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The hydrology of carbonaceous chondrite parent bodies and the evolution of planet progenitors

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Oxygen isotope ratio data from carbonaceous chondrites are explained if aqueous fluids flowed from high to low temperatures within the parent objects of these primitive meteorites. Recognition of ancient hydrological activity on primitive asteroid precursors affords a new class of constraints on the heat sources (ages?) and sizes of planet precursors in the early Solar System based on the geological evolution of planetesimals.

Keywords: carbonaceous chondrites; planetesimals;
thermal modelling; early Solar System

1. Introduction

There is abundant evidence that liquid and/or vaporous water was/were present within the parent bodies of primitive chondrite meteorites (Kerridge & Bunch 1979; McSween & Labotka 1993; Krot *et al.* 1998). There is also evidence that the aqueous fluid phase flowed within parent bodies (Wing & Bada 1991; Young *et al.* 1999). Presence of hydrological activity within planetesimals places limits on their sizes and rates of heat production. To the extent that the relative abundances of the short-lived nuclide ^{26}Al, the most likely source of internal heat within protoplanets, can be related to relative age in the early Solar System, hydrological activity also constrains the formation ages of chondrite progenitors.

Young *et al.* (1999) presented a simple model for fluid flow in a generic carbonaceous chondrite parent body (planetesimal) that accounts for the zero-order mineralogical and oxygen-isotopic features of many carbonaceous chondrites. In this paper we explore some of the constraints imposed by such models on the sizes and thermal histories of carbonaceous chondrite parent bodies and perhaps other planetesimals in the early Solar System.

Results suggest that all of the carbonaceous chondrites could have evolved within planetesimals with diameters of the order of 20 km and rates of internal heat production corresponding to initial ^{26}Al/^{27}Al values of $ca.\,1.0 \times 10^{-5}$. It is found that aqueous alteration of carbonaceous chondrites could have occurred within larger bodies, perhaps typified by asteroid 1 Ceres, if heat production was much lower (initial ^{26}Al/^{27}Al of $ca.\,1.0 \times 10^{-6}$). Metamorphism of carbonaceous chondrites could not have occurred in these same larger bodies.

The two possible settings for aqueous alteration, small hot (young?) bodies and large cool (older?) bodies, represent distinctly different time-scales of geological evolution. Alteration and metamorphism in small bodies would have occurred over an interval spanning less than 1 Myr, while aqueous alteration in a body the size of 1 Ceres would have required more than 5 Myr of heating. These results show that simple thermal models can be used to learn about the sizes and heat contents of meteorite parent bodies, and therefore the dynamics of rocky body formation in the early Solar System. The models provide specific predictions regarding the relative timing of geological events in meteorite parent bodies that are testable with relative chronometers.

2. Models

(a) Overview

The evolution of icy planetesimals is simulated using an explicit finite-difference code that permits simultaneous solution of partial differential equations depicting the flow (or creep) of liquid water, conductive heat flow, oxygen isotope exchange between rock and liquid water, and progress of a model mineralogical reaction. Although more sophisticated solutions to equations depicting heat flow, fluid flow and chemical reaction have been developed for a wide range of terrestrial problems, several aspects of the present calculations are peculiar to the planetesimal problem. Most importantly, water ice within the interiors of planetesimals would have been a readily exhaustible supply of liquid or vapour, since water is likely to have escaped the bodies. Modelling the planetesimals as spheres is required to account for this finite nature of the water supply.

The fluid flow evidenced by the oxygen-isotopic compositions of carbonaceous chondrites has important consequences for simulating the geological evolution of carbonaceous chondrite parent bodies. Fluid–rock reaction is far more limited in intensity, and is more spatially localized, when water flows than when water is stagnant; reaction between rock and water does not generally proceed to completion where fluid flows radially outward. As a result, the thermal runaway effects described by Cohen & Coker (2000) produced by exothermic hydration reactions between rock and a stationary fluid are not observed in the models shown here.

(b) Fluid flow

Consideration of permeabilities appropriate for chondrites and vapour pressures within icy planetesimals predicts that the bodies should have exploded due to vapour overpressures once water ice begins to melt (see, for example, Grimm & McSween 1989; Wilson *et al.* 1999; Cohen & Coker 2000). Although overpressures were likely to have been important in the structural evolution of icy planetesimals, the occurrence of aqueously altered chondrites testifies to the fact that not all icy bodies exploded prior to fluid–rock reaction. In order to account for the presence of alteration in the face of the tendency for explosion, Grimm & McSween (1989) and Cohen & Coker (2000) periodically vented water vapour from the interiors of their model asteroidal bodies (an approach that simulates hydraulic fracturing but detracts from the apparent 'ground truth' associated with using measured permeabilities to define conductance). These studies did not consider the isotopic and mineralogical effects of liquid as well as vapour advection.

In the present calculations liquid water is permitted to flow in the absence of convection as a result of pressure gradients. A gradient in pressure within icy planetesimals will have existed because the vapour pressure of water builds in the interior while the surface is exposed to the vacuum of space. The calculations simulate a growing pocket of water vapour within the body that forces liquid water continuously outward through an interconnected pore space that extends to the surface. This constitutes a zero-gravity approximation to unsaturated liquid flow that might be termed the 'vapour-piston' approach.

The vapour piston is simulated in these calculations by allocating a volume of interconnected pore space to the water vapour released during an interval of ice melting. The amount of space required by the vapour is calculated such that the moles of vapour expand to the equilibrium vapour pressure of liquid water at the temperature of interest. The equation that determines the rate of fluid production by ice melting at a single location within the body is then

$$R_{\text{melting}} = \frac{\mathrm{d}V}{\mathrm{d}t} = k_{\text{melting}} \frac{RT}{P_{\text{eq}} \hat{V}_{\text{H}_2\text{O}}},$$

where V is the volume occupied by new vapour per unit of pore space volume, $\hat{V}_{\text{H}_2\text{O}}$ is the molar volume of water vapour, P_{eq} is the equilibrium vapour pressure of water at temperature T, and k_{melting} is a rate constant (s^{-1}) that specifies the rate at which gas fills void spaces. The value for k_{melting} in these calculations is the same value used by Young *et al.* (1999). The calculations are insensitive to this parameter because the rate limiting step for fluid generation, and thus fluid-flux generation, is the migration of the 273 K isotherm in combination with the thermal buffering effect of the enthalpy of melting.

The problem is greatly simplified if it is assumed that the flowing water is incompressible. In this case the divergence of pore velocity v can be equated with the production rate of liquid water by ice melting in the pores and the rate of liquid loss due to freezing and reaction. The rate of freezing was adjusted in an ad hoc way to permit significant freezing of water at temperatures below 273 K (i.e. suitable values of the rate of water freezing, R_{freezing}, are prescribed by the values for the rate of ice melting, R_{melting}).

Real advection of water within small bodies with temperature gradients will have been driven in part by thermocapillary forces. Thermocapillary forces arise because energy per unit area decreases with temperature. The temperature dependence of surface energy causes a net force that draws plugs of intergranular water from higher to lower temperatures. The importance of surface tension as a driving force for unsaturated flow is known to be enhanced in microgravity environments (see Taitel & Witte (1996) and figure 1). Straightforward application of the Darcy equation for fluid flow is therefore unlikely to yield an accurate picture of the flux of liquid attending unsaturated flow of water through a planetesimal; the Darcy equation quantifies the competing effects of pressure gradients and restrictions to flow imposed by permeability but ignores surface tension.

(c) Heat flow

The thermal evolution of the planetesimals in these calculations was controlled by the competing effects of one-dimensional, spherically symmetrical conduction of
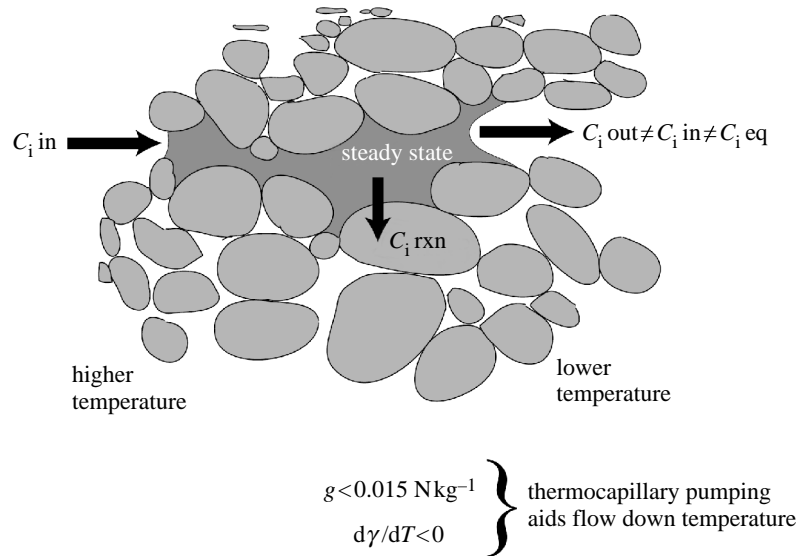
Figure 1. Schematic illustrating the development of a steady-state concentration of a chemical or isotopic species in a liquid-water phase flowing in the pore space of a carbonaceous chondrite parent body. The steady state is the consequence of the competing effects of the flow velocity and exchange with the enveloping rock. Flow in a planetesimal may be influenced by thermocapillary pumping along a negative temperature gradient, since surface energy per unit area (γ) goes down with increasing temperature.

heat, heat generated by decay of ^{26}Al distributed uniformly throughout the bodies (6.4×10^{-13} J per decay), and the enthalpies of reactions. The energy conservation equation to be solved is

$$\frac{\partial T}{\partial t} = \kappa \left(\frac{\partial^2 T}{\partial r^2} + \frac{2}{r} \frac{\partial T}{\partial r} \right) + (1 - \phi) \frac{Q}{C} + R_{\text{melting}} \frac{\Delta H_{\text{melting}}}{C}$$
$$+ R_{\text{freezing}} \frac{\Delta H_{\text{freezing}}}{C} + R_{\xi} \frac{\Delta H_{\xi}}{C},$$

where T is temperature, t is time, r is radial distance from the centre of the body, ϕ is the porosity of the rock, Q is heat production due to decay of ^{26}Al (W kg^{-1}), C is the rock specific heat, $\Delta H_{\text{melting}}$ is the enthalpy of ice melting, $\Delta H_{\text{freezing}}$ is the enthalpy of water freezing, ΔH_{ξ} is the enthalpy of reaction for the model mineralogical reaction, and R_{ξ} is the rate of mineralogical reaction. Values for Q were obtained using the expression $Q = 6 \times 10^{-3} (^{26}\text{Al}/^{27}\text{Al})_0 \exp(-\lambda t)$, where λ is the decay constant for ^{26}Al, $(^{26}\text{Al}/^{27}\text{Al})_0$ is the ratio at the time the body accretes, and the pre-exponential includes a median chondritic Al concentration.

Advection of heat by flowing water has been ignored but should be included in future calculations.

(d) Oxygen isotope exchange between rock and water

Changes in the oxygen-isotopic composition of carbonaceous chondrite rock were calculated by allowing the rock and fluid to exchange oxygen isotopes at a finite

rate driven by departures from equilibrium. The fluid-isotopic composition evolved in response to continued melting of water ice, advection of liquid water, and the exchange of oxygen isotopes between liquid water and rock. The conservation equations solved for each of the rare oxygen isotopes are of the form

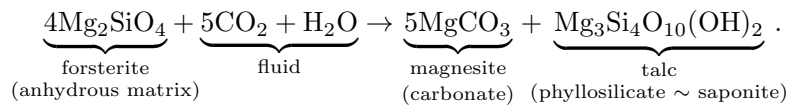$$\frac{\partial C_f}{\partial t} = -\nabla(C_f \upsilon) + C_{f,ice} R_{melting} + C_f R_{freezing} + R_{exchange},$$

where C_f is the concentration of the isotopic species in the fluid phase, $C_{f,ice}$ is the concentration in the water ice, and $R_{exchange}$ is the rate of isotopic exchange between water and rock. The rate of exchange was taken to be

$$R_{exchange} = k_{exchange}((\beta/\alpha)C_s - C_f),$$

where C_s is the concentration in the solid, β is the ratio of oxygens per unit volume in fluid and rock, and α is the equilibrium isotope fractionation factor. A value of $ca. \, 1.0 \times 10^{-13}\,s^{-1}$ was used for the rate constant $k_{exchange}$ in these calculations. This value was chosen because it results in isotopic exchange where new mineral growth is present. Vastly lower values for the rate constant result in mineralogical reactions with no isotopic equilibration, a situation deemed unrealistic. Drastically larger values lead to nearly complete resetting of rocks in the absence of new mineral growth, a scenario also deemed to be unrealistic. An alternative to choosing a fixed rate constant would be to link the rate of exchange directly to the growth of new minerals.

(e) Mineralogical responses to water–rock reaction

In order to monitor the mineralogical effects of aqueous alteration, the carbonaceous chondrite parent bodies were considered to be composed of forsterite olivine. The forsterite was permitted to be converted to secondary hydrous and carbonate minerals by progress of the reaction

$$\underbrace{4Mg_2SiO_4}_{\substack{\text{forsterite} \\ \text{(anhydrous matrix)}}} + \underbrace{5CO_2 + H_2O}_{\text{fluid}} \rightarrow \underbrace{5MgCO_3}_{\substack{\text{magnesite} \\ \text{(carbonate)}}} + \underbrace{Mg_3Si_4O_{10}(OH)_2}_{\text{talc (phyllosilicate} \sim \text{saponite)}} .$$

Talc is used as a convenient analogue for the hydrous minerals (it is a trioctahedral phyllosilicate like the common alteration product saponite). For the isotopic calculations it was assumed that the hydrous minerals were composed of equal parts (in oxygen units) serpentine and saponite, as there is a significant difference between the oxygen-isotopic compositions of these minerals at equilibrium (saponite was assumed to be similar to talc in its affinity for the oxygen isotopes due to structural similarities). Magnesite was used as the analogue for calcite and dolomite in the thermochemical calculations. The isotopic calculations were based on fractionation factors for calcite. The sources for the fractionation factors are given by Young *et al.* (1999).

Progress of the analogue reaction, represented by the reaction progress parameter ξ (mol reaction m^{-3} rock), was determined by the difference between CO_2/H_2O in the aqueous liquid and the values dictated by equilibrium with the rocks at each radial position in the parent bodies. The rate equation is

$$R_\xi = -(5/\phi)k_\xi(-\Delta\mu^{T,*} + RT\ln f_{H_2O} + RT\ln f_{CO_2}).$$

In this equation R_ξ is the rate of mineralogical reaction $(\mathrm{mol}\,(\mathrm{m}^3\,\mathrm{s})^{-1})$, $-\Delta\mu^{T,*}$ is the reaction affinity (i.e. the deviation from a zero change in chemical potentials for the balanced reaction among the pure chemical components at the temperature of interest), k_ξ is the reaction rate constant (s^{-1}), and the fugacities of CO_2 and H_2O, f_{CO_2} and f_{H_2O}, were calculated from the activities of CO_2 and H_2O in the liquid phase at each time-step. The rate constant was taken to be $k_\xi = 2.0/(RT)\exp(-6.3\times10^4/(RT))$. Justification for this phenomenological estimate for the rate constant was given by Young *et al.* (1999). The evolution of fugacities is a result of continued melting of water ice that contains an initial concentration of CO_2 (representing oxidized carbon), advection of the liquid, and the rate of consumption of CO_2 by the formation of carbonate in the rock. The resulting conservation equation for CO_2 in the fluid is

$$\frac{\partial C_{\mathrm{f}}^{CO_2}}{\partial t} = -\nabla(C_{\mathrm{f}}^{CO_2}v) + C_{\mathrm{f,ice}}^{CO_2}R_{\mathrm{melting}} + C_{\mathrm{f}}^{CO_2}R_{\mathrm{freezing}} + R_\xi.$$

The method of relating concentration of CO_2 in the fluid, $C_{\mathrm{f}}^{CO_2}$, to the fugacities was described by Young *et al.* (1999). The resulting rates of reaction are *ca.* 100 times faster than some estimates for the rates of terrestrial weathering reactions.

(f) *Method of solution*

The equations above were solved simultaneously by explicit finite difference. The initial condition for all cases included 0.2 volume fraction water ice, 0.1 volume fraction open pore space, a temperature of 180 K, a bulk ice mole fraction CO_2 of 0.2, rock $\delta^{18}O$ and $\delta^{17}O$ values ($\delta^{18}O$ and $\delta^{17}O$ are the per mil deviations in $^{18}O/^{16}O$ and $^{17}O/^{16}O$ from 'standard mean ocean water') of -3.6 and -4.6, respectively (corresponding to a rock $\Delta^{17}O$ of -2.7, a value equivalent to the matrix of the Allende meteorite, where $\Delta^{17}O$ is the deviation from the terrestrial mass fractionation line along the $\delta^{17}O$ axis in a plot of $\delta^{17}O$ against $\delta^{18}O$), and water $\delta^{18}O$ and $\delta^{17}O$ values of 35.0 and 34.0, respectively (corresponding to a water $\Delta^{17}O$ of 16.0). Ice $\Delta^{17}O$ values substantially greater than rock are consistent with numerous other studies (see, for example, Clayton & Mayeda 1984). The initial values for ice $\delta^{17}O$ and $\delta^{18}O$ are arrived at by iteration for a given isotope exchange rate in order to match the isotopic compositions of carbonaceous chondrite minerals. The outer surfaces of the spherical bodies were maintained at 180 K as a simple and numerically stable approximation to radiative cooling into space.

The original calculations presented by Young *et al.* (1999) have been improved upon by minimizing numerical instabilities and inaccuracies that arose near 'fronts' in the original simulations. Instabilities were removed by 'smearing' the abrupt phase changes involving water. Instead of specifying a single temperature (e.g. 273 K) for the freezing point of water, liquid water was permitted to freeze gradually at a rate controlled by an inverse error function (e.g. $R_{\mathrm{freeze}} = V_{\mathrm{liq}}\,\mathrm{erfc}(0.25(T-245))k_{\mathrm{freezing}}$, where k_{freezing} is a rate constant for freezing measured in s^{-1} such that a 'hard freeze' occurs at 235 K and V_{liq} is the volume of pore-filling liquid at a specified location). In this instance the gradual freezing simulates the presence of 'unfrozen water', a form of intergranular water known to migrate by capillary forces below the freezing point in frozen terrestrial soils (Oliphant *et al.* 1983; Rietmeijer 1985). Similarly, melting of water ice was permitted to occur over a narrow range in temperature (within several degrees of 273 K).
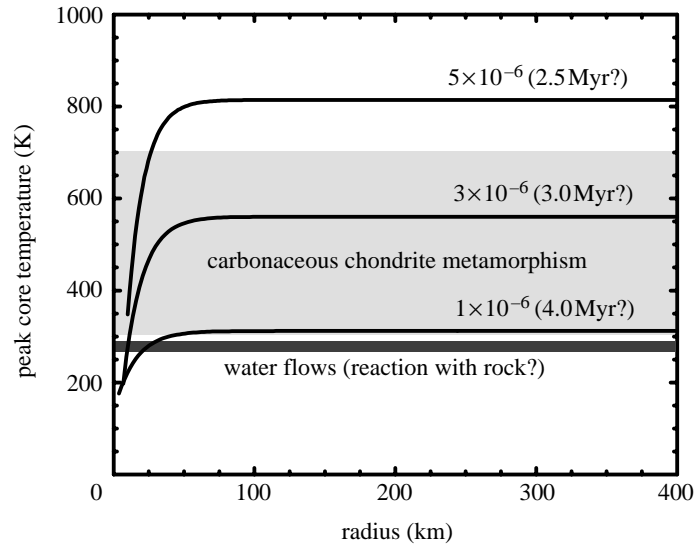
Figure 2. Plot of peak temperature in the core of model planetesimals against radius. The curves correspond to different initial $^{26}Al/^{27}Al$. The labels indicating initial $^{26}Al/^{27}Al$ also show the implied formation ages relative to a canonical initial $^{26}Al/^{27}Al$ of 5.0×10^{-5}. The grey zone represents temperature estimates for metamorphism in carbonaceous chondrites. The horizontal line signifies the temperature of water ice melting and water–rock reaction where it occurs.

3. Results

(a) First-order observations from thermal modelling

Results of even simple thermal models provide first-order checks on hypotheses for the geological evolution of planetesimals assuming they were heated uniformly from within. For example, figure 2 shows that maximum temperatures experienced within carbonaceous chondrite parent bodies depended upon the rate of internal heating (in these models this means initial $^{26}Al/^{27}Al$) but not size if radii exceeded *ca.* 50 km. The flat shapes of the peak T versus radius curves in figure 2 show that the size of a body cannot be deduced from the maximum temperatures recorded in the rocks even if the rate of heat production is well known (a radiation flux boundary condition rather than a fixed temperature boundary condition adds a small positive slope to the curves). The shapes of the curves reflect the fact that bodies with radii of less than 50 km maintain significant temperature gradients throughout, while bodies with radii greater than 50 km have approximately uniform temperatures except near their radiating outer surfaces. As a result, planetesimals with radii less than 50 km provide for a larger range in geological conditions than planetesimals with radii much greater than 50 km (see below).

Figure 2 also shows that icy planetesimals with initial $^{26}Al/^{27}Al$ greater than *ca.* 3×10^{-6} all result in temperatures exceeding the maximum temperatures recorded in carbonaceous chondrites (Krot *et al.* 1995) unless those bodies were smaller than *ca.* 100 km in diameter. This means that a large body like 1 Ceres cannot be composed of carbonaceous chondrites like those in our meteorite collections unless it had bulk initial $^{26}Al/^{27}Al$ approximately one order of magnitude lower than in calcium–

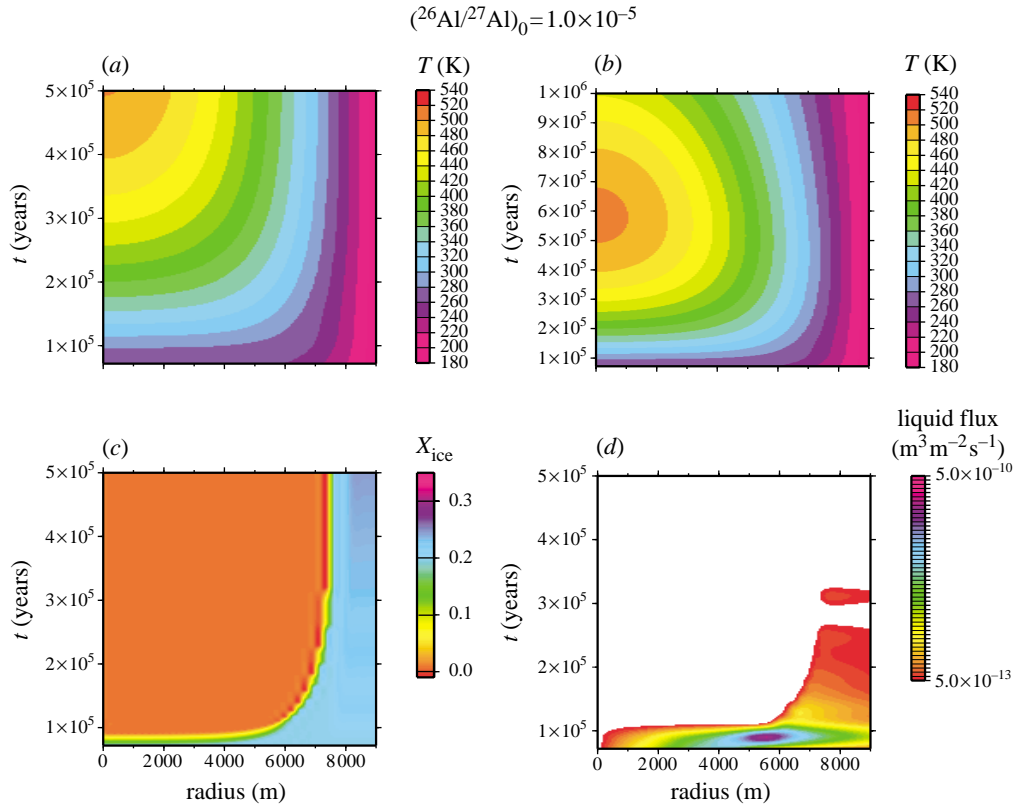$(^{26}\text{Al}/^{27}\text{Al})_0 = 1.0 \times 10^{-5}$

Figure 3. Plots of radius versus time that summarize model calculations for the small body (a radius of 9 km) described in the text. (*a*) The temperature history up to 500 000 yr. (*b*) The temperature history up to 1 Myr. (*c*) The distribution of water ice in volume fraction. The initial water ice fraction was 0.2 throughout the body. (*d*) The flux of liquid water with time and radial position. Note the episodic nature of the flux at radial positions beyond *ca.* 7 km.

aluminum-rich inclusions (CAIs, the oldest known objects of Solar System origin). Assuming that the canonical CAI initial $^{26}\text{Al}/^{27}\text{Al}$ of 5.0×10^{-5} applies for the solar nebula as a whole, the implication is that a body like 1 Ceres must have formed more than 3.0 Myr after the formation of CAIs if it is indeed composed of typical carbonaceous chondrite.

All icy bodies exhaled water from their interiors towards their surfaces once their interiors warmed to *ca.* 273 K, but the timing of exhalation, and the extent to which water and host rock reacted with one another, was controlled by size and heat-production rate, as shown below.

(*b*) *Geological evolution of small icy planetesimals*

Small in this context means those bodies with radii less than the thermal saturation cut-off of *ca.* 50 km shown in figure 2. Results for a body with a radius of 9 km and an initial $^{26}\text{Al}/^{27}\text{Al}$ value of 1×10^{-5} are shown in figures 3, 4 and 5. The initial $^{26}\text{Al}/^{27}\text{Al}$ corresponds to an accretion age of 1.6 Myr after the formation of CAIs, based on a canonical initial $^{26}\text{Al}/^{27}\text{Al}$ value of 5.0×10^{-5}.

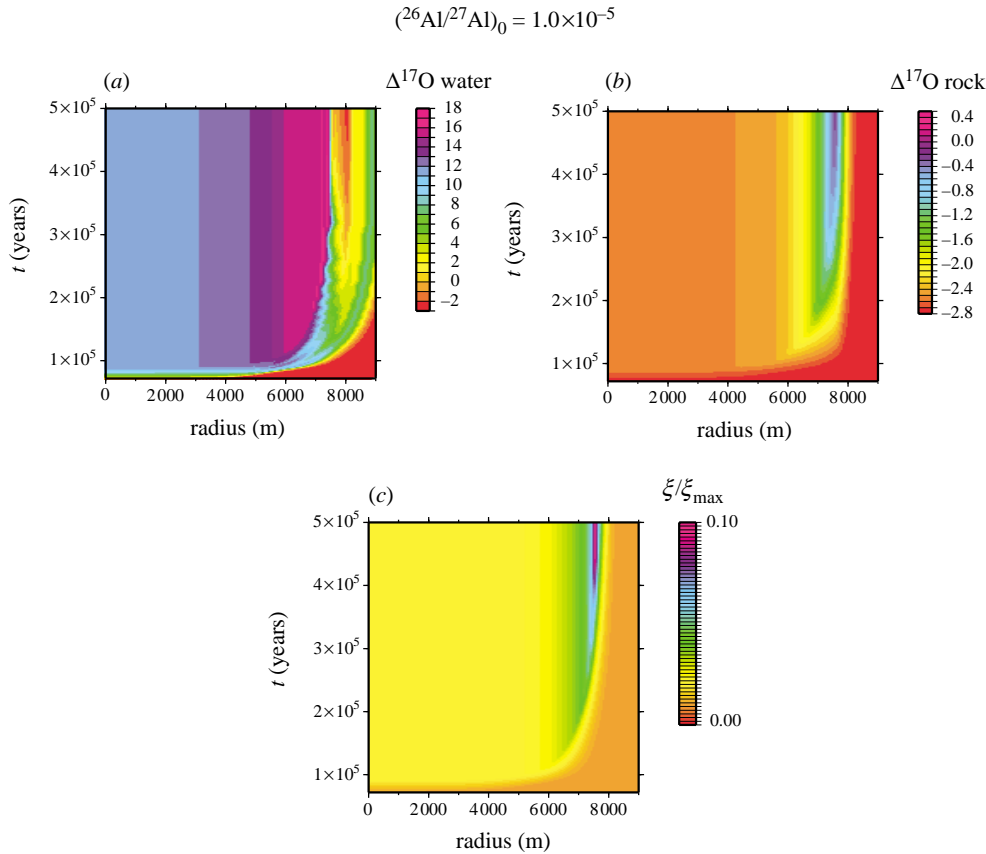$(^{26}\mathrm{Al}/^{27}\mathrm{Al})_0 = 1.0\times10^{-5}$

Figure 4. Plots of time versus radius for the small-body (a radius of 9 km) model described in the text showing the oxygen-isotopic and mineralogical evolution of the body. (*a*) Changes in $\Delta^{17}\mathrm{O}$ in the liquid water with time and position in the body. The initial rock $\Delta^{17}\mathrm{O}$ was -2.7‰ and the initial water $\Delta^{17}\mathrm{O}$ was 16.0. (*b*) The evolution of rock $\Delta^{17}\mathrm{O}$. Note the narrow zone with $\Delta^{17}\mathrm{O}$ values near 0‰. This zone represents the radial position where CI carbonaceous chondrites may have formed. (*c*) The progress of the model mineralogical reaction relative to the maximum progress in mol. Note that the zone of maximum mineralogical alteration coincides with the zone of maximum shift in $\Delta^{17}\mathrm{O}$.

The small body experiences aqueous alteration within the first several hundred thousand years of heating (figure 3). Ice is lost from the interior of the body while excess ice eventually accumulates near the surface (figure 3). Because the mass of ice is least in the centre, ice is rapidly lost near the centre. The melted ice deeper in the body contributes to the flux of liquid further out towards the surface, with the result that a zone of protracted and intense fluid flow occurs between the centre and the frozen outer surface. This zone of prolonged fluid flow gives rise to a region in which the carbonaceous chondrite rock is intensely altered. The alteration is characterized by an increase in $\Delta^{17}\mathrm{O}$ and growth of carbonate and hydrous minerals (figure 4) and resembles the CI chondrites. The range in oxygen isotope ratios and their correlation with mineralogy mimics that in the major carbonaceous chondrite meteorite groups including CVs, CMs and CIs (figure 5).
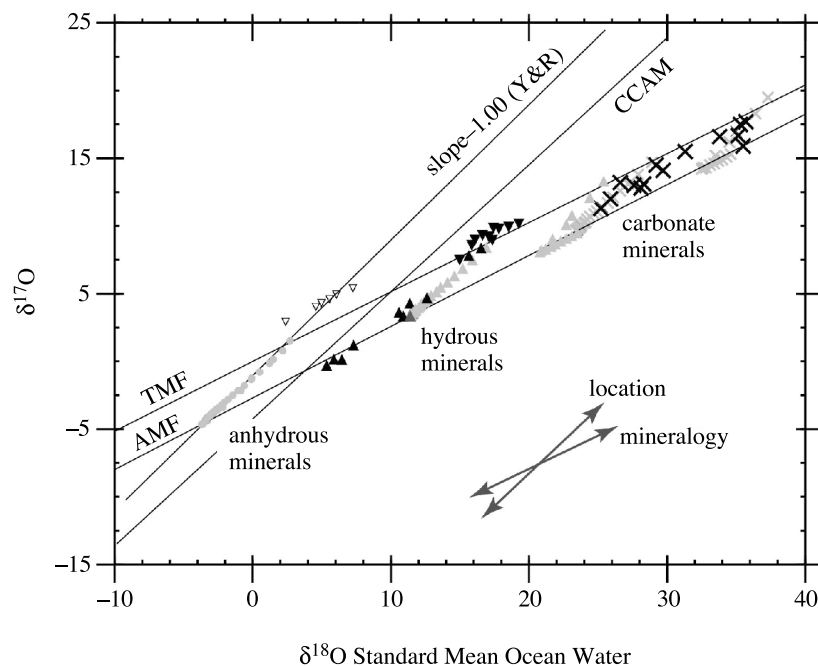
Figure 5. Three-isotope plot for oxygen comparing the model carbonaceous chondrite oxygen isotope ratios (grey symbols) with published measured values (black and open symbols) ($\triangledown$, CI magnetites; $\blacktriangle$, CM hydrous phases; $\blacktriangledown$, CI hydrous phases; $\times$, CM carbonates). These measured values are to be compared with the model anhydrous phases (grey circles), model hydrous phases (grey triangles) and model carbonates (grey crosses). Each model datum represents a single radial position in the parent body following cessation of alteration. The correspondence between the measured data and the model results suggests that the oxygen isotope systematics of carbonaceous chondrites can be explained as the result of sampling different positions in hydrological flow systems (steep vector) and different mineralogies (shallow vector). The measured values were taken from Clayton & Mayeda (1984), Rowe *et al.* (1994) and Benedix *et al.* (2000). Also shown are the terrestrial mass-dependent fractionation line (TMF), the Allende mass-dependent fractionation line (AMF, see Young *et al.* (1999)), the carbonaceous chondrite anhydrous mineral line (CCAM, see Clayton *et al.* (1977)), and the slope-1.00 line (Y&R, see Young & Russell (1998)).

The striking similarity between the model results and the measured oxygen isotope ratios in carbonaceous chondrite minerals suggests that the different types of carbonaceous chondrite meteorites differ in mineralogy and oxygen isotope ratios as a result of having come from different parts of hydrological flow systems like the one modelled here (Young *et al.* 1999). It does not follow that the carbonaceous chondrites all derive from a single parent body, but only that the processes that controlled the geological evolution of the parent bodies of each of the carbonaceous chondrite meteorites could have been similar.

Some carbonaceous chondrites are thought to have suffered metamorphism due to heating to as much as 670 K (Krot *et al.* 1995). Inspection of the temperature contours in figure 3 shows that the metamorphosed carbonaceous chondrites could also have formed in the same bodies as those exposed to aqueous alteration and minimal heating. This is because small bodies heated from within but cooled radiatively at

their surfaces sustain large temperature gradients for hundreds of thousands of years. The models also predict that it is not possible for the same rock to experience intense aqueous alteration and higher-temperature metamorphism (cf. Brearley 1999). This is because temperatures never rise much above the melting point of water ice in the outer portions of the body where alteration is most intense.

The important features of the evolution of small icy bodies is the short time-span associated with their geological evolution (less than 1 Myr), and the presence of protracted temperature gradients that permit coexistence of metamorphosed rocks and aqueously altered rocks.

(c) Geological evolution of large icy planetesimals

Large in this context is defined as objects with radii well beyond the thermal cut-off in figure 2, where maximum core temperatures are insensitive to size. A model for a body the size of asteroid 1 Ceres serves as an illustration of the extreme case. The goal was to explore under what conditions such a body could produce carbonaceous chondrites like those observed.

Like the smaller bodies, large bodies also develop localized zones of alteration. However, if the rate of heating is too high, water exhales from the bodies before substantial alteration can occur. Conversely, if the rate of heating is too low, water ice never melts and there can be no water–rock reaction. In general, heating rates sufficient to melt water ice produce alteration zones within several kilometres of the surface of the large bodies.

Figure 6 shows results for a model based on a radius of 468 km (the size of 1 Ceres) and an initial $^{26}Al/^{27}Al$ value of 6.8×10^{-7}, corresponding to an accretion age of 4.4 Myr relative to CAIs. The maximum temperatures experienced by such a body will not have been much higher than the melting temperature of water. Alteration with isotopic and mineralogical features resembling CI chondrites occurs near the surface, while most of the body experiences a very low level of alteration, comparable, perhaps, with the least altered of the CM chondrites, or perhaps with some of the more altered CV meteorites. The isotopic characteristics of the minerals produced in this model also span the major carbonaceous chondrite groups, as in figure 5.

Alteration is expected to have occurred much later for large bodies relative to the smaller bodies based on the model calculations. In the case of the 1 Ceres-sized object with low initial $^{26}Al/^{27}Al$, alteration begins 5–6 Myr after commencement of heating. This is a general result; alteration in large bodies occurs after several Myr and is concentrated near the surface. Metamorphism never occurs in the large bodies with low initial $^{26}Al/^{27}Al$ (figure 6). If heat production corresponded to an initial $^{26}Al/^{27}Al$ value less than 6×10^{-7}, there would be insufficient heat to melt water ice and the bodies would lie dormant.

Increasing heat production in a large body like that in figure 6 will produce temperatures high enough to account for metamorphism of carbonaceous chondrites. However, with heat production much higher than that in figure 6, water is expelled with such speed that no water–rock reaction can take place. This is shown in figure 7. A 1 Ceres-sized object with an initial $^{26}Al/^{27}Al$ of 5.0×10^{-6} (corresponding to a formation age of 2.35 Myr post CAI) exhales water with little or no water–rock reaction (figure 7).
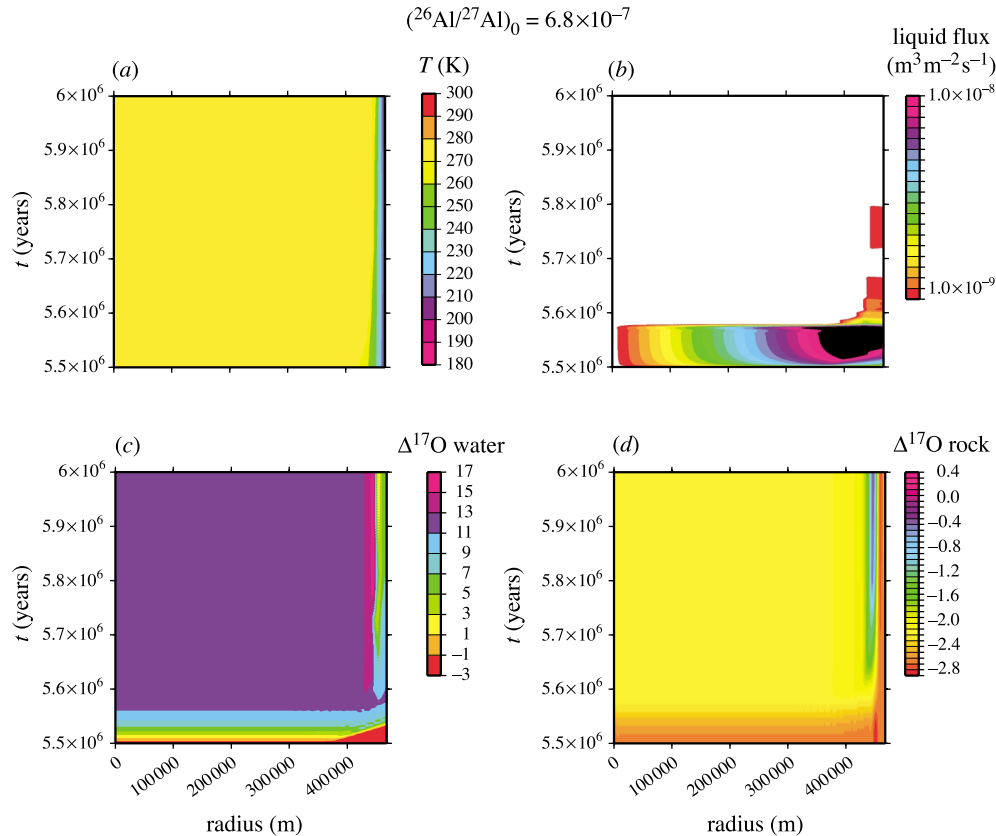
Figure 6. Time versus radius for a model planetesimal the size of asteroid 1 Ceres with low rates of heat production, as described in the text. Part (*a*) shows that the body stays cool throughout its history. (*b*) The distribution of fluid flux with position and time. (*c*) The evolution of $\Delta^{17}O$ of the liquid water with time. (*d*) The distribution of rock $\Delta^{17}O$ with time and position. Note that a zone of alteration forms near the surface. The oxygen isotope and mineralogical characteristics of this zone are similar to the alteration zone shown for the small body in figure 4, and the oxygen isotope ratios for the individual minerals are similar to those in figure 5. Note also that higher-temperature metamorphism would not occur anywhere in this body.

4. Discussion

The full range of possibilities is yet to be explored, but the few results shown here demonstrate that thermal models, combined with the constraints imposed by the mineral-specific oxygen isotope ratios in carbonaceous chondrites, can be used to place limits on the types of planetesimals represented in our meteorite collections.

If aqueous alteration occurred in large bodies like asteroid 1 Ceres, metamorphosed carbonaceous chondrites would have to have formed in different, hotter parent bodies. If the hotter parent bodies were also large, they would have been devoid of aqueous alteration because water would have been expelled too rapidly (figure 7). On the other hand, if the hotter bodies were small (radii less than 50 km), then aqueous alteration is predicted to have been present in the same planetesimals in which the metamorphosed rocks formed (figure 4).
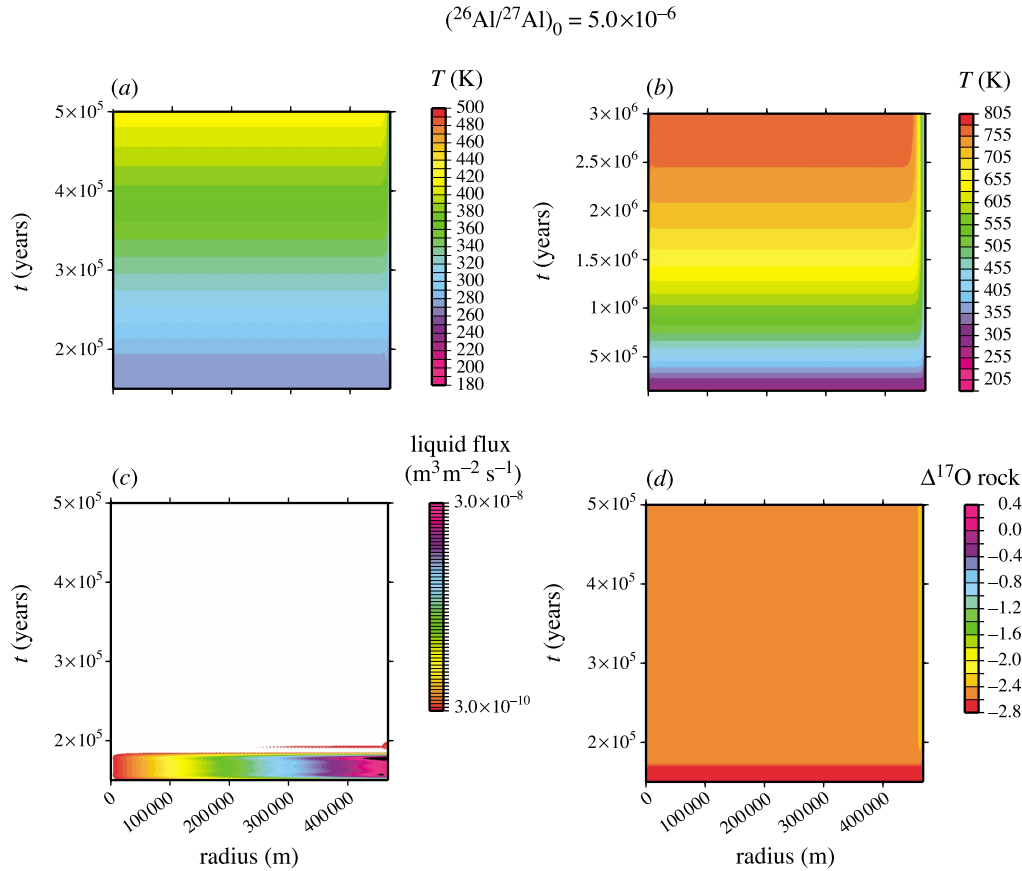
$(^{26}Al/^{27}Al)_0 = 5.0 \times 10^{-6}$

Figure 7. Time versus radius for a model planetesimal the size of asteroid 1 Ceres with a higher rate of heat production, as described in the text. (*a*), (*b*) The temperature history of the body. (*c*) The distribution of fluid flux with time and distance. Note that fluid is expelled early. (*d*) The distribution of rock $\Delta^{17}O$ with time and position. The rock $\Delta^{17}O$ remains nearly constant everywhere with time, showing that there is no zone of intense alteration like that observed in the small-body model (figure 4) or in the cooler large-body model (figure 6). This implies that large bodies capable of producing metamorphism in carbonaceous chondrites cannot be the settings in which carbonaceous chondrites exhibiting the greatest degree of aqueous alteration formed.

Aqueous alteration and metamorphism are predicted to have occurred in less than 1 Myr following planetesimal accretion if carbonaceous chondrite parent bodies had diameters of the order of 20 km. Conversely, aqueous alteration is predicted to have been delayed for *ca.* 6 Myr relative to the time of accretion if the carbonaceous chondrite parent bodies were the size of 1 Ceres (diameters approaching 900 km). Metamorphism of carbonaceous chondrites should have occurred within 2 Myr of accretion if the parent bodies of the metamorphosed samples were the size of 1 Ceres.

Hutcheon *et al.* (1999) report ^{53}Mn–^{53}Cr chronometry data which suggest that aqueous alteration in carbonaceous chondrites occurred over a period of at least 8 Myr in the early Solar System. At face value, such a protracted period of hydrological activity supports the hypothesis that the parent bodies within which alteration

occurred were large (diameters of hundreds of kilometres). Chromium isotope ratio analyses of carbonate indicate that carbonate from the highly altered CI meteorite Orgueil formed 6 Myr after carbonates from the less altered CM carbonates (an age difference of *ca.* 6 Myr). This difference in ages is consistent with diachronous evolution in large bodies but is inconsistent with the rapid evolution of small bodies. Further studies, perhaps involving more short-lived nuclides, are required if this conclusion is to be adopted as evidence that some carbonaceous chondrite parent bodies were hundreds of kilometres in diameter.

5. Conclusions

Oxygen isotope ratios, mineralogy and thermal modelling suggest that all carbonaceous chondrites could represent pieces of small icy planetesimals that evolved geologically in less than 1 Myr. The alternative is that carbonaceous chondrites represent larger bodies like 1 Ceres and evolved diachronously over several million years. In the latter case each of the distinct carbonaceous chondrite groups would have to represent a parent body, or class of parent bodies, with significantly different rates of internal heat production (i.e. different abundances of ^{26}Al, the primary heat-producing nuclide). Heterogeneity in heat production implies formation of parent bodies over an interval of several million years or heterogeneity in the distribution of ^{26}Al.

These predictions can be tested if it is possible to date geological events in carbonaceous chondrites with a relative precision of the order of 1 Myr. Ambiguity surrounding the origin of some of the short-lived nuclides (especially ^{26}Al, see other papers in this issue) represents a setback in this regard. Nevertheless, thermal models capable of reproducing the salient features of the geological evolution of the carbonaceous chondrites provide a means for evaluating the validity of hypotheses for the formation of these primitive meteorites.

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Discussion

J. C. BRIDGES (*Natural History Museum, London, UK*). The work done by us on ordinary chondrites and you on carbonaceous chondrites has clearly shown the significance of low-temperature aqueous processes in controlling the isotopic composition of oxygen in chondrites. I think this is one of the issues that has come out of this meeting. The next step is to determine the fractionation processes that created the ^{16}O-poor aqueous reservoirs and ^{16}O-rich minerals seen in chondrites.

E. D. YOUNG. Some of the differences in ^{16}O content (Δ^{17}O) among minerals arise from reactions between ^{16}O-poor water and ^{16}O-rich rock. This is evident in the models where Δ^{17}O is seen to rise, with alteration in some regions of the planetesimals.

R. ASH (*Department of Earth Sciences, University of Oxford, UK*). If there is a continuum of alteration degree upon the carbonaceous chondrite parent body, as suggested in this talk, why do we have distinct groups of meteorites in terms of petrology and mineralogy as well as O isotopes. There are some meteorites which may be intermediates between types but these are rare.

E. D. YOUNG. The pattern of alteration observed in the models is one of abrupt changes in mineralogy and oxygen isotope ratios (especially Δ^{17}O) with radial position. Random sampling of such bodies would tend to yield populations of altered rocks and populations of unaltered rocks with relatively few samples coming from the boundaries intermediate between altered and unaltered areas.

C. ALLÈGRE (*Institut de Physique du Globe de Paris, Paris, France*). Have you measured D/H? Have you looked at fluid inclusion?

E. D. Young. We have not examined the consequences of our model for D/H nor fluid inclusions. This should be done.

P. Cassen (*NASA Ames Research Center, CA, USA*). In the models, how do you decide how much heat is carried by the fluid compared with that conducted?

Are not the hydration reactions exothermic and do they not cause runaway heating?

E. D. Young. Advective heat transport does shift temperature contours, but calculations thus far show that the shifts in time and space are not large for fluid exhalation. Enthalpies of exothermic reactions are potentially important, but the rates of reactions at low temperatures are apparently low, with the result that the heat is released slowly (the rate of heat production by exothermic reaction amounts to about $0.1\,\mathrm{K}$ per year in the models shown here) in comparison with the rates of heat transport.