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A model for planetesimal meltdown by ^{26}Al and its implications for meteorite parent bodies

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Abstract–The melting of planetesimals heated by ^{26}Al has been modelled using a new finite difference method that incorporates convection. As an example, we consider a planetesimal with a radius of 64 km, which accretes instantaneously at t = 0.75 Myr (after the formation of calcium-aluminum-rich inclusions) from cold (250 K) nebular dust with 50% porosity. At t = 0.9 Myr (T = 700 K), the planetesimal shrinks to a radius of 50 km due to sintering. At t = 1.2 Myr (T = 1425 K), the fully insulated interior, deeper than a few kilometers, starts to melt, and at t = 1.5 Myr (T = 1725 K), with 50% melting, convection starts. By t = 2 Myr, the planetesimal is a globe of molten, convecting slurry inside a thin residual crust. From about t = 2.5 Myr, the crust thickens rapidly as the power of ^{26}Al fades.

Planetesimals probably melt in this manner when they accrete before t = 1.3 Myr and are large enough to insulate themselves (R >20 km for accretion at t = 0, rising to >80 km at t = 1.3 Myr). Melting behavior will also be affected by the level of ^{60}Fe in nebular dust, by the extent of devolatilization reactions and basalt segregation during heating, and by gradual accretion.

The model suggests that a) the parent bodies of differentiated meteorites had accreted before about t = 1.5 to 2 Myr and before most chondritic parent bodies had formed, and b) that molten planetesimals may be a source for chondrule melt droplets.

INTRODUCTION

Half a century ago, Urey (1955) suggested that the short-lived isotope ^{26}Al (half-life 0.73 Myr) was a likely heat source for the metamorphism and melting of meteorite parent bodies (planetesimals). Lee et al. (1976, 1977) later discovered evidence for once-live ^{26}Al in a calcium-aluminum-rich inclusion (CAI) from the Allende meteorite. Furthermore, these authors showed that the ratio of $^{26}Al/^{27}Al$ in the inclusion (about 5×10^{-5}) supported Urey's idea of radioactive heating and melting of planetesimals; if the ratio were generally representative of $^{26}Al/^{27}Al$ in nebular dust, the corresponding decay energy would have caused melting in the insulated interiors of planetesimals made from the dust.

CAIs from different classes of chondritic meteorite are now known to consistently have had initial $^{26}Al/^{27}Al$ equal to 5×10^{-5} (dubbed the canonical value), suggesting that ^{26}Al was indeed distributed uniformly in the nebula (MacPherson et al. 1995; Russell et al. 1996; Huss et al. 2001). Moreover, evidence for ^{26}Al has now been detected in chondrules, plagioclase fragments in chondrites, and basaltic meteorites. The inferred initial $^{26}Al/^{27}Al$ in these last objects is lower than the 5×10^{-5} of CAIs, and it correlates closely with the ages of the objects based on the precise and independent Pb-Pb dating method (e.g., Amelin et al. 2002; Zinner and Göpel 2002; Zinner et al. 2002). The correlation strongly reinforces the view that when CAIs were formed (taken to define the start of the solar system; t = 0), $^{26}Al/^{27}Al$ was equal to 5×10^{-5} throughout the reservoir of dust from which not only CAIs, but also chondrules and meteorite parent bodies were made. It almost certainly means that ^{26}Al was an active heat source in any planetesimal that accreted out of nebular dust within one or two half-lives of the beginning.

While most previous attempts to model the heating of planetesimals by ^{26}Al have dealt with the sub-solidus thermal (and aqueous) metamorphism of chondritic parent bodies (e.g., Miyamoto 1991; Ghosh et al. 1999, 2003; Cohen and Coker 2000; Young 2001; LaTourrette and Wasserburg 1998), a few studies have explored the melting of planetesimals. These include the early work of Fish et al. (1960), Lee et al. (1976), Herndon and Herndon (1977), and Wood (1979). Furthermore, Woolum and Cassen (1999) used the analytical method developed by LaTourrette and Wasserburg (1998) to delimit the accretion time and planetesimal radius necessary for melting, Ghosh and McSween (1998) presented an elegant finite element model for the asteroid Vesta, and Merk et al. (2002) used a numerical model to determine the peak central temperature in planetesimals that accrete gradually.

In this paper, we pursue the investigation of ^{26}Al-induced melting with a new finite-difference model. Finite differencing has great potential for flexible modelling because it allows parameters (like thermal conductivity and density) that must otherwise be assumed constant to vary with time and temperature. In particular, we aim to address the problem of heat transfer after the onset of partial melting. Existing models assume that heat is lost only by conduction, and this leads to impossibly high internal temperatures. Here we propose that convection becomes important once the melt fraction exceeds 50%. Our model also permits variation in a planetesimal's volume. This allows the transition from an initially porous and highly insulating state to one that has no porosity below a residual regolith, and it also allows planetesimals to grow in size gradually rather than in an instant.

ANALYTICAL PROCEDURES

Thermal Modelling: A Preliminary Approach Using an Analytical Method

The aim of our modelling is to determine the temperature, T, within a radioactively heated planetesimal as a function of depth (expressed as radial distance, r, from the center), and as a function of time, t, measured in millions of years (Myr) after the time when CAIs were formed (t = 0). Results are shown graphically, either as T-t plots or as T-r plots. T-t plots display the temperature rise and fall at chosen depths (or chosen fractions of the planetesimal radius, R). T-r plots are temperature profiles from the center to the surface of the planetesimal at chosen times. Planetesimals are assumed to accrete instantaneously.

To determine the temperature as a function of depth and time, the heat equation needs to be solved in radial form. Equation 1 is:

$$\rho C_p \frac{\partial T}{\partial t} = \frac{1}{r^2}\frac{\partial}{\partial r}\left(K r^2 \frac{\partial T}{\partial r}\right) + A_0(r, t) \tag{1}$$

where $T = T(r, t)$ is the temperature as a function of radial distance and time; $K = K(T)$ is thermal conductivity as a function of temperature; $\rho = \rho(T)$ is density as a function of temperature; $C_p = C_p(T)$ is specific heat capacity as a function of temperature; and $A_0(r, t)$ = the heat source term per unit volume per unit time as a function of radial distance and time.

By assuming that ρ, C_p, and K have constant values that are independent of temperature, the heat equation can be expressed in a form that can be solved analytically (Carlslaw and Jäger 1959):

$$T = T_0 + \frac{\kappa A_0}{K\lambda} e^{-\lambda t}\left(\frac{(R \sin r(\lambda/\kappa)^{1/2})}{(r \sin R(\lambda/\kappa)^{1/2})} - 1\right) + \frac{2R^3 A_0}{r\pi^3 K}\sum_{n=1}^{\infty}\frac{(-1)^n}{n(n^2 - \lambda R^2/\kappa\pi^2)}\sin\frac{n\pi r}{R} e^{-\kappa n^2 \pi^2 t/R^2} \tag{2}$$

In Equation 2, κ is the thermal diffusivity ($K/\rho C_p$); t is the elapsed time since the planetesimal formed; A_0 is the heat output per unit volume at the time of planetesimal formation; and λ is the decay constant for ^{26}Al. A_0 is equal to $H_0\rho$ where H_0 is the heat output per unit mass at the time of planetesimal formation. T_0 is the ambient (i.e, starting and surface) temperature.

The values of the parameters used to solve Equation 2 are given in Table 1. LaTourrette and Wasserburg (1998) used Equation 2 in their modelling; the values they chose for ρ, C_p, and K are also used here. For T_0, we use 250 K, being the average of the range of temperatures (100–400 K) estimated by Woolum and Cassen (1999) for the mid-plane of the circum-solar disk at 2.5 AU, and 1 Myr after formation of the sun. For initial H_0 (at t = 0), we use 1.9×10^{-7} W kg^{-1} because we assume that the planetesimal had the composition of CI meteorites, with 0.9% by mass of aluminium (Anders and Grevesse 1989), and we take the decay energy of ^{26}Al as 4 MeV (= 6.4×10^{-13} J) per atom. The corresponding energy stored in the primitive dust (at t = 0) is 6.4 kJ g^{-1}.

Solutions to Equation 2 for nine selected combinations of radius and formation time (R = 5, 20, and 50 km; t = 0, 0.75, and 1.5 Myr) are shown as T-t plots in Fig. 1. Temperature changes are shown for three depths, r/R = 0 (the center), r/R = 0.5 (midway between the center and the surface), and r/R = 0.9 (at a depth corresponding to 10% of the radius). The horizontal dashed line at 1850 K marks, for reference, the approximate temperature for complete melting (the liquidus temperature).

These results suggest that a planetesimal whose radius is 5 km will lose heat almost as quickly as the heat is produced, and so will never become hot enough to melt (Note, however, that this analytical solution does not include the insulating effect of a regolith. For a highly insulating, "fluffy" material Lee et al. [1976] inferred that a radius of 0.6 km would be sufficient to cause melting. In our numerical model below, we too assume initial highly insulating dust, but we also assume sintering and porosity loss at about 700 K). In contrast, a planetesimal with a radius of 50 km will melt and become substantially overheated if it forms at t = 0 or t = 0.75 Myr. For these two cases, the very high peak temperatures shown in Fig. 1 are, of course, not real. They result from the erroneous assumption, mentioned earlier, that heat moves only by conduction, even after melting has occurred. Nevertheless, the peak temperatures do reflect the high level of radioactive energy initially stored in primitive dust. The energy is about four times as much as is necessary to induce melting.

A quick calculation confirms this factor of four. Taking the values of ρ and C_p from Table 1, and taking the latent heat of fusion as 2.56×10^5 J kg^{-1} (see below), an energy input of 1.6 kJ will heat and melt 1 gram of primitive dust, starting from a temperature of 250 K (recall that ^{26}Al at $t = 0$ endows the dust with an energy reserve of 6.4 kJ g^{-1}).

Thermal Modelling: A Numerical Approach

Fortunately, Equation 1 may be solved numerically as well as analytically, thereby raising the prospect of a more realistic model of planetesimal evolution that avoids impossibly high temperatures. To do this, we have employed a finite differencing scheme. Finite differencing involves using a Taylor Series expansion and rewriting the partial differentials as finite changes of the temperature in space and time. The Crank-Nicholson method of calculation (Gerald and Wheatley 1994) has been chosen because it sets no limit to the time step needed to obtain a stable result. With this method, a set of tridiagonal equations is established, and these are solved using the Thomas algorithm (Morton and Mayers 1994). A full listing of the computer code is available on request to the authors.

We tested and confirmed the validity of the finite differencing code by running the program using the same (fixed) values for ρ, C_p, and K as were employed in the analytical approach. The results were identical to those obtained analytically.

Having demonstrated that the numerical model yields appropriate results, at least in this last special case, we had sufficient confidence in its application to introduce the temperature-dependent values of ρ, C_p, and K proposed by Yomogida and Matsui (1984). Their formulations of these parameters enable us to begin with a planetesimal made from a highly insulating, porous aggregate of dust. The formulations incorporate a rapid increase in ρ and a corresponding jump by almost three orders of magnitude in K (due to sintering and porosity loss) at around 700 K, and an increase in C_p from around 650 to 1250 J kg^{-1} K^{-1} over the temperature range considered here (250 K–1850 K). In addition, we include in our model the latent heat of fusion, taken as 2.56×10^5 J kg^{-1} (Young 1991) and assume that this heat is absorbed in five increments between the solidus and liquidus. We take the solidus and liquidus temperatures as 1425 K and 1850 K, respectively, and assume that a temperature of 1725 K corresponds to a partially molten state with equal volumes of liquid and solid (McKenzie and Bickel 1988). We use the values of T_0 and H_0 in Table 1. We arbitrarily assume a porosity of 50% at the time of instantaneous accretion. Finally we assume that the partial melt is "locked up" below 1725 K, endowing the crystal mush with rigidity, and that above 1725 K the partial melt loses its shear strength and starts to flow by convection.

Table 1. The values of the constants used in calculating temperatures by the analytical method (Equation 2).

Term	Symbol	Units	Value
Thermal conductivity	K	W m^{-1}K^{-1}	2.1
Specific heat capacity	C_p	J kg^{-1}K^{-1}	837
Density	ρ	kg m^{-3}	3300
Diffusivity	κ	m^2 s^{-1}	7.6×10^{-7}
Decay constant	λ	yr^{-1}	9.50×10^{-7}
Initial power per unit volume	A_0	W m^{-3}	6.4×10^{-4}
Initial power per unit mass	H_0	W kg^{-1}	1.9×10^{-7}
Ambient temperature	T_0	K	250

RESULTS OF THE NUMERICAL MODELLING

Sintering and Volume Loss

We illustrate the results of our model for a representative planetesimal that accretes at $t = 0.75$ Myr and has a final radius of 50 km (Figs. 2, 3, 4a, and 4b).

Initially the entire planetesimal at 250 K is highly porous and insulating with a radius of about 64 km, but the radius shrinks rapidly to 50 km between $t = 0.88$ Myr and $t = 0.90$ Myr owing to sintering (Yomogida and Matsui 1984) when the uniform internal temperature reaches about 700 K (Fig. 2). The outermost layer of the planetesimal, where the temperature stays below 700 K, remains as a thin insulating blanket. After shrinkage at $t = 0.90$ Myr, continued heating until $t = 1.22$ Myr leads to a steady temperature rise throughout the totally insulated interior. At any given time, a flat thermal profile extends from the center to within less than 10 km of the surface. Beyond this, the temperature profile gradually gets steeper and curves asymptotically into a gradient of about 180 K km^{-1} close to the surface. The gradient then jumps to around 180 K per meter through the outermost two or three meters composed of residual, porous, highly insulating dust below 700 K, to the actual surface at 250 K.

At $t = 1.22$ Myr, the insulated interior begins to melt as the temperature passes the solidus (1425 K). Thereafter, the rate of temperature increase slows down owing to the absorption of latent heat of fusion. By about $t = 1.52$ Myr, the temperature reaches 1725 K and the degree of melting reaches 50%. This stage, we assume, marks the transition in the partially molten interior from a rigid state with interlocking crystals to a cohesionless state capable of flowing.

Before describing the results of further heating beyond 1725 K, we point out a small modification we made to our model at high temperatures. The bounding surface between the dense planetesimal interior and its covering of porous, insulating dust will automatically adjust to the 700 K isotherm during heating. For subsequent modelling, therefore, we have

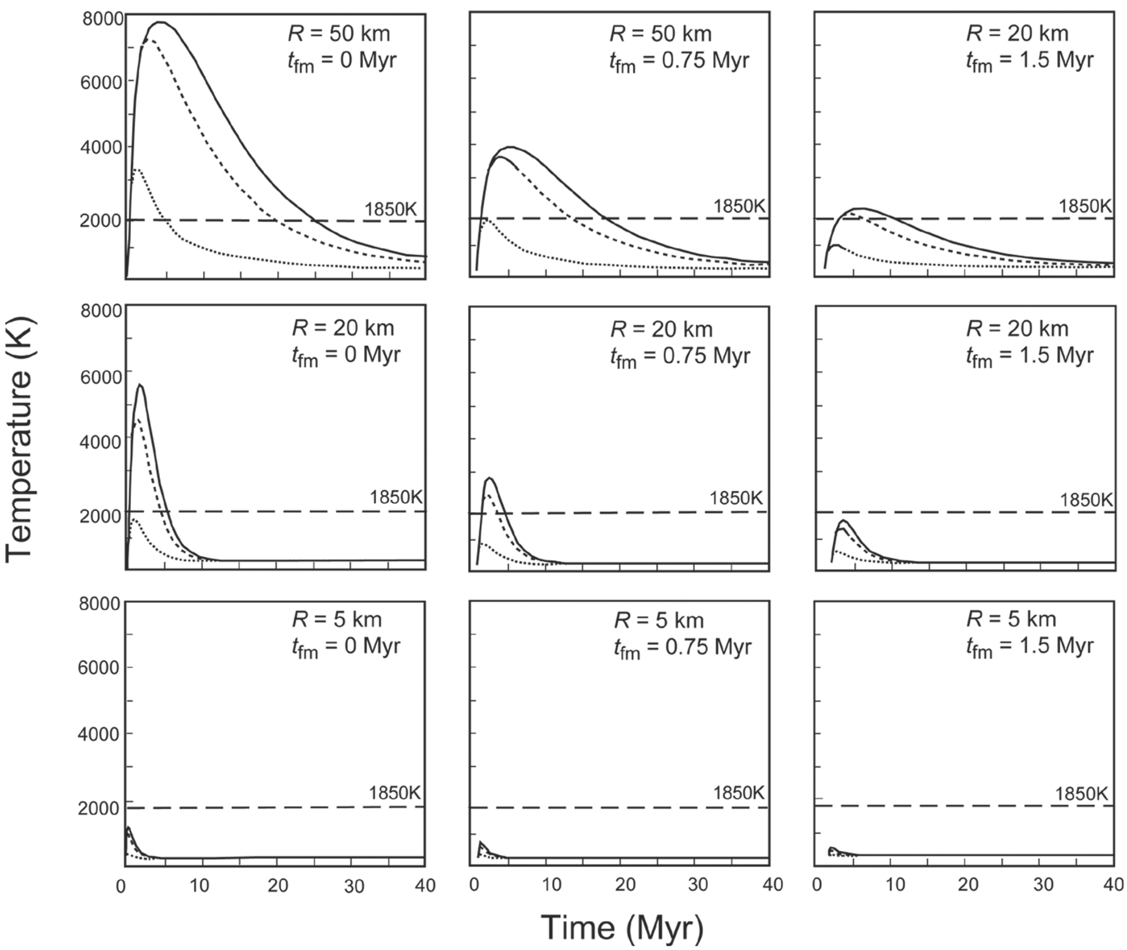

Fig. 1. Temperature-time plots for nine planetesimals showing the effect of radius, R, and time of formation, t_{fm}, on thermal history. 1850 K is the temperature for total melting (the liquidus). The continuous curves, long-dashed, and short-dashed curves trace the temperature at the center, halfway to the center, and one-tenth of the way to the center of the planetesimal, respectively. The curves were calculated using the analytical method (Equation 2), but identical curves were obtained by our numerical method when using the same values for ρ, C_p, and K (Table 1) as were employed in the analytical approach.

let the program skip the initial stage of heating and volume loss to start running from t = 0.88 Myr, with an initial radius of 50 km, zero porosity and a "surface" temperature of 700 K (where the 700 K "surface" is a meter or so beneath the actual surface at 250 K). Between t = 0.88 Myr and t = 1.52 Myr this simplification yields virtually identical results (Fig. 3) to those of the original program (Fig. 2) and it reduces computing time considerably.

Transition to Convection

After t = 1.52 Myr, convection begins. Convection is not easy to model. Here we simulate convection by assuming that at temperatures greater than 1725 K the thermal conductivity, K increases by three orders of magnitude. As a result, heat is transferred rapidly from the convecting interior into the overlying rigid, partially molten zone. This zone in turn becomes cohesionless by increments on being heated above 1725 K and joins the convective regime of the interior. Excess radiogenic heat does not, therefore, raise the temperature of the convecting slurry significantly; rather, it increases the total volume of the interior that is molten and convecting (Figs. 4a and 4b). The volume of melt continues to increase, and the thickness of the overlying rigid crust decreases, until the rate of heat generation in the molten interior balances the rate at which that heat can escape by conduction through the remaining crust.

The crust of our representative planetesimal (Fig. 4a) achieves a minimum thickness of about 2 km and has a temperature gradient (beneath the regolith) of about 500 K km^{-1}, roughly when t = 2 Myr. Thereafter, the crust thickens, reaching about 10 km when t = 5 Myr. The entire planetesimal becomes rigid through to its center by about t = 10 Myr.

As a second example, we show results for a planetesimal of the same size as above (R = 50 km), but which forms 0.75 Myr earlier, when t = 0 (Figs. 5a and 5b). In this case, owing to the short half-life of ^{26}Al, the initial heating rate is doubled. Convection begins at around t = 0.3 Myr, and

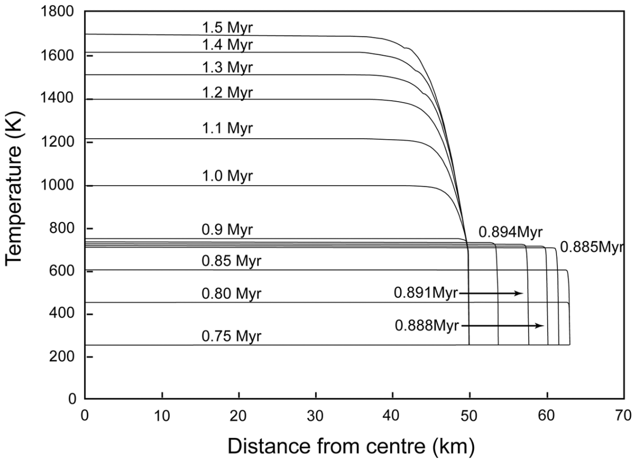

Fig. 2. *T-r* plot for a planetesimal that forms at $t = 0.75$ Myr from CI-like dust at 250 K with initial radius $R = 64$ km and 50% porosity. Temperature profiles at successive times up to $t = 1.5$ Myr are shown. Note the reduction in volume between $t = 0.88$ and $t = 0.9$ Myr, giving a final radius of 50 km.

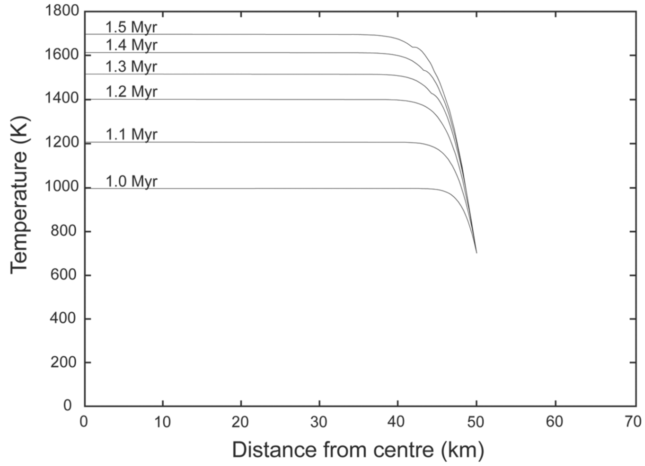

Fig. 3. *T-r* plot for a planetesimal that forms at $t = 0.883$ Myr from CI-like dust at 700 K with zero porosity and radius $R = 50$ km. Note that the temperature profiles are identical to those in Fig. 2, showing that the volume reduction stage in the modelling can be omitted when computing thermal changes at higher temperatures than 700 K.

between about $t = 0.5$ Myr and $t = 1$ Myr, the crust has thinned to a mere 500 meters and has a temperature gradient of about 2000 K km^{-1}. The crust then thickens up, slowly at first then more quickly. When $t = 2$ Myr, it has thickened to about 2 km and the planetesimal's subsequent cooling history is "in step" with the first example (Fig. 4b). In general, the model suggests that all molten planetesimals of a given size eventually cool and solidify in tandem, regardless of the time when they accrete.

A brief consideration of the behavior of larger planetesimals suggests that a body with a radius of 100 km that forms at $t = 0$ will have double the crustal heat flow of a 50 km body, since heat generation goes with the cube of the radius, and heat conduction goes with the square of the radius. The thickness of the crust will be just 250 meters during the early intense phase of heat production, when more than 99% of the volume of the planetesimal will be magma undergoing convection.

The Limits of Planetesimal Size and Formation Time for Significant Melting to Occur

We ran our model for planetesimals with many different values of radius and formation (accretion) time in order to get some idea of the limiting values of these parameters for substantial melting to occur. To reduce computing time, we did not include convection in these runs, but merely sought to determine whether or not melting temperatures are reached. The "will melt" field above and to the left of the bold curve in Fig. 6 has somewhat arbitrarily been picked from a broad spectrum of different degrees of melting. The field delimits combinations of size and formation time for planetesimals whose central temperature eventually reaches or exceeds the liquidus temperature of 1850 K assuming that heat is lost only by conduction. Within the "will melt" field we show contours of the volume percentage of the planetesimal whose temperature exceeds 1725 K at the moment the central temperature reaches 1850 K. These contours give a rough idea of the intensity of melting for different combinations of formation time and radius. Contours outside the "will melt" field show the peak central temperature reached, which here is less than 1850 K.

It is clear that our representative planetesimals, marked by stars on Fig. 6, ($R = 50$ km, accretion at $t = 0$, and at $t = 0.75$ Myr) are situated comfortably within the "will melt" field. We suggest that planetesimals will undergo substantial melting and become thin-skinned molten globes if they plot above the 50% contour in Fig. 6. i.e., if they have a radius >20 km where they accrete at $t = 0$, with the limiting radius rising to about 80 km where they accrete at $t = 1.3$ Myr. Moving below the 50% contour, the extent of melting falls off quite rapidly. Planetesimals that plot just below or just to the right of the "will melt" field (e.g., with radii between about 10 km and 7 km, or forming after $t = 1.3$ Myr but before about $t = 1.5$ Myr) will undergo some convective melting in their deep interiors, but those less than 7 km in radius, or forming after $t = 1.5$ Myr will never reach 1725 K, and will remain rigid throughout (even if partially molten at depth). These limiting radii assume, of course, the presence of an insulating regolith with a steep thermal gradient between 700 K and the surface at 250 K.

Figure 6 says nothing about the time when "melting" actually begins. It merely reports the range of accretion times and radii of planetesimals that are destined eventually to "melt." How long does it take to reach this state? The duration of heating depends mainly on the time of accretion. From the examples given in Figs. 4 and 5, we know that when accretion happens at $t = 0$, convection begins at $t = 0.3$ Myr, and with

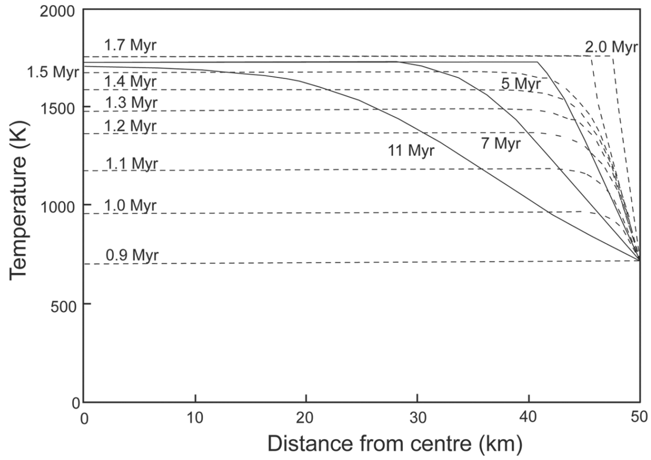

Fig. 4a. *T-r* plot for the same planetesimal as in Fig. 3, showing temperature profiles at times up to $t = 11$ Myr. Convection begins at $t = 1.52$ Myr, and by $t = 2$ Myr, the interior is a molten slurry to within 2 km of the surface. By 11 Myr, the center is still hot and partially molten, but it is "locked up" and unable to flow.

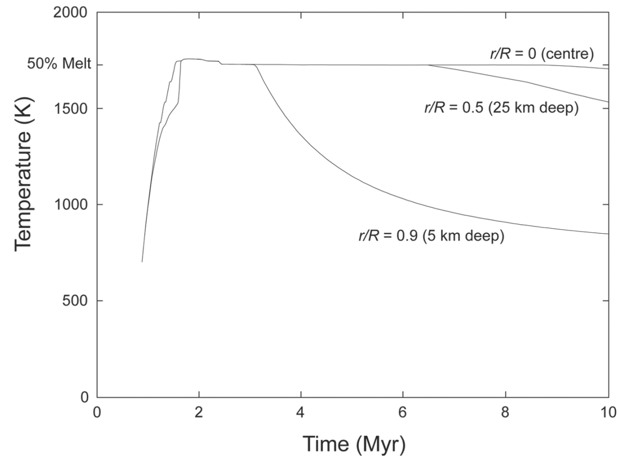

Fig. 4b. *T-t* plot of the same data as shown in Fig. 4a. The three traces (superimposed over part of their length) show the temperature at depths of 50 km, 25 km, and 5 km.

accretion at $t = 0.75$ Myr it begins at $t = 1.5$ Myr. If a planetesimal's size and formation time place it close to the curved margin within the "will melt" field, then the onset of convection will generally be delayed by several million years, and will only affect the deep interior of the planetesimal.

SOME FACTORS THAT MAY AFFECT THE MODEL

Additional Heating by ^{60}Fe

^{26}Al was not the only source of radioactive heating at the start of the solar system. ^{60}Fe, which decays to ^{60}Ni with a half-life of 1.5 Myr, may also have contributed to the heat budget. The recent discovery of excess ^{60}Ni in grains of troilite in Semarkona suggests an initial ^{60}Fe/^{56}Fe in the troilite of 9.2×10^{-7} (Mostefaoui et al. 2005). If the troilite has the same age as the chondrules in Semarkona (Kita et al. 1998), then the solar system initial value of ^{60}Fe/^{56}Fe may have been roughly 1.8×10^{-6}, broadly similar to the value inferred from measurements on a CAI by Birck and Lugmair (1988). In this case, the total radiogenic energy available from ^{60}Fe in primitive (i.e., CI-like) dust at the start of the solar system would have amounted to about 1.6 kJ g^{-1}. This is about one-quarter of the energy available in ^{26}Al at that time. However, because of its longer half-life, ^{60}Fe would have had an initial ($t = 0$) heat output of only about one-eighth of that due to ^{26}Al.

Unfortunately the initial ^{60}Fe/^{56}Fe of the solar system is still rather uncertain. Estimates are both greater (3×10^{-6}, Moynier et al. 2005; 4.4×10^{-6}, Quitté et al. 2005) and smaller (about 3×10^{-7}, Tachibana and Huss 2003) than the 1.8×10^{-6} cited above, which means that the contribution of ^{60}Fe to planetesimal heating is poorly constrained. For now, we note that if ^{60}Fe were present at a level anywhere in the range of the current estimates, the additional heat output would probably not significantly change the results of our model. For a given set of starting conditions, melting would begin marginally earlier, and cooling would be delayed for somewhat longer than with ^{26}Al alone. However, it may be significant that, after melting, iron will sink downward and any ^{60}Fe, as well as extending the duration of heating, will also release its energy deep inside the planetesimal (Yoshino et al. 2003).

Thermal Implications of Highly Volatile Species

The thermal implications of reactions involving volatile species are ignored in the model, but may be very important. The CI chondrite composition that we assume for a newly accreted planetesimal includes around 30% by weight of highly volatile components, chiefly H_2O and SO_3. These are chemically bound within the solid phases. Most of them will be released by devolatilization reactions and be driven out of the interior of a planetesimal by the time the temperature there reaches 1000 K, bearing in mind the low ambient internal pressure.

Reactions that release volatile components are generally endothermic and involve substantial absorption of heat. A rough idea of the importance of endothermic reactions comes from a brief consideration of the latent heat associated with H_2O. The energy needed to melt one gram of ice and to evaporate the water is about 2.6 kJ g^{-1}, which is significantly more than the 1.6 kJ required to raise one gram of cold primitive dust to its liquidus temperature. Future development of the model will clearly need to consider the thermal effects of reactions involving highly volatile species, including the melting and evaporation of ice.

The thermal evolution at subsolidus temperatures will, in addition, be complicated by the possibility that any water released by melting ice may first react with olivine in a highly

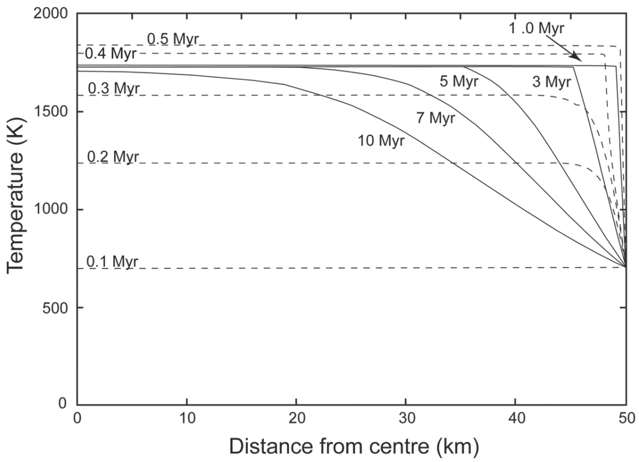

Fig. 5a. *T-r* plot for a planetesimal that forms at t = 0.06 Myr (rounded up to 0.1 Myr on the diagram) from CI-like dust at 700 K with zero porosity and radius R = 50 km. This is equivalent to formation at t = 0 and at 250 K, the ambient nebular and surface temperature. Convection begins at about t = 0.3 Myr, and between t = 0.5 and 1 Myr the molten, convecting interior extends to within about 0.5 km of the surface.

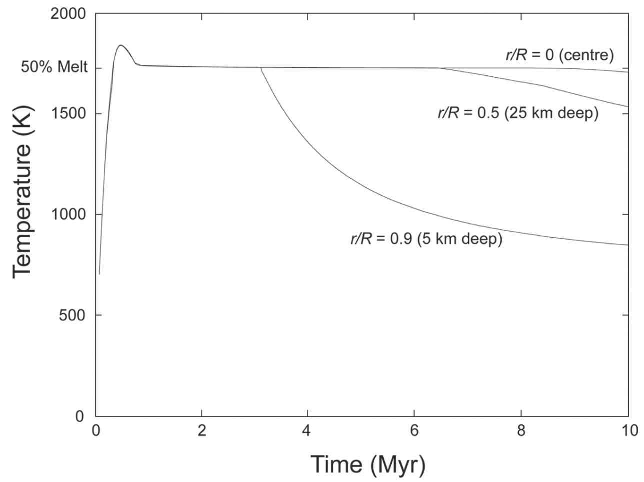

Fig. 5b. *T-t* plot of the same data as shown in Fig. 5a. The three traces (superimposed over part of their length) show the temperature at depths of 50 km, 25 km, and 5 km. Note that the cooling history is identical to that in Fig. 4b, suggesting that molten planetesimals of the same size cool in tandem, regardless of the time when they accrete.

exothermic reaction to produce serpentine. However, the resulting rise in temperature will be short-lived, as the endothermic effect of losing the H_2O again will be felt when the serpentine becomes dehydrated. This happens at or below about 800 K.

A very important aspect of having volatile components in the starting material is that their loss automatically leads to an increase in the concentration of ^{26}Al in the remaining solid matter. Whereas CI meteorites contain about 0.9% by weight of aluminium, the devolatilized equivalent (i.e., dry primitive dust) contains around 1.3%. Thus, the heat production per unit mass will be higher when the volatiles have been removed.

If planetesimals are initially accreted from dry primitive dust instead of CI-like dust, their radioactive energy content will be over 40% greater. At t = 0, the total stored radioactive energy due to ^{26}Al will be about 9 kJ g^{-1} rather than the 6.4 kJ g^{-1} in CI-like dust. The corresponding heat output will be 2.7×10^{-10} W g^{-1} instead of 1.9×10^{-10} W g^{-1}.

How would this significantly higher energy content and rate of heating affect our model? It would simply add 0.36 Myr to every value of time, t. Thus, the thermal profiles in Figs. 4a and 5a would now relate to times that are 0.36 Myr later than those indicated. The *T-t* curves shown in Figs. 4b and 5b would all shift by 0.36 Myr to the right; they would now relate to planetesimals that accrete at t = 1.11 Myr and t = 0.36 Myr, respectively, rather than 0.75 Myr and 0 Myr. The "will melt" field (Fig. 6) will also shift 0.36 Myr to the right, meaning that a large planetesimal accreting as late as t = 1.7 Myr will have the potential to reach 1850 K in its center.

Woolum and Cassen (1999) presented a diagram similar to Fig. 6, but showing the field for planetesimals whose peak central temperature exceeds about 1750 K rather than 1850 K. Their equivalent "will melt" field is larger than that in Fig. 6, extending to about t = 2 Myr. This is not only because their heat output at t = 0 assumes dry dust, producing 2.7×10^{-10} W g^{-1}, but also because they assume a higher value than we do for heat capacity and a lower one for conductivity, and they do not consider latent heat of fusion. All these assumptions act together to expand their "will melt" field to later times.

If a planetesimal is initially formed from CI-like material then, by the time it has lost its original H_2O and other highly volatile components, its thermal behavior will be identical to that of a planetesimal that accretes at a later time from dry primitive dust. Just how much later can be calculated simply as the time needed to release the energy for removing the volatiles.

The Possibility of Basalt Segregation During the Early Stages of Partial Melting

Taylor et al. (1993) suggest that the "basaltic" melt fraction in a partially molten planetesimal >50 km in radius will possibly be transported to near-surface regions in substantially less than 1 Myr. If this suggestion were correct, then basalt segregation would transfer aluminium (and hence the ^{26}Al heat source) from the interior to the surface. This would effectively forestall further ^{26}Al heating and melting in the interior. Wood (2000) assumes such a case in his own analysis of planetesimal melting. Clearly, if basalt were to move to the planetesimal's surface before the temperature gets high enough to initiate convection, our model is seriously flawed.

Uncertainties in estimating the rate of melt migration are considerable. The rate quoted above is a maximum; it assumes a grain size of 1 cm, dilational cracking during melting which leads to dyking, and the possible expulsion of melt driven by volatile expansion. With a smaller grain size (say, 3 mm) and an absence of dyking and volatile release, Taylor et al. (1993) indicate that the time needed to remove the melt could be more than 1 Myr. The latter period is much longer than the time needed to heat a planetesimal from the solidus to a state of 50% molten. For example, in the case of the planetesimal of Fig. 4, it takes only 0.3 Myr to go from the start of melting at 1425 K to the onset of convection at 1725 K. The same critical rise in temperature takes a mere 0.1 Myr when accretion happens at t = 0 Myr (Fig. 5). It is certainly possible, therefore, that the rate of heating could outstrip the rate of melt migration, so that convection would begin while the melt fraction was still slowly moving upward.

We are aware that melt segregation did occur on asteroid 4 Vesta, which is surfaced by basalt. However, two features of Vesta, absent from our model planetesimals, would strongly favour the segregation of basalt. First, the radius of Vesta is about 250 km, which would increase melt buoyancy roughly by a factor of five compared to buoyancy in a 50 km body. Second, by the time the first basalts erupted on Vesta, which was probably 3 or 4 Myr after CAIs (Srinivasan et al. 2000; Wadhwa et al. 2005) the rate of ^{26}Al heating was about an order of magnitude lower than it had been at t = 1 Myr. The slower rate of heating on Vesta at the time of eruption, compared to the rate in our model at about t = 1 Myr, would have given much more time for the basaltic partial melt to escape. Indeed, at this late time, the peak temperature may have hovered in the partially molten interval and failed to overstep the 1725 K threshold needed to initiate convection. The relatively late melting of Vesta, incidentally, implies that this asteroid accreted rather late, possibly close to the boundary of the “will melt” field in Fig. 6.

Iron meteorites and pallasites provide strong independent evidence that their parent bodies did not lose early basalt, but became heated beyond the onset of convection, above about 1725 K. Taylor et al. (1993) argue (later in their paper) that the IVB iron meteorites (with only 0.6–1.2 wt% sulphur) crystallized from liquid iron at >1770 K, a temperature that implies at least 60% partial melting in the overlying silicate. They also argue that the absence of pyroxene in pallasites implies a melt fraction above about 50% and that the Mg-rich composition of the olivine suggests it was close to 70%.

Taylor et al. (1993) also, incidentally, show that the Rayleigh number and the Reynolds number both exceed substantially the critical values necessary for turbulent convection in the “magma oceans” on the parent bodies of iron meteorites and pallasites. Indeed, their cartoon depicting the processes operating in an asteroidal magma ocean (their Fig. 7) bears a close resemblance to the internal state of the kind of molten planetesimal predicted by our model.

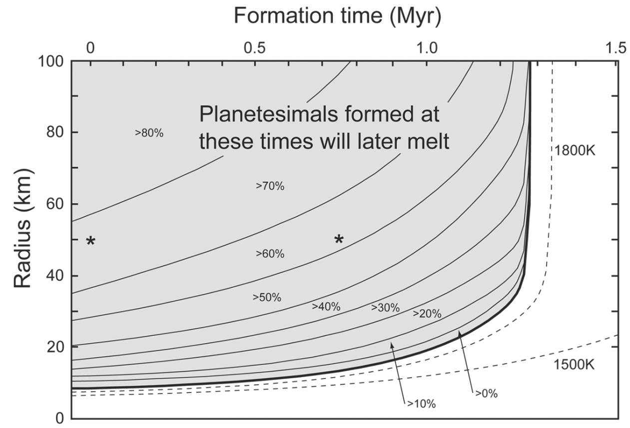

Fig. 6. Combinations of planetesimal radius and formation time that will lead to eventual melting. “Melting” is defined for this figure only as reaching or exceeding a central temperature of 1850 K without allowing for convection. The shaded “will melt” field, above and to the left of the heavy curved line, is contoured in the volume percent of the planetesimal that exceeds 1725 K at the instant 1850 K is reached in the center. Dashed contours outside this field show the maximum central temperature reached.

A Radiation Boundary Condition at the Surface

The model assumes that the surface temperature of a planetesimal is the same as the starting temperature (the ambient nebular temperature). This is not strictly the case because the surface temperature must rise above the ambient temperature for heat to be radiated away. We determined the scale of this effect and found it to be negligible. Our calculations show that by including a radiation boundary condition in the model, the temperature at the planetesimal surface will rise by less than 1 K.

The Fate of Thin Crust

For large planetesimals at the peak of melting, the model portrays a thin, brittle crust enclosing the molten interior. Being more dense than the magma flowing beneath it, the crust will be prone to foundering, triggered, perhaps, by small impacts or convective drag. With the submersion of segments of the carapace, the incandescent magma will reach the surface and, for a brief moment, greatly increase the rate of heat loss. Furthermore, the newly quenched surface will no longer have an insulating layer of dust. These factors together will hasten the loss of heat from the planetesimal interior, but no attempt has been made here to quantify their importance.

The Effect of Gradual Accretion on Thermal History

We have extended our model at a reconnaissance level to investigate the effect on thermal history of gradual accretion. Results are presented for two 50 km bodies that begin to

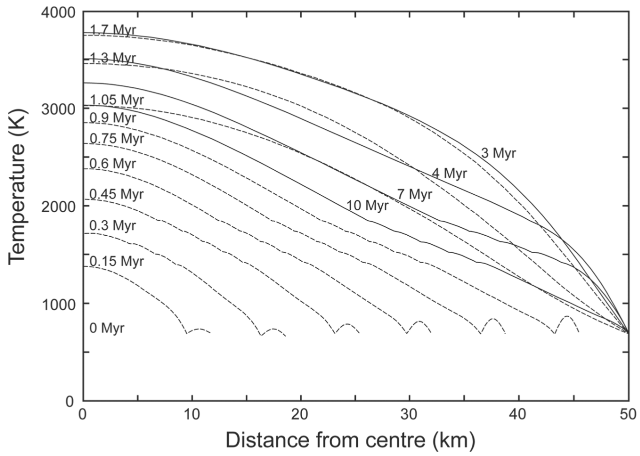

Fig. 7a. *T-r* profiles through a planetesimal whose radius increases in 10 regular increments of 5 km between t = 0 Myr and t = 1 Myr. Convection is not considered.

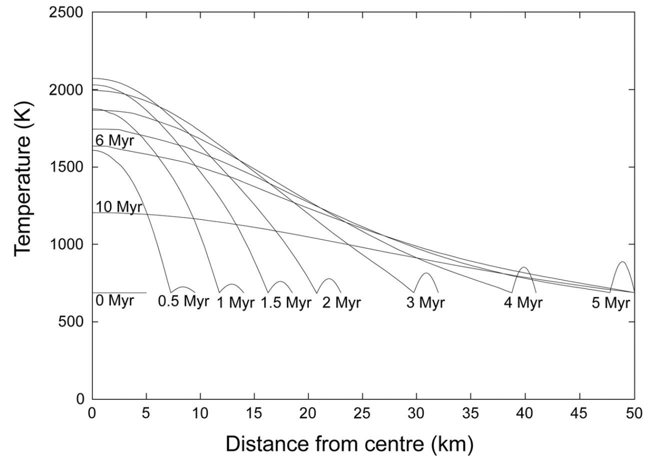

Fig. 7b. *T-r* profiles through a planetesimal whose radius increases in 10 regular increments of 5 km between t = 0 Myr and t = 5 Myr. Convection is not considered.

accrete at the beginning, t = 0, and grow in size by ten regularly spaced increments, adding 5 km to the radius with each step over a duration of 1 Myr and 5 Myr, respectively (Figs. 7a and 7b). These results do not include convection. Consequently, meaningless high temperatures above 1850 K have been computed, as they were in Fig. 1. Nevertheless, some useful observations may be made.

With accretion lasting 1 Myr (Fig. 7a), the center begins to melt at t = 0.3 Myr. Substantial melting is initially held back by the successive overlays of cold dusty shells, but after 1 Myr it is likely that the molten interior will expand outward as convection begins. In the second example, where accretion takes 5 Myr (Fig. 7b), a small hot central volume with some melting remains deeply buried beneath a cool, solid cover between 30 and 40 km thick.

We assume that the radius increases linearly with time, but in reality, the rate of accretion is probably a non-linear function of time. However, the question of whether accretion is instantaneous or gradual is probably less important for melting than the question of whether most of a planetesimal's mass is in place before about 1.5 Myr, while ^{26}Al is still sufficiently potent to cause meltdown.

IMPLICATIONS OF THE RESULTS FOR METEORITE PARENT BODIES

The Timing of Accretion of Meteorite Parent Bodies

Our results suggest that the parent bodies of iron meteorites and other differentiated meteorites formed (i.e., accreted) inside the "will melt" field of Fig. 6, before about t = 1.3 Myr. Accretion could have been up to 0.36 Myr later than this if the dust were initially "dry" (i.e., free of highly volatile species like H_2O and SO_3) or became devolatilized after accretion, and later still if, in addition, ^{60}Fe contributed to the heat budget, but it would probably have been no later than about t = 2 Myr.

Meibom and Clark (1999) inferred that the world's meteorite collections include samples derived from some 135 separate parent bodies. Altogether 108 (or 80%) of these parent bodies must have melted because they are represented by differentiated meteorites, most of them ungrouped irons. It follows, therefore, that four-fifths of these original parent bodies appear to have accreted within 2 Myr of the beginning of the solar system.

Whether or not these parent bodies still exist as asteroids today is unclear. Perhaps many of them were disrupted long ago, and only fragments of them survive today within asteroidal "rubble piles" that were assembled at some time after the early phase of intensive melting.

The Parent Bodies of Chondrites

The corollary of the above argument is that the chondritic parent bodies accreted outside the "will melt" field in Fig. 6. Some of them, like the H chondrite parent body (Trieloff et al. 2003), evidently had a large radius (above 30 km). Since they did not melt, such bodies could not have accreted during the first 1.3 Myr, and possibly not even during the first 2 Myr. Their accretion was presumably delayed until after the heat production from ^{26}Al had fallen to an appropriately low level.

The late accretion of chondritic parent bodies is, of course, consistent with the young ages of around t = 2 Myr reported for many chondrules (Kita et al. 2000; Huss et al. 2001; Amelin et al. 2002).

The idea that chondritic parent bodies formed after the parent bodies of differentiated meteorites is, perhaps, a little disconcerting for those who regard chondritic meteorites, with their primitive chemistry and the great antiquity of their component CAIs and chondrules, as samples of the earliest planetesimals. However, if we accept the presence in nebular dust of canonical ^{26}Al and assume that many of the parent planetesimals were larger than about 30 km in radius, then we are led to conclude that many chondritic meteorites survive from the closing stages of a phase of planetesimal accretion and melting that lasted for the first two or three million years of the solar system. Differentiated meteorites, on the other hand, although they may finally have crystallized and cooled at an even later time, were evidently generated in planetesimals whose accretion history dates from well before that of the chondritic parent bodies.

Lundgaard et al. (2004) also conclude that chondritic parent bodies are probably younger than achondritic parent bodies, based on the premise that ^{26}Al was the dominant planetesimal heat source. Kleine et al. (2005) reach a similar conclusion, with evidence for early meltdown (and hence early accretion) from the tungsten isotope systematics of metal in meteorites. A strongly negative ^{182}W anomaly in nearly all iron meteorites is taken to indicate early separation of molten metal from silicate, probably before the accretion of chondritic parent bodies.

The Survival of CAIs

The preservation of CAIs with excess ^{26}Mg poses an interesting conundrum. Following their formation at the start of the solar system, CAIs were evidently quickly incorporated into planetesimals where they were stored, safe from being destroyed by drifting into the early sun. However, the planetesimals that protected the CAIs were presumably also being heated by ^{26}Al. Had the temperature exceeded 650 °C, the record of excess ^{26}Mg in the CAIs would have been erased by diffusion (LaTourrette and Wasserburg 1998). It has been suggested, therefore, that CAIs were captured by small planetesimals (e.g., <5 km radius) or were shallowly buried close to their host planetesimal's cold outer surface. Another possibility is that the disruption of planetesimals by mutual collisions may have served to recycle CAIs, along with other debris, into later generations of planetesimals, and each new generation may have started from a cold beginning and also been small.

Yet another factor that may have been important in maintaining the necessary low temperatures within early formed parent bodies is the accretion of water ice along with primitive dust. Most kinds of carbonaceous chondrite display evidence for aqueous alteration, and the source of the water may well have been ice grains. Ice constitutes a major fraction of the primitive material in comets, so its incorporation into accreting planetesimals is entirely plausible. Since the energy needed to melt ice and evaporate H_2O is of the same order as the energy needed to heat and melt cold primitive dust, ice could have been extremely efficient in buffering low temperatures and preserving CAIs during the critical early period when ^{26}Al heating was most active. Of course, any endothermic alteration of olivine to serpentine would have lessened the buffering of low temperatures by H_2O.

Recently, Cuzzi et al. (2003) proposed a mechanism whereby turbulence in the nebula counteracts the inward spiral of CAIs and keeps them in the disk for periods of the order of a million years. In this case, there would be no need to invoke early accretion into a planetesimal host to keep the CAIs from destruction.

Implications of the Model for Chondrule Formation

The model implies that substantially molten planetesimals would have been commonplace in the protoplanetary disk between $t = 0.3$ Myr and about $t = 2$ Myr. It seems unlikely that these planetesimals evolved completely independently; from time to time they probably collided one with another. After all, the transition from a disk with many small planetesimals to one with a few large planets is widely assumed to have involved a succession of many accretionary collisions. If the molten planetesimals were protected only by a thin brittle crust, then even low-encounter velocities may have led to impact plumes with a very high proportion of melt droplets, and these droplets conceivably cooled to form chondrules.

Sanders (1996) and Lugmair and Shukolyukov (2001) favor the production of chondrules by this mechanism, but many details of the process remain to be clarified. A more extensive discussion of the strengths and weaknesses of this particular way of making chondrules is presented by Sanders and Taylor (2005).

SUMMARY AND CONCLUSIONS

The case for a uniform canonical ratio of $^{26}Al/^{27}Al$ in nebular dust is sufficiently well-established to know that the short-lived radioactivity embedded in early formed planetesimals was enough to melt their well-insulated interiors. Our finite difference model for simulating the thermal history of such planetesimals points to volume loss as porosity collapses. Heating and partial melting follow. The model incorporates heat transfer by convection once the melt fraction exceeds about 50% at 1725 K. With further heating, the planetesimal becomes a molten sphere undergoing turbulent convection within a thin rigid shell. By 2 or 3 Myr after CAIs, with radioactive power waning, planetesimals begin to freeze steadily from the outside in. Planetesimals melt and evolve in this way only where they form before $t =$ about 1.5 Myr and where their radius exceeds 20 to 80 km, depending on the time of accretion.

The potential to follow this path will be helped if the raw dust is volatile-free and if ^{60}Fe is an additional heat source; it will be hindered if water ice accretes with the dust or if the heat source migrates to the surface at an early stage of melting.

The results of the model help to constrain a picture of the unfolding events in the protoplanetary disk over its first three million years of history. They imply that the parent bodies of differentiated meteorites accreted within the first 1.5, or at most 2 Myr. In contrast, the parent bodies of chondritic meteorites, unless they were small or ice-laden, must have accreted at a later time, which is consistent with the measured ages of the chondrules within them. Chondrules themselves may possibly be frozen droplets of melt released by disruption of the molten planetesimals during this poorly understood time interval.

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