H-chondrite parent asteroid: A multistage cooling, fragmentation and re-accretion history constrained by thermometric studies, diffusion kinetic modeling and geochronological data

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Abstract

We present a detailed thermometric study and cooling history analysis of selected H-chondrites from the petrologic types 4–6 on the basis of their mineralogical properties, and integrate these data with other available constraints on the cooling rates to develop a comprehensive model for the cooling, fragmentation and re-accretion history of the parent asteroid. Temperatures have been determined on the basis of two-pyroxene (2-Px) and spinel (Spnl)–orthopyroxene (Opx)/olivine (Ol) thermometers using the average of line scans and distributed spot analysis of coexisting pairs in each set. All of these minerals have been found to be compositionally homogeneous from C24 to 2l m from the interface within the resolution of microprobe analysis. The thermometric results for the H5 (Allegan and Richardton) and H6 (Guarena and Kernouve) samples are very similar. Also, while the 2-Px temperature increases by “90 °C from H4 to H5/6, a reverse trend is observed for the Spnl–Opx/Ol temperatures implying compositional resetting of these pairs during cooling. For the H4 sample (Forest Vale) all thermometric results are similar. The cooling rates calculated from numerical modeling of the compositional profiles in Opx–Cpx pairs in H5 and H6, corrected for the spatial averaging or convolution effect in microprobe analysis, are ~25–100 °C/ky, which are 3–4 orders of magnitude higher than the cooling rates implied by in situ cooling in an onion-shell parent body model. Similar numerical simulation of the compositional profile in Opx–Spnl pair in H4 yields a cooling rate ~50 °C/ky, which is in very good agreement with recent metallographic cooling rate of this sample and geochronological constraints on the cooling T–t path. Numerical simulation suggests that the slow cooling of the H5/6 samples at a rate of ~15 °C/My, as deduced by recent metallographic study, could not have commenced at a temperature above ~700 °C since, otherwise, the simulated compositional profile fails to match the observed profile. For the H5 samples, the T–t path constructed on the basis of Hf–W age of peak metamorphism and two stage cooling model satisfies the constraints imposed by the Pb–Pb ages of the phosphates and Ar–Ar ages of the feldspars vs. their respective closure temperatures, whereas that for H6 samples constructed using the same approach fails to satisfy these geochronological data. A second stage of even slower cooling at ~3 °C/My needs to be invoked to satisfy the geochronological age vs. closure temperature relations. We present a model of fragmentation and re-accretion history of the parent body that could lead to the reconstructed T–t paths of the H-chondrite samples studied in this work, and discuss some of their broader implications. The cooling rates retrieved from the available data on the Fe–Mg ordering states of orthopyroxenes of some other H5 and H6 samples are orders of magnitude faster than the metallographic cooling rates that

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are applicable to similar low temperature domain. It is, thus, likely that all samples of the same petrologic type did not share a common cooling history.

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1. INTRODUCTION

Known falls of chondritic meteorites, which are generally accepted to be very primitive and undifferentiated objects of the solar system, are grouped into three chemical classes, carbonaceous, ordinary and enstatite chondrites. On the basis of their iron and metal contents, the ordinary chondrites are further subdivided into three chemical classes, H, L and LL, and petrologic types 3, 4, 5 and 6, depending on the intensity of metamorphism and textural maturity that increase in the above sequence (e.g. Wasson, 1985). It has been argued that since they share common properties, such as oxygen isotope composition, oxidation state and major element chemistry, the samples of a given chemical group were derived from a specific parent asteroid (Wasson, 1985; Trieloff et al., 2003).

The structure and cooling history of H-chondrites have been subjects of considerable debate for many years (Taylor et al., 1987; Lipschutz et al., 1989; Ganguly and Tirone, 2001; Trieloff et al., 2003; Kessel et al., 2007). Recently, however, the so-called “onion shell” model, in which the progressively higher petrologic types were supposed to have been buried at increasing depths (Fig. 1), seems to have gained greater acceptance in the planetary science community, especially after two detailed thermochronometric studies (Trieloff et al., 2003; Kleine et al., 2008). These studies showed that the closure temperature ($T_c$) vs. age data of the metamorphic types 4–6 by several geochronological methods fall on the cooling paths calculated for different depths in a parent asteroid of 50–100 km radius that was heated by the decay of $^{26}$Al($^{26}$Al → $^{26}$Mg + $\beta$) and cooled by heat conduction (Fig. 1) (Trieloff et al., 2003; Kleine et al., 2008).

On the basis of mineralogical data, Kessel et al. (2007) recently suggested that the cooling rate of H6 chondrites could be considerably higher than that suggested by the thermochronological data. Their conclusion is based on modeling the extent of resetting of temperature derived from olivine (Ol)–spinel (Spnl) Fe–Mg exchange thermometer of Sack and Ghiorso (1991a,b) relative to earlier data (Bunch and Olsen, 1974) on temperatures obtained from empirical calibration of orthopyroxene (Opx)–clinopyroxene (Cpx) “solvus thermometry”. The high temperature cooling rates could be constrained by modeling diffusion induced compositional zoning near the interface of mineral pairs (olivine–spinel, orthopyroxene–spinel/clinopyroxene) and such compositional zonings are also expected in the higher petrologic types if these have indeed cooled as slowly as deduced in the earlier studies (Trieloff et al., 2003; Kleine et al., 2008). However, this expectation is not compatible with the few compositional profiles shown by Kessel et al. (2007) that show almost uniform mineral compositions as function of distance from the interface of olivine–spinel pairs. Furthermore, on the basis of their study on the Fe–Mg ordering states of orthopyroxenes, Folco et al. (1996) concluded that the petrologic types 4–6 of H-chondrites had cooled in similar environments in the low temperature regime as they have similar cation ordering states reflecting (ordering) closure temperatures of ~400–500 °C that have no correlation with the petrologic types. This observation
also seems to be incompatible with the onion-shell parent body model and in situ cooling of the samples that is expected to produce an inverse correlation between $T_c$ and petrologic type. We, thus, undertook a detailed study to measure compositional properties of coexisting mineral phases in several H4–H6 chondrites in order to constrain their peak metamorphic temperatures and the high temperature cooling rates. In addition, we also model the available data on the Fe–Mg ordering states of orthopyroxenes (Folco et al., 1996) to constrain the relatively low temperature cooling rates of the different petrologic types. Finally we show that a multistage cooling history model for the H-chondrite parent body (or bodies) is required to reconcile the diverse set of petrological–kinetic constraints, as deduced in this study, and the available geochronological data.

2. COMPOSITIONAL PROPERTIES OF MINERALS AND THERMOMETRY

We studied in detail five H-chondrite samples, namely Guarena (H6), Kernouve (H6), Allegan (H5), Richardton (H5), and Forest Vale (H4). These samples were also used for thermochronological studies (Trieloff et al., 2003; Kleine et al., 2008). The coexisting mineral pairs selected for thermometric studies and compositional profile measurements are orthopyroxene–clinopyroxene (Opx–Cpx), orthopyroxene–spinel (Opx–Spnl) and olivine–spinel (Ol–Spnl). We have analyzed a number of these coexisting pairs as well as isolated grains. The conditions of the microprobe analyses are as follows: 20 kv accelerating voltage, 20 nA beam current with a focussed beam of $\sim$1 $\mu$m diameter, 20 s counts on the peak and 20 s on the background. Most of the microprobe analyses were carried out in two Cameca SX-50 models. (All microprobe data could be made available upon request to the senior author.)

Regardless of the meteorite sample, grain size or coexisting phase, every grain of olivine, clinopyroxene, orthopyroxene and spinel that we have analyzed was found to be almost homogeneous in composition within the resolution of microprobe analysis. Several representative compositional profiles from the three petrologic types, H4, H5 and H6, are illustrated in Fig. 2, with the data for Mg# in enlarged scales. The spinel grains in all samples were found to be very chromium rich ($X(Cr)\sim0.80–0.85$). Clinopyroxenes (Cpx) grains are very small (a few tens of microns or less) and as a consequence we have not been able to detect them optically. However, we succeeded in detecting these grains on the basis of Ca X-ray maps in an electron microprobe.

Detailed examination of BSE images of a thin section of Richardton in Cameca-SX100 suggests a reaction relation between orthopyroxene and a groundmas of alkali rich plagioclase composition to form clinopyroxene. The SX-100...

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**Fig. 2.** Representative compositional profiles of orthopyroxene (Opx)–clinopyroxene (Cpx) and orthopyroxene–spinel (Spnl) pairs approximately normal to the respective interfaces. Mg#: Mg/(Mg + Fe), shown in enlarged scales. The points indicate spot analyses. (a) Guarena (H6), Opx–Cpx; (b) Kernouve (H6), Opx–Cpx; (c) Allegan (H5) Opx–Cpx and Opx–Spnl with Mg# and Fe# (1 – Mg#) of Opx–Spnl pair shown in an enlarged scale in lower right panel; (d) Forest Vale (H4) with Mg# for Opx and Spinel shown in enlarged scales.
has much better counting statistics and resolution than SX-50 and thus yields clearer BSE image. A BSE image of the textural relation is shown in Fig. 3. Clinopyroxene grains are present as thin laths within the groundmass and also as rinds on orthopyroxene cores. The groundmass composition is variable and is roughly Ab$_{50-85}$Kspar$_{2-40}$An$_{5-20}$. It is also present as small inclusions within some clinopyroxene grains. (Upon reviewing this paper, Prof. Ozawa commented that these textural relations suggest that Cpx crystallized from interstitial melt that cooled at a high rate with later morphological modification during cooling.)

Compositions obtained by averaging the data for line scans in each coexisting pair of Opx–Cpx, Spnl–Opx and Spnl–Ol, but excluding the data within 2 µm from the interfaces, which suffer from convolution effects, and sometimes a few outlier data points, have been used to calculate equilibration temperatures of the respective mineral pairs. In a few cases, additional temperatures were calculated using data from distributed spot analyses of the mineral pairs. The primary thermometric formulations used in this study are:

(a) Anderson et al. (1993) for two-pyroxene (2-Px) thermometry,
(b) Liermann and Ganguly (2003, 2007) for Opx–Spnl Fe–Mg exchange thermometry (Eq. (5)), and
(c) Ballhaus et al. (1991) for Ol–Spnl Fe–Mg exchange thermometry.
Each of these thermometric formulations has been based on a large body of experimental data, and account for the compositional effects arising from non-ideal thermodynamic mixing properties. The Ol–Spnl thermometric formulation of Ballhaus et al. (1991) incorporated the experimental data of O’Neill and Wall (1987). The results of these thermometric calculations are summarized in Table 1 and illustrated in Fig. 4. Temperatures estimated for individual mineral pairs according to a thermometric method are listed below the Table 1. The temperature en-

Table 1
Summary of thermometric and age data for the meteorite samples studied in this work. Hf–W ages are relative to CAI formation age. The parenthetical values represent 1 standard deviation (1σ). All temperature calculations are based on average mineral compositions from line scans (e.g. Fig. 2) and distributed spot analyses of coexisting minerals. Estimated temperatures for each mineral pair are listed below the table. The average temperatures are listed in the table. The Ar–Ar ages could be older be 30 My because of a possible change of the decay constant of K.

<table>
<thead>
<tr>
<th>Meteorite</th>
<th>Type</th>
<th>T, °C</th>
<th>Ages</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>2-Px</td>
<td>Spnl–Opx</td>
</tr>
<tr>
<td>Forest Vale</td>
<td>H4</td>
<td>757 (25)</td>
<td>746 (12)</td>
</tr>
<tr>
<td>Forest Vale</td>
<td>H4</td>
<td>757 (25)</td>
<td>746 (12)</td>
</tr>
<tr>
<td>Allegan</td>
<td>H5</td>
<td>817 (8)</td>
<td>643</td>
</tr>
<tr>
<td>Richardton</td>
<td>H5</td>
<td>877 (7)</td>
<td>602</td>
</tr>
<tr>
<td>Guarena</td>
<td>H6</td>
<td>848 (5)</td>
<td>653</td>
</tr>
<tr>
<td>Kernouvé</td>
<td>H6</td>
<td>848 (6)</td>
<td>653</td>
</tr>
</tbody>
</table>

2-Px: Forest Vale: 764(29), 738(50); Allegan: 808(17), 804(19), 819(34), 818(24), 827(12); Richardton: 925(36), 884(11), 941(36), 883(17), 893(21), 843(39), 869(26), 877(48), 847(27), 852(35), 863(42); Guarena: 853(56), 873(38), 853(18), 820(14), 847(6), 858(9); Kernouvé: 821(19), 861(10), 845(8).


differences in the thermometric formulations and possibly also quality of microprobe data. For example, Bunch and Olsen (1974) determined the temperatures of Guarena (H6) and Kernouve (H6) as 780 and 920 °C, respectively. But their analytical data yield 842 (±26) °C for Guarena and 823 (±22) °C for Kernouve when used in the program of Anderson et al. (1993). The latter set of temperatures is in good agreement with the results of this study (Table 1).

Some important results of thermometric calculations are discussed below.

(a) The peak metamorphic temperatures determined for the petrologic types 5 (Richardton and Allegan) and 6 (Guarena and Kernouve) by 2-Px thermometry are higher than the peak temperature of type 4 (Forest Vale), while, surprisingly, the reverse trend is exhibited by Spnl-Opx/Ol Fe–Mg exchange thermometric results.

(b) The 2-Px and Spnl–Opx/Ol temperatures of H5 (Richardton and Allegan) are comparable to those of H6 (Guarena and Kernouve), with the 2-Px temperature of Allegan ~30 °C lower and of Richardton ~30 °C higher than that of the H6 samples. Thus, there is no thermometric reason for the classification of these H-chondrites into two petrologic types.

(c) The Spnl–Opx temperatures are somewhat higher than those of Spnl–Ol temperatures for all samples, with an average discrepancy of ~50 °C (range 38–67 °C). Similar discrepancy between the two thermometers was also noted by Liermann and Ganguly (2003, 2007) for terrestrial ultramafic samples at temperatures above 900 °C.

As discussed above, Kessel et al. (2007) determined the temperatures of a number of H-chondrites using the Ol–Spnl thermometric formulation of Sack and Ghiors (1991a,b). These temperatures are much higher than the Ol–Spnl and Opx–Spnl temperatures derived in this study for the same petrologic types. Thus, we calculated the temperatures for three samples from the three petrologic types (H4, H5, H6) using the Sack & Ghiors formulation and find that the temperatures are similar to those calculated by Kessel et al. (2007). These temperatures are also illustrated in Fig. 4, which show the Ol–Spnl temperatures to be still much lower than Opx–Cpx temperatures, except for the petrologic type H4 (Forest Vale) for which all temperature estimates are similar. However, because of the problem with the Ol–Spnl thermometric calibration, we did not make any use of the temperatures derived from this mineral pair in this study.

3. NUMERICAL SIMULATION OF THE COMPOSITIONAL PROFILES IN COEXISTING MINERAL PAIRS AND RETRIEVAL OF COOLING RATES

3.1. Method and data

We have used the thermometric data summarized in Table 1 and the compositional profiles of Opx–Cpx pairs...
in H5 and H6 and of Opx–Spnl pair in H4 to constrain the cooling rates of the H-chondrite samples. The choice of Opx–Spnl pair for H4 is dictated by the fact that diffusion in spinel is much faster than that in clinopyroxene so that a cooling rate that preserves the peak temperature deduced by Opx–Spnl thermometry must also do so for Opx–Cpx pair with similar record of peak temperature. Additionally, as discussed later, a satisfactory formulation of the temperature dependence of the Fe–Mg distribution coefficient between Opx and Cpx is still lacking for the type of Cpx composition encountered in the H4 sample.

The observed homogeneity of the compositions across the interface of the coexisting mineral pairs does not necessarily imply that the mineral compositions are completely homogeneous. If the element fractionation between the mineral pairs causes their Mg/(Mg + Fe) ratios, henceforth referred to as Mg#, to recede from one another during cooling, then the spatial averaging or convolutive effect stemming from the size of the excitation volume of the sample in microprobe analysis could damp the zoning profiles very close to the interface. We present below numerical simulations for the development of compositional profiles in Opx–Cpx pairs of Guarena (H6) and Allegan (H5), and Spnl–Opx pair in Forest Vale (H4), with and without the convolution effects. The convolution effect has been treated according to the mathematical theory developed by Ganguly et al. (1988).

To determine the evolution of compositional profile during cooling, we solve the 1D diffusion equation for a binary system in a Cartesian coordinate,

$$\frac{\partial C_i}{\partial t} = \frac{\partial}{\partial x} \left( D(Fe - Mg) \frac{\partial C_i}{\partial x} \right)$$

where $C_i$ is the concentration of a diffusing component ($i$ and $D(Fe - Mg)$ is the binary interdiffusion coefficient for Fe and Mg. We assumed homogeneous initial concentrations of Fe and Mg in both minerals ($x$ and $\beta$) in a coexisting pair, as determined by microprobe analyses of the samples. The compositions of the two minerals at the interface were assumed to satisfy the distribution coefficient for the Fe–Mg exchange reaction, which is given by

$$K_D(Fe - Mg) = \frac{\text{Fe/Mg}{x}}{(\text{Fe/Mg}){\beta}} = f(T)$$

An additional boundary condition is given by the requirement that the flux of a component out of one phase must equal that into the other phase:

$$D(Fe - Mg) \frac{\partial C_i}{\partial x} = D(Mg - Fe) \frac{\partial C_i}{\partial x}$$

There is no analytical solution of the diffusion equation (Eq. (1)) for the above initial and boundary conditions. We, therefore, developed a finite difference scheme according to the widely used Crank–Nicolson implicit method (Crank, 1975) that is known to be unconditionally stable. The temperature dependence of $K_D(Fe - Mg)$ for Opx–Cpx pair is derived in this study by combining the $K_D$ expressions for Grt–Opx and Grt–Cpx pairs (Ganguly et al., 1996), and is as follows.

$$\ln K_D^{\text{Opx} - \text{Cpx}}(Fe - Mg) = \frac{2119}{T(K)} - 1.43$$

This equation has been tested by comparing the temperatures, which results from its application to several samples, with those derived from solvus thermometry for Opx–Cpx (Lindsay, 1983; Anderson et al., 1993). Using the same analytical data sets for Guarena (H6), Kernouve (H6), Richardton (H5) and Allegan (H5), the mean discrepancy between the two sets of temperatures is 37°C (range: 20–50°C). For Forest Vale (H4), the $K_D$ thermometry yields absurdly high temperature that is ~250°C above the 2-Px solvus temperature. The reason for this discrepancy for the Forest Vale sample is not clear, but at least part of it is most likely due to the relatively lower Ca content of Cpx in this sample ($x(CaSiO_3) = 0.39$) as compared to the $x(CaSiO_3) = 0.46-0.47$ for the others. The effect of lowering the Ca content of Cpx is to increase its Fe + Mg contents, with perhaps a somewhat greater increase of Fe than Mg since Ca tends to substitute preferentially for Fe. It can be shown that the net effect of lowering the Ca content is to lower $K_D(Fe - Mg)$ in an Opx–Cpx pair and a corresponding increase in the estimated value of temperature from Eq. (4). For the Opx–Spnl pair, the temperature dependence of $K_D$ has been taken from Liermann and Ganguly (2003), and is as follows.

$$\ln K_D^{\text{Spnl} - \text{Opx} (\beta)}(Fe - Mg) = 1174 - 1863(x_{\text{Al}}^{\text{Spnl}}) + 2309(y_{\text{Spnl}}^{\text{Cpx}}) + 3037(y_{\text{Spnl}}^{\text{Ti}}) - 0.296$$

where $x_{Al}^{Spnl}$ stands for the mole fraction of Al in the octahedral site of orthopyroxene, and $y_{\text{Spnl}}^{Cpx}$ and $y_{\text{Spnl}}^{Ti}$ stand for the mole fraction of Cr and Ti, respectively, in the octahedral site of spinel. The compositional data for Opx do not suggest any significant octahedral Al, and thus the correction for the effect of $x_{Al}^{Spnl}$ has been neglected. In the numerical simulations of Opx–Spnl pairs, we have treated the $y_{\text{Spnl}}^{Cpx}$ and $y_{\text{Spnl}}^{Ti}$ terms to be fixed by their initial values. In other words, only Fe and Mg were allowed to exchange between Opx and Spnl as the system cooled.

The diffusion coefficients for orthopyroxene, diopside and spinel have been obtained from Ganguly and Tazzoli (1994), Zhang et al. (2010) and Liermann and Ganguly (2002), respectively. The appropriate diffusion data to use for our calculations are those for Fe–Mg interdiffusion. However, according to the results of Liermann and Ganguly (2002), the Fe and Mg self-diffusion in spinel are quite similar, with $D(Fe)$ being slightly greater than $D(Mg)$. Thus, instead of carrying out detailed calculations of $D(Fe-Mg)$ from the self-diffusion data, we approximated it by $D(Mg)$ since the value of an interdiffusion coefficient is biased towards the diffusion coefficient of the dilute component ($X_{Mg}(spnl) \sim 0.2$; see Fig. 2), as discussed, for example, by Ganguly (2002). We increased the $D(Mg)$ of Liermann and Ganguly (2002) by a factor of 10 to account for the effect of Cr substitution, according to the results of Suzuki et al. (2008). There are no $D(Fe-Mg)$ data for diopside, nor any $D(Fe)$ data that can be combined with $D(Mg)$ to calculate $D(Fe-Mg)$. Thus, we approximated $D(Fe-Mg)$ by $D(Mg)$ for diopside. However, based on the difference between $D(Fe-Mg)$ and $D(Mg)$ in other minerals (Ganguly et al., 1998; Dohmen et al., 2007), this approximation of
$D(\text{Fe-Mg})$ by $D(\text{Mg})$ in diopside is not expected to have an effect on the calculation of cooling rates that could lead to significant modification of our general conclusions.

3.2. Results

Fig. 5 shows the results of selected numerical simulations for the development of (quenched) compositional profiles in Opx–Cpx pairs in Guarena (H6) and Allegan (H5), and in a Spnl–Opx pair in Forest Vale (H4). For each pair, we have illustrated the quenched convolved profiles corresponding to two different cooling rates, one showing a very good match and the other showing a slight but measurable misfit to the measured profiles. In each case, the (black) dashed line shows the simulated compositional profiles, whereas the solid (red) line shows the corresponding convolved profile. The numerical simulations for the following cooling rates show very good agreement with the measured profiles. Guarena (H6): $\sim 100 \text{ °C/ky}$; Allegan (H5) $\sim 15 \text{ °C/ky}$; Forest Vale (H4): $50 \text{ °C/ky}$. The cooling rate for Kernouve (H6) is similar to that of Guarena. We have not specifically modeled the flat concentration profile.

Fig. 5. Numerical simulations of the development of compositional zoning in coexisting mineral pairs in three petrologic types for prescribed cooling rates. In each panel, the dashed black line shows the calculated compositional profile for the slower cooling rate whereas the solid black line shows the corresponding convolved profile; the solid red line shows the convolved profile for the faster cooling rate, matching the analyzed data shown by open red circles. (a) Opx–Cpx pair in Guarena (H6) (b) Opx–Cpx pair in Allegan (H5) and (c) Opx–Spnl pair in Forest Vale (H4), with the simulation in the spinel side shown in an enlarged scale. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)
of an Opx–Cpx pair in Richardton, but from simple scaling argument (t changes roughly as 1/D), the cooling rate for Richardton should be ~175 °C/ky because of its higher peak temperature (877 °C) relative to Guarena (~850 °C). These cooling rates are 3–4 orders of magnitude faster than those predicted for in situ cooling of the samples in an onion shell parent body (Fig. 1).

The simulated concentration profile (Mg#) in Opx–Cpx pair in Guarena for 20 °C/My cooling rate, conforming to the in situ cooling in an onion shell parent body of 100 km radius (Fig. 1), is compared with the measured data in Fig. 6. The large difference between the simulated and measured concentration profiles strongly argues against the scenario of in situ cooling of Guarena in an onion shell parent body.

3.3. The Problem of Resetting of Opx–Spnl Temperatures

We have explored if the observed resetting of Opx–Spnl Fe–Mg exchange thermometer could be achieved by readjustment of composition of the two minerals during cooling. For this purpose, we have recalculated Mg# values for Opx–Spnl pair of Allegan (H5) corresponding to an initial temperature of 820 °C, and subjected it to a two stage cooling process, as deduced above, that satisfy the Opx–Cpx compositional profiles and the lower temperature metallographic cooling data. The imposed cooling rates are 25 °C/ky from 820 °C to 700 °C and 15 °C/My at lower temperature. The result of the simulation is shown in Fig. 7. The final spinel composition is found to match the observed composition, but there is a slight but measurable mismatch between the observed and simulated compositions of orthopyroxene. The complete resolution of the problem of resetting of Opx–Spnl thermometer has to await further analysis, but it is not critical to the main objectives of this paper.

4. ADDITIONAL CONSTRAINTS ON COOLING RATES

4.1. Metallographic data

Krot et al. (2012) have recently determined the cooling rates from metallographic data for a number of H3–H6 chondrites that include the H-chondrites investigated in this work. The inferred cooling rates, which are strictly applicable within a limited temperature domain around 550 °C, are as follows. Forest Vale (H4): 104 °C/My; Allegan (H5): 15 °C/My; Guarena (H6): 15 °C/My; Kernouve (H6): 10 °C/My. The metallographic cooling rate for Forest Vale is found to be in very good agreement with the cooling rate (~5 × 104 °C/My) calculated above from modeling flat concentration profile of Mg in Opx–Spnl pair. The apparent discrepancy for the H5 and H6 samples implies that the cooling rates of these samples drastically slowed down by major changes of their physical environments during cooling.
cooling later. The likely scenario for such changes has been discussed later. Here we explore if the simulated compositional profiles at high temperature deduced in this study to conform to the observed profiles could be preserved by a second stage of slow cooling beginning at some temperature above 550 °C. The threshold temperature above which the compositional profile gets disturbed by the second stage of slow cooling determines the upper limit of temperature for the domain of validity of the metallographic cooling rate.

Fig. 8 shows two-stage cooling profiles for Mg# in Opx-Cpx pair in Guarena (H6), starting with a cooling rate of 100 °C/kay at the peak temperature of 850 °C, as deduced above (Fig. 5a), and changing to 15 °C/My at 650 °C or 750 °C. The convolution of the simulated concentration profile matches the measured profile if the transition to slow cooling takes place at 650 °C, whereas it deviates slightly, but measurably, from the observed profile near the interface if the transition temperature is set to 750 °C. We, thus, conclude that the transition to slow cooling for Guarena (H6) and Allegan (H5), which have been deduced from metallographic studies, had taken place at a temperature no higher than ~700 °C.

4.2. Fe–Mg ordering data in orthopyroxene

In orthopyroxenes, Fe2+ (henceforth referred to as Fe) and Mg fractionate between two nonequivalent crystallographic sites, M2 and M1, with a preference for the former by Fe and for the latter by Mg (Ghose, 1962). The fractionation is temperature sensitive, and can be described by an intracrystalline exchange reaction

\[ \text{Fe}(M1) + \text{Mg}(M2) \leftrightarrow \text{Fe}(M2) + \text{Mg}(M1), \]

(a)

which is driven progressively to the right with decreasing temperature. Building upon the chemical rate theory of order–disorder by Mueller (1967), Ganguly (1982) developed a method of retrieving cooling rates of host rocks from the quenched Fe–Mg ordering states of orthopyroxenes. The method has subsequently been applied to a number of terrestrial and meteorite samples (e.g. Ganguly et al., 1994; Molin et al., 1994; Ganguly and Domenechetti, 1996) and also verified against independent geochronological constraints (Stimpfl et al., 2005).

Folco et al. (1996) determined the quenched Fe–Mg ordering states of orthopyroxenes of a number of unshocked equilibrated H-chondrites by single crystal structure refinement. They calculated the final equilibration or closure temperatures of ordering \( T_{c(ordering)} \) of orthopyroxenes from metamorphic types 4, 5 and 6, but did not find any correlation between this parameter and the metamorphic type, which is in contrast to the inverse correlation that is expected for an onion-shell parent body model (Fig. 1). Folco et al. (1996) thus concluded that the H-chondrite samples investigated by them cooled in similar environments, regardless of the degree of metamorphism.

Using the most thorough calibration of the intracrystalline distribution coefficient, which is defined according to (a) as \( k_D = (\text{Fe}/\text{Mg})^{M2}/(\text{Fe}/\text{Mg})^{M1} \), as a function of temperature (Stimpfl et al., 1999), we have recalcu...
by metallographic method (Krot et al., 2012) since both methods are applicable to limited temperature range around 550°C. We find that while the cooling rates for the H4 samples are similar to those determined by the metallographic method, those for the H6 samples are three orders of magnitude larger than the metallographic rates. The reason for this discrepancy is not clear, but the metallographic cooling rates were determined for H6 samples that are different from those studied by Folco et al. (1996). Thus, one cannot rule out the possibility that even at temperatures around 500°C, some H6 samples cooled at a rate much faster than those studied by the metallographic method. A comparison between the results obtained by the two methods for the same sample would be instructive.

5. THERMOMETRIC AND COOLING RATE CONSTRAINTS VS. GEOCHRONOLOGICAL DATA AND IMPLICATIONS

5.1. Hf–W ages vs. peak temperatures of H5 and H6 chondrites

The Hf–W age of Richardton (H5) is older than that of Kernouve (H6) by 4 ± 1.4 Ma, but the peak metamorphic temperatures of these samples, as calculated by 2-Px thermometer, are similar (Table 1); if there is any real difference, then the T(peak) of Richardton, calculated to be ~875°C, is probably higher than that of Kernouve (~850°C). Kleine et al. (2008) argued that the Hf–W decay system records the peak metamorphic ages of H5/6-chondrites. In that case the above data for T(peak) and Hf–W ages pose an obvious paradox, if the two samples were derived from the same parent body, in that the calculated T(peak) of Richardton requires that it was buried at about the same depth as or somewhat deeper than Kernouve, whereas its older peak metamorphic age requires that it was derived from a shallower level than Kernouve (Fig. 1). If 26Al decay was essentially the sole reason for heating of parent asteroids, then the resolution of this paradox requires that either the two samples were derived from different parent bodies, with that of Richardton probably accreting somewhat earlier than the parent body of Kernouve, or that there was inhomogeneous distribution of 26Al in the single parent body in which Richardton was buried in a region that was shallower than the source region of Kernouve but had higher 26Al concentration. The second model retains the commonly held notion (e.g. Wasson, 1985; Trieloff et al., 2003) that because of their chemical similarities, the samples belonging to the same chemical group (H, L and LL) were derived from the same parent body, but argues against the common practice of calculating the thermal evolution of the parent body assuming homogeneous distribution of the heat producing isotopes.

5.2. Comparison of the thermochronological data with the inferred cooling rates of H-chondrites and a multistage cooling model

We now compare the thermometric data and cooling rate constraints derived above with the available thermochronological data to develop a comprehensive understanding of
the cooling history of the H-chondrites through a large span of temperature and then explore the implications in terms of parent body processes. 

![Diagram](image)

**Fig. 11.** Integrated cooling paths of H4 (Forest Vale), H5 (Allegan and Richardton) and H6 (Guarena and Kernouve) samples, as constrained by the high temperature cooling rates, deduced in this study, metallographic cooling rates and geochronological data, which are summarized in Table 1. The Pb–Pb ages of the two H5 samples, Allegan and Richardton, are indistinguishable in the plot. The closure temperatures of the Pb–Pb ages have been calculated in this study using the inferred cooling rates and observed average grain size ($r = 20 \mu m$). Smaller effective diffusion distance leads to lower $T_c$ ($\sim 25^\circ C$ lower for $r = 10 \mu m$). The spread in the Ar–Ar ages of each sample stems from the uncertainty of the decay constant of $K$.

As evident from Fig. 11, the $T$–$t$ path of Forest Vale (H4) based on a constant cooling rate of 50 $^\circ C$/ky satisfies all data, viz. compositional profiles in Cpx–Opx pairs, metallographic cooling rate at 550 $^\circ C$, Pb–Pb age data for phosphates and Ar–Ar age data for feldspars. For H5, the $T$–$t$ path constructed by imposing the metallographic age, but instead reflects the timing of an earlier high temperature event, i.e. chondrule formation. Thus, for H4 (Forest Vale), we have used the phosphate Pb–Pb age since, as shown below, no significant resetting of the Pb–Pb age could have been possible during the very rapid cooling of this sample, as determined in this study and by metallographic method (Krot et al., 2012). However, for clarity of presentation, we have raised the Pb–Pb age of Forest Vale by 3 My, which is within the 2$\sigma$ error limit (i.e. 4 My) of the age data. For H5 and H6, we have used the Hf–W ages determined for Richardton and Kernouve, respectively, instead of the average ages for these metamorphic types, as we have also calculated the $T$–$t$ paths of these samples.

Using the theoretical formulations of Dodson (1973) and Ganguly and Tirone (1991, 2001), Pb diffusion data in apatite by Cherniak et al. (1991), the observed average grain radius of $\sim 20 \mu m$ (Göpel et al., 1994), and the inferred cooling rates, we have calculated the $T_c$–$t$ for Pb diffusion in phosphates in the H-chondrite samples. For Forest Vale (H4), the $T_c$ is found to be essentially the same as the peak temperature (745 $^\circ C$ vs. $T_\text{p} = 750 ^\circ C$), whereas for H5 and H6, the $T_c$ is $\sim 455 ^\circ C$ (Fig. 11). In these calculations, we have used a spherical geometry that maximizes diffusive flux out of a mineral grain, and as a result the $T_c$ values might have been somewhat underestimated. The calculated $T_c$ for H4 is $270 ^\circ C$ higher than the conventional value whereas for H5/6, it is similar to the latter.

As evident from Fig. 11, the $T$–$t$ path of Forest Vale (H4) based on a constant cooling rate of 50 $^\circ C$/ky satisfies all data, viz. compositional profiles in Cpx–Opx pairs, metallographic cooling rate at 550 $^\circ C$, Pb–Pb age data for phosphates and Ar–Ar age data for feldspars. For H5, the $T$–$t$ path constructed by imposing the metallographic...
cooling rate (15 °C/My) from 700 °C satisfies the age vs. \( T_c \) data for both Pb–Pb and Ar–Ar systems. However, for the H6 samples, the metallographic cooling rate of 15 °C/My commencing at 700 °C does not satisfy the Pb–Pb age data for phosphates and Ar–Ar age data for feldspars. To satisfy these data, we need at least one more stage of slower cooling commencing at some temperature below 550 °C. In Fig. 11, we illustrate a \( T-t \) path using a cooling rate of 3 °C/My that satisfies, within errors, all low temperature age vs. \( T_c \) data. It is also possible that the effective diffusion distance (EDC) of Pb within the phosphates is smaller than their grain sizes. In that case the \( T_c \) for the Pb–Pb ages would be lower than those calculated above. For example, if the EDC is half of the observed grain size (~10 μm), then the \( T_c \)-s for H5 and H6 samples are reduced by ~25 °C. This permits a better match between the age vs. \( T_c \) data and the cooling \( T-t \) path.

6. A FRAGMENTATION-REASSEMBLY MODEL OF THE H-CHONDRITE PARENT BODY

The multistage cooling model for the H-chondrites illustrated in Fig. 11 cannot be explained by simple monotonic cooling at different depths in an onion-shell parent asteroid, with the higher grade rocks buried at progressively greater depths (Fig. 1), but instead requires a complex history of fragmentation and reassembly of the parent body. A possible model of such parent body process is presented below. In accordance with the data on chemical similarities of the H-chondrites, as discussed above, we assume that the samples were derived from a single parent body. However, the complexity of the model is not reduced by invoking different parent bodies.

(a) The initial very rapid cooling rates of H4–H6 samples could be achieved by the excavation of the parent body by an impact, thereby leading to the exposure of the samples to near surface conditions. The much faster cooling of the H4 (Forest Vale) sample compared to the H5/6 samples might have been due to its ejection to the surface or the penetration of the excavation depth to the H4 layer in an onion-shell parent body.

(b) The slow cooling of ~15 °C/My commencing at ~700 °C for the H5/6 samples (Fig. 11) could have been due to accretion/accumulation of material on the excavated rocks at this stage or reassembly of the material into a new parent body following their ejection after one or more additional impacts. In the reassembled parent body, the H6 samples (Guarena and Kernouve) might have been buried deeper, thus permitting slower cooling and greater resetting of their Pb–Pb phosphate age relative to the H5 samples.

(c) The final stage of even slower cooling of a few degrees per My for the H6 samples (Fig. 11) might imply another episode of accretion of material on the H6 samples or burial under a thick regolith blanket that has low thermal conductivity with the regolith material resulting from an impact.

Nakamura et al. (2011) have recently reported the mineralogy and mineral chemistry of the dust particles scooped from the Asteroid Itokawa. They found the mineralogical properties to be indicative of high temperature metamorphism at ~800 °C and classified the dust particles as LL5 and LL6. Since such high temperature requires deep burial in an Asteroidal parent body heated by \( ^{26}\text{Al} \) decay, Nakamura et al. (2011) suggested that the surface material of Itokawa was originally buried deep in a much larger asteroid, which was catastrophically disaggregated by one or more impact events, and that some of the impact pieces were then re-accreted into the greatly diminished rubble-pile asteroid that is now Itokawa. This scenario and the discovery of metamorphic types 5 and 6 on the surface of an asteroid give credence to the fragmentation-reassembly model of the parent body that we have presented above. Additionally, there is no obvious reason as to why the process of re-accretion should be restricted to the materials ejected from a single parent body; it is quite possible that there was mixing of materials derived from different parent bodies to form new asteroids.

The \( T-t \) path presented in Fig. 11 is qualitatively similar to that developed by Ganguly et al. (1994) for the parent body of mesosiderites, which is characterized by very rapid cooling of several degrees per ky from ~1150 °C, followed by very slow cooling at ~1 °C/My at ~725 °C. This similarity may suggestive of some common scenario in the fragmentation and re-accretion/regolith blanketing in the early history of asteroids.

7. SUMMARY AND CONCLUSIONS

The important conclusions that have emerged from the analysis of mineralogical properties and their comparison with metallographic cooling rates and geochronological constraints may be summarized as follows.

(a) The H4 sample, Forest Vale, has distinctly lower peak temperature than the H5 (Richardton and Allegen) and H6 (Guarena and Kernouve) samples, whereas there is no significant difference between the peak temperatures of the last two groups (Fig. 4). (This study should also prompt careful determination of the peak metamorphic temperatures of other H5 and H6 samples.)

(b) While the 2-Px and Spnl–Opx /Ol thermometric temperatures are mutually compatible for the H4 sample, they diverge in opposite directions for the H5 and H6 samples (Fig. 4). The discrepancy between the two types of thermometric results for the H5/6 samples is due to the resetting of spinel compositions during slow cooling below 700 °C, but a completely satisfactory model of the compositional readjustment of the thermometric pairs is still lacking.

(c) For the H4 sample all cooling rate constraints derived from modeling compositional zoning and metallographic data are mutually compatible and the consequent \( T-t \) path also agrees with all geochronological data (Pb–Pb and Ar–Ar ages vs. \( T_c \)-s). The very rapid cooling of ~50 °C/ky inferred for this
sample suggests cooling under near surface environment, effectively through the entire temperature range.

(d) The very rapid cooling rates of 25–100 °C/ky of the H5/6 samples down to ~700 °C that are constrained by modeling compositional profiles of coexisting Opx–Cpx pairs are incompatible with the commonly held notion of in situ slow cooling of the different petrologic types at different depths in an onion-shell parent body.

(e) Assuming that the decay of 26Al was the principal mechanism of heating of an asteroid, the relative peak metamorphic temperatures and Hf–W ages of Richardson (H5) and Kernoué (H6) seem to imply that either the two samples were derived from different parent bodies, with that of Richardson probably accreting somewhat earlier, or that there was inhomogeneous distribution of 26Al within a single parent body of the two samples in which Richardson was buried at a shallower depth than Kernoué (H6), but in a region with higher 26Al concentration.

(f) The T–t paths constrained by thermometric and cooling rate constraints of the H5 (Allgen and Richardson) and H6 (Guarena and Kernoué) chondrites, as determined in this study and from metallographic data (Krot et al., 2012), could be reconciled with the available thermochronological data, except for the whole rock Pb–Pb age data, through a process of multistage cooling that could have been due to the following processes:

(i) Impact induced fragmentation/deep excavation of the parent body shortly after the attainment of the peak temperatures, followed by

(ii) accumulation or accretion of material on the excavated rocks when the latter had cooled to ~700 °C, or impact-ejection at this stage and reassembly into a separate parent body leading to slow cooling, and

(iii) subsequent re-accretion of asteroidal material on, or impact induced regolith blanketing of, the H6 samples.

Our model is not incompatible with onion-shell parent body structure at the initial stage, but it represents a major paradigm shift in the interpretation of the low temperature thermochronological data in terms of in situ cooling in an onion-shell parent body. The general form of the cooling paths for H5 and H6 samples, characterized by very rapid cooling at high temperature (~25–100 °C/ky) followed by much slower cooling commencing at ~700 °C (~15 °C/My) is similar to that deduced for mesosiderites, which are stony-iron meteorites, by Ganguly et al. (1994).

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