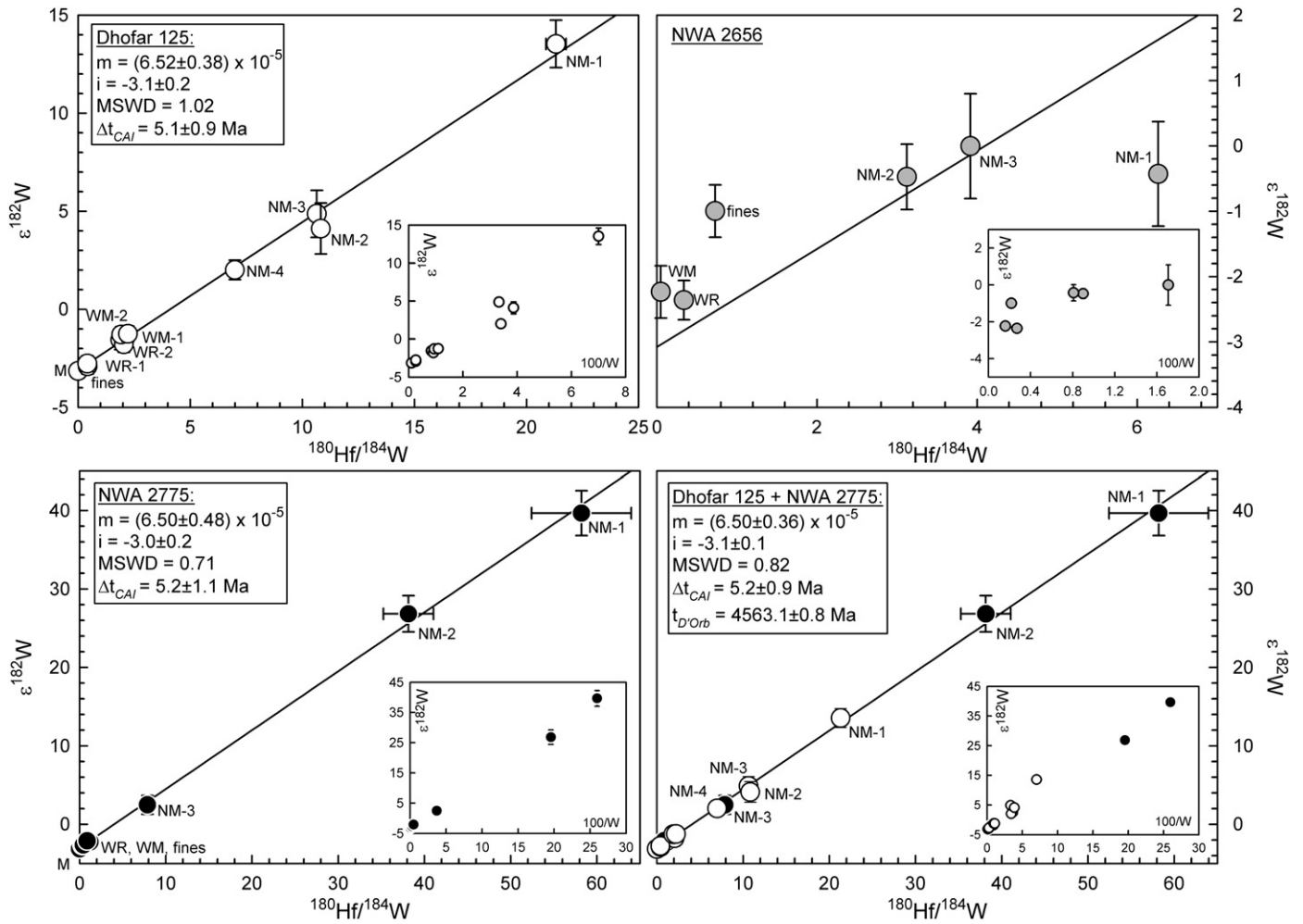


Fig. 1. $\epsilon^{182}\text{W}$ versus $^{180}\text{Hf}/^{184}\text{W}$ for acapulcoites Dhofar 125, NWA 2656, NWA 2775. m = initial $^{182}\text{Hf}/^{180}\text{Hf}$, i = initial $\epsilon^{182}\text{W}$. Regressions are calculated using the model 1 fit of Isoplot (Ludwig, 1991). Δt_{CAI} is the formation age relative to CAIs; the absolute age t is calculated relative to the angrite D'Orbigny ($^{182}\text{Hf}/^{180}\text{Hf} = (7.20 \pm 0.21) \times 10^{-5}$; $^{207}\text{Pb}/^{206}\text{Pb}$ age = 4564.42 ± 0.12 Ma). Linear regressions of the Hf–W data for Dhofar 125 and NWA 2775 yield identical slopes and intercepts and the regression of the combined data for these two acapulcoites provides the best estimate for the initial $^{182}\text{Hf}/^{180}\text{Hf}$ and $\epsilon^{182}\text{W}$ at the time of the last Hf–W equilibration in acapulcoites. No correlation between $\epsilon^{182}\text{W}$ and $^{180}\text{Hf}/^{184}\text{W}$ is observed for NWA 2656, indicating that the Hf–W system in this sample is disturbed.

nebulizer. The measurement protocol was identical to that described in Kleine et al. (2008). Prior to measurement, the samples were re-dissolved and dried several times in a few drops of 14 M HNO_3 and 30% H_2O_2 to remove organic compounds and then taken up in a 0.56 M HNO_3 –0.24 M HF mixture. Tungsten isotope compositions of metals and whole-rocks were typically measured with a signal intensity of ~ 2 V on ^{182}W , which was obtained for a ~ 20 ppb W solution. For these samples, 60 ratios (3 blocks of 20 ratios) were measured resulting in within-run statistics of the order of 0.2 ϵ units (2 σ). Owing to the low W contents in some NM fractions, their W isotope compositions were measured in 1 or 2 blocks of 20 ratios each with signal intensities of ~ 0.2 to 1 V on ^{182}W . The within-run statistics of these measurements were typically between 0.5 and 1.5 ϵ units. Instrumental mass bias was corrected relative to $^{186}\text{W}/^{183}\text{W} = 1.9859$ using the exponential law. Small isobaric interferences of Os on masses 184 and 186 were corrected by monitoring ^{188}Os and were negligible.

The $^{182}\text{W}/^{184}\text{W}$ and $^{183}\text{W}/^{184}\text{W}$ ratios of all samples were determined relative to two standard runs bracketing the sample run and are

reported in $\epsilon^{181}\text{W}$ units, which is the deviation of the $^{181}\text{W}/^{184}\text{W}$ ratio from the terrestrial standard value in parts per 10,000. The reproducibility of the ~ 20 ppb standard during one measurement day is typically equal to or better than ~ 0.3 ϵ units (2 SD) for the $^{182}\text{W}/^{184}\text{W}$ ratio and ~ 0.2 ϵ units (2 SD) for the $^{183}\text{W}/^{184}\text{W}$ ratio. The external reproducibility of the W isotope measurements typically is 0.3–0.4 ϵ units (2 SD) for the $^{182}\text{W}/^{184}\text{W}$ ratio and 0.2–0.3 ϵ units (2 SD) for the $^{183}\text{W}/^{184}\text{W}$ ratio (Kleine et al., 2008). The uncertainties for the W isotope measurements of the W-poor NM-1 and -2 fractions of Dhofar 125 and NWA 2775 were assessed by repeated measurement (1 block of 20 ratios each) of ~ 2 and ~ 10 ppb W standard solutions that yielded an external reproducibility of ~ 2 and ~ 0.8 $\epsilon^{182}\text{W}$ (2 SD), respectively. This is similar to the within-run statistics obtained for the measurements of the W-poor NM fractions (Table 1). Note that the major source of uncertainty in the $^{182}\text{W}/^{184}\text{W}$ of the W-poor NM fractions is the blank correction.

The accuracy of the measurements was monitored by analyzing several carbonaceous chondrites, which all yielded the previously determined value of -1.9 ± 0.1 $\epsilon^{182}\text{W}$ (Kleine et al., 2004). Furthermore,

Notes to Table 1

M = metal fraction, WM = weakly magnetic fraction, NM = non-magnetic fraction, WR = whole rock, corr. = corrected for blank, meas. = measured. The quoted $\pm 2\sigma$ uncertainties for measured $\epsilon^{182}\text{W}$ and $\epsilon^{183}\text{W}$ are analytical errors. Uncertainties on blank-corrected $\epsilon^{182}\text{W}$ values are calculated by propagating the external reproducibility of the isotope measurements and a $\pm 50\%$ uncertainty on the blank correction. The regression calculations use an external reproducibility of ± 0.4 ϵ except for the W-poor fractions (Dhofar 125 NM-1, 2 and 3; NWA 2656 NM-1 and 2; NWA 2627 NM-1), for which the uncertainties as given in the table were used (see text for details).

$^{183}\text{W}/^{184}\text{W}$ ratios were used as a monitor for accurate measurements and agree for most samples to within $\pm 0.2 \epsilon$ units with the terrestrial value (Table 1). Elevated $^{183}\text{W}/^{184}\text{W}$ ratios for some fractions from the NWA 2775 acapulcoite and the NWA 4663 and NWA 2627 lodranites are attributed to an organic interference on mass 183 that was successfully removed for all other samples by treatment with $\text{HNO}_3\text{-H}_2\text{O}_2$. Elevated measured $^{183}\text{W}/^{184}\text{W}$ ratios have been observed before during W isotope measurements of some eucrites, carbonaceous chondrites, CAIs and H chondrites and for these samples the $^{182}\text{W}/^{184}\text{W}$ ratio normalized to $^{186}\text{W}/^{184}\text{W} = 0.92767$ agrees with $^{182}\text{W}/^{184}\text{W}$ ratios for other samples of these groups (Kleine et al., 2002, 2004; Burkhardt et al., 2008; Kleine et al., 2008). This indicates that only ^{183}W is affected, such that for the fractions with elevated measured $^{183}\text{W}/^{184}\text{W}$ the reported $\epsilon^{182}\text{W}$ values were calculated from the $^{182}\text{W}/^{184}\text{W}$ ratio normalized to $^{186}\text{W}/^{184}\text{W}$.

3. Results

The Hf and W concentrations and the W isotope compositions of metals, whole-rocks and non-magnetic fractions analyzed for this study are reported in Table 1 and shown in Figs. 1–3. The acapulcoite and lodranite metals have W concentrations ranging from ~850 to 1400 ppb and have indistinguishable $\epsilon^{182}\text{W}$ values of ~-3.0. All metal

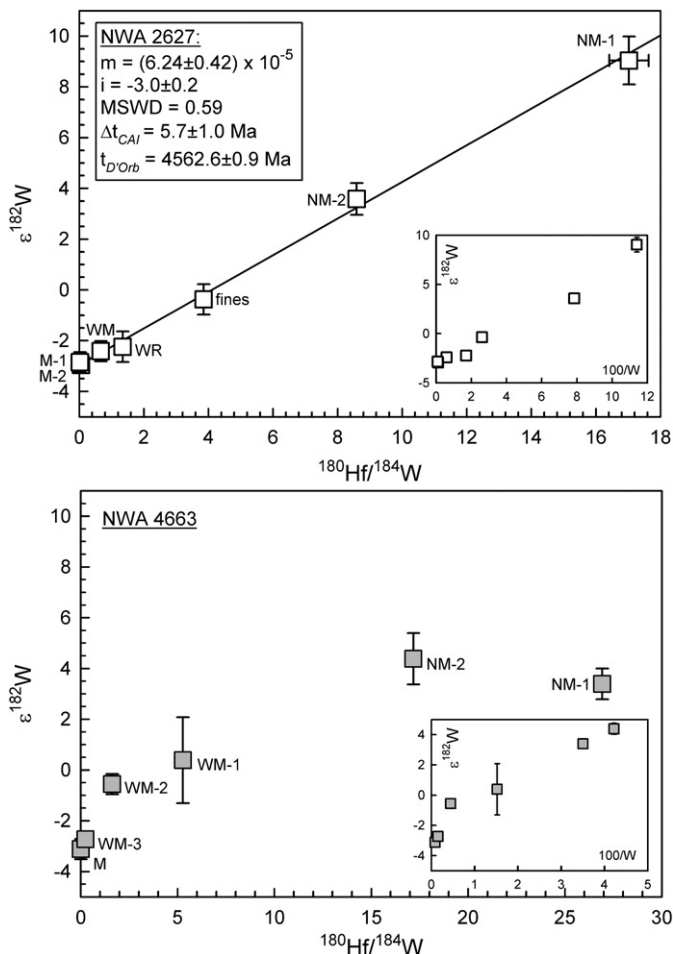


Fig. 2. $\epsilon^{182}\text{W}$ versus $^{180}\text{Hf}/^{184}\text{W}$ for lodranites NWA 2627 and NWA 4663. m = initial $^{182}\text{W}/^{180}\text{Hf}$, i = initial $\epsilon^{182}\text{W}$. Regressions are calculated using the model 1 fit of Isoplot (Ludwig, 1991). Δt_{CAI} is the formation age relative to CAIs; the absolute age t is calculated relative to the angrite D'Orbigny. Linear regressions of the Hf–W data for NWA 2627 yield a precise isochron but no correlation between $\epsilon^{182}\text{W}$ and $^{180}\text{Hf}/^{184}\text{W}$ is observed for NWA 4663, indicating that the Hf–W system in this sample is disturbed.

separates have low $^{180}\text{Hf}/^{184}\text{W}$ ratios of less than ~0.05, indicating that pure metal separates were obtained. The acapulcoite and lodranite whole-rocks have $^{180}\text{Hf}/^{184}\text{W}$ ratios ranging from ~0.3 to ~2 and $\epsilon^{182}\text{W}$ values ranging from ~-2.4 to ~-1.5 but no systematic difference between acapulcoites and lodranites is apparent. The WM fractions have W contents and $\epsilon^{182}\text{W}$ between those of the whole-rocks and metals in most cases, most likely reflecting a higher abundance of metal in the WM fractions compared to the whole rocks. The NM fractions have variable Hf and W contents ranging from ~90 to ~660 ppb Hf and from ~4 to ~124 ppb W, respectively. Their $^{180}\text{Hf}/^{184}\text{W}$ ratios are between ~3 and ~60, resulting in elevated $^{182}\text{W}/^{184}\text{W}$ ratios ranging from ~0 to ~40 $\epsilon^{182}\text{W}$. Owing to the low W contents and radiogenic $^{182}\text{W}/^{184}\text{W}$ of some of the NM fractions, blank corrections were significant and typically ranged from <1 to ~4 $\epsilon^{182}\text{W}$ (Table 1).

As shown in Figs. 2 and 3, the $^{180}\text{Hf}/^{184}\text{W}$ ratios and $\epsilon^{182}\text{W}$ values correlate for the acapulcoites Dhofar 125 and NWA 2775, and for the lodranite NWA 2627, such that precise regression could be obtained for these samples (MSWD < 1). The uncertainties on the slopes of these regressions are better than ~7%, resulting in uncertainties for the ages of better than $\pm 1 \text{ Ma}$, if these regressions are interpreted as isochrons. In contrast, no regression could be obtained for NWA 2656 and NWA 4663 and this most likely reflects disturbance of the Hf–W systematics due to terrestrial weathering. In spite of their different $^{180}\text{Hf}/^{184}\text{W}$ ratios, the NM fractions of NWA 2656 have a terrestrial W isotope composition, suggesting that this sample has been contaminated with terrestrial W during weathering in the desert. These effects are less pronounced in the metal and whole-rock fractions because these have much higher W contents. Note that for NWA 2656 no metal grains could be separated from the specimen investigated for this study, indicating substantial oxidation of the metals during terrestrial weathering. Similarly, in spite of very high $^{180}\text{Hf}/^{184}\text{W}$ ratios the NM fractions of NWA 4663 have much lower $\epsilon^{182}\text{W}$ values compared to fractions from other acapulcoites and lodranites with similar $^{180}\text{Hf}/^{184}\text{W}$. A plot of $\epsilon^{182}\text{W}$ vs. $1/W$ reveals that the NM fractions of NWA 4663 plot on a mixing line with terrestrial W, suggesting that their low $\epsilon^{182}\text{W}$ values are due to addition of terrestrial W.

4. Discussion

4.1. Hf–W isochron ages for acapulcoites and lodranites

To define an isochron the minerals of a sample must once have been in W isotope equilibrium, i.e., they must have had the same W isotope composition initially. Given that the acapulcoite and lodranite fractions were obtained mainly based on their magnetic susceptibility, the correlation of $\epsilon^{182}\text{W}$ with $^{180}\text{Hf}/^{184}\text{W}$ could potentially represent a mixing line between W-rich metal and virtually W-free silicates. Such a mixing line would have no chronological significance if the two endmembers had different initial $^{182}\text{W}/^{184}\text{W}$ ratios. However, as shown in Fig. 3, the variations in $^{180}\text{Hf}/^{184}\text{W}$ ratios among the fractions of the analyzed acapulcoites and lodranites require the presence of at least three independent components for Hf and W among the coexisting phases. The major host of W is metal while the major host of Hf is high-Ca pyroxene. The third component includes olivine and low-Ca pyroxene and is characterized by low Hf and W contents. These two minerals are considered here as one component because no pure olivine and low-Ca pyroxene separates were obtained. Both olivine and low-Ca pyroxene are not capable of incorporating significant amounts of neither Hf nor W (Richter and Shearer, 2003), such that their presence mainly causes dilution of the high Hf content of high-Ca pyroxene. Therefore, the variability in the Hf contents of the various NM fractions of one sample most likely results from different proportions of high-Ca pyroxene.

The presence of at least three independent components with regard to Hf and W among the coexisting phases of acapulcoites and lodranites reveals that the correlation between $\epsilon^{182}\text{W}$ and $^{180}\text{Hf}/^{184}\text{W}$

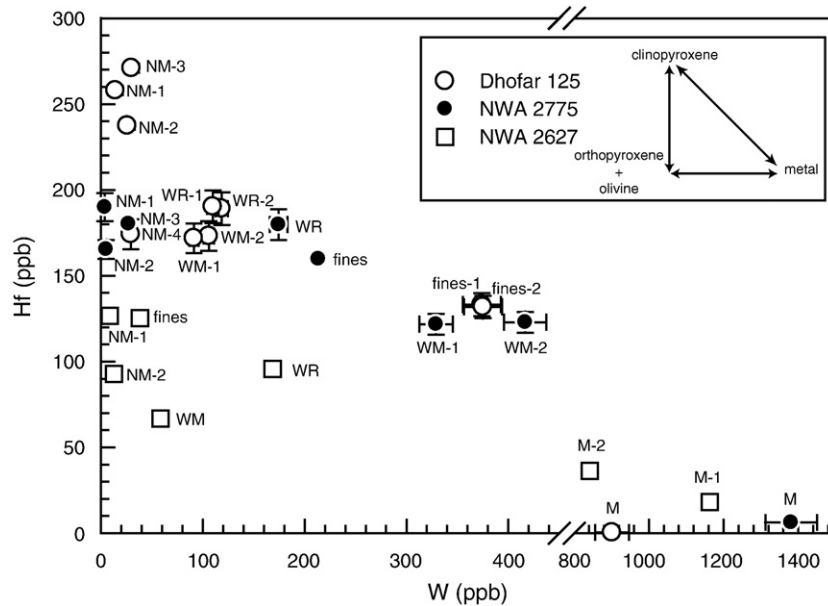


Fig. 3. Hf versus W contents for the different fractions of the analyzed acapulcoites and lodranites. The Hf and W concentrations in the coexisting phases of these acapulcoites and lodranites are not collinear, indicating the presence of at least three independent components for Hf and W among the coexisting phases. These components are high-Ca pyroxene, olivine + low-Ca pyroxene, and metal.

observed for the fractions of each of the acapulcoites and lodranites cannot reflect simple binary mixing between W-rich metal and virtually W-free silicates. This is also apparent from plots of $\epsilon^{182}\text{W}$ versus $1/\text{W}$, in which binary mixtures should form straight lines. This is not the case for any of those samples that exhibit linear correlations in the $\epsilon^{182}\text{W}$ versus $^{180}\text{Hf}/^{184}\text{W}$ plots. Each of the fractions, therefore, evolved to radiogenic $\epsilon^{182}\text{W}$ according to their $^{180}\text{Hf}/^{184}\text{W}$. Hence, the Hf–W data for the separates of acapulcoites and lodranites define isochrons and can be interpreted to have chronological significance.

Relative Hf–W ages, Δt_{CAI} , were calculated from the initial $^{182}\text{Hf}/^{180}\text{Hf}$ ratios obtained from the slopes of the isochrons relative to an initial $^{182}\text{Hf}/^{180}\text{Hf} = (9.72 \pm 0.44) \times 10^{-5}$ for CAIs (Burkhardt et al., 2008) and refer to the time of Hf–W closure in a sample elapsed since crystallization of type B CAIs. Acapulcoites Dhofar 125 and NWA 2775 have indistinguishable initial $^{182}\text{Hf}/^{180}\text{Hf}$ ratios of $(6.52 \pm 0.38) \times 10^{-5}$ and $(6.50 \pm 0.48) \times 10^{-5}$, respectively (Fig. 1). All data from these two acapulcoites combined define a single isochron with an initial $^{182}\text{Hf}/^{180}\text{Hf}$ of $(6.50 \pm 0.36) \times 10^{-5}$, corresponding to an age of $\Delta t_{\text{CAI}} = 5.2 \pm 0.9$ Ma. The initial $^{182}\text{Hf}/^{180}\text{Hf}$ of lodranite NWA 2627 of $(6.24 \pm 0.42) \times 10^{-5}$ is indistinguishable from the value for the acapulcoites and corresponds to an age of $\Delta t_{\text{CAI}} = 5.7 \pm 1.0$ Ma (Fig. 2).

The comparison of relative Hf–W ages to absolute ages (e.g., Pb–Pb ages) requires conversion of Hf–W formation intervals to an absolute timescale, which in turn requires knowledge of the initial $^{182}\text{Hf}/^{180}\text{Hf}$ and the absolute age of Hf–W closure in a given sample. Due to differences in closure temperatures of different chronometers, the ideal samples to obtain such information are angrites because (i) they cooled rapidly, such that differences in closure temperatures do not result in resolvable age differences, and (ii) they exhibit high U/Pb ratios, such that precise Pb–Pb ages are available (Lugmair and Galer, 1992; Amelin, 2008; Connelly et al., 2008). Precise initial $^{182}\text{Hf}/^{180}\text{Hf}$ ratios (Kleine et al., 2009) and Pb–Pb ages are available for several angrites and all of these can be used to calculate absolute Hf–W ages. Using the Pb–Pb age of 4564.42 ± 0.12 Ma (Amelin, 2008) and initial $^{182}\text{Hf}/^{180}\text{Hf} = (7.20 \pm 0.21) \times 10^{-5}$ for the angrite D’Orbigny (Markowski et al., 2007; Kleine et al., 2009) results in an absolute Hf–W age of 4563.1 ± 0.8 Ma for the acapulcoites Dhofar 125 and NWA 2775, and 4562.6 ± 0.9 Ma for lodranite NWA 2627. Identical results are obtained if absolute ages are calculated relative to the angrites Sahara 99555, NWA 4590, or NWA 4801 (Burkhardt et al., 2008; Kleine et al., 2009).

4.2. Closure temperature for the Hf–W system in acapulcoites and lodranites

To evaluate the significance of the Hf–W ages for constraining the thermal evolution of the acapulcoite–lodranite parent body, it is necessary to know the closure temperature, T_c , for W diffusion in the appropriate silicate–metal mixture. Here we use the approach employed by Kleine et al. (2008) to calculate T_c as a function of grain size and cooling rate for H chondrites. This approach is based on the models of Van Orman et al. (2001, 2006) to estimate the diffusion parameters of W in high-Ca pyroxene, the major host of Hf and hence radiogenic ^{182}W in acapulcoites and lodranites, and to numerically simulate the diffusion behavior of W in a high-Ca pyroxene–metal system. Details regarding this model are given in Kleine et al. (2008) and Van Orman et al. (2006).

Calculation of T_c as a function of cooling rate requires knowledge of the peak temperature, the grain size of high-Ca pyroxene, and the high-Ca pyroxene/metal ratio. Acapulcoites show evidence for melting at the FeNi–FeS eutectic, indicating that peak temperatures must have been higher than ~ 1000 °C. There is also evidence for limited melting of silicates (Zipfel et al., 1995; Mittlefehldt et al., 1996), such that ~ 1150 °C is a reasonable assumption for the peak temperature of acapulcoites. In contrast, lodranites exhibit depletions in troilite and plagioclase, suggesting mobility of FeNi–FeS and basaltic melts. This requires temperatures of up to ~ 1300 °C (Mittlefehldt et al., 1996; McCoy et al., 1997a), which was chosen here as the peak temperature for lodranites. The grain size of high-Ca pyroxene and the high-Ca pyroxene/metal ratios in the acapulcoites and lodranites investigated here were determined using their back scattered electron (BSE) images and are 100–200 μm and 350–450 μm , respectively.

Fig. 4 shows closure temperatures calculated as a function of cooling rate. For grain sizes of high-Ca pyroxenes of ~ 100 – 200 μm , typical for the acapulcoites investigated here, the assumed initial temperature is 1150 °C. For grain sizes from ~ 350 μm to ~ 450 μm , the assumed initial temperature is 1300 °C, as appropriate for lodranites. Fig. 4 reveals that T_c for the Hf–W system in acapulcoites ranges from ~ 900 °C for slow cooling ($dT/dt \sim 10$ °C/Ma) to ~ 1050 °C for fast cooling ($dT/dt \sim 1000$ °C/Ma). Due to their larger grain sizes the Hf–W closure temperature for lodranites are slightly higher and range from ~ 975 °C for slow cooling ($dT/dt \sim 10$ °C/Ma) to ~ 1100 °C for fast

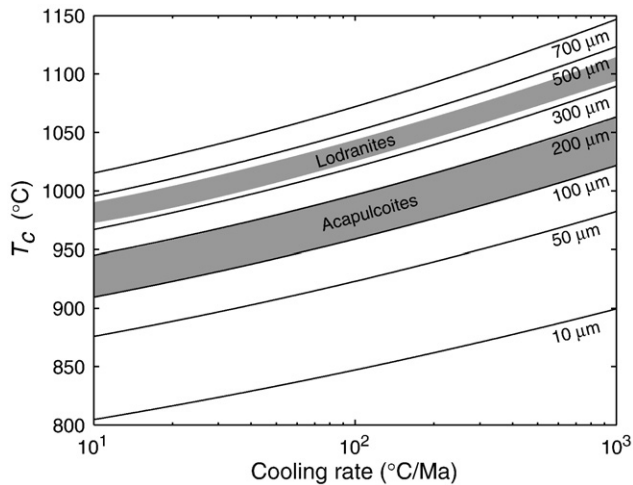


Fig. 4. Closure temperature of the Hf–W system as a function of grain size and cooling rate. The assumed initial temperature is 1150 °C for grain sizes of 10–200 μm and 1300 °C for larger grain sizes. Grey shaded areas indicate grain sizes of high-Ca pyroxenes in acapulcoites (100–200 μm) and lodranites (350–450 μm). Note that cooling rates for acapulcoites and lodranites at 900–1100 °C are on the order of ~100 °C/Ma (for details see text).

cooling ($dT/dt \sim 1000$ °C/Ma). As we shall see below, the cooling rate of acapulcoites and lodranites at high temperature is on the order of ~100 °C/Ma, and for this cooling rate Hf–W closure temperatures for acapulcoites and lodranites are 975 ± 50 °C and 1025 ± 50 °C. Note that these estimates are not very sensitive to variations in the cooling rate unless cooling is one order of magnitude slower or faster.

The comparison of the Hf–W ages to other radiometric ages provides a consistency test for the validity of the T_c estimates for the Hf–W system. However, age information for acapulcoites and lodranites is sparse and the samples investigated here have not been studied with other high precision chronometers. Nevertheless, the Hf–W ages for the acapulcoites Dhofar 125 and NWA 2775 may be compared with a Pb–Pb age for Acapulco phosphates (Göpel et al., 1992). It seems reasonable to assume that the Hf–W age of Acapulco does not differ significantly from the Hf–W ages of Dhofar 125 and NWA 2775, because metals from these three acapulcoites have indistinguishable W isotope compositions, which are also indistinguishable from those of the lodranite metals. The Hf–W age for acapulcoites of 4563.1 ± 0.8 Ma is ~6 Ma older than the 4557 ± 2 Ma Pb–Pb age for phosphates from Acapulco. The closure temperature of Pb diffusion in apatite (grain sizes from 100 to 300 μm) range from ~490 °C for slow cooling ($dT/dt \sim 10$ °C/Ma) to ~655 °C for fast cooling ($dT/dt \sim 1000$ °C/Ma) (Cherniak et al., 1991). Hence, T_c for the Pb–Pb system in Acapulco phosphates is always lower than T_c for high-Ca pyroxene in acapulcoites, which is consistent with the observation that the Hf–W age for Dhofar 125 and NWA 2775 is older than the Pb–Pb age for Acapulco phosphates.

4.3. Accretion and cooling history of the acapulcoite–lodranite parent body

The Hf–W ages and T_c estimates for acapulcoites and lodranites in conjunction with thermochronological information from other chronometers and peak temperature estimates based on petrology can be used to constrain the accretion age and cooling history of the acapulcoite–lodranite parent body. The available radiometric ages for acapulcoites and lodranites are summarized in Table 2. The acapulcoites and lodranites that were investigated for their Hf–W systematics in this study have not been dated with other chronometric methods. However, as argued above, the Hf–W age of Acapulco is probably not very different from those of the acapulcoites and lodranites investigated here, such that the Hf–W ages for acapulcoites

Dhofar 125 and NWA 2775 in conjunction with Pb–Pb, Ar–Ar and Pu fission ages of Acapulco can be used to constrain the cooling history of Acapulco. The cooling history of other acapulcoites remains less well constrained and must await the acquisition of Pb–Pb ages for phosphates from Dhofar 125 and NWA 2775. Likewise, lodranite NWA 2627 has not been dated with a chronometer other than the Hf–W system but the similarity of available Ar–Ar ages of lodranites and acapulcoites together with their indistinguishable Hf–W ages suggest that the cooling histories of acapulcoites and lodranites were not substantially different.

Temperature profiles for spherical asteroids heated by energy released from ^{26}Al decay (Carslaw and Jaeger, 1959; Miyamoto et al., 1981) were calculated and the parameters used are identical to those in Kleine et al. (2008). In addition to ^{26}Al , decay of ^{60}Fe (half-life ~1.5 Ma) may have been an important heat source for early-formed planetary objects. However, the initial abundance of ^{60}Fe and its distribution throughout the solar system is poorly constrained. In-situ Ni isotope analyses of sulfides in ordinary chondrites yield estimates of the initial $^{60}\text{Fe}/^{56}\text{Fe}$ at the time of chondrule formation ranging from ~1 to $\sim 9 \times 10^{-7}$ (Tachibana and Huss, 2003; Mostefaoui et al., 2005). Likewise, in-situ Ni isotope analyses of chondrules from ordinary chondrites yield initial $^{60}\text{Fe}/^{56}\text{Fe}$ ratios at the time of chondrule formation of $\sim 5\text{--}10 \times 10^{-7}$ (Tachibana et al., 2006). In contrast, Ni isotope analyses of bulk carbonaceous chondrites and iron meteorites by MC-ICPMS do not show clear evidence for the presence of ^{60}Fe and seem to require relatively low $^{60}\text{Fe}/^{56}\text{Fe}$ ratios at the beginning of the solar system of $< 6 \times 10^{-7}$ (Dauphas et al., 2008) and $< 1 \times 10^{-7}$ (Regelous et al., 2008). The currently available data, therefore, suggest that ^{60}Fe may have been present in appreciable amounts in some meteorite parent bodies, but almost absent in others. Consequently, the initial ^{60}Fe abundance in the acapulcoite–lodranite parent body, for which no Ni isotope data exist, is poorly constrained. Hevey and Sanders (2006) noted that the presence of ^{60}Fe at a level anywhere in the range of current estimates does not change the results of thermal models based on ^{26}Al decay as the sole heat source significantly. However, the thermal evolution of differentiated objects may have been different, once most of the Fe became concentrated in the metal core (Hevey and Sanders, 2006; Sahijpal et al., 2007). Investigating whether such processes were important in the acapulcoite–lodranite parent body must await a determination of its initial ^{60}Fe abundance.

In our model, we assume instantaneous accretion and do neither include the insulating effects of a regolith (Akridge et al., 1998) nor temporal and local variations in physical and thermal parameters (e.g., changes in thermal conductivity due to a decrease in porosity) (Bennett and McSween, 1996). For instance, if accretion had taken place over a timescale similar to the ^{26}Al half-life, then a body would have started

Table 2

Compilation of radiometric ages and closure temperatures for acapulcoites and lodranites.

Sample	System	Minerals	Age (Ma) $\pm 2\sigma$	T_c (°C)	References
<i>Acapulcoites</i>					
NWA 2775 + Dhofar 125	$^{182}\text{Hf}\text{--}^{182}\text{W}$	Metal-silicate	4563.1 ± 0.8	975 ± 50	This study
Acapulco	$^{207}\text{Pb}\text{--}^{206}\text{Pb}$	Phosphates	4557 ± 2	550 ± 100	(1)
Acikapulco	$^{39}\text{Ar}\text{--}^{40}\text{Ar}$	Whole rock; plagioclase	4514 ± 16	277 ± 50	(2)
Acapulco	^{244}Pu fission	Whitlockite	4410 ± 16	117 ± 50	(2)
Acapulco	U–Th/He	Apatite	4538 ± 32	120 ± 50	(3)
<i>Lodranites</i>					
NWA 2627	$^{182}\text{Hf}\text{--}^{182}\text{W}$	Metal-silicate	4562.6 ± 0.9	1025 ± 50	This study
Gibson	$^{39}\text{Ar}\text{--}^{40}\text{Ar}$	Whole rock; plagioclase	4490 ± 40	277 ± 50	(4)

(1) Göpel et al. (1992); (2) Pellas et al. (1997); (3) Min et al. (2003); (4) McCoy et al. (1997a,b).

retaining the heat produced by ^{26}Al decay before reaching its terminal mass and peak temperatures would have been reached earlier than estimated when assuming instantaneous accretion (Merk et al., 2002; Ghosh et al., 2003). A thick insulating regolith would have resulted in slower cooling rates and hence higher temperatures in the interior,

compared to the model used here (Akridge et al., 1998). In spite of these simplifications, to a first order the model used here is useful for calculating cooling curves for acapulcoites and lodranites and for obtaining some information on the accretion time, size and internal structure of the acapulcoite–lodranite parent body.

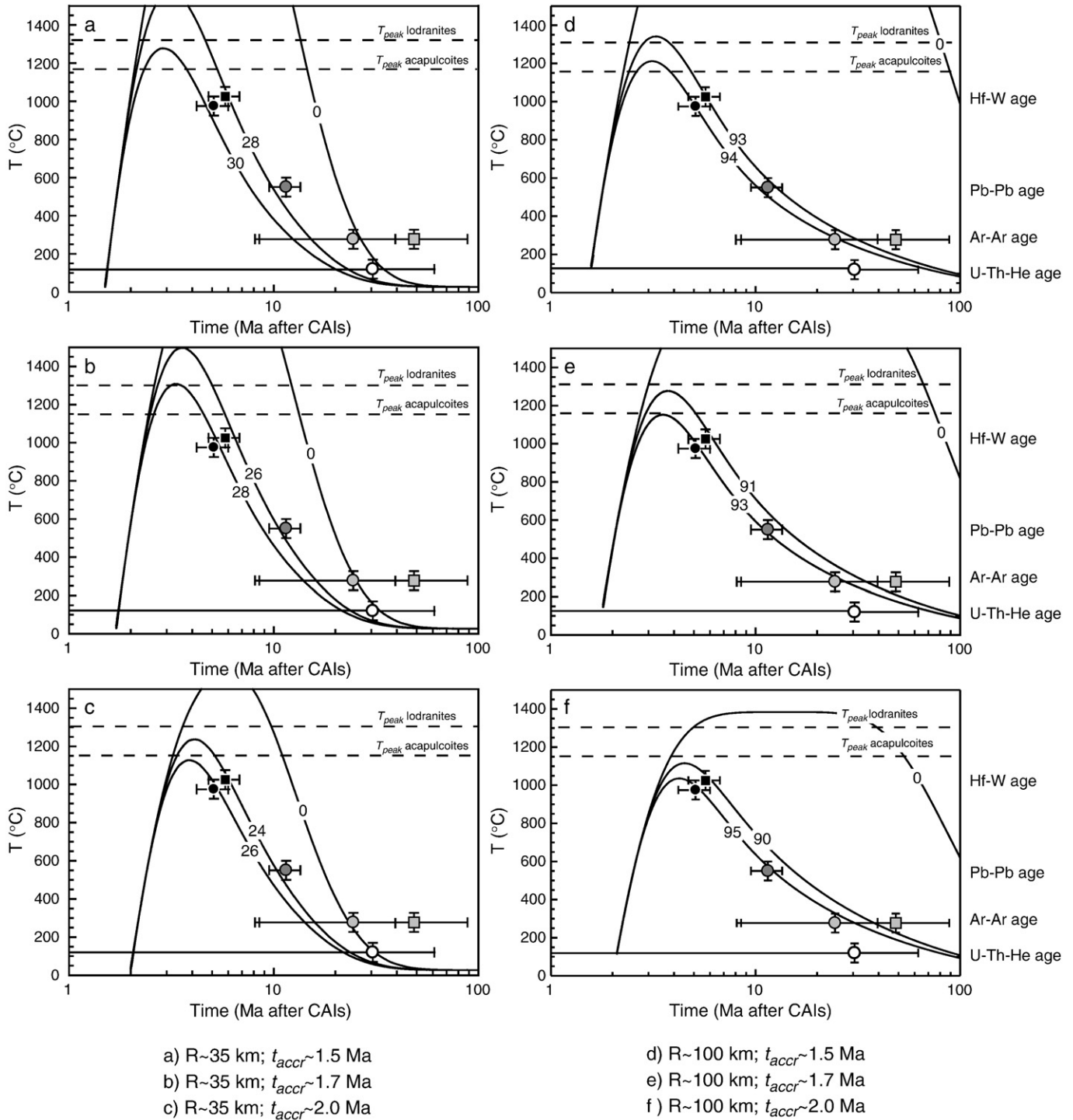


Fig. 5. Cooling curves for acapulcoites and lodranites. Solid lines indicate calculated temperature profiles for different depths in spherical bodies with radii of 35 and 100 km accreted at different times. Numbers indicate distance in km from the center. Parameters are identical to those used in Kleine et al. (2008) and are: thermal conductivity $K = 1.0 \text{ W m}^{-1} \text{ K}^{-1}$; thermal diffusivity $\kappa = 5.0 \times 10^{-7} \text{ m}^2 \text{ s}^{-1}$; density $\rho = 3.2 \times 10^3 \text{ kg m}^{-3}$; heat generation $A = 11.67 \times (^{26}\text{Al}/^{27}\text{Al}) \text{ W m}^{-3}$; emissivity $h = 1.0 \text{ m}^{-1}$. The assumed ambient temperature is $T_0 = 300 \text{ K}$ and the initial $^{26}\text{Al}/^{27}\text{Al}$ ratios are 1.38×10^{-5} , 1.14×10^{-5} , and 0.85×10^{-5} , corresponding to accretion at 1.5, 1.7, and 2 Ma after CAI formation. Hf–W ages and closure temperatures are from this study, all other ages are from the literature and are summarized in Table 2. Black = Hf–W ages; dark grey = Pb–Pb ages; light grey = Ar–Ar ages; white = U–Th–He ages. Circles represent acapulcoites, squares are for lodranites. Ar–Ar ages are shifted by ~30 Ma due to the proposed revision in the ^{40}K decay constant (Renne, 2000; Trierloff et al., 2001).

In Fig. 5, calculated temperature profiles for different depths in spherical asteroids with radii of 35 and 100 km are shown for accretion ages of 1.5, 1.7, and 2 Ma after CAI formation. In these models the acapulcoite–lodranite parent body must have had a radius of more than ~35 km because for smaller radii, cooling is faster than indicated by the differences in Hf–W and Pb–Pb closure temperatures and ages for acapulcoites. In the thermal model used here, the best fit to the Hf–W and Pb–Pb constraints is obtained for the uppermost ~10 km of a body with radius of ~35 to ~100 km. However, if the acapulcoite–lodranite parent body had a thick insulating regolith, then cooling at a given depth would have been slower, such that the thermochronological data might also be consistent with a parent body radius of less than ~35 km.

The thermal models also reveal that for accretion at ~2 Ma the peak temperatures of the lodranites of ~1300 °C are only reached near the center of the parent body but these domains would have cooled below ~1025 °C (i.e., T_c of the Hf–W system in lodranites) later than ~10 Ma after CAI formation, which is inconsistent with the ~5.7 Ma Hf–W age for lodranites. This conclusion holds true even if the acapulcoite–lodranite parent body had a thick regolith because in this case cooling at a given depth would have been even slower (Akridge et al., 1998). Conversely, for accretion as early as ~1.5 Ma, peak temperatures for lodranites and acapulcoites would have been ~1400 °C and ~1200 °C, respectively, which is somewhat too high. These observations suggest that the acapulcoite–lodranite parent body accreted later than ~1.5 Ma but before ~2 Ma.

The thermochronological data are thus consistent with a parent body of ~35–100 km radius that accreted at 1.7 ± 0.3 Ma after CAI formation (Fig. 5). In this model, acapulcoites formed ~7 km below the surface and lodranites originate from a layer that is ~2 km deeper. Peak temperatures in the acapulcoite and lodranite regions were reached at ~3–4 Ma after CAI formation and were ~1150 °C and ~1280 °C, respectively, consistent with petrological constraints. If this model is correct, then large parts of the acapulcoite–lodranite parent body (i.e., the entire body except the uppermost ~10 km) must be differentiated into a metal core and silicate mantle because this domain must have maintained temperatures in excess of ~1300 °C for several millions of years. If any of these materials were delivered to Earth, it has yet not been identified in the meteorite collections.

Cooling rates for acapulcoites and lodranites can be obtained from the slope of the cooling curves and for acapulcoites cooling rates decrease from ~120 °C/Ma at ~1000 °C (i.e., just below their peak temperature) to ~50 °C/Ma at ~600 °C. Over the same temperature range the cooling rate of lodranites decrease from ~110 °C/Ma to ~40 °C/Ma (Fig. 6). This thermal history is different from those of H5 and H6 chondrites, which cooled at a much slower rate (Fig. 6). This reflects a deeper burial depth of H5 and H6 chondrites compared to acapulcoites and lodranites.

The thermal history at lower temperatures is less well constrained. Fig. 5 reveals that the thermal model presented here is consistent with Ar–Ar ages for acapulcoites and lodranites (McCoy et al., 1996; Mittlefehldt et al., 1996; McCoy et al., 1997a; Pellas et al., 1997; Renne, 2000) as well as U–Th–He ages for Acapulco phosphates (Min et al., 2003). However, the uncertainties on the Ar–Ar and U–Th–He ages are large and would be equally consistent with relatively slow cooling of <20 °C/Ma for temperatures below ~400 °C (as would be the case for the thermal model shown in Fig. 6) as well as with almost immediate cooling from ~550 °C to ambient temperature. Evidence for very slow cooling at low temperatures comes from the determination of ^{244}Pu fission track densities in orthopyroxene, whitlockite and apatite, which indicate a cooling rate of 1.7 °C/Ma in the temperature range of 280–90 °C (Pellas et al., 1997). This is somewhat lower than the cooling rates obtained from the model presented here, which decrease from ~7 °C/Ma at ~280 °C to ~1 °C/Ma at ~100 °C. Based on the (U–Th)/He age of 4538 ± 32 Ma, Min et al. (2003) argued that cooling below ~120 °C must have been rapid, inconsistent with the slow cooling rates

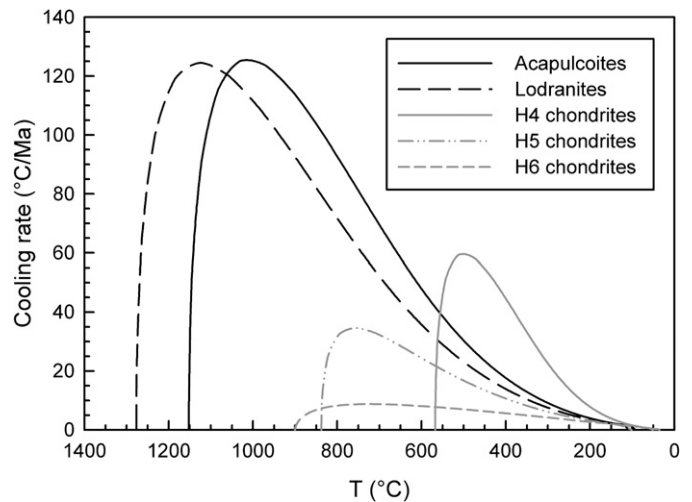


Fig. 6. Cooling rates for acapulcoites and lodranites as calculated from the slope of the cooling curves shown in Fig. 5. Cooling rates for H chondrites are shown for comparison and are from Kleine et al. (2008).

obtained from the ^{244}Pu fission track data. These authors suggest that track annealing by later thermal disturbance resulted in spuriously young ^{244}Pu fission track ages. Metallographic cooling rates for acapulcoites and lodranites (McCoy et al., 1996, 1997a,b) and estimates based on Ca zonation in Acapulco olivines (Zipfel et al., 1995) indeed suggest relatively fast cooling of >1000 °C/Ma for temperatures below ~600 °C. The shift from initially moderate cooling at ~50–100 °C/Ma (from peak temperatures down to ~600 °C) to rapid cooling at >1000 °C/Ma (at temperatures below ~600 °C) might be related to excavation of acapulcoites and lodranites, possibly caused by impacts on their parent body (McCoy et al., 1996, 1997a,b). Alternatively, it is also possible that one of the chronometric methods for determining cooling rates is not entirely reliable.

In the thermal models discussed above it is assumed that ^{26}Al decay is the sole heat source for melting and metamorphism of acapulcoites and lodranites. However, it has recently been suggested that impact heating could have been the dominant heat source (Rubin, 2007). While impact heating of asteroids is difficult to exclude as a heat source, the data presented here suggest that energy release by ^{26}Al decay must have been important in the thermal evolution of the acapulcoite–lodranite parent body. This is because the Hf–W age of 5.2 ± 0.9 Ma for acapulcoites refers to the time of cooling below ~975 °C, such that for reasonable cooling rates, which are on the order of 100 °C/Ma as constrained by the difference between the Hf–W and Pb–Pb ages, acapulcoites must have reached their thermal peak of ~1150 °C at ~3–4 Ma after CAI formation. Obviously their parent body must have accreted before this time and thermal modeling indicates that in bodies accreted that early ^{26}Al decay was an important heat source.

4.4. Bulk Hf–W systematics of the acapulcoite–lodranite parent body—nebular versus parent body processes

The $^{180}\text{Hf}/^{184}\text{W}$ ratios and $\epsilon^{182}\text{W}$ values of the acapulcoite and lodranite whole-rocks reported in this study show wide variations and are most likely distinct from the composition of carbonaceous chondrites (Table 1). This may reflect real differences in the Hf–W systematics caused by Hf/W fractionation related to melting and segregation of metals in the acapulcoite–lodranite parent body. However, the variations in the measured $^{180}\text{Hf}/^{184}\text{W}$ ratios among the different whole-rocks may also be due to sample heterogeneities. This seems likely given that in acapulcoites and lodranites almost all the W is located in the metal, such that small variations in the metal

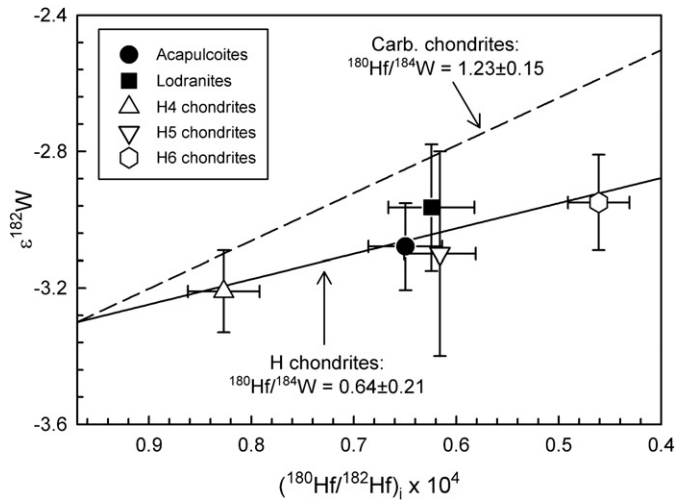


Fig. 7. W isotope evolution diagram for acapulcoites and lodranites. The evolution line for carbonaceous chondrites is defined by their present-day $\epsilon^{182}\text{W}$ of -1.9 ± 0.1 (Kleine et al., 2004) as well as the initial $^{182}\text{Hf}/^{180}\text{Hf} = (9.72 \pm 0.44) \times 10^{-5}$ and initial $\epsilon^{182}\text{W} = -3.28 \pm 0.12$ of Allende CAIs (Burkhardt et al., 2008). These values indicate that carbonaceous chondrites evolved with $^{180}\text{Hf}/^{184}\text{W} = 1.23 \pm 0.15$, which is consistent with $^{180}\text{Hf}/^{184}\text{W} = 1.25 \pm 0.08$ measured for carbonaceous chondrites (Kleine et al., 2004, in press). The analyzed acapulcoites and lodranites depart from the evolution line for carbonaceous chondrites but are consistent with the W isotope evolution of H chondrites. The low $^{180}\text{Hf}/^{184}\text{W}$ of acapulcoites–lodranites has been established before ~ 2 Ma after CAI formation and thus must reflect Hf/W fractionation in the solar nebula.

abundance in the analyzed aliquots will lead to variations in the $^{180}\text{Hf}/^{184}\text{W}$ ratio.

The $^{180}\text{Hf}/^{184}\text{W}$ ratios of bulk acapulcoites and lodranites are more reliably determined by their time-integrated W isotope evolution. In a plot of initial $\epsilon^{182}\text{W}$ vs. initial $^{182}\text{Hf}/^{180}\text{Hf}$, the acapulcoites and lodranites analyzed for this study plot below the evolution line for carbonaceous chondrites, indicating that they evolved with a $^{180}\text{Hf}/^{184}\text{W}$ ratio lower than that of carbonaceous chondrites (Fig. 7). The latest possible time at which this low $^{180}\text{Hf}/^{184}\text{W}$ ratio could have been established is obtained by assuming $^{180}\text{Hf}/^{184}\text{W} \sim 0$ for acapulcoites and lodranites. In this case, acapulcoites and lodranites departed from the carbonaceous chondrite evolution line at ~ 2.5 Ma. However, the acapulcoites–lodranites must have $^{180}\text{Hf}/^{184}\text{W} > 0$, indicating that the Hf–W fractionation event must have taken place well before this time. Given that metal melting in the acapulcoites and lodranites most likely occurred at ~ 3 Ma after CAI formation (see above and Fig. 5), their low $^{180}\text{Hf}/^{184}\text{W}$ ratios cannot reflect Hf–W fractionation during metal segregation in the parent body but rather reflect processes prior to parent body accretion. Remarkably, acapulcoites and lodranites plot on the same W isotope evolution line as H chondrites (Kleine et al., 2008). This is consistent with other compositional similarities between H chondrites and acapulcoites–lodranites (Palme et al., 1981) and suggests that the precursor materials of their parent bodies had similar chemical compositions that were established by fractionation and/or mixing processes in the solar nebula.

The similarity in time-integrated $^{180}\text{Hf}/^{184}\text{W}$ ratios of the acapulcoites and lodranites implies that the FeS–FeNi melts in lodranite NWA 2627 did not migrate, such that the chemical composition of this sample remained unaffected by the partial melting processes that are typical for lodranites (McCoy et al., 1997b). Whether this also applies to other lodranites remains to be determined by future Hf–W studies on a more comprehensive set of lodranites.

5. Conclusions

The Hf–W ages for acapulcoites and lodranites presented here are 5.2 ± 0.9 Ma and 5.7 ± 1.0 Ma after CAI formation and are the most

ancient ages yet reported for this group of meteorites. This confirms the high closure temperature of the Hf–W system in these meteorites, which for both acapulcoites and lodranites is less than ~ 200 °C below their peak temperatures. The Hf–W ages, therefore, provide constraints on the high-temperature thermal history of the acapulcoite–lodranite parent body, information that is not obtainable from other chronometers due to their much lower closure temperatures. The results presented here indicate that the acapulcoite–lodranite parent body accreted later than ~ 1.5 Ma but before ~ 2 Ma after CAI formation and that ^{26}Al was an important heat source in its thermal history. Cooling rates for both acapulcoites and lodranites were ~ 100 – 120 °C/Ma just below their thermal peak and decreased to ~ 40 – 50 °C/Ma at 600 °C. Thus, cooling of acapulcoites and lodranites was roughly one order of magnitude faster than for the most metamorphosed H chondrites (i.e., H6 chondrites most of which cooled at ~ 10 °C/Ma in the temperature interval from ~ 900 to ~ 200 °C). This difference results from a deeper burial depth of H6 chondrites compared to acapulcoites–lodranites. For instance, acapulcoites and lodranites could have been located in the uppermost ~ 10 km of a parent body with a radius of ~ 35 to 100 km that accreted at ~ 1.7 Ma after CAI formation, whereas H6 chondrites were located ~ 40 – 60 km below the surface of an asteroid with a radius of ~ 100 km that accreted ~ 2.5 Ma after CAI formation (Kleine et al., 2008).

Constraining the thermal evolution of asteroids does not only provide information on parent body size and burial depths of a suite of samples but also is key for constraining the parameters that controlled the evolution of meteorite parent bodies. The acapulcoite–lodranite parent body accreted between 1.5 and 2 Ma after CAI formation and, hence, later than the parent bodies of magmatic iron meteorites (Kleine et al., 2005; Schersten et al., 2006; Burkhardt et al., 2008; Qin et al., 2008) but earlier than most (or all) chondrite parent bodies (Kita et al., 2000; Kunihiro et al., 2004; Rudraswami and Goswami, 2007; Kleine et al., 2008). Thus, the accretion age of asteroids is inversely correlated with the peak temperatures reached in their interiors. Differentiated asteroids (i.e., the parent bodies of magmatic iron meteorites) accreted within the first ~ 1 Ma after CAI formation, partially differentiated bodies—such as the acapulcoite–lodranite parent body—formed between 1.5 and 2 Ma, and undifferentiated asteroids (i.e., the chondrite parent bodies) accreted later than ~ 2 Ma after CAIs. This provides further evidence for earlier conclusions that the different thermal histories of meteorite parent bodies primarily reflect variations in their initial ^{26}Al abundance, which, owing to the short ^{26}Al half-life, is determined by their accretion time (Grimm and McSween 1993; Kleine et al., 2005, 2009).

Acknowledgements

We thank A.J. Irving and T.E. Bunch for providing the samples for this study and for several discussions regarding the petrology of acapulcoites and lodranites. T.E. Bunch is also thanked for providing the grain size estimates for high-Ca pyroxene in the NWA samples. C. Göpel is thanked for providing the Acapulco metal separate. We thank L. Qin and M. Trieloff for their reviews and R. Carlson for his editorial efforts, which improved our paper.

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