



## Thermal evolution and differentiation of planetesimals and planetary embryos

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### ABSTRACT

In early Solar System during the runaway growth stage of planetary formation, the distribution of planetary bodies progressively evolved from a large number of planetesimals to a smaller number of objects with a few dominant embryos. Here, we study the possible thermal and compositional evolution of these planetesimals and planetary embryos in a series of models with increasing complexities. We show that the heating stages of planetesimals by the radioactive decay of now extinct isotopes (in particular <sup>26</sup>Al) and by impact heating can occur in two stages or simultaneously. Depending on the accretion rate, melting occurs from the center outward, in a shallow outer shell progressing inward, or in the two locations. We discuss the regime domains of these situations and show that the exponent  $\beta$  that controls the planetary growth rate  $\dot{R} \propto R^\beta$  of planetesimals plays a crucial role. For a given terminal radius and accretion duration, the increase of  $\beta$  maintains the planetesimals very small until the end of accretion, and therefore allows radioactive heating to be radiated away before a large mass can be accreted. To melt the center of ~500 km planetesimal during its runaway growth stage, with the value  $\beta = 2$  predicted by astrophysicists, it needs to be formed within a couple of million years after condensation of the first solids. We then develop a multiphase model where the phase changes and phase separations by compaction are taken into account in 1-D spherical geometry. Our model handles simultaneously metal and silicates in both solid and liquid states. The segregation of the protocore decreases the efficiency of radiogenic heating by confining the <sup>26</sup>Al in the outer silicate shell. Various types of planetesimals partly differentiated and sometimes differentiated in multiple metal–silicate layers can be obtained.

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### 1. Introduction

Evidence from <sup>182</sup>Hf–<sup>182</sup>W systematics of meteoritic samples points to a rapid accretion of terrestrial planetary bodies and their early differentiation into a metallic core and silicate mantle (Kleine et al., 2002; Yin et al., 2002). This applies not only to present-day planets but also to smaller bodies, a prime example being asteroid Vesta (e.g., Ruzicka et al., 1997). The inferred differentiation time-scales for the parent bodies of some magmatic iron meteorites are as low as ~1 My after the crystallization of CAIs (Kleine et al., 2009). Such a rapid differentiation most likely required a separation of metal from silicates in a partially or fully molten state. This could be accomplished by percolation of molten iron through solid silicates or by liquid–liquid separation in an early magma ocean, and most likely occurred by both mechanisms over different time windows (Stevenson, 1990).

The two important sources of energy in the early Solar System during planetesimal growth are radioactive heating by short-lived radionuclides and accretion heat brought by impacts. The former heat source is important for small growing bodies when the impact heating is still insignificant, within a few half-lives of the decaying radionuclides. The impact heating becomes the dominant energy source later as the planetesimal becomes more massive, and after the radionuclides have sufficiently decayed.

Separation of the denser molten metal from a porous silicate matrix may occur at temperatures in between the melting temperature of the metallic phase and the silicate solidus. Static percolation theory relates permeability at small melt fraction to the magnitude of dihedral angle formed between liquid metal and solid silicate grains (e.g., von Bagen and Waff, 1986). For a dihedral angle below the critical value of 60° an interconnected network of melt tubules are formed and porous flow is possible even for tiny melt fractions. On the other hand, in the case of larger angles, the melt tends to form isolated drops and a minimum melt fraction is necessary to get interconnection. For this reason, this issue has been the focus of many experimental studies. At high pressure,

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large dihedral angles have been found (Minarik et al., 1996; Shannon and Agee, 1996; Terasaki et al., 2005) and percolation was generally considered as an inefficient core formation mechanism. However, recent experimental results of Terasaki et al. (2008) suggest that dihedral angles fall below  $60^\circ$  at conditions of high oxygen fugacity and pressures below 2–3 GPa, and percolation may thus be the dominant segregation mechanism in planetary bodies with a radius less than  $\sim 1300$  km. Moreover, even for angles above the critical value, the threshold for connectivity has been constrained from *in situ* electrical conductivity measurements (Yoshino et al., 2003, 2004; Roberts et al., 2007) to be 3–6 vol.%, even though some studies put the threshold higher (Walte et al., 2007; Bagdassarov et al., 2009). In any case, the typical reduced metal content of chondrites (up to 8% by volume for ordinary chondrites and even higher for some carbonaceous, enstatite and other chondrites; e.g., Krot et al., 2003; Scott and Krot, 2003), is above this threshold and the total metal content that forms planetary cores, about 20% by volume, is well above this threshold. Once the interconnectivity is established, the liquid iron can separate by percolation and its volume fraction decreases down to a “pinch-off” fraction, which is somewhat smaller than the percolation threshold (von Bagen and Waff, 1986; Terasaki et al., 2005).

Another aspect of the problem that has been completely left out of the discussion up to now concerns the deformation of the solid matrix. Indeed, even if permeability is sufficient to allow for Darcy flow with in- and out-flow, the situation is quite different in the case of a fixed volume. For the metal to go down and form the core, the partially solid silicate matrix needs to move up and therefore deform viscously, a process termed compaction. The high viscosity of the solid matrix can be the main limiting factor for core formation in planetesimals particularly at high permeability. Some early experiments (Takahashi, 1983; Walker and Agee, 1988) of partial melting of meteorites did not observe metal migration, which was interpreted as indication for the necessity of silicate melting (Taylor, 1992). However, the main reason is that compaction at that scale is very difficult to achieve even when the metal network is well connected. A crude estimate of the timescale for phase separation by this process can be obtained by balancing the available gravitational deviatoric stress  $\Delta\rho g d$ , with  $\Delta\rho \sim 5000 \text{ kg m}^{-3}$  being the density difference between metal and silicate,  $g$  the acceleration of gravity and  $d$  the length-scale at which the process must occur, and the viscous resistance of the matrix,  $\eta v/d = \eta/t$ , with  $\eta \sim 10^{18} \text{ Pa s}$  the silicate viscosity,  $v$  the velocity of its deformation and  $t$  the timescale. For a laboratory experiment, where  $d \sim 1 \text{ cm}$  and  $g = 10 \text{ m s}^{-2}$ , one gets  $t = \eta/(\Delta\rho g d) \sim 100 \text{ My}$  whereas for a planetesimal where  $d \sim 500 \text{ km}$  and  $g \sim 1 \text{ m s}^{-2}$ ,  $t \sim 10 \text{ yr}$ . This explains well why segregation of metal through a deforming solid matrix under Earth gravity is not possible to observe in the laboratory but is a viable mechanism for planetesimals.

In order to establish whether the melting temperature of the metallic component can be reached in a planetesimal, various authors considered  $^{26}\text{Al}$  and  $^{60}\text{Fe}$  heating of an instantaneously

formed planetesimal (e.g., Yoshino et al., 2003; Walter and Tronnes, 2004; Rubie et al., 2007). They concluded that planetesimals of radius larger than  $\sim 30 \text{ km}$  experienced significant melting of the metallic component. However, as Merk et al. (2002) showed using models where the planetesimal radius increases linearly with time, the assumption of instantaneous formation at its final size is a crucial limiting factor in this analysis. The maximum temperature reached in the planetesimal’s interior depends strongly on the accretion model. In order to properly constrain the thermal evolution of a planetesimal, one must consider the interplay between the accretion rate, generally variable in time, the final accreted size, and the possible delay of accretion after the element synthesis in the solar nebula.

In this study we investigate the thermal evolution of accreting planetesimals and planetary embryos (the continuous evolution described in the paper, from small to large bodies, makes it difficult to use a strict vocabulary, and we use loosely and interchangeably the terms planetesimal and embryo). In particular, we explore the effect of different choices for the accretion law, and the time interval over which accretion occurred. Unlike Merk et al. (2002), we include the thermal and gravitational energies brought by the impactors responsible for the growth of the planetesimal and employ a multiphase model that accounts for the separation of the metal from the silicates.

The sequence of models is constructed from the simplest scenario, i.e., a volumetrically-heated fixed-radius body with no differentiation, to more realistic cases. This way, the effects that various physical aspects of the final model have on the thermal evolution can be shown and discussed individually.

## 2. Thermal evolution without differentiation

### 2.1. Fixed radius, $^{26}\text{Al}$ heating

As a first estimate of planetesimal temperatures we consider a body of fixed radius heated by short-lived radionuclides. At this point we do not consider the accretion of the planetesimal or the segregation of metal. The thermal evolution is controlled by diffusion, heat production,  $H(t)$  (in  $\text{W kg}^{-1}$ ), and potentially latent heat release  $Q_L$ . The temperature  $T(r, t)$  in the spherically symmetric body follows:

$$\overline{\rho C} \frac{\partial T}{\partial t} + Q_L = \frac{1}{r^2} \frac{\partial}{\partial r} \left( \overline{k_T r^2} \frac{\partial T}{\partial r} \right) + \overline{\rho H(t)}, \quad (1)$$

where  $t$  is time and  $r$  is the radial coordinate. We start with a simple homogeneous mixture of silicate material and metal. The volumetric heat capacity  $\overline{\rho C}$ , the thermal conductivity  $\overline{k_T}$  and heat sources  $\overline{\rho H(t)}$  are calculated as volume fraction-weighted averages of the respective metal and silicate values (e.g.,  $\overline{k_T} = k_T^{\text{met}} \phi + k_T^{\text{sil}} (1 - \phi)$  where  $\phi$  is the volume fraction of metal, considered to be 18%). The important nuclides are  $^{26}\text{Al}$  with half-life  $\tau_{1/2}^{\text{Al}} = 0.717 \text{ My}$

**Table 1**

The initial heat production (at CAI formation time) for each radionuclide,  $H_0^{\text{Al}}$  and  $H_0^{\text{Fe}}$ , is calculated from  $H_0 = C \mathcal{F} N_A \lambda \varepsilon / \mathcal{M}$ .  $C$  is the chondritic weight fraction of the element (Lodders and Fegley, 1998),  $\mathcal{F}$  is the initial (CAI) isotopic ratio (fraction of radionuclide for element; Huss et al., 2009; Ouellette et al., 2009),  $\varepsilon$  is the nuclear decay energy per atom (Castillo-Rogez et al., 2009),  $\lambda = (\ln 2)/\tau_{1/2}$  is the decay constant,  $\mathcal{M}$  is the molar mass, and  $N_A = 6.02210^{23}$  is Avogadro’s number.

Quantity	Symbol	Unit	Al, $^{26}\text{Al}$ values	Fe, $^{60}\text{Fe}$ values
Chondritic abundance (element/chondrite)	$C$	kg/kg	0.00865	0.182
Initial isotopic fraction ( $^{\text{rad}}\text{X}/\text{X}$ where X is Al or Fe)	$\mathcal{F}$	N/N	$5 \times 10^{-5}$	$(3-10) \times 10^{-7}$
Molar mass of radionuclide	$\mathcal{M}$	$\text{kg mol}^{-1}$	0.026	0.060
Half-life of radionuclide	$\tau_{1/2}$	My	0.74	2.62
Decay energy	$\varepsilon$	J/atom	$5.1 \times 10^{-13}$	$4.3 \times 10^{-13}$
Decay constant	$\lambda = \frac{\ln 2}{\tau_{1/2}}$	$\text{s}^{-1}$	$3.0 \times 10^{-14}$	$1.5 \times 10^{-14}$
Initial radiogenic heat production ( $\text{W kg}^{-1}$ chondrite)	$H_0 = \frac{C N_A \lambda \varepsilon}{\mathcal{M}}$	$\text{W kg}^{-1}$	$1.5 \times 10^{-7}$	$(0.3-1.2) \times 10^{-8}$

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