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Fluid flow and chemical alteration in carbonaceous chondrite parent bodies

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ABSTRACT

Carbonaceous chondrites are a class of primitive meteorites that likely represent early solar system materials, thus providing critical insights about planetary evolution. However, evidence shows that they have undergone considerable processing. Elucidating the complex formation history of aqueously altered chondrite groups has been a principal objective of cosmochemistry. Fulfilling this objective requires understanding the nature of the water–rock interactions involved. Many current alteration models are generally founded upon an assumption of uniform alteration in a closed system. In these models, different chondrite groups must originate on separate parent bodies. Contrary to this common view, we show that chondrite data are consistent with non-uniform alteration in a convecting system so that multiple chondrite groups can form within a single body.

1. Introduction

Carbonaceous chondrites are a suite of primitive meteorites. However, despite their primitive characteristics, mineralogical and isotopic data indicate that most experienced various degrees of alteration in their parent bodies (Kerridge and Bunch, 1979; McSween, 1979; Bunch and Chang, 1980; Zolensky and McSween, 1988; Rowe et al., 1994). In particular, the compositions of the CI and CM chondrite groups provide abundant evidence of processing by aqueous alteration (Clayton, 1993). Detailed studies of these chondrites reveal a complex formation history involving periods of mineral dissolution, precipitation, and oxygen isotope exchange due to interactions with fluids (Johnson and Prinz, 1993; Riciputi et al., 1994; Zolensky et al., 2008). Consequently, the primary mineralogy and oxygen isotopic compositions of the chondrites have been erased by aqueous alteration in the parent body (Richardson, 1978; McSween, 1979; Bunch and Chang, 1980; Endress and Bischoff, 1993, 1996; Endress et al., 1996; Leshin et al., 1997).

Our current understanding of parent body processing is largely derived from carbonaceous chondrite samples. However, there are a number of challenges to deciphering the information provided by these samples (McSween and Richardson, 1977; McSween, 1987; Zolensky et al., 1987; Rowe et al., 1994; Zolensky et al., 1997; Clayton and Mayeda, 1999a; Young et al., 1999; Mittlefehldt, 2002; Airieau et al., 2005; Brandon et al., 2005). Controversy over the nature of water–rock interactions on the parent bodies of these primitive meteorites further complicates the interpretation of carbonaceous chondrite data (Young et al., 2003). An ongoing debate has been whether fluid flow in a parent body can be reconciled with mineralogical and isotopic data from CI and CM chondrites. The implications for a parent body vary notably between fluid flow and static pore water models (Clayton and Mayeda, 1984; Grimm and McSween, 1989; Young et al., 1999). In a convecting body, reaction kinetics, temperature, fluid flow, permeability, and rock and fluid compositions depend sensitively upon each other (Norton and Knapp, 1977; Curewitz and Karson, 1997; Hulen and Lutz, 1999; Durst, 2002; Tenthorey and Gerald, 2006). The feedback among these processes produces a heterogeneously altered body. In this case different groups of carbonaceous chondrites could be derived from distinct locations within the same parent body. Conversely, static pore water results in homogeneous alteration of the parent body and different groups of chondrites cannot share the same parent body. Although the primitive bulk chemistry of carbonaceous chondrites has been cited (McSween, 1979) to argue in favor of a static pore water model, convective flow appears more consistent with isotopic water–rock ratios and heat-loss constraints (Grimm and McSween, 1989; Grimm, 2007). Correctly reconstructing the origin of CI and CM chondrites requires determining the feasibility of convective flow models.

Despite the importance of determining whether a convective flow model can be reconciled with chondritic mineralogical and isotopic data, there have been no studies of asteroidal bodies that combine fluid circulation with mineral alteration and oxygen isotope exchange (Clayton and Mayeda, 1984; Grimm and McSween, 1989; Young et al., 1999). Here, we present a new numerical model of hydrothermal

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circulation in a carbonaceous chondrite parent body that tracks heat and fluid transport, geochemical reactions, and isotope exchange. Out of necessity, the geochemical reactions have been simplified (see discussion in Section 3) and therefore do not reproduce the complex mineralogy observed in carbonaceous chondrites. Nevertheless, our combined approach allows us to interpret the degree of mineral alteration and isotopic composition in the context of a convective flow model. A comparison of our results to mineralogical and isotopic data from CI and CM chondrites demonstrates that a fluid circulation model can be consistent with these data. Furthermore, we establish a potential genetic link between different groups of carbonaceous chondrites as well as provide constraints on temperatures of alteration, initial compositions, and timing of aqueous activity. Our work also predicts that, as additional samples are obtained and analyzed, more carbonaceous chondrites containing traits intermediate between different groups or belonging to multiple groups will be found.

2. Numerical approach

2.1. Flow and transport modeling

The hydrothermal evolution of a parent body is determined by solving the conservation equations for mass, energy, and momentum. The equation for the conservation of fluid momentum in porous media is derived from Darcy's Law. The Darcy velocity, $\underline{u}$, can be expressed as

$$\underline{u} = -\frac{k}{\mu}\left(\nabla p - \rho_{\mathrm{w}}\,\underline{g}\right) \tag{1}$$

where k is permeability, μ is the viscosity of liquid water, ρ_{w} is the density of liquid water, p is the water pressure, and g is the acceleration of gravity. Gravity in a spherical (parent) body depends on depth according to (Travis and Schubert, 2005)

$$\underline{g} = -\gamma\,\underline{r}_{\mathrm{b}} \tag{2}$$

where $\underline{r}_{\mathrm{b}}$ is the radius vector from the center of the body and γ is a gravity parameter written as

$$\gamma = \frac{4}{3}\pi\rho_{\mathrm{b}}G \tag{3}$$

with ρ_{b} the bulk density and G the gravitational constant. The bulk density is given by

$$\rho_{\mathrm{b}} = \phi[\rho_{\mathrm{w}}\sigma + \rho_{\mathrm{ice}}(1-\sigma)] + (1-\phi)\rho_{\mathrm{r}} \tag{4}$$

where ϕ is porosity, ρ_{r} is the density of rock, ρ_{ice} is the density of ice, and σ is the fraction of pore space occupied by liquid water.

The conservation of fluid mass is written as

$$\nabla\cdot(\rho_{\mathrm{w}}\,\underline{u}) = 0. \tag{5}$$

Eq. (5) has been derived using the common assumption that both the fluid and rock matrix are essentially incompressible. The expressions for Darcy velocity (Eq. (1)) and gravity (Eq. (2)) can be substituted into Eq. (5) to obtain the expression for fluid flow in porous media

$$\nabla\cdot\left(\frac{k}{\mu}\rho_{\mathrm{w}}\nabla p\right) = -\frac{1}{r_b^2}\frac{\partial}{\partial r}\left(\frac{k}{\mu}\gamma\rho_{\mathrm{w}}^2 r_{\mathrm{b}}^3\right). \tag{6}$$

The equation for the mass conservation of a chemical solute that includes both transport and chemical reactions in porous media is (Bear, 1979)

$$\frac{\partial(\phi\rho_{\mathrm{w}}C_i)}{\partial t} + \nabla\cdot(\underline{u}\rho_{\mathrm{w}}C_i) = \nabla\cdot(\phi\sigma\rho_{\mathrm{f}}D\nabla(n_i)) + R_i \tag{7}$$

where C_i is the mass concentration of some species i in the fluid, D is the combined dispersion–diffusion coefficient, ρ_{f} is the fluid density (water plus solutes), n_i is the molar concentration of some species i, and R_i is the total reaction rate of species i. The first and second terms on the left represent the rate of change and advection of solute mass, respectively. The right side of the equation contains the transport of mass by dispersion/diffusion and the generation of a species due to the mineral–fluid reactions. The total rate at which the concentration of species i changes is related to the rate of precipitation or dissolution of mineral j, r_j, and the stoichiometric coefficient, ν_{ij}, by

$$R_i = -\sum_{j=1}^{N}\nu_{ij}r_j \tag{8}$$

where N is the number of minerals. Following from transition state theory, a general form of the rate law expressing mineral dissolution and precipitation is (Appelo and Postma, 2005)

$$r_j = A_j k_j\left(\frac{m}{m_0}\right)_j^n\left(1-\Omega_j^{\beta}\right) \tag{9}$$

where m_0 is the initial moles of solid, m is the moles of solid at a given time, Ω_j is the saturation ratio, k_j is the rate constant, and A_j is the reactive surface area of mineral j. The coefficient β is related to the stoichiometry of the reaction when an activated complex is formed and is typically equal to 1.0 (Aagaard and Helgeson, 1982; Delany et al., 1986). The mole ratio, $(m/m_0)^n$, accounts for changes in surface area, where n is 2/3 for dissolving or growing spheres (Appelo and Postma, 2005).

The oxygen isotopic evolution of rock and water is controlled by both the mechanisms and rates of isotopic exchange and transport. Laboratory measurements indicate that the major spatial transport of oxygen isotopes is by the moving fluid (i.e., advection) and other means of isotope transport are assumed negligible (Norton and Taylor, 1979; McKibbin and Absar, 1989). The conservation equation for oxygen isotopic species follows directly from Eq. (7) by replacing C_i with an expression for the isotopic composition (Norton and Taylor, 1979)

$$\frac{\partial}{\partial t}(\phi\rho\eta\delta O)_{\mathrm{w}} + \sum_{j=1}^{N}\frac{\partial}{\partial t}[\phi\rho\eta\delta O]_j = -\nabla\cdot(\underline{u}\rho\eta\delta O)_{\mathrm{w}} \tag{10}$$

where η is the mass fraction of exchangeable oxygen in water or mineral j. In order to describe the mineral composition as a function of water–rock oxygen isotope exchange, Eq. (10) is combined with a first-order rate law (Baumgartner and Valley, 2001; Brikowski et al., 2001). The values A, B, and C are experimentally determined constants (Faure and Mensing, 2005).

The equation for the conservation of energy is

$$\begin{aligned}&\frac{\partial}{\partial t}[(1-\phi)\rho_{\mathrm{r}}c_{\mathrm{r}}T + \phi\sigma\rho_{\mathrm{w}}I_{\mathrm{w}} + \phi(1-\sigma)\rho_{\mathrm{ice}}(c_{\mathrm{ice}}T + Q_{\mathrm{ice}})] + \nabla\cdot(\underline{u}\rho_{\mathrm{w}}E_{\mathrm{w}})\\&\quad = \nabla\cdot(\lambda_T\nabla T) + (1-\phi)\rho_{\mathrm{r}}\frac{fJ}{\tau}e^{-t/\tau} + \sum_{j=1}^{N}E_{\mathrm{rxn},j}r_j\end{aligned} \tag{11}$$

where c_{r} is the isobaric specific heat of rock, c_{ice} is the isobaric specific heat of ice, T is temperature, I_{w} is the internal energy of water, E_{w} is the enthalpy of liquid water, λ_T is the temperature-dependent volume averaged thermal conductivity, Q_{ice} is the latent heat of melting/freezing ice, f is the mass fraction of rock that is radiogenic, J is the energy release through decay of unit mass of radiogenic material, τ is the mean lifetime of the radiogenic element, and $E_{\mathrm{rxn},j}$ is the enthalpy of reaction. The heat of reaction for hydration reactions is also included in Eq. (11) (typically ~200 kJ kg^{-1}; Robie and Hemmingway, 1995).