6

Silicate Melting and Volatile Loss During Differentiation in Planetesimals

ROGER R. FU, EDWARD D. YOUNG, RICHARD C. GREENWOOD, AND LINDA T. ELKINS-TANTON

6.1 Introduction

Planetesimals ranging between tens and hundreds of kilometers in size populated the inner solar system during its first several million years (Wood, 1964; Tang and Dauphas, 2012). Direct observations of asteroids and the meteoritic record provide abundant evidence that radiogenic heating of such bodies drove the production of aqueous fluids from ice and, for some objects, the melting of metals and silicates (Castillo-Rogez and Young, Chapter 5, this volume; Kleine and Wadhwa, Chapter 11, this volume).

Progressive internal heating led to a diverse array of physical and chemical changes in the host planetesimal. The potential for pore fluids to migrate over length scales greater than hand-sample size potentially led to the net transport of labile elements and isotopes in meteoritic samples (Young *et al.*, 1999; Young *et al.*, 2003; Bland *et al.*, 2009; Palguta *et al.*, 2010). Subsequent advection of silicate melts may have been an efficient mode of heat transport and holds strong implications for the survival of chondritic material in both meteorites and modern asteroid surfaces (Wilson and Keil, 2012; Fu and Elkins-Tanton, 2014; Neumann *et al.*, 2014). Understanding the prevalence and effects of both volatile and silicate melt migration during radiogenic heating is therefore key to interpreting both meteoritic and remote-sensing data.

Despite its importance, the extent of both volatile and silicate melt migration remains poorly constrained. Although the mineralogy of primitive chondrites implies the past existence of aqueous fluids, predictions about the mobility of such fluids must cope with large uncertainties in the material permeabilities, the interplay of forces driving fluid migration, and the unknown effect of large-scale fracturing. Because the volatile content of silicate melt strongly affects buoyancy, these uncertainties regarding aqueous fluid flow are propagated into models of silicate melt mobility.

In this chapter, we consider in chronological sequence the behavior of volatiles and, ultimately, silicate melts during progressive internal heating of planetesimals. We review the expected thermal and volatile budget of early-forming planetesimals (Section 6.2) and summarize theoretical and observational constraints on the transport and evolution of aqueous fluids (Section 6.3), emphasizing the potential implications of fluid–rock reactions and transport for the interpretation of parent-body and nebular processes based on meteorite observations. After reviewing the behavior of silicate melts on planetesimals (Section 6.4), we discuss how this sequence of events informs our interpretations of meteoritic and remote sensing observations (Section 6.5).

6.2 Radiogenic Heating and Volatile Contents of Early Planetesimals

Extinct radioactive isotopes that may have contributed to heat generation in early-forming bodies include ^{26}Al, ^{60}Fe, ^{53}Mn, ^{146}Sm, and ^{182}Hf (Urey, 1955; Lee *et al.*, 1976). In addition, the most important producers of radiogenic heat today, ^{238}U, ^{235}U, ^{40}K, and ^{232}Th, were more abundant early in the solar system. Aluminum, however, dominates the heat budget. With a half-life of 717 000 years and at the canonical level of $5.23 \pm 0.13 \times 10^{-5}$ ^{26}Al/^{27}Al (Jacobsen *et al.*, 2008), the energy output of decaying ^{26}Al is sufficient to cause silicate melting in planetesimals that accrete to more than ~20 km radius in less than ~2 Myr after calcium–aluminum inclusions (CAIs), assuming thermal conductivities corresponding to intact rock (Elkins-Tanton *et al.*, 2011). We note, however, that porosity may lower dramatically the minimum radius necessary for silicate melting and initial ^{26}Al contents of planetesimals may have also varied, implying a range of required sizes and accretion times to achieve differentiation (Schiller *et al.*, 2015; Henke *et al.*, 2012).

Some early-accreting planetesimals also contained significant quantities of H_2O. CM and CI chondrites contain between 9 and 17 wt% water (Jarosewich, 1990). Although most ordinary chondrites today contain far less water, the presence of hydrated phases in very low metamorphic grade samples indicates that their parent bodies accreted with a higher water content (Alexander *et al.*, 1989). Although other volatiles such as CO and CO_2 may be abundant in comets (Ootsubo *et al.*, 2012), their lower condensation temperatures imply that H_2O dominated the volatile budgets of bodies forming near the location of the modern asteroid belt (Öberg *et al.*, 2011).

6.3 The Fate of Volatiles

The mobility of aqueous fluids in planetesimals is potentially an important mechanism for the advection of both dissolved materials and heat (Grimm and

McSween, 1989). In addition, the concentration of volatiles retained in the interior of intensely heated planetesimals strongly affects the buoyancy of any silicate melts ultimately produced.

Upon the melting of accreted ice to form interstitial aqueous fluids, the density contrast between aqueous fluids and the surrounding rocky matrix may lead to compaction of pore space and fluid ascent. In this scenario, the velocity of fluid migration (v_{Darcy}) may be described by the classic Darcy flow expression (Turcotte and Schubert, 2002):

$$v_{Darcy} = \frac{k\, g(\rho_{rock} - \rho_{fluid})}{\mu\phi} \tag{6.1}$$

where k is the permeability, g is the gravitational acceleration, μ is the dynamic viscosity, Φ is the porosity, and ρ_{rock} and ρ_{fluid} are the densities of the rocky matrix and the pore fluids, respectively. Flow described by Eq. (6.1) is driven by compaction of the denser silicate matrix, which expels pore fluids towards the surface. Given the rates of flow predicted by Eq. (6.1), this type of flow is likely the most efficient mechanism for devolatilizing the interiors of bodies where the initial fluid volume fraction exceeds the final pore-space volume (see below for discussion of fluid and pore-space volumes). For interior fluids to migrate to the surface of ~100-km-radius bodies in ~1 million year (Myr) timescales, the flow rate must be greater than ~10 cm per year. In a body with $g \approx 0.05$ m s^{-2} and a porosity of 0.1, fluids near the H_2O critical point would achieve this flow velocity if the permeability of the rock matrix exceeded 10^{-17} to 10^{-16} m^2 (Fu and Elkins-Tanton, 2014). Bodies with radius much smaller than 100 km (e.g. ~40 km for a dry olivine rheology) may not compact viscously over the timescales of radiogenic heating. For such objects, the velocity given in Eq. (6.1) represents an overestimate, implying that the interior may not be efficiently devolatilized via compaction-driven flow alone.

In addition to Darcy flow, surface tension constitutes another possible driver for fluid migration in the microgravity environments relevant to small bodies (Taitel and Witte, 1996). As an illustration, the height (h) of a column of water held by surface tension acting against gravity can be written as:

$$h = \frac{2\ \ \gamma\ \ cos\,\theta}{r\ \ \ g\ \ \rho_{fluid}} \tag{6.2}$$

where γ is the temperature-dependent surface tension (0.073 N m^{-1} for water at room temperature), θ is the wetting angle (approximately zero for wetted surfaces), and r is the radius of the column of fluid. In a planetesimal with the parameters given above, water can stand ~70 m given pore-space diameters of ~100 μm. The implication is that imbibition of water by porous material away from the site of ice melting can occur on at least this scale given sufficient wetting of surfaces.

The scale of capillary flow is augmented where temperature gradients exist. Because capillary forces at lower temperature exceed those at higher temperature, slugs of water will migrate down a temperature gradient. Finally, in addition to compaction and capillary forcing, thermal convection may take place in regions with relatively high permeabilities greater than ~10^{-13} m^2 (Young *et al.*, 2003).

Irrespective of the driving mechanism, uncertainties in the permeabilities of chondritic material complicate our understanding of fluid flow in early planetesimals. Parent-body processes such as the growth of secondary minerals or impact-driven compaction imply that measured permeabilities of chondrite samples may not reflect those in their parent body's early history (Xu *et al.*, 2004). At the same time, passage of meteorites through the Earth's atmosphere may remove samples with high porosity (Sears, 1998), thereby biasing collected samples towards those of lower permeability.

Despite these open questions, both model calculations and direct measurements have been used to constrain the permeability of early planetesimals. Gas permeability experiments show that, at the ~1-cm scale, chondritic permeabilities span a wide range between $<10^{-21}$ and 10^{-14} m^2 (Figure 6.1; Sugiura *et al.* (1984); Corrigan *et al.* (1997)). Calculated permeabilities of chondritic matrix, based on direct measurement of the matrix grain size distribution in the primitive carbonaceous chondrite Acfer 094, lie between 10^{-19} and 10^{-17} m^2 (Bland *et al.*, 2009). These model values fall at the lower end of experimental data for carbonaceous chondrites, which may be due to the authors' use of fine matrix grain sizes to calculate permeability. Taking these laboratory and computed estimates of permeabilities at face value, compaction-driven porous flow would permit fluid ascent at >10 cm yr^{-1} through most carbonaceous chondrites, some ordinary chondrites, but

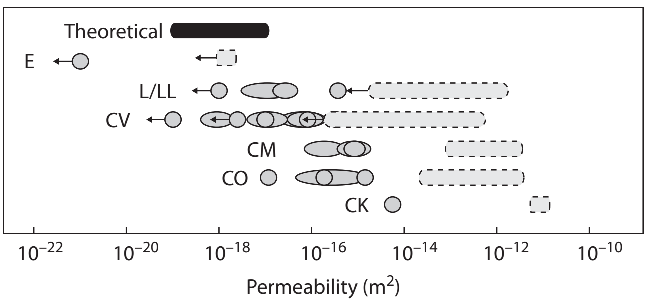

Figure 6.1 Compilation of measured and modeled chondrite permeabilities. Symbols with solid outlines denote hand-sample scale values while dashed symbols have been scaled up by 10^4 to approximate the effect of fracture networks at the global scale. The black bar represents modeled permeability from Bland *et al.* (2009). Experimental references: Corrigan *et al.* (1997) and Sugiura *et al.* (1984).

not enstatite chondrites. Chondrite parent bodies or regions thereof with very low permeabilities and negligible porous flow may participate in "mudball" convection where fine particles and fluid flow in a single velocity field (Bland *et al.,* 2013).

However, measured and modeled permeabilities at the <1-cm scale represent only a lower bound to those relevant to global fluid transport. Fracture networks at the greater than hand-sample scale may arise in chondrite parent bodies due to impacts, self-gravity, or the expansion of gas during progressive heating (Fu *et al.*, 2014; Young *et al.*, 2003). For bodies whose interiors heated to above approximately 200 °C due to radiogenic heating, production of water vapor produces pressures of up to several tens of MPa. This value exceeds the tensile strength of intact basalts and likely readily exceeds that of chondrites (Schultz, 1993), thereby providing a robust mechanism to create large-scale fracture networks in chondritic parent bodies. In terrestrial rock formations, the presence of fracture networks enhances regional permeabilities by three to four orders of magnitude relative to the values measured at the hand-sample scale (Clauser, 1992; Huenges *et al.*, 1997). Therefore, global-scale permeabilities for most non-enstatite, H_2O-bearing chondrite parent bodies whose interiors were heated to >200 °C may have reached the 10^{-14} to 10^{-12} m^2 range (Figure 6.1), potentially permitting channelized single-pass flow or thermal convection.

The uncertainties in permeabilities of chondritic material at both the hand-sample and global scales as outlined above motivate close examination of the meteoritic record for direct evidence regarding the extent and nature of fluid flow. In the remainder of this section we review mineralogic, chemical, and isotopic observations of chondrites that potentially reflect early aqueous activity.

The widespread occurrence of hydrated minerals provides prima facie evidence of past fluids in chondritic parent bodies. However, the implications of secondary mineral assemblages for fluid mobility are less clear. Thin rims (<<1 mm) of alteration products found around primary igneous minerals have been interpreted to suggest that fluid flow was limited to similar length scales (Bland *et al.*, 2009). On the other hand, despite evidence for aqueous fluid-assisted metamorphism and production of H_2O in alteration reactions in the past, the Allende CV chondrite does not contain detectable H_2O (Krot *et al.*, 1998; Jarosewich, 1990). Therefore, H_2O in Allende was apparently mobile over the length scale of the meteorite fall (~1 m).

The observed elemental compositions of chondrites also permit a range of interpretations depending on the meteorite class. The agreement between the elemental abundances of CI chondrites and the solar photosphere suggests that significant net transport of labile elements did not take place in these samples (Bland *et al.*, 2009). Several characteristics of CI chondrites may have hindered the mobility of their interstitial fluids. The lack of a coarse grain fraction (e.g. chondrules and chondrule fragments) have contributed to very low permeabilities while low metamorphic temperatures limited the potential for fracture network formation due to gas expansion.

The CI chondrites are the only meteorites with a solar composition. In non-CI chondrites, volatile elements such as Zn and K are depleted relative to more refractory elements such as Mg (Kallemeyn and Wasson, 1981). Such trends are commonly interpreted to reflect high temperature, volatility-controlled processes in the solar nebula. However, elemental volatility and solubility are closely correlated (Young *et al.*, 2003). As such, aqueous elemental transport on chondritic parent bodies may also contribute to the observed fractionations. Future simulations of secondary mineral phase growth that incorporate effects such as latent heat, two-phase flow, and reaction kinetics will be necessary to appreciate fully the chemical effects of fluid flow. If such simulations are successful in reproducing observed fractionations in chondrites, water–rock interactions on parent bodies must then be considered in conjunction with volatility-controlled nebular processes to understand the compositions of early planetesimals.

Isotopic studies of chondrites provide insight into the quantity and provenance of past interstitial water. The results of the study of oxygen isotope exchange during aqueous alteration of carbonaceous chondrites require volumetric water/rock ratios of 0.6 to 1.2 (Clayton and Mayeda, 1984, 1999). In a closed system such water/rock ratios correspond to porosities of 38 to 55%. These are minimum estimates based on the assumption that every molecule of pore-filling water reacted with the rock, a situation that rarely, if ever, occurs in terrestrial environments. Such large porosities, which are comparable to "water-saturated sand" (Clayton and Mayeda, 1999), contrast with the much smaller porosities of 5 to 29% exhibited by carbonaceous chondrites today (Britt and Consolmagno, 2003). In support of the oxygen isotopic data, thermodynamic simulations of closed-system carbonaceous chondrite alteration also suggest large water/rock ratios with values ranging from 0.6 to 14 (Rosenberg *et al.*, 2001; Zolensky *et al.*. 1989). These fluid volumes correspond to porosities of 40% to more than 90% and are again minimum estimates that far exceed rock porosities.

In terrestrial settings, water/rock ratios that exceed pore volumes are interpreted as indicators of fluid flow (Gregory and Criss, 1986). In the case of carbonaceous chondrites there are, therefore, two possible interpretations of the high water/rock ratios. First, aqueous alteration of the most altered rocks (CMs, CIs) may have occurred while they were soil-like materials with pore volumes of 40 to 50% or more. Second, the large water volumes may be the result of fluid flow by analogy to terrestrial environments.

The oxygen isotopic compositions of chondrites can potentially distinguish between these interpretations. In the first case, minerals alter in a closed system and exchange oxygen atoms with only the immediately surrounding fluid. In the second case, the minerals alter in an open system into which fresh, isotopically distinct water is continually introduced. These two scenarios predict different values for the observed $\Delta^{17}O$ of the unaltered and altered mineral grains, where $\Delta^{17}O$ can be thought of as a fingerprint for a given oxygen reservoir (see Figure 6.2).

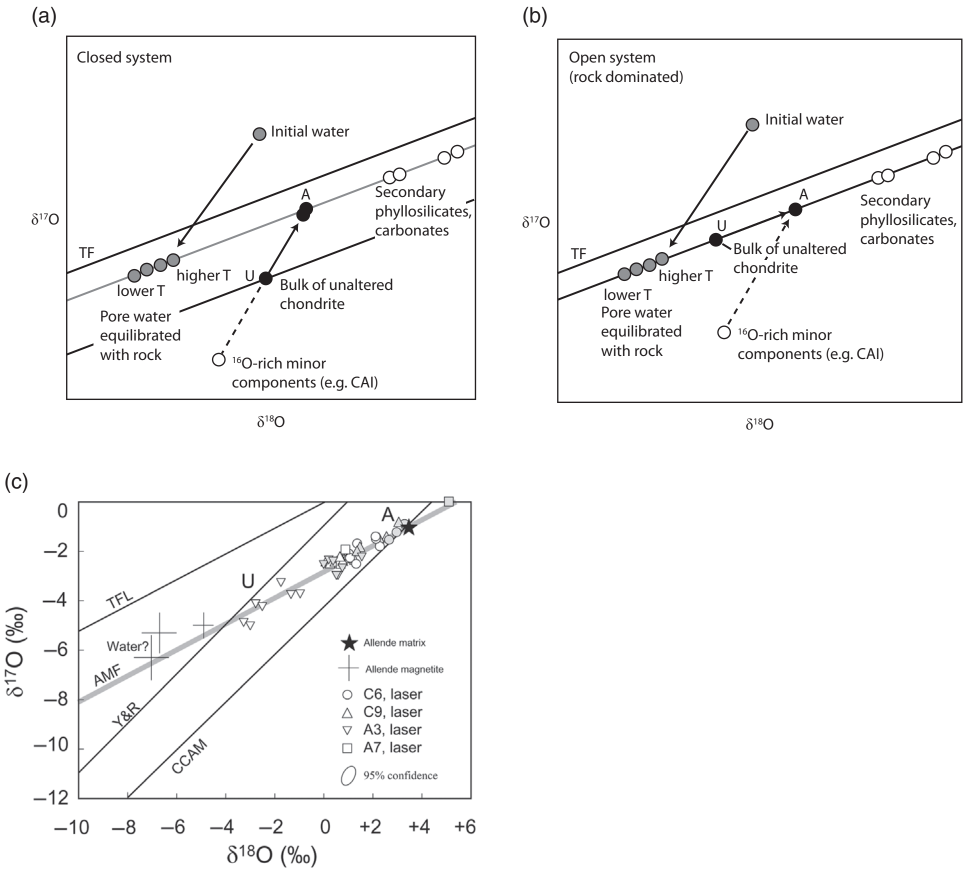

Figure 6.2 Schematic illustration of oxygen isotope exchange between water and rock in (a) a closed system and (b) an open system. A line between unaltered (U) and altered (A) rock that is steeper than the mass-dependent fractionation lines (e.g. TFL, AMF) is diagnostic of closed-system behavior, while the buffering effect of rock in an open system results in a characteristic slope 0.52 line between the unaltered and altered components. (c) Laser ablation oxygen isotope data for Allende chondrules lacking refractory ^{16}O-rich components. Grey points show signs of alteration. White points show little or no signs of alteration. Also shown are Allende magnetites (Choi *et al.*, 1997), Allende matrix (Clayton and Mayeda, 1999), the CCAM (Clayton *et al.*, 1977) and Y&R (Young & Russell 1998) mixing lines, the terrestrial fractionation line (TFL), and the Allende mass fractionation line (AMF) (Young *et al.*, 1999). Isotopic fractionations are relative to standard mean ocean water (SMOW).

Because mass-dependent fractionation curves have slopes close to a value of 0.52, a widely quoted formulation of $\Delta^{17}O$ is: $\Delta^{17}O = \delta^{17}O - 0.52\ \delta^{18}O$ (e.g. Clayton and Mayeda (1999)).

Water in the early solar system likely had a higher initial $\Delta^{17}O$ than most rocks (Choi *et al.*, 1998; Sakamoto *et al.*, 2007). In such cases, mass balance in a closed system requires that the oxygen isotopic compositions of both the water and altered rock migrate towards an equilibrium mass fractionation line that lies between the initial water and rock $\Delta^{17}O$ values on the three-isotope plot (grey line through point A in Figure 6.2A). The precise $\Delta^{17}O$ of the new equilibrium mass fractionation line is controlled by the water/rock ratio (Clayton and Mayeda, 1999). Therefore, for closed systems, lines connecting unaltered and altered material will always deviate from mass-dependent fractionation lines.

On the other hand, open-system exchange of oxygen between water and rock can lead to lines between unaltered and altered rock with slopes of nearly 0.52 (Figure 6.2B). Such shallow slopes occur because flowing water exchanges continuously with the host rock until its $\Delta^{17}O$ is buffered to the initial value of the rock. The result is that the equilibrium mass fractionation line for the water and rock has the same $\Delta^{17}O$ as the initial rock in rock-dominated portions of the flow system. In effect, the rock dominates the system so that the bulk-system $\Delta^{17}O$ is nearly the same as the initial-rock $\Delta^{17}O$ (Young *et al.*, 1999). At the same time, the $\Delta^{17}O$ that the water acquires during flow depends on the rock through which it flows.

Measurements of both unaltered and altered mineral phases in chondrites can reveal the relationship between these constituents in three-isotope space and therefore the implied degree of fluid mobility. Young *et al.* (1999) used the technique of ultraviolet laser ablation combined with gas chromatography isotope ratio monitoring (GCirms) to show that the oxygen isotope systematics of the Allende meteorite resemble those expected for open systems (Figure 6.2C). Specifically, the altered material (grey data points) in a given chondrule has higher $\delta^{18}O$ than the unaltered material (white data points) in the same chondrule but the $\Delta^{17}O$ of both components are the same (i.e. they plot along a slope ~0.52 line in three-isotope space).

In addition to distinguishing between open- and closed-system alteration, the $\Delta^{17}O$ data from Allende and other chondrites may also constrain the quantity of water involved in isotopic exchange with the rock. With protracted oxygen exchange, rock loses its capacity to buffer the isotopic composition of flowing water. Rocks downstream of the spent rock react with fluids that have pristine $\Delta^{17}O$ values similar to those of the initial water. In this scenario, also known as a fluid-dominated system, rock $\Delta^{17}O$ can begin to change. In the case of carbonaceous chondrites, $\Delta^{17}O$ would increase towards the initial isotopic composition of pore water.

In light of these predictions, measurements performed by Clayton and Mayeda (1999) suggest that both rock- and water-dominated environments existed on carbonaceous chondrite parent bodies (Figure 6.3). Specifically, when compared with the predicted evolutionary paths in three-isotope space (Young *et al.*, 1999; Young, 2001), COs, CVs, and some CMs have oxygen isotopic compositions that fall on a mass-dependent fractionation line indicative of rock-dominated alteration. Meanwhile, CIs and some CMs display a wider range of $\Delta^{17}O$ consistent with water-dominated interactions (Young, 2001; Young *et al.*, 1999). If this interpretation is correct, these isotopic data would imply a diversity of fluid-flow histories among carbonaceous chondrite parent bodies and even within the same parent body. Furthermore, this model requires the occurrence of open-system fluid flow on these parent bodies. With more oxygen isotopic analyses of chondrite matrix, combined with assessment of their degree of alteration, it will be possible to test whether or not the trends in Figure 6.3 are robust and therefore indicative of a process common to carbonaceous chondrite parent bodies.

Further evidence for the diversity of hydration histories in a single parent body arises from water/rock ratios inferred from oxygen isotopic data. A common misconception is that inferred water/rock ratios are direct measures of the physical amount of water that resided in the parent body. Instead, water/rock ratios calculated on the basis of mineralogical, chemical, or isotopic changes are a measure of reaction progress, not a measure of the bulk water content. The difficulties with

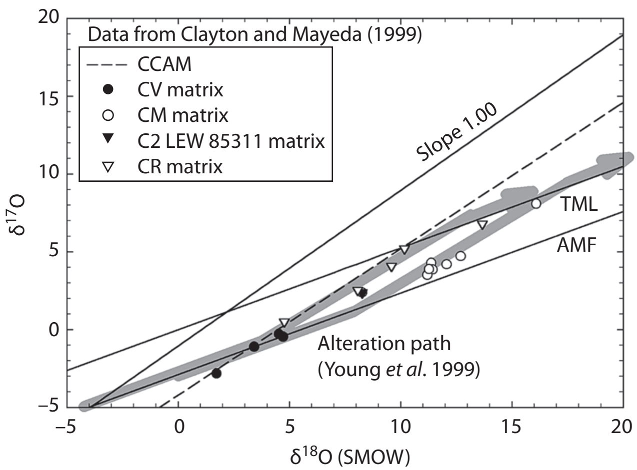

Figure 6.3 Oxygen three-isotope plot comparing published matrix compositions of carbonaceous chondrites and the alteration path proposed by Young *et al.* (1999). The predicted alteration path from rock- to water-dominated compositions is represented by the two grey arrows, which depict different temperature gradients. Abbreviations as in Figure 6.2.

interpreting water/rock ratios as parent-body water concentrations may be illustrated by comparing oxygen isotopic compositions of carbonates from CM, CI, Tagish Lake, and Sutter's Mill meteorites with closed-system models that assume equilibration of static water and rocks (Figure 6.4). Taken literally, each datum in $^{17}O/^{16}O$ vs. $^{18}O/^{16}O$ space must represent a different water–rock value and temperature. The carbonate data require variations in water/rock ratios from about 0.1 to 0.5 with variable temperatures. Much of this variability applies to CMs alone. Because such a broad range of accreted water–rock values is unlikely on a single parent body, the inferred water/rock ratios likely reflect a diversity of reaction progress.

The preceding discussion of oxygen isotopic variation highlights the potential diversity of alteration and fluid flow histories, even within the same parent body. Textural and mineralogical observations of chondrites provide further clues on such heterogeneities. Carbonaceous chondrites commonly contain a small complement of millimeter-to-centimeter-sized lithic clasts, generally referred to as "dark inclusions" (DIs; Johnson *et al.*, 1990)), that tend to be more altered than their host meteorites (Brearley and Krot, 2012). Chondrules in Allende DIs are often

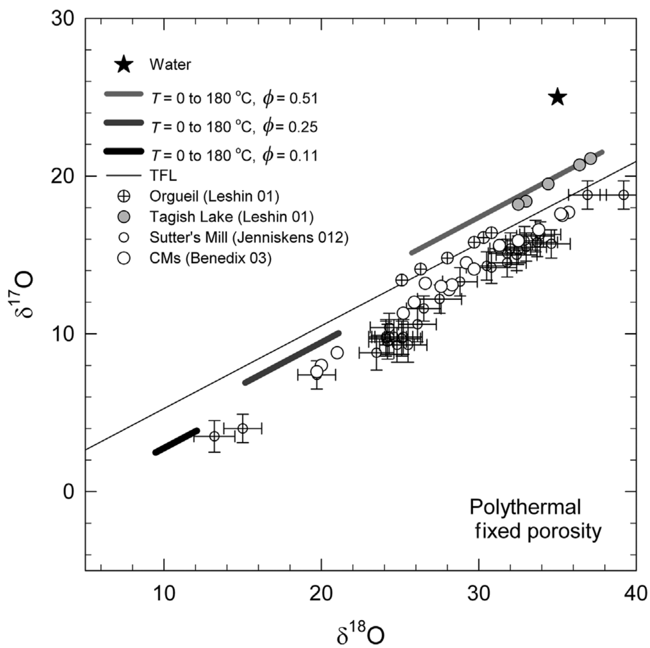

Figure 6.4 Classical closed-system water/rock calculations attempting to explain the range in carbonate oxygen isotope data for carbonaceous chondrites. References: Jenniskens *et al.* (2012), Benedix *et al.* (2003), Leshin (2001). Isotopic fractionations are relative to SMOW. A black and white version of this figure will appear in some formats. For the colour version, please refer to the plate section.

completely pseudomorphed by Fe-rich olivine ($Fa_{35\text{-}45}$) and nepheline, with opaque nodules replaced by Fe-rich olivine, Fe-rich diopside, and Fe,Ni sulfides (Brenker and Krot, 2004).

In terms of their oxygen isotopic compositions, Allende DIs display relatively restricted variation, plotting close to the CCAM line (Clayton and Mayeda, 1999). In contrast, DIs in the less altered CV3 reduced subgroup (Vigarano, Leoville, and Efremovka) show a much wider compositional range, typically displaying much greater heavy-isotope enrichments than those in Allende. This apparent paradox, whereby DIs in the more pristine CV3 reduced meteorites appear to have experienced greater levels of aqueous alteration than those in Allende, suggests a multi-stage alteration history. In this model, the first stage involved extensive hydrothermal alteration with replacement of primary chondrule minerals in DIs by a secondary assemblage of Fe-rich olivine, nepheline, sodalite, and Ni-rich sulfides as described above. The second phase of alteration followed excavation of the DI material and incorporation within the matrix of the preserved meteorite (Brenker and Krot, 2004), where DIs in Allende experienced greater thermal processing (Bonal *et al.*, 2006). Based on their mineralogy, DIs in CV chondrites appear to represent zones of more extensive alteration on their parent body, highlighting the heterogeneous nature of aqueous alteration and potentially reflecting differences in accreted ice/rock ratios or the occurrence of channelized fluid flow.

Synthesizing the lines of evidence presented above, a variety of interpretations exist in the current literature regarding the extent of fluid flow in carbonaceous chondrites. While theoretical models generally invoke flow on scales of kilometers or tens of kilometers [e.g. Young *et al.* (1999)], detailed mineralogical studies suggest that elemental transport was limited to the scale of centimeters at most (Brearley and Krot, 2012). Bland *et al.* (2009) point to low permeabilities and lack of either depletions or enrichments in water-soluble elements in highly altered chondrites as further evidence against significant fluid flow on their parent bodies. However, Brearley and Krot (2012) suggest that such extreme positions can be reconciled if there were wide variations in the local-scale geochemical environment, for example in pH. Thus, while extensive fluid flow may have taken place in chondrites, soluble elements would be transported only short distances before being reprecipitated when a geochemical reaction front was encountered. It is clear that more extensive geochemical modeling will be required to resolve these seemingly contradictory perspectives concerning the extent of fluid flow.

Beside carbonaceous chondrites, ordinary chondrites also exhibit evidence for extensive interactions with liquid water (Alexander *et al.*, 1989). Evidence for fluid mobility is less clear. Most recently, Dyl *et al.* (2012) found oxygen isotopic evidence for episodic pulses of hydrothermal activity on the parent body of the

Villalbeto de la Peña meteorite, although pulses of water derived from exogenous sources cannot be ruled out.

Beyond the evidence for fluid mobility in meteorites, water mobility in planetesimals is implied by the existence of water–rock differentiation in the solar system. Castillo-Rogez and Schmidt (2010) have shown that the Themis family asteroids exhibit ranges in physical properties indicative of ice–rock differentiation. Specifically, the core of the Themis parent body resembles CI and/or CM carbonaceous chondrites whereas outer regions are dominated by water ice. At the same time, the outer shape of (1) Ceres, the largest asteroid, indicates central condensation into what is most likely an ice-rich shell overlying a silicate-rich interior (Castillo-Rogez, 2011). The structures of these large asteroids therefore indicate the radial migration of water early in the history of the parent body.

6.4 The Fate of Silicate Melts

The establishment of fracture networks in intensely heated meteorite parent bodies as described in Section 6.3 implies that vapor phase and supercritical water was likely able to migrate from the planetesimal interior for most chondrite parent bodies. Furthermore, as mentioned above, observations of metamorphosed chondrites and heating experiments show that H_2O-bearing mineral phases undergo dehydration upon heating to below ~900 °C (Akai, 1992), which is well below the first production of silicate magmas (Agee *et al.*, 1995; McCoy *et al.*, 2003). Primordial interstitial volatiles and those released from hydrated silicates were therefore unavailable to play a critical role in silicate magma mobility. Even so, volatile exsolution may drive melt ascent on planetesimals if sufficient quantities of key species such as H_2O, CO_2, CO, N_2, and Cl can be retained in nominally anhydrous silicate phases up to the generation of mobile silicate melts.

We therefore evaluate the probable volatile contents of silicate melts at the temperature when they first become mobile. Melting experiments on enstatite chondrites suggest that efficient melt migration initiates upon heating to ~1250 °C or between 5 and 10% melt fraction (McCoy *et al.*, 1997). Progressive heating experiments on chondrites from all major groups show that CO_2 and N_2 are efficiently released at much lower temperatures, <800 °C, while the amount remaining at 1250 °C is much less than the experimental detection threshold of ~100 ppm (Muenow *et al.*, 1995; Muenow *et al.*, 1992; Akai, 1992). Cl may be retained to higher temperatures up to >1300 °C; however, only ~100 ppm, or a quarter of the initial abundance, appears to remain at such temperatures.

Estimates of retained H_2O contents are complicated by its small but potentially significant solubility in olivine and pyroxene, which are two of the most abundant anhydrous phases in chondrites. The solubility of H in olivine at 300 MPa varies between 20 and 1 ppm, depending on the oxygen fugacity (fO_2; Bai and Kohlstedt, 1993). Given that melting on most planetesimals took place under reducing conditions below the iron-wüstite (IW) buffer, the most relevant solubility of H_2O in olivine is likely near or below 1 ppm. The partitioning coefficient (D) between olivine and melt has been experimentally determined to ~0.0015 for temperatures of ~1250 °C (Hauri *et al.*, 2006). The concentration of H_2O in the incipient melt on reduced bodies is therefore expected to be <600 ppm while melts on oxidized bodies may contain ~13 000 ppm. Finally, the effect of CO on silicate melt migration depends on the viability of the C–FeO smelting reaction, which may produce significant amounts of CO at temperatures above 1200 °C (Walker and Grove, 1993; McCoy *et al.*, 1997). In the absence of the smelting reaction, however, CO is efficiently degassed below 1255 °C (Muenow *et al.*, 1992, 1995).

Volatile exsolution occurs in magma if the concentration of retained volatiles exceeds their solubility at the ambient pressures. For pressures of ~4 MPa, corresponding to the base of an approximately 10-km-thick conductive chondritic lid on a 150-km-radius body (Šrámek *et al.*, 2012), the solubilities of H_2O, CO_2, and Cl in basaltic melts are approximately 1800, 20, and 800 ppm (Papale, 1997; Webster, 1997; Wilson *et al.*, 2010). Comparing these concentrations with those expected of silicate magmas on differentiated planetesimals, Cl is not likely to have driven melt ascent via exsolution, while the role of CO_2 depends on the residual content at super-solidus temperatures, which is <<100 ppm. Given the likelihood of reducing conditions during silicate melting, most meteorite parent bodies probably did not retain sufficient H_2O in nominally anhydrous phases to cause exsolution. However, relatively oxidized melts such as those of the angrites may have been buoyant due to the effect of retained H_2O (Sarafian *et al.*, 2015), while the onset of the C–FeO smelting reaction in other parent bodies including the asteroid (4) Vesta and the ureilite parent body may have also permitted explosive volcanism (McCoy *et al.*, 2006; Wilson *et al.*, 2010; Walker and Grove, 1993).

Carbon- or water-bearing magmas forming inside planetesimals may erupt at high speed, driven by expanding volatiles (Wilson and Keil, 2012). They may coat the surface of the planetesimal, obscuring any persistent chondritic crust, or they may erupt with sufficient speed to escape the gravity of the body (Wilson and Keil, 1991). Ghosh and McSween (1998) describe an end-member model for Vesta in which all melt from the interior erupts onto the surface and another end-member model in which no melt extrudes. Their efforts demonstrate the difficulty of arguing completely for one or another eruptive scenario.

For bodies whose magmas avoided significant volatile exsolution, the melt densities may be calculated by considering the major element composition and temperature. Based on partial melting experiments on chondrites, Fu and Elkins-Tanton (2014) evaluated the densities of mobile magmas on the H, CV, CM, and

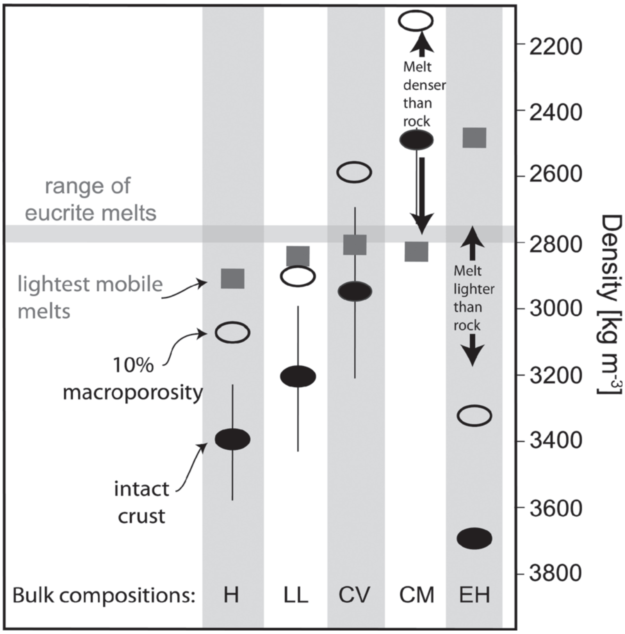

Figure 6.5 Estimates for the densities of meteoritic hand samples (solid ovals), the bulk chondritic lid (open ovals), and mobile magmas (gray rectangles) for a range of chondritic parent bodies.

EH chondrite parent bodies. These estimated densities can then be compared with those of the bulk overlying unmolten lid. Accounting for both macroporosity of the lid and the viscous removal of porosity at depth, carbonaceous chondrite melts, in part due to their higher FeO content, are negatively buoyant relative to the overlying lid while the opposite relationship is expected for ordinary and enstatite chondrites (Figure 6.5). Consistent with these predictions, aubrites, which are achondrites linked to enstatite chondrite-like precursors, may have undergone extensive eruption on the E-type asteroids (Keil, 2010; Gaffey *et al.*, 1992). We note that even in the case of carbonaceous chondrite melts, the bulk density of the mantle source region is expected to be in excess of 3200 kg m^{-3}, corresponding to the grain densities of anhydrous olivine and pyroxene. As such, carbonaceous chondrite melts are expected to rise from the deep interior and accumulate at the base of the primitive lid (Wilson and Keil, Chapter 8, this volume).

6.5 Implications

Volatile migration may have occurred in planetesimals and especially in the most vigorously heated bodies. If near-surface volatiles are removed, by impacts or by venting, currently dry bodies like Vesta may have accreted with a complement of volatiles.

Some minor bodies such as Vesta (HED suite), the Moon, and the angrite parent body display extreme levels of volatile depletion compared with Earth and Mars (Sarafian *et al.*, 2013). However, the lack of K isotope anomalies in either HEDs or lunar rocks provides evidence against the possibility that volatiles were simply boiled off into space from an essentially molten body (Humayun and Clayton, 1995).

Alternatively, extreme volatile depletion from minor bodies has also been explained as a feature inherited from the precursor materials, reflecting incomplete condensation from a nebular gas (Bland and Ciesla, 2010). In the case of Vesta, one potential problem for such a "hot nebula" model is the fact that at a distance of 2.4 AU the temperatures required by this process may be unrealistically high (Ciesla, 2008). In the case of the Moon, extreme volatile depletion was most likely the result of processes that took place in the >5000 K vapor disc following a giant impact (Canup, 2012). Such an energetic scenario is unlikely in the case of either Vesta or the angrite parent body because they accreted extremely early in solar system history in the presence of a gaseous nebula, which dampened impact velocities (Asphaug *et al.*, 2011).

These considerations motivate the pursuit of a devolatilization mechanism that avoids significant isotopic fractionation and occurs without rock vaporizing temperatures. As discussed in Section 6.3, rapid heating of early-formed planetesimals likely resulted in outward fluid migration, which would have left the bulk of the body essentially dry, with the possible exception of a thin outer layer where fluids ponded or froze. An outer hydrous zone could be lost by venting to space if fluid or gas pressures in this region exceeded lithostatic pressure and the rock's tensile strength. Outward migration of fluids and subsequent breakdown of hydrated minerals as internal temperatures rose may have led to significant leaching of relatively soluble elements, such as Na, K, F, Cl, I, C and S. If these species were transported in the fluid phase and then vented to space, this process may represent a viable mechanism to account for the extreme volatile depletion of early-forming bodies, such as Vesta and the angrite parent body.

The preservation of an unmelted chondritic lid overlying a differentiated interior holds strong implications for the interpretation of paleomagnetic data collected from chondritic meteorites. Paleomagnetic experiments from as early as 1972 indicate that the Allende CV chondrite carries a unidirectional component of magnetization removed upon heating to ~300 °C (Sugiura and Strangway, 1985). Carporzen *et al.* (2011) argued that the acquisition of this remanence required a stable and strong (~20 μT) magnetic field, which is most likely due to the presence of a magnetic core dynamo on the CV parent body. This model implies that the CV chondrites sample the unmolten near-surface layers of a differentiated body, which is consistent with the negative buoyancy of silicate melts on the CV parent body as hypothesized above. In the case of the CM parent body, which is also predicted to host negatively buoyant melts, paleomagnetic data suggest that a magnetic core dynamo may have also existed, although improved age controls are required to confirm this interpretation (Cournede *et al.*, 2015).

Finally, the existence of preserved chondritic lids on partially differentiated bodies provides a potential solution to the puzzling scarcity of igneous surface compositions observed in the modern asteroid belt. The frequency of iron and achondritic meteorites both indicate that many planetesimals melted early in solar system evolution. The lack of differentiated material in the asteroid belt may be explained at least in part by partial differentiation, in which a previously molten interior is masked under a chondritic crust. One possible example is the asteroid (21) Lutetia, whose high density strongly suggests that its deep interior underwent substantial heating and sintering. Even so, Lutetia's surface appears to be chondritic, implying that Lutetia may represent the first known partially differentiated asteroid (Weiss *et al.*, 2012). Models by Formisano *et al.* (2013) and Neumann *et al.* (2013) show that partial differentiation is possible if Lutetia accreted in less than 0.7 to several million years after CAIs.

Families of asteroids provide a tempting target to look for evidence of differentiation that was hidden beneath a chondritic crust prior to breakup. This search has not revealed compelling evidence for differentiation, possibly because the parent bodies were too small or because fragments were re-coated with chondritic material in the chaos of breakup. Recently, however, the Nysa–Polana complex, a group of bodies near (135) Hertha and the likely source of the OSIRIS-REx mission target (101955) Bennu, have been found to display both S-type and X-type spectra. One possible explanation is that these objects are the result of the breakup of a partially differentiated body, and so the group contains both chondritic and non-chondritic members (Dykhius and Greenberg, 2015).

6.6 Summary

The possible complexities of silicate melting and aqueous fluid reaction and migration in planetesimals are just being appreciated. Evidence regarding the extent of fluid migration supports a wide range of behaviors from minimal elemental transport in CI chondrites, to indirect isotopic and compositional signs of fluid flow in higher metamorphic grade chondrites, to likely rock–ice differentiation in the case of Ceres. Well-coupled physical and chemical models of fluid flow are necessary to infer the history of fluid migration from experimental evidence. If such models establish that fluid–rock reactions, dry or wet magmatic activity, and fluid flow are capable of explaining the observed mineralogical, chemical, and isotopic features of early solar system materials, then these parent-body effects must be accounted for when using the meteoritic record or direct observations of asteroids to understand earlier nebular processes.

Acknowledgments

We thank C. M. O. Alexander, P. A. Bland, A. J. Brearley, M. A. Hesse, A. N. Krot, M. I. Petaev, A. R. Sarafian, L. Wilson, and M. Y. Zolotov for discussions that improved this manuscript and broadened its scope.

References

Agee, C. B., Li, J., Shannon, M. C., *et al.* 1995. Pressure–temperature phase diagram for the Allende meteorite. *Journal of Geophysical Research*, **100**, 17725–17740.

Akai, J. 1992. T–T–T diagram of serpentine and saponite, and estimation of metamorphic heating degree of Antarctic carbonaceous chondrites. *Proceedings of the NIPR Symposium on Antarctic Meteorites*, **5**, 120–135.

Alexander, C. M. O., Barber, D. J., and Hutchison, R. 1989. The microstructure of Semarkona and Bishunpur. *Geochimica et Cosmochimica Acta*, **53**, 3045–3057.

Asphaug, E., Jutzi, M., and Movshovitz, N. 2011. Chondrule formation during planetesimal accretion. *Earth and Planetary Science Letters*, **308**, 369–379.

Bai, Q. and Kohlstedt, D. L. 1993. Effects of chemical environment on the solubility and incorporation mechanism for hydrogen in olivine. *Physics and Chemistry of Minerals*, **19**, 460–471.

Benedix, G., Leshin, L. A., Farquhar, J., *et al.* 2003. Carbonates in CM2 chondrites: Constraints on alteration conditions from oxygen isotopic compositions and petrographic observations. *Geochimica et Cosmochimica Acta*, **67**, 1577–1588.

Bland, P. A., Travis, B. J., Dyl, K. A., *et al.* 2013. Giant convecting mudballs of the early solar system. *Lunar and Planetary Science Conference,* **45**, 1447.

Bland, P. A., Jackson, M. D., Coker, R. F., *et al.* 2009. Why aqueous alteration in asteroids was isochemical: High porosity $\neq$ high permeability. *Earth and Planetary Science Letters*, **287**, 559–568.

Bland, P. A. and Ciesla, F. J. 2010. The impact of nebular evolution on volatile depletion trends observed in differentiated objects. In Lunar and Planetary Science Conference, **41,** 1817.

Bonal, L., Quirico, E., Bourot-Denise, M., and Montagna, G. 2006. Determination of the petrologic type of CV3 chondrites by Raman spectroscopy of included organic matter, *Geochimica Cosmochimica Acta*, **70**, 1849–1863.

Brearley, A. J. and Krot, A. N. 2012. Metasomatism in the early solar system: The record from chondritic meteorites. In *Metasomatism and the Chemical Transformation of Rock*, ed. D. E. Harlov and H. Austrheim. Berlin: Springer-Verlag, 659–789.

Brenker, F. E. and Krot, A. N. 2004. Late-stage, high-temperature processing in the Allende meteorite: Record from Ca,Fe-rich silicate rims around dark inclusions. *American Mineralologist*, **89**, 1280–1289.

Britt, D. and Consolmagno, G. J. 2003. Stony meteorite porosities and densities: A review of the data through 2001. *Meteoritics & Planetary Science*, **38**, 1161–1180.

Canup, R. M. 2012. Forming the Moon with Earth-like composition via a giant impact. *Science*, **338**, 1052–1055.
Carporzen, L., Weiss, B. P., Elkins-Tanton, L. T., *et al.* 2011. Magnetic evidence for a partially differentiated carbonaceous chondrite parent body. Proceedings of the National Academy of Sciences of the United States of America, **108**, 6386–6389.
Castillo-Rogez, J. C. 2011. Ceres: Neither a porous nor salty ball. *Icarus*, **215**, 599–602.
Castillo-Rogez, J. C. and Schmidt, B. E. 2010. Geophysical evolution of the Themis family parent body. *Geophysical Research Letters*, **37**, L10202.
Choi, B.-G., McKeegan, K. D., Leshin, L. A., *et al.* 1997. Origin of magnetite in oxidized CV chondrites: *in situ* measurement of oxygen isotope compositions of Allende magnetite and olivine. *Earth and Planetary Science Letters*, 337–349.
Choi, B.-G., McKeegan, K. D., Krot, A. N., *et al.* 1998. Extreme oxygen-isotope compositions in magnetite from unequilibrated ordinary chondrites. *Nature*, **392**, 577–579.
Ciesla, F. J. 2008. Radial transport in the solar nebula: Implications for moderately volatile element depletions in chondritic meteorites. *Meteoritics & Planetary Science*, **43**, 639–655.
Clauser, C. 1992. Permeability of crystalline rocks. *Eos Transactions of the AGU*, **73**, 233–238.
Clayton, R. N. and Mayeda, T. K. 1984. The oxygen isotope record in Murchison and other carbonaceous chondrites. *Earth and Planetary Science Letters*, **67**, 151–161.
Clayton, R. N. and Mayeda, T. K. 1999. Oxygen isotope studies of carbonaceous chondrites. *Geochimica et Cosmochimica Acta*, **63**, 2089–2104.
Clayton, R. N., Onuma, N., and Grossman, L. *et al.* 1977. Distribution of the pre-solar component in Allende and other carbonaceous chondrites. *Earth and Planetary Science Letters*, **34**, 209–224.
Corrigan, C. M. *et al.* 1997. The porosity and permeability of chondritic meteorites and interplanetary dust particles. *Meteoritics & Planetary Science*, **32**, 509–515.
Cournede, C., Zolensky, M. E., Dahl, J., *et al.* 2015. An early solar system magnetic field recorded in CM chondrites. *Earth and Planetary Science Letters*, **410**, 62–74.
Dykhius, M. J. and Greenberg, R. 2015. Collisional family structure within the Nyse–Polana complex. *Icarus*, **252**, 199–211.
Dyl, K. A., Bischoff, A., Ziegler, K., *et al.* 2012. Early solar system hydrothermal activity in chondritic asteroids on 1–10-year timescales. Proceedings of the National Academy of Sciences of the United States of America, **109**, 18306–18311.
Elkins-Tanton, L. T., Weiss, B. P., and Zuber, M.T. 2011. Chondrites as samples of differentiated planetesimals. *Earth and Planetary Science Letters*, **305**, 1–10.
Formisano, M., Turrini, D., Federico, C., *et al.* 2013. The onset of differentiation and internal evolution: the case of 21 Lutetia. *Astrophysical Journal*, **770**, 50.
Fu, R. R., Hager, B. H., Ermakov, A. I. *et al.* 2014. Efficient early global relaxation of asteroid Vesta. *Icarus*, **240**, 133–145.
Fu, R. R. and Elkins-Tanton, L.T. 2014. The fate of magmas in planetesimals and the retention of primitive chondritic crusts. *Earth and Planetary Science Letters*, **390**, 128–137.
Gaffey, M. J., Reed, K.L., and Kelley, M. S. 1992. Relationship of E-type Apollo asteroid 3103 (1982 BB) to the enstatite achondrite meteorites and the Hungaria asteroids. *Icarus*, **100**, 95–109.
Ghosh, A. and McSween, H. Y. 1998. A thermal model for the differentiation of asteroid 4 Vesta based on radiogenic heating. *Icarus*, **134**, 187–206.

Gregory, R. T. and Criss, R. E. 1986. Isotopic exchange in open and closed systems. *Reviews in Mineralology and Geochemistry*, **16**, 91–127.

Grimm, R. E. and McSween, H. Y. 1989. Water and the thermal evolution of carbonaceous chondrite parent bodies. *Icarus*, **82**, 244–280.

Hauri, E. H., Gaetani, G. A., and Green, T. H. 2006. Partitioning of water during melting of the Earth's upper mantle at H_2O-undersaturated conditions. *Earth and Planetary Science Letters*, **248**, 715–734.

Henke, S., Gail, H.-P., Trieloff, M., *et al.* 2012. Thermal history modelling of the H chondrite parent body. *Astronomy & Astrophysics*, **545**, p.A135.

Huenges, E., Erzinger, J., Kück, J., *et al.* 1997. The permeable crust: Geohydraulic properties down to 9101 m depth. *Journal of Geophysical Research*, **102**, 18,255–18,265.

Humayun, M. and Clayton, R.N. 1995. Potassium isotope cosmochemistry: Genetic implications of volatile element depletion. *Geochimica et Cosmochimica Acta*, **59**, 2131–2148.

Jacobsen, B., Yin, Q.-Z., Moynier, F., *et al.* 2008. ^{26}Al–^{26}Mg and ^{207}Pb–^{206}Pb systematics of Allende CAIs: Canonical solar initial ^{26}Al/^{27}Al ratio reinstated. *Earth and Planetary Science Letters*, **272**, 353–364.

Jarosewich, E. 1990. Chemical analyses of meteorites: A compilation of stony and iron meteorite analyses. *Meteoritics*, **25**, 323–337.

Jenniskens, P., Fries, M. D., Yin, Q-Z. *et al.* 2012. Radar-enabled recovery of the Sutter's Mill meteorite, a carbonaceous chondrite regolith breccia. *Science*, **1583**, 1583–1587.

Johnson, C. A., Prinz, M., Weisberg, M. K., *et al.* 1990. Dark inclusions in Allende, Leoville, and Vigarano: Evidence for nebular oxidation of CV3 constituents. *Geochimica et Cosmochimica Acta*, **54**, 819–830.

Kallemeyn, G. W. and Wasson, J. T. 1981. The compostional classification of chondrites – I. The carbonaceous chondrite groups. *Geochimica et Cosmochimica Acta*, **45**, 1217–1230.

Keil, K. 2010. Enstatite achondrite meteorites (aubrites) and the histories of their asteroidal parent bodies. *Chemie der Erde*, **70**, 295–317.

Krot, A. N., Petaev, M. I., Scott, E. R. D., *et al.* 1998. Progressive alteration in CV3 chondrites: More evidence for asteroidal alteration. *Meteoritics & Planetary Science*, **33**, 1065–1085.

Lee, T., Papanastassiou, D. A., and Wasserburg, G.J. 1976. Demonstration of ^{26}Mg excess in Allende and evidence for ^{26}Al. *Geophysical Research Letters*, **3**, 41–44.

Leshin, L. A., Farquhar, J., Guan, Y., *et al*. 2001. Oxygen isotopic anatomy of Tagish Lake: relationship to primary and secondary minerals in CI and CM chondrites. *Lunar and Planetary Science Conference*, **32**, 1843.

McCoy, T. J., Keil, K., Muenow, D. W., *et al.* 1997. Partial melting and melt migration in the acapulcoite–lodranite parent body. *Geochimica et Cosmochimica Acta*, **61**, 639–650.

McCoy, T., Mittlefehldt, D. W., and Wilson, L. 2003. Asteroid differentiation. In *Meteorites and the Early Solar System II*, ed. D. Lauretta and H. Y. McSween. Tucson, AZ: University of Arizona Press, 733–745.

McCoy, T. J., Ketcham, R. A., Wilson, L., *et al.* 2006. Formation of vesicles in asteroidal basaltic meteorites. *Earth and Planetary Science Letters*, **246**, 102–108.

Muenow, D. W., Keil, K., and Wilson, L. 1992. High-temperature mass spectrometric degassing of enstatite chondrites: Implications for pyroclastic volcanism on the aubrite parent body. *Geochimica et Cosmochimica Acta*, **56**, 4267–4280.

Muenow, D. W., Keil, K., and McCoy, T. J. 1995. Volatiles in unequilibrated ordinary chondrites: Abundances, sources and implications for explosive volcanism on differentiated asteroids. *Meteoritics & Planetary Science*, **30**, 639–645.

Neumann, W., Breuer, D., and Spohn, T. 2013. The thermo-chemical evolution of Asteroid 21 Lutetia. *Icarus*, **224**, 126–143.

Neumann, W., Breuer, D., and Spohn, T. 2014. Differentiation of Vesta: Implications for a shallow magma ocean. *Earth and Planetary Science Letters*, **395**, 267–280.

Öberg, K., Murray-Clay, R., and Bergin, E. A. 2011. The effects of snowlines on C/O in planetary atmospheres. *Astrophysical Journal Letters*, **743**, L16.

Ootsubo, T., Kawakita, H., and Hamada, S., *et al.* 2012. AKARI near-infrared spectroscopic survey for CO_2 in 18 comets. *Astrophysical Journal*, **752**, 1–12.

Palguta, J., Schubert, G., and Travis, B. J. 2010. Fluid flow and chemical alteration in carbonaceous chondrite parent bodies. *Earth and Planetary Science Letters*, **296**, 235–243.

Papale, P. 1997. Modeling of the solubility of a one-component H_2O or CO_2 fluid in silicate liquids. *Contributions to Mineralology and Petrology*, **126**, 237–251.

Rosenberg, N. D., Browning, L., and Bourcier, W. L. 2001. Modeling aqueous alteration of CM carbonaceous chondrites. *Meteoritics & Planetary Science*, **36**, 239–244.

Sakamoto, N., Seto, Y., Itoh, S., *et al.* 2007. Remnants of the early solar system water enriched in heavy oxygen isotopes. *Science*, **317**, 231–233.

Sarafian, A. R., Roden, M. F., and Patiño-Douce, A. E. 2013. The volatile content of Vesta: Clues from apatite in eucrites. *Meteoritics & Planetary Science*, **48**, 2135–2154.

Sarafian, A. R., Nielson, S. G., Berger, E. L., *et al.* 2015. Wet angrites? A D/H and Pb–Pb study of silicates and phosphates. Lunar and Planetary Science Conference, **46,** 1542.

Schiller, M., Connelly, J. N., Glad, A. C., *et al.* 2015. Early accretion of protoplanets inferred from a reduced inner solar system ^{26}Al inventory. *Earth and Planetary Science Letters*, **420**, 45–54.

Schultz, R. A., 1993. Brittle strength of basaltic rock masses with applications to Venus. *Journal of Geophysical Research*, **98**, 10,810–883,895.

Sears, D. W. G. 1998. The case for rarity of chondrules and calcium–aluminum-rich inclusions in the early solar system and some implications for astrophysical models. *Astrophysical Journal*, **498**, 773–778.

Šrámek, O., Milelli, L., Ricard, Y., *et al.* 2012. Thermal evolution and differentiation of planetesimals and planetary embryos. *Icarus*, **217**, 339–354.

Sugiura, N., Brar, N. S., and Strangway, D. W. 1984. Degassing of meteorite parent bodies. *Journal of Geophysical Research*, **89**, B641–B644.

Sugiura, N. and Strangway, D.W. 1985. NRM directions around a centimeter-sized dark inclusion in Allende. Lunar and Planetary Science Conference, **15**, C729–C738.

Taitel, Y. and Witte, L. 1996. The role of surface tension in microgravity slug flow. *Chemical Engineering Science*, **51**, 695–700.

Tang, H. and Dauphas, N. 2012. Abundance, distribution, and origin of ^{60}Fe in the solar protoplanetary disk. *Earth and Planetary Science Letters*, **359**–360, pp.248–263.

Turcotte, D. L. and Schubert, G. 2002. *Geodynamics*. New York: Cambridge University Press.

Urey, H. C. 1955. The cosmic abundances of potassium, uranium, and thorium and the heat balances of the Earth, the Moon, and Mars. *Proceedings of the National Academy of Sciences of the United States of America*, **41**, 127–144.

Walker, D. and Grove, T. L. 1993. Ureilite smelting. *Meteoritics*, **28**, 629–636.

Webster, J. D. 1997. Chloride solubility in felsic melts and the role of chloride in magmatic degassing. *Journal of Petrology*, **38**, 1793–1807.

Weiss, B. P., Elkins-Tanton, L. T., Barucci, M. A., *et al.*, 2012. Possible evidence for partial differentiation of asteroid Lutetia from Rosetta. *Planetary and Space Science*, **66**, 137–146.

Wilson, L. and Keil, K. 1991. Consequences of explosive eruptions on small solar system bodies: The case of the missing basalts on the aubrite parent body. *Earth and Planetary Science Letters*, **104**, 505–512.

Wilson, L. and Keil, K. 2012. Volcanic activity on differentiated asteroids: A review and analysis. *Chemie der Erde*, **72**, 289–321.

Wilson, L., Keil, K., and McCoy, T. J. 2010. Pyroclast loss or retention during explosive volcanism on asteroids: Influence of asteroid size and gas content of melt. *Meteoritics & Planetary Science*, **45**, 1284–1301.

Wood, J. A. 1964. The cooling rates and parent bodies of several iron meteorites. *Icarus*, **3**, 429–459.

Xu, T., Sonnenthal, E., Spycher, N., *et al.* 2004. TOUGHREACT – A simulation program for non-isothermal multiphase reactive geochemical transport in variably saturated geologic media: Applications to geothermal injectivity and CO_2 geological sequestration. *Compututers & Geosciences*, **32**, 145–165.

Young, E. D. 2001. The hydrology of carbonaceous chondrite parent bodies and the evolution of planet progenitors. *Philosphical Transactions of the Royal Society of London A*, **359**, 2095–2110.

Young, E. D. and Russell, S. S., 1998. Oxygen reservoirs in the early solar nebula inferred from an Allende CAI. *Science*, **282**, 452–455.

Young, E. D., Ash, R. D., England, P., and Rumble, D. III. 1999. Fluid flow in chondritic parent bodies: Deciphering the compositions of planetesimals. *Science*, **286**, 1331–1335.

Young, E. D., Zhang, K., and Schubert, G., 2003. Conditions for pore water convection within carbonaceous chondrite parent bodies: Implications for planetesimal size and heat production. *Earth and Planetary Science Letters*, **213**, 249–259.

Zolensky, M. E., Bourcier, W. L., and Gooding, J. L. 1989. Aqueous alteration on the hydrous asteroids: Results of EQ3/6 computer simulations. *Icarus*, **78**, 411–425.