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Nebular Evolution of Thermally Processed Solids: Reconciling Models and Meteorites

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Abstract. We relate current protoplanetary nebula process models to the observed properties of chondrites and their individual constituents. Important nebula properties and processes that affect the evolution of small solid particles include the nebula temperature and pressure, the generally inward radial drift of particles under gas drag, and nebula gas turbulence. We review these nebula properties and describe how they affect particle evolution, emphasizing the primary accretion stage whereby the first primitive meteorite parent bodies are accumulated. We then turn to chondrite properties and discuss how they constrain the models. We treat physical properties (chondrule and refractory inclusion size distributions, fine-grained and coarse-grained accretionary rims, coarse-grained igneous rims), chemical and mineralogical properties (Wark-Lovering rims, redox state, and elemental fractionations), and isotopic compositions (primarily oxygen). We note how currently inferred accretion and melting ages of asteroidal bodies seem to imply that primary accretion of existing 100-km-sized objects was delayed by 1 Myr or more relative to Ca,Al-rich inclusions, and sketch scenarios for primary accretion in a temporally evolving protoplanetary nebula which allows for this hiatus. We advance a new perspective on explaining non-equilibrium mineralogy in terms of evolutionary timescales for small particles across nebula radial thermal gradients, which should be testable using meteoritical data, and present a specific application to Wark-Lovering rims.

1. Introduction

Understanding any geological formation requires understanding context, and meteorites are no exception. Context leads to insights into current puzzles, and to testable hypotheses. It is widely accepted that meteorite parent bodies formed in the protoplanetary nebula, at least tens of millions of years before the Earth formed (e.g., Kleine et al. 2002; Yin et al. 2002). Chondritic meteorites (chondrites) consist of four major components: refractory inclusions [Ca,Al-rich inclusions (CAIs) and amoeboid olivine aggregates (AOAs)], chondrules, Fe,Ni-metal, and fine-grained matrix materials; almost all of them appear to have been thermally processed to various degrees in the protoplanetary nebula prior to accretion into their parent bodies (Alexander and Scott & Krot, this volume and references therein). The simplest hypothesis is that all chondrite components (except for the traces of stardust and organic material in the matrix) formed at high temperatures in the solar nebula and accreted into their parent bodies. Chondritic components may have experienced low temperature alteration in the nebula, but most alteration of this type occurred in an asteroidal setting (e.g., Brearley 2003).

A traditional context has been that there was, at some significant point in time, a "minimum mass nebula" with some radial temperature (and associated compositional) gradient, containing locally some fraction of "cosmic abundance" which could be solid, and that primitive parent bodies represent spatial samples of that material, at that time (see, e.g., Wood & Morfill 1988, Morfill & Wood 1989). In this context, the diversity of well-defined chondrite groups, and the radial zoning of asteroid types, suggests rapid accretion of chondritic components before their constituents can be mixed significantly. This context lends itself to models in which planetesimal formation begins everywhere at a specific time, perhaps at the midplane, perhaps by some sort of instability.

In recent years, this scenario has been increasingly challenged by the growing body of evidence that the formation of meteorite components, and their accretion into parent bodies, extended over several million years (MacPherson et al. 1995; Amelin et al. 2002, 2004; Bizzarro et al. 2004; Kunihiro et al. 2004; Kurahashi et al. 2004; Russell et al. 2005; Kita et al. this volume, and references therein). It is known from theory and observations that, over a period of millions of years, protoplanetary nebulae evolve dramatically, with significant changes in surface mass density, temperature, and other properties (Calvet, Hartmann, & Strom 2000; Hartmann, this volume; D'Alessio et al. this volume). This context of extended evolution must be incorporated into scenarios of the primary accretion of parent bodies, and its implications explored. For instance, under these circumstances, primary accretion of some particular body – perhaps of an entire meteorite group - represents a *temporal* sample of the nebula as well as a *spatial* sample. Even if the nebula were radially homogeneous at a given time (an unlikely situation), its temporal evolution might contribute to or even dominate the varying properties of the chondrite groups. In this chapter, we emphasize a context in which the nebula remains weakly turbulent for a significant fraction of the evolution of thermally processed solids into primitive parent bodies. This has a number of implications for meteoritics, and makes a number of predictions that should be testable. Some initial tests of the predictions seem to be encouraging, but more testing and observations are required. Theoretical arguments for or against weak nebula turbulence remain unsettled (e.g., Stone et al. 2000; Cuzzi & Weidenschilling 2005, and references therein), and are subjects of active research. Nonetheless, the potential advantages it offers are, we think, significant. We also note the natural predictions of nonturbulent scenarios, and press both scenarios for implications with the intent of stimulating new ways of thinking about the possibilities.

In section 2, we sketch the types of models in question, referring to prior work for all details. In section 3, we discuss how some different aspects of meteoritic observations (physical and mineralogical properties, chemical and isotopic compositions) can constrain the models better. In section 4, we summarize the things that puzzle us most, and note a few key potential observational and theoretical advances that would be pivotal.

2. Models: A Brief Review

2.1. The Gaseous Nebula

The properties of the evolving nebula are discussed in more detail by Hartmann and by D'Alessio et al. in this volume. Here, we touch on the aspects of greatest relevance to the evolution of thermally processed solids. The nebula is almost entirely hydrogen gas, with abundances of rock-forming elements measured in the fractions of percent on average. Nebula pressures vary with gas density and temperature (thus with radius at any given time, and also with time). Time can be measured from any resolvable event which is on a shorter timescale than those presented here – for instance, the first gravitational collapse of the protostar, on timescales of tens of thousands of years. In the hot inner nebula during the era when refractory minerals (CAIs) apparently formed, pressures might have been as high as 10^{-3} bar. One or two Myr later, in the cooler and lower density asteroid belt region, the pressures might have been 10^{-6} bar (D'Alessio et al, this volume; Cuzzi & Weidenschilling 2005). As the disk evolves and its mass density decreases, its accretion rate decreases. Observations indicate that mass accretion rates decrease on about the same timescale as disk masses (Fig. 1). So in some sense, disk accretion rate is a metric for disk age - high when young, low when old. However, considerable scatter exists in observations of accretion rate as a function of time, most of which is apparently real (Calvet et al. 2000; Hartmann and D'Alessio et al. this volume).

The nebula starts with all its solids in submicron interstellar grains, which are very effective (per unit mass) at absorbing optical and infrared radiation. As accretion proceeds, the opacity (area per unit mass) of nebula solids decreases roughly inversely as the typical particle size increases; thus the opacity of cm-size particles is four or five orders of magnitude smaller than that of interstellar grains. The most likely scenario is that some small fraction of the mass can remain in very small grains, keeping the opacity at least moderate, for a long time − consistent with observations of T Tauri disks (Dullemond & Dominik 2005). Because of this dust at high altitudes, astronomical observations often detect only radiation from the cool upper surface of a disk. Woolum & Cassen (1999) derived the much warmer *midplane* temperatures for a number of T Tauri-like disks using their observed properties and assuming a small-grain opacity. Because the mass of the gaseous disk is also inferred using this assumption, a larger grain size (lower opacity) would raise the disk mass to which it applies, leaving the midplane temperature, which depends on both the opacity and disk mass, only weakly affected. In Figure 2, we plot the midplane temperatures of Woolum & Cassen (1999) as a function of disk accretion rate (their Table 2). There are two sets of symbols, because the two different observational groups on which Woolum & Cassen based their analysis adopted different disk masses, and inferred different accretion rates, for the same set of objects; this scatter indicates the uncertainty in the results. Objects with disk masses less than 10^{-2} solar masses are not shown.

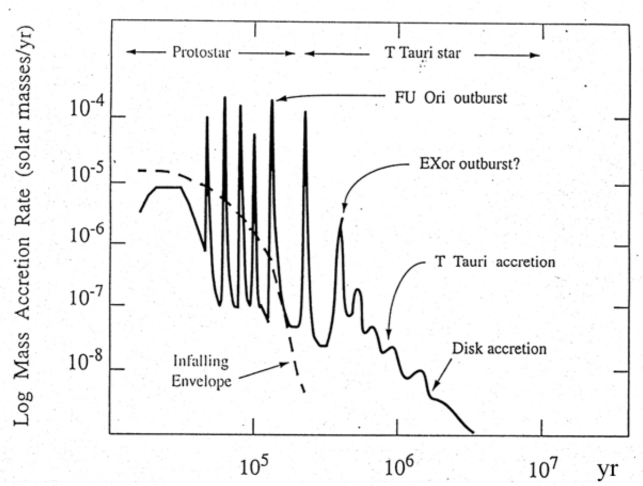

Figure 1. Schematic of nebula mass accretion rate *dM/dt* as a function of time, or age, in years. The mass accretion rate onto the star (solid line) decreases significantly from the early days of the nebula, when CAIs and AOAs formed (on the 10^5 year timescale), to the time when asteroids formed (several Myr later). The dashed line shows how the remnant envelope around the star-disk system disappears, revealing the star. From Calvet et al. (2000).

We adopt a nominal forsterite evaporation temperature of 1400K. We note that some recent observations show evaporation of solid material at this temperature (Muzzerolle et al. 2003). Inspection of Figures 1 and 2 suggests to us that, early in nebula history (when mass accretion rates were larger than a few times 10^{-7} M_sun/yr), midplane temperatures in the terrestrial planet region of actively accreting disks are adequate to evaporate ferromagnesian material, in contrast to some earlier thinking (e.g., Wood & Morfill 1988). Some recent models also indicate this (e.g., Bell et al. 1997). While the midplane temperatures for the *observed* systems of Figure 2 do not quite fit the criterion, the same systems probably *did* fit the criterion just a little earlier in their evolution, when their accretion rate was only slightly higher (*cf.* Fig. 1), but while they were still obscured.

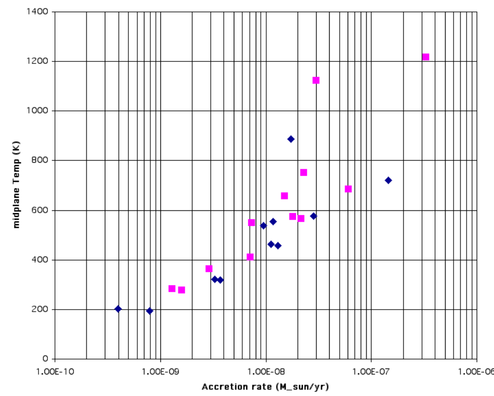

Figure 2. Nebula midplane temperatures at 1 AU as inferred by Woolum & Cassen (1999) from two different observational datasets of revealed T Tauri objects, which had a wide range of accretion rate *dM/dt.* Disks less massive than 0.01 M_sun are not shown. Given the trend and variance of the data, midplane temperatures > 1400K, the evaporation temperature of forsterite, seem quite plausible for accretion rates > 2-3$\times 10^{-7}$ M_sun/yr.

The mechanism by which this accretion is implemented is not presently known (Hartmann, this volume, and references therein). Fundamentally, for matter to accrete onto the star, it must lose angular momentum. The debate surrounds the mechanism by which this angular momentum transport occurs. Turbulent viscosity by itself has been thought unlikely to provide this transport on dual grounds: (1) the radial gradient in rotation rate makes the nebula stable against turbulence, so it is hard to maintain, and (2) turbulence, even if it does occur, transports angular momentum the wrong way. However, a number of recent efforts are focussing on turbulence-generating nonlinear effects that are absent from current simulations, but might come into play at the very large velocities and spatial scales characteristic of the nebula [see e.g., Sreenivasan & Stolovitsky (1995) for a basic overview, and references therein]. Generation of turbulence by the magnetorotational instability, which is known to be robust in well-ionized regions, is problematic in the dense gas of the terrestrial planet region - maybe between 1 and 10 AU (Gammie 1996; Sano et al. 2000). In the young nebula, when ample dust opacity maintains strong thermal gradients, the baroclinic instability might generate turbulence (Klahr & Bodenheimer 2003), but as accretion proceeds and opacity decreases, this might become problematic. Nevertheless, accretion is proceeding and gravitational energy is being released. A simple calculation indicates that only a tiny fraction of this accretional energy, converted into turbulent gas motions, maintains weak turbulence at levels adequate to dominate particle evolution (Cuzzi et al. 2001; Cuzzi & Weidenschilling 2005). We can somewhat simplistically relate observations, which determine the mass accretion rate *dM/dt,* to theoretical models, which parameterize turbulent intensity by the parameter α, by

$$\alpha = (dM/dt)/(3\pi\sigma_g cH)$$

where σ_g is the nebula surface mass density, c is the sound speed, and H is the vertical scale height (*cf.* Hartmann this volume; D'Alessio et al. this volume). For typical values of the parameters in the above equation, $\alpha = 10^{-4}$ to 10^{-2}. By comparison, the turbulent wind fluctuations on a mild, breezy day (average winds of 5-30 mph with 10% fluctuations), have α approximately 10^{-4} to 10^{-3}.

Accretion and stellar winds are correlated, perhaps even causally in the sense that accretion might power the winds (Hartmann this volume). How the alternative theories for stellar winds (the x-wind theory and the disk wind theory) relate to meteoritics and what their key parameters are, are discussed by Hartmann (this volume) and by Cuzzi & Weidenschilling (2005).

2.2. Solids: Incremental Accretion and Decoupling from the Gas

One of the most important take-home messages of this paper is that solids evolve quite differently from the gas in which they are embedded. This applies both to small particles like chondrules and refractory inclusions, and, in a different way, to meter-size boulders. This has important implications for not only the mass distribution of parent body precursors, but also for their chemistry, mineralogy, and isotopic composition. Here we introduce the physics of this evolution, and in section 3 we discuss the meteoritical implications. All of the interaction between a particle and the gas is captured by its gas drag stopping time

$$t_s = r\rho_s/c\rho_g$$

where r and ρ_s are the radius and density of a particle, and c and ρ_g are the sound speed and density in the gas. Particles with r or ρ_s in the range observed for chondrules and CAIs, for instance, are firmly coupled to the gas and have very small relative velocities. This expression is valid for all meteorite constituents; expressions for m-size and larger particles vary, as discussed in more detail by Cuzzi & Weidenschilling (2005) and references therein.

We separate the evolution of solids into several stages. In the first stage, from interstellar grain size to perhaps meter size, the combination of low relative velocities and crushy, porous, velcro-like surfaces probably ensures that grains can grow by sticking. This growth process, which we call *incremental accretion*, is probably not perfectly efficient, and a lot of uncertainty surrounds sticking coefficients and erosional effects even in this size range, but overall we do not think it presents an insurmountable barrier even in the presence of weak turbulence (see Blum 2004 and Cuzzi & Weidenschilling 2005 for discussions). As particles grow and collisional energies slowly increase, particles become more compacted, probably reaching densities of order unity. However, this growth might only be temporary; the energy of most colliding meter-sized compacted aggregates in even weak turbulence at relative velocity $c\alpha^{1/2}$ is probably enough to destroy them, releasing their constituents to retrace the accumulation process (Weidenschilling 1988; Cuzzi & Hogan 2003; Cuzzi & Weidenschilling 2005).

2.2.1. Radial Drift due to the Nebula Headwind

As long as the nebula gas persists, particles have a tendency to drift radially inwards. This tendency for radial drift, first noted by Whipple (1972), arises because the gas, in general, does not rotate at the same speed that particles do. This in turn is because the gas experiences radial pressure gradients - usually inward - leading to outward radial forces which offset a small part of solar gravity. Thus, in general, the gas rotates more slowly than the particles, which lose angular momentum to this headwind and drift inwards (Adachi et al. 1976; Weidenschilling 1977; Cuzzi & Weidenschilling 2005). The headwind, and the drift velocity, increase with particle size but are only weakly dependent on nebula location (Fig. 3).

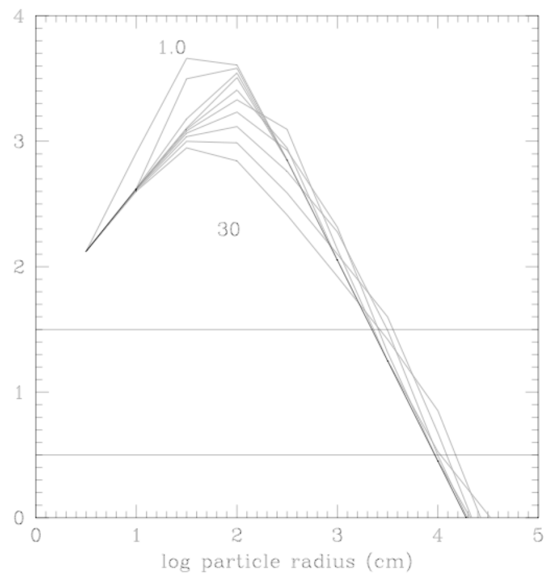

Figure 3. Logarithm of radial drift velocity in cm/sec (plotted vertically) for particles with internal density of 1 g cm^{-3}, plotted at distances from 1 AU to 30 AU (varying by factors of 1.4), for a nebula with gas surface density proportional to R^{-1}. The horizontal lines bracket the range of nebula gas advection, or systematic drift, velocity under viscous angular momentum transport, assuming $\alpha = 10^{-3}$. For larger (smaller) density, the curves shift to the left (right) proportionally. Adapted from Cuzzi & Weidenschilling (2005).

2.2.2. Vertical and Radial Diffusion by Turbulence

In weak turbulence, particles of cm-size and smaller are unable to "settle to the midplane" because turbulence keeps them stirred up (Dubrulle, Morfill, & Sterzik 1995; Cuzzi & Weidenschilling 2005). Figure 4 shows vertical profiles of the ratio of the actual particle mass density $\rho_d(z)$ to its global average value. The profiles are determined by the ratio S of the particle Stokes number Ωt_s to the nebula value of

α. Larger particles (with larger t_s), in weaker turbulence, settle into denser layers. For example, mm-radius particles, with $\alpha = 10^{-4}$, lie on the curve labeled S = 5, showing only a small tendency to settle towards the midplane. This vertical diffusivity has a parallel in the radial direction as well (as discussed below). This *eddy* diffusion is 6-9 orders of magnitude larger than the *molecular* diffusion associated with molecular random motions in a macroscopically stagnant gas, which is entirely negligible for any of the processes mentioned here.

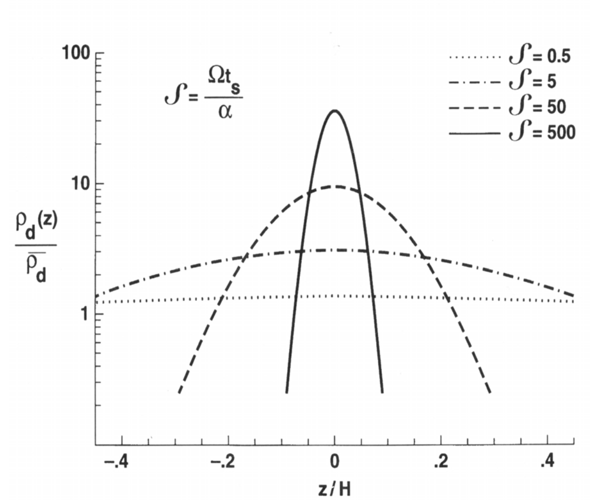

Figure 4. Vertical profiles of steady state particle mass density ρ_d, for several different values of the parameter $S = \Omega t_s/\alpha$. Particles of different stopping times t_s are prevented from settling to the nebula midplane by turbulence with intensity α. The thickness of a layer h_p (and thus the density it may achieve) is given by $h_p/H = (\Omega t_s/\alpha)^{1/2}$, where H is the gas scale height and Ω is the orbit frequency. At 2.5 AU, where $\Omega = 5\times10^{-8}$, a mm-size particle has $t_s \sim 1$ hour, so the curve labeled S = 5 would prevail if the nebula had turbulent intensity $\alpha = 10^{-4}$. From Cuzzi et al. (1996).

With turbulence keeping the volume density of "feedstock" low, meter-sized particles grow slowly, and drift radially so rapidly and by such large distances (1AU/century), that those which escape destruction by mutual collisions reach warmer, inner regions of the nebula where they evaporate. We discuss this more in section 2.3. Thus, we envision a first stage, which might be quite extended and persist as long as does weak turbulence, during which the particle size distribution remains stalled in the micron-to-meter size range. Radial drift is important throughout this stage. During this time, a mm-sized particle might diffuse radially in turbulence, be accreted into a meter-sized particle, drift radially inwards, and be released in a collision - a number of times. All this time, particles are sweeping up smaller ones due to their small, but finite, velocity differences; some features, such as fine-grained dust rims on chondrules and refractory inclusions, might be produced this way (section 3.1.2).

Turbulence allows for quite extensive radial mixing for particles of different types, both by inward radial drift, as mentioned above, and by radial gradient diffusion. Small particles like chondrules and CAIs, and certainly finer grains, are mostly coupled to the nebula gas, and diffuse in turbulence nearly as well as gas molecules during this stage. The mass flux of a particle of a specific kind is proportional to the radial gradient of *its concentration*. As one very relevant example, if a certain kind of particle is only produced at small radii, there will be a net diffusive flux of them outwards. This effect may resolve the longstanding puzzlement as to how CAIs can be produced 1-3 Myr earlier than chondrules, and yet survive their net inward radial "headwind" drift (Cuzzi et al. 2003). A prediction of this theory is that, because large inclusions drift inwards faster than small ones (Fig. 3) and are thus harder for diffusion to preserve, chondrites containing large refractory inclusions (CVs) must be older than other groups which lack them (see section 3.4.1). In a similar way, crystalline grains might be mixed outwards into cold comet-formation regions (Nuth 1999; Nuth et al. 2000; Bockelee-Morvan et al. 2002) and unprocessed presolar grains and volatile organics found in matrix might be mixed inwards into parent body accretion regions (Boss 2005; Keller & Messenger and Huss et al. this volume; see also section 3.4.3).

However, no studies to date have addressed the fact that these radially diffusing particles must run a gauntlet of large particles, onto which they may be swept up or accreted as they diffuse outwards. Once accreted onto a large particle, the diffuser is either immobilized (if its host is large enough) or even quickly transported radially inwards at a rate much faster than its outward diffusion rate! So, there is an effective "loss term" for outwardly diffusing particles which will lead to a steeper radial gradient than found by these previous studies. This process needs to be studied in more detail. An example of its application is noted in section 3.2.1.

Finally, stellar winds might account for outward mixing of tiny grains, as well (Shu, Shang, & Lee 1996; *cf.* also Cuzzi & Weidenschilling 2005).

One aspect of radial mixing is the ability of particles to escape the parcel of gas in which they formed, before it cools sufficiently to alter the mineralogy of the particle. The presence of non-equilibrium minerals in CAIs, AOAs, and other objects has led to the belief that these particles must be removed, or isolated, from the gas in some way before alteration can occur (Krot et al. 2002a; Wood 2004). Departure from the region of formation is one way to achieve this. Burial in a large, immobile planetesimal has also been suggested – but presents very serious problems (Wood 2004; also section 2.6). The main question is, on what timescale must this removal occur? Radial diffusion in turbulence, and radial ejection in a stellar wind, to take two specific examples, occur on very different timescales. It may be that mineralogical tests can distinguish between these different scenarios (see section 3.3.1).

In a nebula which is *not* turbulent, perhaps at the end of a prior stage which was turbulent, particle growth can be quite different. Even fairly small particles can settle into a layer that is considerably denser than the nebula gas before the very presence of the layer itself generates enough (highly localized) turbulence to halt further settling. What qualifies as nonturbulent depends on the particle size: for meter size particles $\alpha < 10^{-4}$, but for mm-size particles $\alpha < 10^{-7}$ (Cuzzi & Weidenschilling 2005; see Fig. 4)! These values are much smaller than estimates based on accretion rates (section 2.1). In such dense layers, the particles dominate the motion of the gas rather than *vice versa*, and radial drift gets much slower. At the same time, relative velocities between particles get smaller. Smaller relative velocities, combined with higher feedstock densities, allow growth to proceed quite quickly beyond the meter-size range (Cuzzi et al. 1993; Weidenschilling 1997). It is this regime for which various midplane instabilities have been proposed, as discussed by Cuzzi & Weidenschilling 2005). Nonturbulent growth scenarios – both incremental growth and instabilities - share the characteristic that growth occurs relatively rapidly, without significant recycling or radial mixing of solids (see below). Overall, turbulence is fundamental to the growth process, and one expects that there should be observable differences between the predictions of turbulent and nonturbulent scenarios. In section 3.1 we mention several candidate observations. Further complexities (which we will not discuss here) might include intermittent turbulence, allowing alternating periods of extended dispersal and planetesimal formation.

Once large (> 100 m) objects are able to form, radial drift slows or ceases because of mass/area effects (Fig. 3); these growing planetesimals can accrete smaller, drifting material as well as outwardly diffusing and condensing vapor. This might be regarded as the final stage of accretion. We discuss this stage further in sections 2.3 and 2.5.

2.3. Evaporation Fronts

Inward radial drift, especially of meter-size particles, can lead to very significant inward mass flow rates of solids. The fate of these particles has largely not been considered, and it is often merely assumed that they are lost into the Sun. However, a more interesting (if less dramatic) fate may await them. Meter-size particles reside near the midplane unless turbulence is implausibly large, and the temperature at the midplane increases inwards at any given time. Drifting particles become warmer and, at some point, enter a region where the temperature exceeds the evaporation temperature of some or all of their material. Within a radial span that depends on the evaporation rate, but is generally smaller than the distance to the Sun, the particle evaporates (it drifts more slowly as it evaporates so the process culminates rapidly). Once the material is vaporized, it is coupled to the local nebula gas and can evolve only by advection with that gas, or by gradient diffusion. Since gas advection is much slower than the radial drift of most solids (Fig. 3), the local abundance of material in the vapor builds up until diffusion and advection can balance the inward flow. This enhancement of material in the *vapor* form *inside* its evaporation/condensation boundary will characterize *all volatiles*; it has been studied in the context of silicates by Cuzzi et al. (2003) and in the context of water by Cuzzi & Zahnle (2004) and Ciesla & Cuzzi (2005). The effect complements and differs from enhancement of solids *outside* the evaporation/condensation boundary (the cold-finger effect) as studied for water by, for instance, Stevenson & Lunine (1988) and Cyr, Sears, & Lunine (1998), which rely on some immobile traps outside the condensation boundary. An early study by Morfill & Völk (1984) included both effects, but was applied to particles too small to show much of an effect.

Cuzzi & Zahnle (2004) identify three distinct regimes of this process (Fig. 5): in *regime 1,* shortly after the inward flow of solids becomes significant, a “plume” of gaseous volatile material can appear just inside the evaporation boundary. In steady state, after some time t_{ss}, the entire region inwards of the evaporation front can fill up with an enhanced abundance of the volatile (*regime 2*). In this simple model, the enhancement parameter $E_o = f_L/\alpha$, where f_L is the mass fraction of solids in meter-sized rubble. Both of these enhancement regimes would lie within the first stage of accretion mentioned above. However, if immobile trapping planetesimals form outside the condensation boundary, and accrete both inwardly drifting meter-sized rubble and outwardly diffusing condensable vapor, the inner nebula can be dried out onto this cold finger (*regime 3*; Stevenson & Lunine 1988). The value of t_{ss} is given by either the inward advection or diffusion time, depending on nebula viscosity and diffusivity (they are different); in the context of the "waterline" (evaporation front of water ice), t_{ss} is in the range of a few times 10^5 to a few times 10^6 years - thus very interesting for compositional variation during nebula evolution of chondrite constituents and parent body formation. A similar scenario applies to silicates earlier in solar system history, when the inner nebula is hot enough for an evaporation front to exist for silicates. While the simple, analytical steady state solutions of Cuzzi & Zahnle (2004) indicate that enhancements can reach several orders of magnitude, the numerical solutions of Ciesla & Cuzzi (2005) find them to be typically more modest (for the water evaporation front), a factor of several up to a factor of 30 or so in regime 1, because of finite supply considerations.

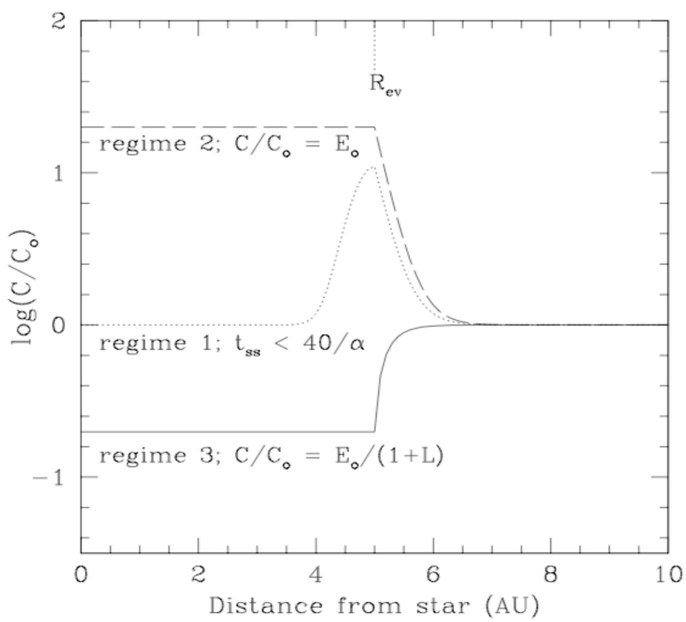

Figure 5. Enhancement of the concentration of some volatile C above its cosmic abundance C_o, as a function of distance from the Sun. A water evaporation front is assumed at R_{ev} = 5 AU. Transient regime 1 evolves into steady state (regime 2) by time t_{ss}. A substantial sink of immobile planetesimals just outside the condensation boundary can serve as a "cold finger" and dry out the inner nebula in regime 3. From Cuzzi & Zahnle (2004).

This process, together with the radial drift described in section 2.2, has a number of meteoritical and even astronomical implications. First, the concept of "cosmic abundance" has meaning only in the most globally averaged sense. This was first emphasized by Stepinski & Valageas (1997), who noted that some nebula regions could be greatly enhanced in solids at the expense of other regions. It now appears that this enhancement extends to the vapor phase, which can affect the chemistry, mineralogy, redox state, and even isotopic composition of objects which *are* solid (section 3.4). Various aspects of meteorite constituents (redox states of major silicates in AOAs, Type I and Type II chondrules, and enstatite chondrites), seem to call for non-cosmic abundances (both enhancements and depletions) of *oxygen*, which could be related to the *water* evaporation front rather than, or in addition to, fractionation and evaporation of *silicate* material (which *is* required to stabilize silicate *melts*). Several authors have noted the difficulties of matching the observations merely by enhancing CI material (Wood & Hashimoto 1993; Fedkin & Grossman 2004; Ebel & Alexander 2005). Sulfur is another obvious player in this game (Pasek et al. 2005). Something else that must be considered in general is ongoing and possibly irreversible refinement of nebula solids as this process unfolds. For instance, the most primitive nebula solids, like cometary dust, contain a significant amount (20-30 wt%) of refractory carbon phases, which is significantly higher than in CI chondrites (Fomenkova 1997; Wooden 2002). Perhaps carbon was combusted away as primitive silicate-carbon rubble evaporated, with the carbon becoming stable CO and the silicates being recycled. In another application, perhaps the isotopic content of outer solar system solids (meter-sized snowballs) was carried inwards to alter the oxygen isotopic content of inner solar system chondritic silicates (Yurimoto & Kuramoto 2004; Krot et al. 2005a). We discuss these ideas in section 3.4.2.

In the discussion above regarding water, it was assumed that solids were made of pure water ice and that water vapor was able to readily escape from their surfaces. In reality, solids in the outer nebula would be mixtures of ices, silicates, and refractory organic material. As these objects migrate inside the snow line, water near the surface would likely vaporize and escape easily, resulting in a more refractory rind forming on the outside of the body. Some of the fine silicate and/or organic material which had been embedded in this frost might also escape into the local nebula gas; whether it accumulates there along with the water vapor, or reaccretes on a larger migrating body to be removed, needs more careful modeling. While temperatures inside these objects continue to climb, the buried water may vaporize, but only slowly diffuse outwards through the rind. This could allow the water vapor pressure in the interior of these objects to increase to the point where liquid water becomes stable, resulting in hydration reactions between the water and the silicates/organics. The extent to which this reaction proceeds and the amount of hydration that occurs has been calculated for silicates in large bodies (Cohen & Coker 2000), but it has not been studied in detail for objects 1-10 meters in size. If hydration takes place on these bodies before they are rearranged by collisions, then some of that water will be stabilized in silicates to quite high temperatures (possibly helping us explain how the Earth got its own water). However, we feel it is even more likely that, for objects of these sizes, collisions are frequent enough to continually expose the buried water so that it can then be reintroduced to the nebula. More detailed study is needed.

2.4. The Chondrule Formation Processes

Several chapters are devoted to this subject, and we will not dwell on it. We do want to note that there might have been *different* chondrule formation processes responsible for different chondrite classes, which imply different formation environments. For instance, chondrules in the CH and CB carbonaceous chondrites are looking increasingly like they formed not only much later than chondrules in most other chondrites (Amelin et al. 2004), but in a very different environment and probably by a very different process (Wasson & Kallemeyn 1990; Rubin et al. 2003; Krot et al. 2002a, 2004a, 2005c).

We note that chondrule formation processes - which to most people mean chondrule *melting* processes (Desch et al, this volume) - might have other interesting applications. As one example, relative velocities between chondrules in plausible turbulence (meters per second) seem too feeble to break chondrules (velocities are even lower in nonturbulent scenarios), and yet a significant fraction of chondrite material appears to be "clastic fragments" of once-molten, larger, droplet chondrules. A process is needed to break up chondrules which is independent of ambient turbulence; efficiency of hypotheses would favor formation mechanisms, such as shocks, which can also do this. Finally, we note that chondrule-forming processes must allow for significant post-formation and post-cooling evolution to mix together objects which *were* melted with those which apparently *escaped* melting (section 3.3.2).

2.5. Primary Accretion

We define *primary accretion* as the process by which the parent bodies of the primitive chondrites were put together, from what appear to be individual mineral objects which floated around in the nebula. These parent bodies are "sedimentary rocks" in that their constituents derive from a variety of distinct environments but are brought together, perhaps over long distances and times, by some common process. We envision as an archetypal chondrite parent body the parent body of the H chondrites, which is inferred to be a single object, roughly 100 km in radius, composed initially of H3.0-like material with a homogeneous distribution of chondrules, refractory inclusions, and matrix (McSween et al. 2003; Ghosh et al. 2003; Trieloff et al. 2003). In any region of the nebula, a number of objects with comparable, but not identical, properties, might be forming at the same time. We seek to understand how such objects can form. Most chondrites also show the effects of planetary processing - collisions (sometimes violent and disruptive), thermal and aqueous alteration, and even melting or dissolution which obliterate the primitive fabric of their predecessors. These *planetary processes* would constitute yet a third stage (following early incremental accretion and true primary accretion; Weidenschilling & Cuzzi 2005). However, we need to peer back through these subsequent processes to glimpse the primary processes which brought the parent bodies together in the first place.

We distinguish the primary accretion stage from the incremental accretion which probably started with the earliest days of the nebula, but which might have stalled at meter size for a long time. Particles can be stored in meter-size chunks for indefinite times without any irreversible effects, but once they become part of something larger than a few km or so, their properties might be noticeably altered (sections 2.6 and 3.1.3). Incremental growth *might* lead to primary accretion, if nebula turbulence vanishes and objects can bypass the meter size barrier in a dense midplane layer. Also if nebula turbulence vanishes, one or more of the various midplane instabilities might play a role (but probably only if the density of solids were enhanced over cosmic by at least an order of magnitude). It is in the nature of all these midplane accretion scenarios that accretion happens quickly once the appropriate conditions are met in some part of the nebula. Incremental growth occurs on the 10^3-10^4 year timescale, and 10^5 years or so after the onset of nonturbulent conditions, tens of thousands of 100-km radius bodies can have formed in the asteroid belt region, which can actually be problematic as discussed further in section 3.1.3. Midplane instabilities are discussed in more detail by Ward (2000) and Cuzzi & Weidenschilling (2005).

If nebula turbulence is ubiquitous in space and time, at an intensity > $\alpha \sim 10^{-5}$ or so, only one primary accretion process has been advanced so far which might overcome the meter-size barrier - *turbulent concentration* (Cuzzi et al. 2001). In this process, particles that have a stopping time in the gas equal to the overturn time of the smallest eddy in turbulence are selected for concentration, by orders of magnitude, relative to their average abundance. This possibility was actually first imagined by John Wood (1963a,b). This theory makes some encouraging predictions about the properties of concentrated particles (section 3.1.1) but is incomplete, in the sense that it has not yet been shown that it leads to anything more than particle-rich zones in the nebula which are still of far lower density than an actual planetesimal. Work is in progress on delineating the subsequent steps between high-particle-density zones and actual planetesimals. While details are beyond the scope of this chapter, below we sketch the direction of current work.

A look ahead: Particle-rich, high concentration zones favor regions where the vorticity, or angular velocity, of the gas is a local minimum (Cuzzi et al. 2001). Our current thrust is to determine the volume fraction in the nebula which contains zones of sufficiently high particle density, in regions of sufficiently low angular velocity, for the self gravity of the particle-rich zone to become important. It appears that the role of gas pressure in precluding actual gravitational instability has been widely overlooked (Sekiya 1983; Cuzzi & Weidenschilling 2005) and will need to be closely explored in future work in the particle size regime of interest.

Furthermore, a region having particle density orders of magnitude larger than the gas density will experience solar gravity as a unit and begin to move as a unit relative to the surrounding gas (which is orbiting at a different speed). The ensuing relative velocity leads to a ram pressure on the clump which can disrupt it, at least in the absence of other forces. For example, consider a drop of liquid falling in a less dense liquid - say, ink in water. The pressure induced by the terminal velocity of the dense liquid disrupts it and it dissolves in the less dense liquid. However, where the dense liquid has a surface tension - say water in oil - the situation is different. Surface tension holds the drop together against ram pressure, and it settles as a coherent unit at terminal velocity. Self gravity in the clump may play a role analogous to surface tension; expressions can be derived which constrain the particle mass density needed to stabilize dense clumps against the ram pressure associated with their relative velocity through the gas. Dense clumps, once stabilized by self gravity and either collapsing quickly under self gravity, or even shrinking more slowly because of gas pressure effects, might undergo an extended coagulation scenario with each other in the nebula as they evolve. Parameters of the problem indicate that the initial stable agglomerates are much more massive than the problematic meter-size range; thus, in principle, this scenario might allow us to leapfrog directly to stable planetesimal size objects even in ubiquitous nebula turbulence. Numerical simulations of this process require modeling more complex physics (compressible entrained gas) than presently available. Further studies are needed to quantify whether particle-rich regions can indeed become actual planetesimals, and if so, at what rate.

While the later stages of this process remain speculative at this time, it seems to have some potential for leading to *sporadic* but *ongoing* production of planetesimals even while turbulence persists. At any given time and place, all the newborn planetesimals would share the same physical characteristics (i.e., particle size, as selected by turbulent concentration), so they could collide and merge without noticeable changes in their properties except possibly increasing compression. Each parent object (perhaps a few hundred meters across) would represent a sample of local particles as constrained by their aerodynamic stopping time (section 2.2). Even if radial mixing homogenizes the nebula radially over several to ten AU lengthscales, this local mix might change chemically or isotopically with time by virtue of a change in its gaseous environment, its temperature, or other global-scale property, and these variable properties might be preserved as *temporal* snapshots of an evolving nebula mixture. In this scenario, primary accretion could occur continually - and sporadically – even while stage 1 type diffusion and drift of meter-and-smaller objects continues.

2.6. Modeling Thermal Evolution of Parent Bodies

Objects larger than a certain size can build up heat produced by their radioactive materials faster than they can radiate it away; their internal temperatures increase, producing first thermal alteration and eventually melting of their silicates. The literature on this subject contains a number of increasingly detailed thermal evolutionary models of the internal temperature of large (100 km size) objects, most recently including growth over time (Ghosh et al. 2003). Woolum & Cassen (1999) find that objects larger than 5 km radius which contained the fractional abundance of radiogenic ^{26}Al/^{27}Al found in typical CAI material, would melt in, at least, their central regions. Parameters include the unknown thermal conductivity of the subsurface, and of the porous, insulating, regolith layer. Complications include whether or not early melt (basalt) moves to the surface, carrying with it an inordinate amount of radiogenic atoms. La Tourrette & Wasserburg (1998) state that objects larger than 15 km radius, even if formed 0.5 Myr *after* CAIs, would reset Al-Mg systematics in more than half their volume. Because the heat production increases with volume, and the heat loss with area, one can estimate that an object of, say, 10 times this large in radius − about 100 km, or the size of asteroid 6 Hebe (the putative H chondrite parent body) would have to wait to accrete until the ^{26}Al had decayed to 10% of its canonical abundance ($^{26}\mathrm{Al}/^{27}\mathrm{Al} = 5\times10^{-5}$) − about three half-lives, or 2 Myr. This very crude estimate is consistent with more sophisticated models by Grimm & McSween (1993), McSween et al. (2003) and Ghosh et al. (2003). Presence of water ice in the initial body would complicate the evolution further, leading to less melting. However, for most meteorite parent bodies, water seems to have been present in only small to moderate abundance, and its effects may not change the picture significantly. The role of ^{60}Fe must be allowed for in future models of this sort, as its abundance becomes more well constrained (Yoshino et al. 2003) . Thermal models of this sort are essential for interpreting well-defined closure ages in terms of accretion or formation ages (section 3.1.3).

3. Chondrites: What They Do, or Could, Tell Us About the Models

3.1. Physical Properties

3.1.1. Size Distributions

The most striking physical property of most chondrites is that they are composed mostly of *chondrules* and *chondrule fragments* (rimmed with fine dust, interspersed with fine-grained matrix material, and accompanied by a sprinkling of other kinds of particles as discussed below). As mentioned in section 2.5, there is reason to believe that all primitive chondrite bodies were initially composed *entirely* of this same sort of mixture – with variations on this theme defining the different groups. Furthermore, several studies have found that certain very primitive chondrites, or primitive clasts within chondrite breccias, seem to have escaped much of the collisional fragmentation, grinding, and physical stirring on the parent body which characterized subsequent planetary processing in the asteroid belt. In these rare meteorites or clasts, *only* dust-rimmed chondrules seem to be present – nestled against each other as if only gently compacted. This material is referred to as "*primary texture*" (Metzler, Bischoff, & Stöffler 1992, Brearley 1993 – see Figure 6), and we suggest it corresponds directly to primary accretion. While some chondrites are essentially solid rocks, many chondrites can be disaggregated with only mild effort – freeze/thaw cycles, acoustic vibration, *etc*. with entire chondrules emerging intact, complete with their fine-grained rims.

The well-defined and highly characteristic size distribution of chondrules is one of their more remarkable features. In a number of studies, Hughes (1980a,b) and Martin & Hughes (1980) noted that both rounded, complete, droplet chondrules and broken fragments of originally *larger* chondrules share the same size distribution. Also, tiny "microchondrules" are found embedded within chondrule rims along with other fine grained material (Krot et al. 1997). The average chondrule size varies between chondrite groups, with CO having the smallest (CH chondrites are not considered here) and CV the largest (Grossman 1988), but in the five ordinary and carbonaceous chondrites explored carefully by disaggregation, the size distributions, once normalized to the median size, were essentially identical (Fig. 7; Paque & Cuzzi 1997; Cuzzi et al. 2001). Thus, the chondrule *formation* process produced a wider range of molten chondrule sizes than normally observed in an individual chondrite, and some subsequent process imposed the same narrow size distribution on the accreted *chondrule fragments* as on the unbroken droplet chondrules (Fig. 8 shows some examples of chondrule fragments). Specifically, the very good agreement shown in Figure 7 between the size distribution predicted by turbulent concentration (which has, essentially, no free parameters regarding its shape) and the observed size distributions in a number of different meteorites (when appropriately normalized) is what we call one of the "fingerprints" of turbulent concentration. The meteorite data, originally obtained in terms of the measured chondrule radius-density product $r\rho_s$, was simply renormalized to a Stokes number (section 2.2) $St_\eta = \Omega t_s = 1$ at the peak concentration, as predicted by turbulent concentration (Cuzzi et al. 2001) so we are really seeing the constancy of the *shape* of the distribution here. The fingerprint is found in ordinary and carbonaceous chondrites. More observations along these lines would be quite valuable.

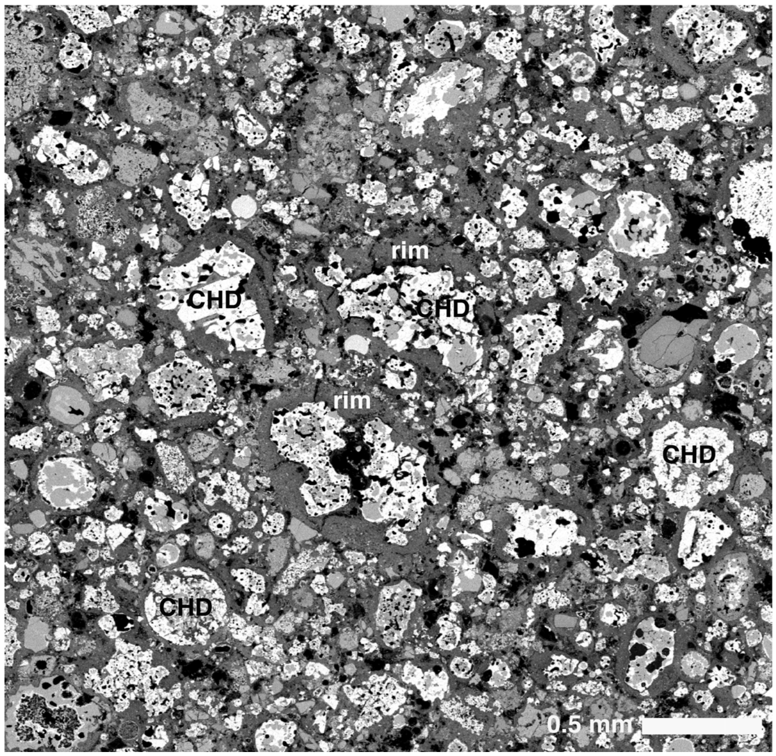

Figure 6. Elemental map in Mg X-rays of the ungrouped carbonaceous chondrite MAC88107. The meteorite is unbrecciated and shows primary accretionary texture; *i.e.*, all coarse-grained components (CAIs, AOAs, chondrules, and chondrule fragments) are surrounded by continuous, fine-grained accretionary rims. The Mg-rich chondrules (CHD) are the bright and light-grey objects. The dark-grey material is the more Fe-rich fine-grained rims, which surround all the larger objects. Black regions are Fe,S opaque assemblages and empty space.

Other evidence for aerodynamic processing of chondrite components was reviewed in Cuzzi & Weidenschilling (2005), and will not be repeated here in detail. If aerodynamics ruled, everything should scale with the particle stopping time t_s (section 2). Metal grains (except in the unusual CB chondrites) are typically smaller than chondrules by an amount consistent with their larger density (Skinner & Leenhouts 1993; Keubler et al. 1999), but their irregular shape confuses the situation. Size distributions have never been systematically measured for CAIs or AOAs (*cf.* May, Russell, & Grady 1999). Because CAIs and AOAs are very metal-poor, they have lower density than chondrules and would be expected to be larger in general, if aerodynamically sorted together with chondrules. Scott & Haack (1993) note that density difference alone is insufficient to explain the large size differences between chondrules and CAIs in CV chondrites. However, many CAIs and AOAs are also porous or "fluffy"; this factor should be addressed. Sometimes one finds lumps of material called dark inclusions, which are matrix-rich but contain tiny chondrules and which appear to have been gently emplaced – sometimes after acquisition of fine-grained rims (next section); the sizes of these lumps are comparable to those of other objects in the same chondrite (*cf.* Krot et al. 2004e).

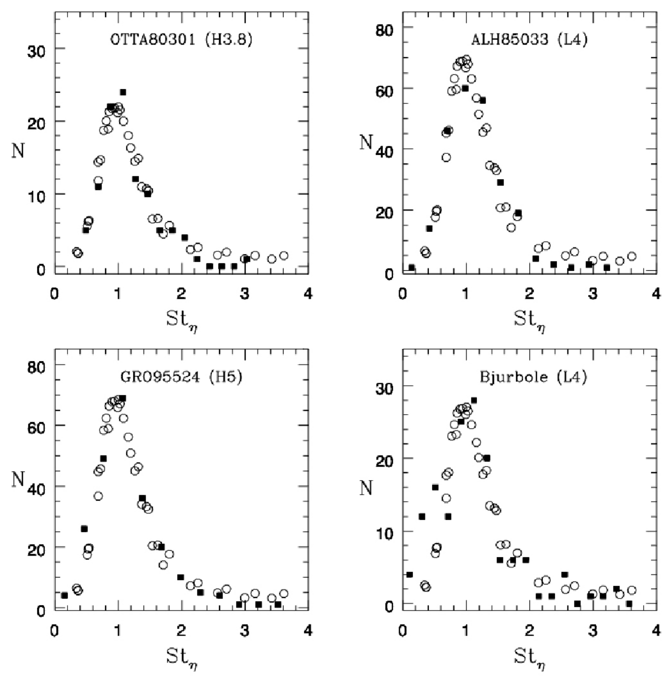

Figure 7. Chondrule size distributions from four different ordinary chondrites (filled symbols) compared to predictions of turbulent concentration (open symbols). The quantity N plotted is simply the number of disaggregated chondrules observed in each size bin; the chondrule radius-density product was converted to a Stokes number (horizontal axis). The theoretical profile, obtained from numerical simulations, has no free parameters (Hogan & Cuzzi 2001). See also Paque & Cuzzi (1997) for Allende (CV3).

Establishing whether CAI/AOA/matrix lump size distributions were, or were not, aerodynamically the same as chondrules and fragments in the same meteorite would be a valuable test. These types of particles are often irregular in shape and porous besides, so determining their size-density product is tricky. In a general sense, chondrite groups with large chondrules (CVs) also have large CAIs, and groups with small chondrules (COs) tend to have small CAIs (Scott et al. 1996). However, CRs also have large chondrules, like CVs, but virtually lack large CAIs (Krot et al. 2002a; Aléon et al. 2002). Part of the explanation might be temporal evolution. Even with turbulent diffusion, large (5-10 mm) CAIs and AOAs are harder to preserve against gas drag drift than are smaller ones, so groups containing them must be older than groups lacking them (Cuzzi et al. 2003). Results of Amelin et al. (2002, 2004) and Bizzarro et al. (2004) indicate that chondrules in CVs are older than those in CRs; so, perhaps there were simply not any large AOAs or CAIs left around when CRs accumulated.

3.1.2. Fine-Grained and Coarse-Grained Rims around Chondrules and CAIs

Chondrules and refractory inclusions in primitive chondrites are typically surrounded by fine-grained rims (Metzler et al. 1992; Figs. 6 and 8). Intriguing relationships have been found between rim thickness and core chondrule size (Metzler et al. 1992;

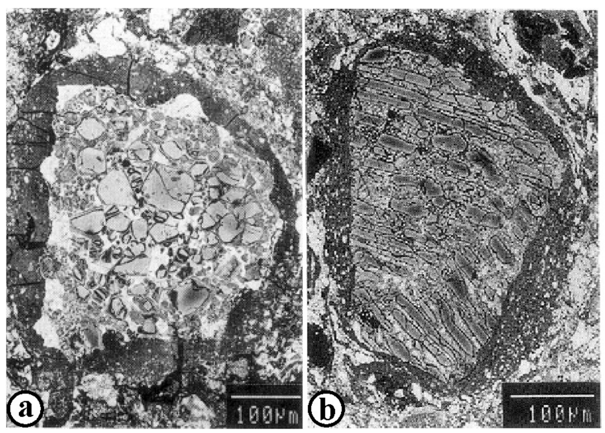

Figure 8. Backscattered electron images of fine-grained accretionary rims around broken fragments of chondrules in CM chondrites (Metzler et al. 1992). Notice how the fine-grained material fills in irregularities in the surfaces of the chondrules.

Paque & Cuzzi 1997). Morfill et al. (1998) and Cuzzi (2004) presented models of how fine-grained rims could form in the nebula, and showed that rim thicknesses are diagnostic of ambient dust density and the time duration between melting of a chondrule and its accretion into a parent body. Wasson et al. (2005) present concerns about rim porosities which can be produced this way. Although Bunch et al. (1991) and Symes et al. (1998) have argued that fine-grained rims formed as a result of impact processing on asteroids, studies of distributions of noble gases (Nakamura et al. 1999) and radiation tracks (Metzler 2004) have shown that rims did not form on asteroids. Bland, Prior, & Hough (2003) have detected a layered structural fabric of ferrous olivine grains in fine-grained rims around chondrules in the Allende meteorite, which seems to support a gentle, extended, accretion evenly from all sides, as the chondrule floated and tumbled through the nebula gas for an extended period. We note, however, that ferrous olivine in the Allende matrix (and thus perhaps also fine grained rims) probably formed during extensive alteration on the CV asteroidal body (Krot, Petaev, & Bland 2004; Nuth, Brearley, & Scott this volume). Thus, it is important to extend these observations to unaltered meteorites.

In many chondrites, there is also interstitial fine-grained matrix material, which is chemically and mineralogically similar to fine-grained rims, but often more porous and coarser-grained (Scott et al. 1996). The discovery of clasts, or entire meteorites, containing primary accretionary texture (Metzler et al. 1992; section 3.1.1) suggests that fine-grained dust in the very first sizeable bodies may have been accreted mostly, or only, on the surfaces of chondrules and other large particles, and that the more common, interstitial matrix *per se* might be a product of grinding, abrasion, and aqueous alteration on parent bodies. In pristine carbonaceous chondrites, rims and interstitial matrix material are essentially identical in mineralogy and composition (Brearley 1993; see Nuth et al. this volume). CI chondrites are exceptional as they lack chondrules, perhaps due to overwhelming aqueous alteration, and consist almost entirely of matrix material.

Some chondrules are surrounded by coarse-grained *igneous* rims (Fig. 9), thought to result from melting of fine-grained rim precursors by some "flash heating" event too weak to melt the entire chondrule (Krot & Wasson 1995). We note, however, that fine-grained accretionary rims are dominated by ferrous silicates, whereas coarse-grained igneous rims are compositionally similar to the enclosed chondrules; e.g., Type I chondrules are surrounded by magnesian silicates, whereas Type II chondrules are surrounded by ferrous silicates (Krot & Wasson 1995). In addition, there are no mineralogically transitional rims between the fine-grained accretionary and coarse-grained igneous rims. Based on these observations, we infer that there is no simple genetic relationship between fine-grained accretionary rims and coarse-grained igneous rims.

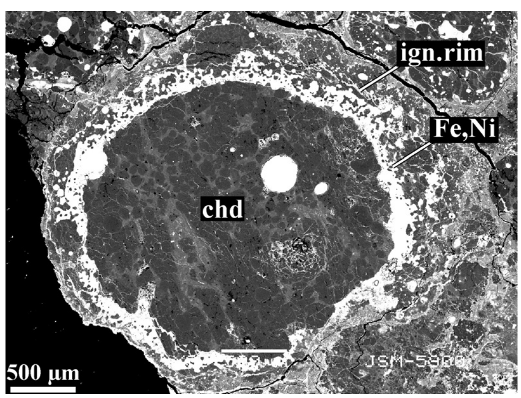

Figure 9. Backscattered electron image of a porphyritic olivine-pyroxene magnesian chondrule (chd) from the CR carbonaceous chondrite EET87747. The chondrule is surrounded by a continuous layer of Fe-Ni metal (Fe,Ni) and a coarse-grained igneous rim (ign. rim) composed of low-Ca pyroxene, high-Ca pyroxene, glassy mesostasis, and Fe,Ni-metal.

If CAIs were transported from the hot, inner solar nebula to the cooler region where chondrules formed, one would expect that accretionary rims around CAIs would have recorded spatial or temporal differences between the CAI- and chondrule-forming regions. MacPherson, Hashimoto, & Grossman (1985) described accretionary rims around Allende CAIs. These rims consist mainly of crystalline ferrous olivine and Ca-Fe-rich silicates [wollastonite ($CaSiO_3$), andradite ($Ca_3Fe_2Si_3O_{12}$), hedenbergite ($CaFeSi_2O_6$)] and often show mineralogical layering which was thought to reflect variable physico-chemical conditions in the solar nebula. It was later shown that these minerals resulted from alteration on the CV parent body (Krot et al. 1998a,b). Krot et al. (2001b, 2004b) described less altered, coarse-grained, forsterite-rich accretionary rims around CAIs in the reduced CV chondrites Efremovka, Leoville, and Vigarano. These rims lack secondary minerals and consist of forsterite, Fe,Ni-metal, Al-diopside, anorthite, and minor spinel. Mineralogical and isotopic observations and thermodynamic analysis suggest that these rims are aggregates of condensates from an ^{16}O-rich nebular gas, much like AOAs, suggesting formation in the CAI-AOA forming region. The forsterite-rich accretionary rims are surrounded by fine-grained accretionary rims, which appear to be similar to fine-grained rims on chondrules. However, systematic studies of primitive chondrites are required to check this.

3.1.3. Melting/Parent Body Formation Timescales

If ^{26}Al was the primary heating source and was homogeneously distributed in the protoplanetary disk (section 2.6), igneously-differentiated asteroids larger than 50 km in diameter or so must have accreted less than 2.5 Myr after formation of CAIs, and, conversely, the parent asteroids of (unmelted) chondrites must have accreted at a later time (Fig. 7b; McSween et al. 2003; Woolum & Cassen 1999). Current radiometric ages of a number of chondrites and achondrites are consistent with this scenario, within their stated uncertainties (see reviews by Keil 2000 and Shukolyukov & Lugmair 2003).

For instance, detailed modeling studies involving various closure ages have derived accretion ages for Vesta, the putative HED parent body, and for Hebe, the putative H chondrite parent body. Even allowing for gradual accretion, Hebe had to accrete no sooner than 2.7 Myr after CAIs (Ghosh et al. 2003). McSween et al. (2003) have Vesta forming initially at about 2.9 Myr after CAIs, and differentiating at 4.6 Myr after CAIs. These results are consistent with ^{53}Mn-^{53}Cr isotope dating (Shukolyukov & Lugmair 2003); different *absolute* ages for CAIs derive from different isotope systems, but the age *differences* inferred are both consistent with a few Myr hiatus between CAIs and chondrules. Other achondrites dated fall in the same age range, or later (Shukolyukov & Lugmair 2003). Formation of an iron core is a key aspect of asteroid accretion and evolution. Because of their relative strengths there are more examples of asteroid iron core fragments than of igneous asteroid surfaces (the associated achondrites are preferentially destroyed in transit); there are about 60 different asteroid cores represented amongst the iron meteorites (Burbine, Meibom, & Binzel 1996). Sugiura & Hoshino (2003) have estimated using a thermal model that about 1.7-1.9 Myr had to pass before formation of the 15-20 km radius parent body of the IIIAB iron meteorites. It could be important to extend studies like this to the other iron meteorite groups, to see if any contain evidence for extremely early accretion (interesting results will have uncertainties of less than a few times 10^5 years). The role of ^{60}Fe as an additional heat source should also be allowed for.

In addition to radioisotope age dating of meteoritic samples, the relative abundances of melted and unmelted asteroids, as determined telescopically, can be applied to theories of parent body formation. Studies of the reflection spectra of asteroid surfaces (Gaffey et al. 2003; Binzel 2003) and the relative numbers of achondrites (Meibom & Clark 1999; Keil 2000) suggest that, while a significant proportion of both asteroids and meteorites show *some* evidence for melting, or at least heating, only a small fraction show evidence for near-total melting. However, the evolution of the surfaces of these objects is very complex (see, e.g., Keil 2003; Scott 2003), and more work – telescopic observations, sample return, and modeling – is needed to infer their initial melting state from their current reflectivity.

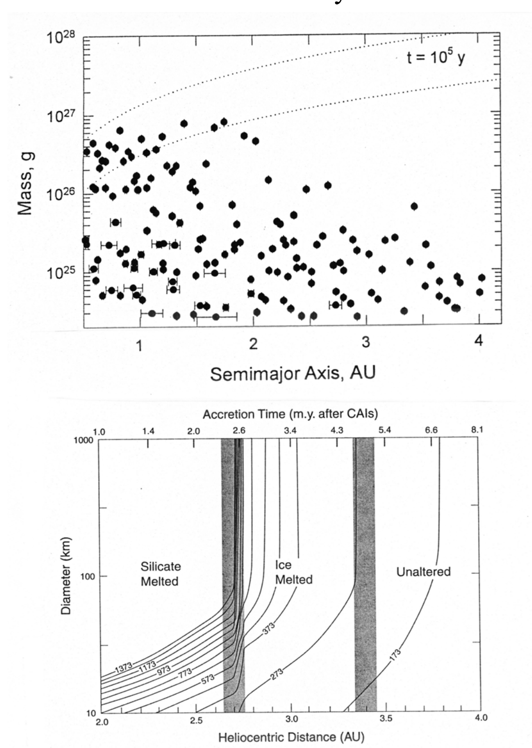

Figure 10. top - Model results of incremental accretion in a nonturbulent nebula; the dots are individual objects, and abundances increase in a powerlaw fashion to even smaller sizes. Our moon has a mass of about 10^{26} g, and Ceres has a mass of about 10^{24} g. A 100 km radius Hebe-sized parent body has mass of about 10^{22} g. Dotted lines represent a range of estimates of the mass a growing object can reach before it isolates itself from surrounding material. These results are comparable to those in Ghosh et al. (2003), who assumed a nebula with a steeper density falloff. In either model there are many, many objects the size of Hebe after only 10^5 years of accretion. bottom - Results from McSween et al. (2003), showing how silicate bodies larger than about 80 km diameter would melt from ^{26}Al heating if ^{26}Al were homogeneously distributed and they formed less than 2.5 Myr after CAIs. Contours are labeled by their central temperatures (degrees K).

The simplest explanation of the achondrite *and* asteroid data, taken together, seems to be that most asteroid formation was delayed by 1-2 Myr or more – consistent with the age spreads of chondrules; this is readily explained if the nebula were weakly turbulent (sections 2.2 and 2.5). However, clearly, puzzles remain. Moreover, not all achondrites (or chondrites) need to have formed at the same time. The general properties of achondrites (e.g., oxygen isotopes, Clayton & Mayeda 1996) suggest that some of their parents may have formed from components much like those in some known chondrites. CV chondrites contain what seem to be the oldest chondrules; how large could the CV parent have been, without melting? Perhaps other, larger, objects which formed from this same material did melt and become achondrites. Formation of chondrite (and achondrite) parent bodies might have continued for quite a long time, and in parallel. More, and more precise, studies of crystallization ages on more samples would be useful.

These constraints have important implications for the primary accretion process and its timing. Figure 10a shows results of an incremental accretion model after only 10^5 years in a nonturbulent nebula for a nebula density profile proportional to R^{-1} (Weidenschilling, unpublished), which is only slightly different from that published in Ghosh et al. (2003) for a steeper density profile. Based on the number of 10^{25} g objects seen in these results, and assuming equal mass per decade (Weidenschilling 1977, 2000, 2003), one estimates that tens of thousands of Hebe-sized objects also had formed by this time. Models of asteroid belt formation (Chambers 2004) suggest that the current belt is a randomly selected, but tiny, fraction of the original belt – perhaps 1%. If, as in Figure 10a, all these Hebe-sized objects formed in such a short amount of time (10^5 years after CAIs), they would all have completely melted.

From the modeling standpoint, to go beyond the current sophistication of thermal models and derive more precise *accretion times* from observed *crystallization times* is a challenge, but perhaps not an insurmountable one. The onset of melting must be characterized under a range of parameters and conditions. The effect of core formation and the removal of early melt must be considered. Formation of these two kinds of parent bodies may well have been going on simultaneously, after some initial delay, with the larger and/or earlier ones becoming achondrites and the smaller/later ones remaining chondrites.

3.2. Chemical Properties

Some of the most fundamental properties of chondrites, which help to classify them into groups, are chemical differences (Krot et al. 2004c and references therein). Some of these relate to major rock-forming elements such as Fe, Si, and Mg. There are strong signatures of fractionation of various elements – enhancement or depletion relative to "solar" (as represented in practice by CI chondrites). Some of these fractionations are clearly related to volatility of the element (temperature at which it enters solid minerals). Even this is somewhat complicated by the fact that local pressure and even chemistry can affect volatility. Condensation of sulfur, as one example, is strongly affected by the oxidation state of the gas (Larimer 1975; Pasek et al. 2005). In turn, this might affect the abundance of the chalcophile elements in different groups (see, e.g., Sears & Dodd 1988, their Fig. 1.1.10).

In both this section and section 3.3, a critical role will be seen for differential evolution of the particles which ended up in meteorities, from their parent nebula gas. Meteoriticists allude to this possibility by distinguishing between chemical/mineralogical evolution in *closed* and *open* systems, where closed system behavior would imply that a growing and cooling solid always remained in perfect chemical communication with its original parent volume of gas, and open system behavior implies that the nature of the gas surrounding a cooling (or heating) particle changes independently over time.

3.2.1. Volatility Fractionation

One fractionation that is of obvious importance to nebula evolution of solids is the fractionation of moderately volatile elements (Palme & Jones 2004). It is generally believed that volatile depletions resulted from nebula-scale thermal processing of dust and gas in the protoplanetary disk, when high-temperature solids accreted before more volatile elements could condense, and all the while the gas, containing these volatile elements, was being removed (Wai & Wasson 1977; Cassen 1996; see, however, Yin, this volume, for a different view). Whatever the cause of this depletion, it must represent the physicochemical processing of individual constituents, including gas-solid mixing of some kind, because at some local temperature an element will be partitioned between the gas and one or more condensed phases, with the fraction of element present in the gas (while dependent upon the ambient pressure and chemical composition) being very well determined.

Several kinds of mixing which may change the local chemical composition are possible. Perhaps the simplest is spatial transport of solids relative to the gas – by turbulent diffusion, gas-drag driven inward drift, or stellar wind ejection. Another kind of mixing is temporal – as modeled most recently by Cassen (1996); here, volatiles are gradually removed with the (somehow) dissipating nebula gas, so the last-condensing solids are depleted relative to the earliest-condensing solids. This approach requires that the condensation timescale (thermally driven) must be somehow aligned with the gas removal timescale; Cassen (1996) envisioned that removal of the gas decreased the opacity and caused nebula cooling. Another form of mixing, which decouples gas removal from condensation, is the CWPI model of Petaev et al. (1998). In this model, solids larger than a certain size become isolated from reaction with the gas. The size required for this isolation depends on the timescale on which the gas is ultimately cooled and removed (see section 3.3), but can be mm-size. "Mixing" can be accomplished in this model either by physical mixing of subsequent condensates, or by *partial* alteration of condensates (near their surfaces) by ongoing gas reactions, perhaps as regulated by temperature-dependent diffusivity (section 3.3.1).

Interestingly, ordinary and carbonaceous chondrite groups differ in their *refractory* lithophile contents by an amount which is larger than can be explained by the greater enhancement of CAIs in carbonaceous chondrites. Furthermore, CI chondrites show no enhancement in refractory elements relative to solar abundances. Can this be used to rule out augmentation of the CI source region in any outwardly diffusing CAI particles, as has been suggested by Cuzzi et al. (2003) for other chondrite groups? The models of Cuzzi et al. (2003) show outward radial diffusion to be a powerful homogenizing process spatially. However, these models do not account for loss of outwardly-diffusing particles to accretion by drifting, meter-sized rubble, which could limit their propagation to large distances (section 2.5). This sort of modeling might eventually lead to some constraints on the location of the CI source region, the efficiency of diffusion, or both.

3.2.2. Oxidation States

Oxidation states of chondrite minerals vary widely. CAI silicate properties are consistent with formation in a gas of solar composition (Beckett et al. 1988; Krot et al. 2000b and references therein). Certain refractory metal grains *within* CAIs, were thought to require locally enhanced oxygen fugacity to explain observed depletions in Mo and W (Fegley & Palme 1985), but these depletions might also reflect later, low-temperature alteration (Krot et al. 1995; Campbell et al. this volume – see their Fig. 8). AOAs and Type I chondrules imply about 50× solar oxygen enhancement (Petaev & Wood, this volume), and Type II chondrules require an even higher degree of oxygen enhancement. Type I and Type II chondrules are often found in the same meteorite, probably indicating mixing of materials from two distinct environments separated in space and/or in time. Enstatite chondrites appear to have formed under extremely reduced conditions – but the common presence of relict ferrous silicates inside enstatite chondrite chondrules (Lusby, Scott, & Keil 1987; Fagan, Krot, & Keil 1999) indicates that these reduced conditions postdated more oxidizing conditions (e.g., Hutson & Ruzicka 2000).

Regarding the evidence for substantial redox-state variations, it has long been thought that evaporation of silicates – perhaps locally, related to the chondrule-forming events themselves - provides the needed enhancement in oxygen fugacity. Wood & Hashimoto (1993) and Ebel & Grossman (2000) have modeled enhancement by various separate components – rock, oxygen-rich ice, and carbon-rich tar. Petaev & Wood (this volume) assume enhancement in full solar relative abundance of everything except hydrogen and helium. These models have produced encouraging matches with observations, but their various starting conditions are as yet unconnected with any nebula scenario.

Some of the evaporation front (EF) effects discussed in section 2.3 might be helpful in this regard. For instance, an obvious implication of the evolution of the water evaporation front (section 2.3) is that oxygen fugacity can be initially nominal, and later *enhanced* in the vapor everywhere inside the EF, but to varying degrees depending on the distance from the evaporation front and on time, and then *depleted* once a “cold finger” forms (Stevenson & Lunine 1998). For this effect to occur, rapidly drifting meter-sized particles must make up a non-negligible fraction of the total mass budget of solids, and must be able to drift long distances without growing. Enhancements of an order of magnitude – or usually less - are being found in ongoing numerical models of this effect (Ciesla & Cuzzi 2005). The larger oxygen fugacities inferred from the minerals must then rely on an additional effect – perhaps the density and pressure increases associated with passing shocks. The possibility of time evolution of nebula oxygen fugacity – first increasing and then decreasing – may have been recorded by chondrules in enstatite chondrites (Lusby et al. 1987; Fagan et al. 1999; Hutson & Ruzicka 2000). This late-stage decrease in oxygen fugacity may have resulted from the partial removal of water from the inner solar nebula during regime 3 of the evaporation front model (Fig. 5; Stevenson & Lunine 1988). Evaporation front effects of this sort are most naturally accomplished in the framework of a mildly turbulent nebula. In a non-turbulent nebula, even if meter-sized particles did drift past an EF, there is nothing to redistribute the vapor (molecular difffusion is negligible), so it would just sit there as a very intense, localized, enhancement ring rather than diffusing and advecting through the inner nebula. This evolution might then also suggest that Type I chondrules form earliest, perhaps in regions which were hotter and/or further from the water EF, followed by Type II chondrules. Because FeO-poor relict grains are rather common in Type II chondrules and vice versa, mixing of Type I and Type II chondrule precursors is required. The extent of the cross-mixing of these types might help us understand the radial and temporal extent of mixing processes, and the survival/wandering time of typical chondrules before accretion (see section 3.4.2 for more discussion).

Another element with powerful redox potential is carbon (Connolly et al. 1994). Even the most primitive CI chondrites are depleted in carbon by about an order of magnitude compared to either solar or cometary abundances. A significant fraction (about 20-30 wt%) of all primordial carbon is in some refractory form (Jessberger & Kissel 1991; Lawler & Brownlee 1992; Fomenkova 1997; Wooden 2002), which was seen in comet Halley and Wilson dust, and which must be stable at several hundred degrees K. How did most of that carbon escape chondrites and where did it all go? Could it have all been oxidized to CO during chondrule formation events, or prior to those events? If carbon had been widely present in very early solids, it might help explain the CAI redox state (Cuzzi et al. 2003), but if it were still abundant in chondrule precursors several Myr later, it might make the presence of some oxidized iron in even Type I chondrules hard to understand. A scenario might be constructed where chondrite precursor material was simply hot enough very early on to evaporate (or combust) its refractory carbon *locally* prior to the era of chondrule formation and parent body accretion (Huss et al. this volume and references therein). Another possible explanation is that a large fraction of primordial carbon in all chondrite (and chondrule!) precursors had already been stripped away, transformed into chemically inert CO by a massive reprocessing of drifting rubble in the *innermost solar system* (Cuzzi et al. 2003), at a silicate evaporation front existing well inwards of, and long before, the place and time of chondrule formation, recondensing and diffusing outwards, to be recycled millennia later into chondrites. A possible intermediate situation might be an evaporation front for carbon-rich organics, with location external to the asteroid belt. The redox story is likely to resolve into a battle between carbon, oxygen, and hydrogen, with each playing the dominant role as their abundances ebb and flow.

3.3. Mineralogy

Here, we focus primarily on one, very general, aspect of chondrite mineralogy: the fact that many particles, and minerals within particles – even while showing all the aspects of nebula condensation - never achieved complete equilibrium with a cooling gas of solar composition. This points to an important role of kinetics during nebular processing of solid material. Although a number of papers dealing with kinetic aspects of gas phase reactions (Lewis & Prinn 1980; Fegley & Prinn 1989; Prinn &

Fegley 1989; Mendybaev et al. 1985) and gas-solid reactions (e.g., Fegley 1988; 2000; Hong & Fegley 1998; Lauretta et al. 1998; Mendybaev et al. 1989, 2002) relevant to the solar nebula has been published, unfortunately none of them addresses the kinetics of gas-solid condensation reactions of interest here. Moreover, there are no experimental data relevant to the high-temperature gas-solid reactions taking place during the evolution of a nebula of solar composition.

Equilibrium condensation sequences have been extremely valuable (e.g., Grossman 1972); however, discrepancies exist which are indicative of unmodeled nebula processes or properties. Refractory inclusions (CAIs and AOAs) are prime examples of this puzzle. Indeed they *do* consist of the refractory oxides and silicates that condense first from a cooling gas (or crystallize from melts of early condensates). The ubiquitous difference in the order in which spinel and melilite condense have been explained recently by Petaev & Wood (this volume) by disequilibrium condensation, where the observed sequence is seen at pressures less than a few times 10^{-4} bar. Incidentally, this pressure is somewhat lower than the 10^{-3} bar one might expect close to the star in the energetic early years of the nebula, but significantly higher than expected in the asteroid belt during the low-accretion rate "revealed" T Tauri stage (see section 2.1). However, if inclusions were to remain in that same cooling gas, they would continue to react with it, absorbing silicon, iron, magnesium, *etc.,* and would transform into a mixture of Ca,Al-rich minerals and ferromagnesian silicates. The very presence of primary refractory oxides shows that this ongoing reaction sequence did not go to completion (see, e.g., Wood 2004 and MacPherson et al. this volume). Similar arguments apply to AOAs in which forsterite (Mg_2SiO_4) would expect to be extensively replaced by enstatite ($MgSiO_3$), which is, however, extremely rare and volumetrically insignificant (Krot et al. 2004c).

This interrupted reaction sequence requires that the particle in question be isolated from its parent gas "quickly", before alteration can proceed to completion. "Quickly" means that the isolation timescale must be faster than the alteration timescale. The isolation of condensates is a complex physical and chemical process, with the actual mechanisms still remaining to be understood. Petaev & Wood (1998) discussed a number of possible physical and chemical mechanisms, focussing on isolation by "burial": (1) the growth of mineral grains, (2) the envelopment of grains of earlier condensate in layers of later condensates, and (3) the aggregation of mineral grains to form larger objects. An example of the case (1) is the condensation of zoned metal grains from the CH and CB_b chondrites (Petaev & Wood, this volume; Campbell et al. this volume). An application of case (2) is discussed below.

Another way to think of isolation, is by separation or removal of the grain from its parent gas parcel without further growth. At least three factors determine the rate at which a mineral can be altered, even assuming it is in full contact with the nebula gas. First, energetics implies that there might be an energy barrier to overcome before the reaction can proceed; this might require supercooling. Second, supply of altering atoms from the surrounding gas can, in principle, be slow enough to limit the reaction rate. Third, diffusion of altering atoms into the lattice must occur in order for the minerals there to be altered. The first of these three factors is hard to assess because of the lack of experimental data. The very fact that the alteration of melilite to anothite ± other minerals is widespread, and has proceeded to a different extent in different inclusions, suggests that the activation energy must have been overcome somehow. So, we assume that it is not a rate-limiting step. The second is fairly easy to assess; if all requisite elements are indeed present in cosmic abundances, the supply rate can be calculated from the gas density and molecular sound speed. It turns out that for midplane densities and cosmic abundances, not to mention dust-enriched systems, the supply rate of altering atoms is not a problem. These atoms, after hitting the surface of a particle, need to diffuse into the mineral lattice to alter pre-existing minerals, and this may be the rate-limiting step in cases of most interest. Crudely, since diffusion is a random walk process, an atom will diffuse a distance L in time t if its diffusion coefficient in the mineral is $\mathcal{D}$, where $L^2 = \mathcal{D}\, t$. Diffusion is faster along cracks and through porous media, and things can quickly get complicated. In the simplified case, one can estimate alteration times for various depths from (limited) diffusion coefficient data available in the literature, and compare these with removal timescales associated with different nebula mechanisms.

From the standpoint of closed or open system behavior raised first in section 3.2, closed system behavior ensues for a particle of radius r when $r^2/\mathcal{D} < t_d$, where t_d is some dynamical timescale for the particle – either physical removal from its parent gas or burial, and open system behavior ensues when this is not the case. Specifically, the good agreement of refractory silicate mineralogy with closed system, equilibrium models is due to the high temperatures (and $\mathcal{D}$) which characterize these objects, which ensures that they remain in chemical communication with the surrounding gas on timescales long compared to their evolution time. At lower temperatures, as $\mathcal{D}$ decreases, open system behavior becomes more likely and the disagreement with equilibrium models grows.

3.3.1. A Simple Model for Nebula Alteration

To test this simple idea, we selected one aspect of non-equilibrium mineralogy that we thought was appropriately simple. Virtually all CAIs are surrounded by Wark-Lovering rim layers, which appear to have resulted from interaction between solid CAIs and nebular gas. In the simplest case, as seen in the reduced CV chondrites which experienced only minor alteration on the parent body, Wark-Lovering rims have three layers: an inner spinel±hibonite-perovskite layer - evaporative residue (Wark 2004; Wark et al. 2005), overlain by a middle melilite condensation layer, which is often altered to anorthite completely or partially, and which is itself overlain by the condensation layers of Al-diopside and forsterite (Fig. 11).

The removal process which prevents particles from achieving mineralogical equilibrium is simply that the particle leaves its parent gas volume at some temperature T, moving (perhaps irregularly, but inexorably) to regions of lower temperature. As the temperature decreases, the diffusion coefficient $\mathcal{D}$, and thus the rate at which atoms can penetrate the lattice, decreases exponentially as:

$$\mathcal{D}(T) = \mathcal{D}_0 \exp(-d_0/T).$$

Moreover, at some point, the abundance of altering atoms falls off because the particle enters gas regions where they are either already condensed out, or quickly condense on pre-existing fine-grained dust rather than alter the wandering particle itself; we do not treat that complication here. For the system in question, the altered mineral is the middle layer of initial melilite, which is altered by silicon and oxygen diffusing inwards through the outer pyroxene layer. The slow diffusion rate of Si and O in pyroxene limits the absorption depth. Diffusion times to several depths, at constant temperature, are given in Table 1 below (using D_o and d_o for Si and O in pyroxene from Dimanov et al. 1996):

Table 1. Timescales t for diffusion to depth L at various temperatures.

L(μm)	t(1500K)	t(1300K)	t(1100K)
2	10 yr	300 yr	10^5 yr
20	10^3 yr	3×10^4 yr	10^7 yr
200	10^5 yr	3×10^6 yr	10^9 yr

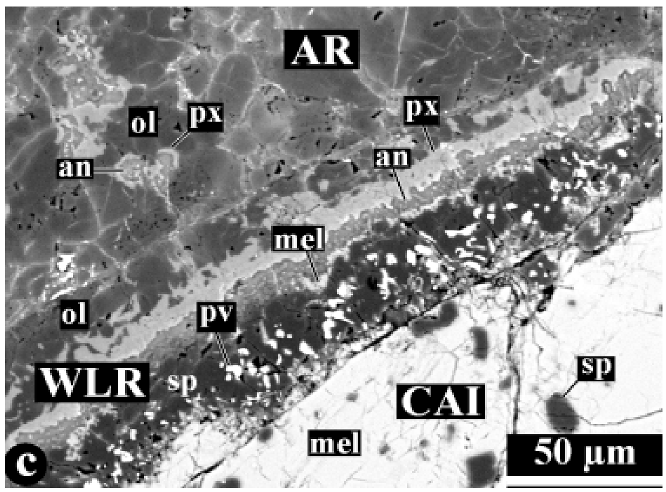

Figure 11. Backscattered electron image of an outer portion of a Type B CAI from the CV chondrite Efremovka. The CAI is made of melilite (mel), Al,Ti-diopside, spinel (sp), and anorthite (an), and is surrounded by Wark-Lovering rim (WLR) layers composed of spinel, perovskite (pv), anorthite, Al,Ti-diopside (px), and forsterite (ol); the latter is surrounded by a forsterite-rich accretionary rim (AR) composed of forsterite and CAIs composed of spinel, Al-diopside, and anorthite.

Removal on shorter timescales than these would prevent alteration. Because of how rapidly diffusion coefficients (and perhaps reaction kinetics) decrease with temperature, the particles do not actually have to change their temperature very much – or move very far – depending on the nebula thermal gradient, in order for diffusion to become so slow that the particle decouples from the gas chemically.

To explore this, we have modeled a viscously heated and evolving solar nebula using analytical expressions derived from those of Lynden-Bell & Pringle (1974), and released into it at T = 1500 K a number of tracer "CAI's" which were allowed to diffuse using a crude random walk formulation. Their thermal histories are followed and, for the survivors (about 10^{-3} of the initial population), converted into probability distribution functions giving the time spent at greater than some temperature. Typical results are shown in Figure 12.

It can be seen from Figure 12 that the time spent by a typical particle (the 50% contour) at temperatures over 1300 K is a few thousand years, sufficient to result in alteration to a depth of some tens of microns (Table 1) but not more. To some extent this is a selection effect; particles which evolve *outwards* rapidly are most likely to escape being swept into the forming sun with the bulk of the nebula gas. Thus, outward radial diffusion is a natural explanation for the alteration observed in the melilite-anorthite layer. The few CAIs with Wark-Lovering rims containing *unaltered* melilite layers could be rare, fast escapees from the alteration region (the 99% contour of figure 12). As objects diffuse outwards from hotter formation regions into cooler regions, not only is the diffusion of altering atoms through the lattice slower, but the supply of altering atoms goes down because there is less altering material in the vapor form. The particle is not remaining in a slowly cooling, "chemically adiabatic", or closed gas. Thus, the removal timescale that is consistent with the observed alteration could be longer in reality. Another interesting potential application of a diffusion (random walk) model might be in explaining certain reversely zoned, fine-grained refractory inclusions (melilite-rich mantle, anorthite-rich core) as described by Krot et al. (2004b). Such particles might have experienced cooling and alteration of anorthite throughout, followed by re-heating and alteration of melilite in the outer layers, before escaping.

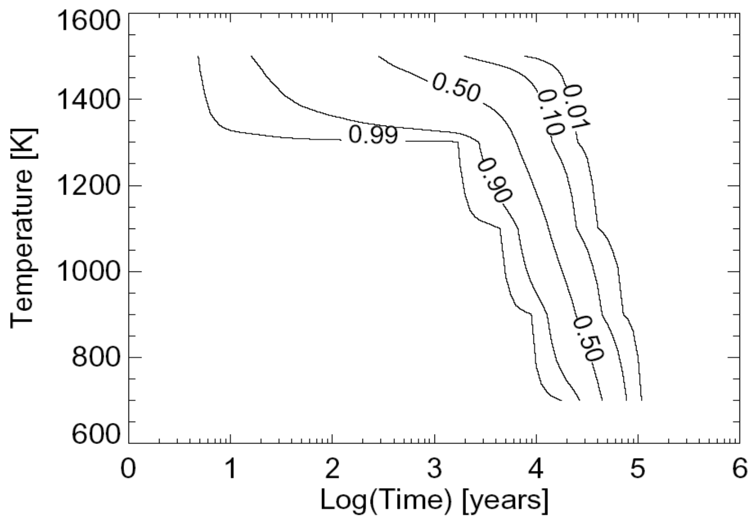

Figure 12. Averaged probability density functions for the fraction of particles which spend more than some amount of time at greater than a certain temperature. These temperatures get somewhat higher than seen in Figure 2, because this process occurs at earlier times and accretion rates, when nebula temperatures were higher than those in Figure 2. This is allowed for in the model.

By comparison, ejection by a stellar wind is on the orbital timescale (years), is a one-way trip, and is also at fairly low gas density which is problematic from a supply standpoint. It might be difficult to achieve the observed alterations under these conditions, without simply postulating the needed residence time in regions suitable for alteration, for the appropriate time, prior to ejection.

Other objects might be treated in similar ways. For instance, corundum → hibonite alteration in CAIs or individual grains (Simon et al. 2002; Krot, Huss, & Hutcheon 2001; Nakamura et al. 2005), macro-zoning in fine-grained inclusions (Krot et al. 2004b) and forsterite → enstatite alteration in AOAs (Krot et al. 2004c) might be profitably studied in this way. Can fluffy Type A CAIs, largely composed of melilite, be explained by the alteration of more refractory condensates? Can this occur without any evidence of surface alteration to anorthite? These speculations merit some thought. More study could be devoted to quantifying alteration and evolution timescales of small mineral particles in protoplanetary nebulae. More experimental measurements of diffusion of different atoms in different minerals – both clean lattices and disordered, fractured ones, would be very helpful.

3.3.2. A CAI for Every Chondrite Group?

A concept that has recently received considerable attention is: to what extent does each meteorite class have its own distinct population of refractory inclusions, and what might the implications of this be (e.g., Krot et al. 2002a; McKeegan et al. 2004; Gounelle & Russell, this volume; Alexander, this volume)? We have already discussed the possibility that the unique presence of *large* CAIs and AOAs in CV chondrites can be readily explained if they accreted the earliest (section 2.2). The next question is whether chemical or mineralogical differences between CAIs in different groups are diagnostic of differently *produced* populations, or merely differently *evolved* populations. One might first ask the extent to which parent body processes (aqueous alteration) might be responsible. Next one might ask the extent to which the differences involve non-equilibrium nebular alteration, which is size, time, and temperature dependent. Groups accreting mainly small inclusions might well find all of them more heavily altered from their original mineralogy even before accretion. Of course, to make sense of this, we need to understand what the original mineralogy *was* for all these different inclusions. We cannot solve this problem, but merely suggest issues that might need to be considered by others.

The presence in the same meteorite of components which suffered an extensive melting event locally (chondrules) and components of the same size which did not (some CAIs, most AOAs) presents a puzzle for how melting events are separated from accretion by nebula transport and mixing. However, are we sure that the apparently unmelted objects were really not present in the chondrule-forming regions? Some isotopically-normal (non-FUN) CAIs show no evidence for ^{26}Al. How sure are we that these *do not* represent a population that was, at least, reheated in the chondrule formation region? This idea can be tested by detailed study of the oxygen isotopic composition of ^{26}Al-depleted CAIs (section 3.4.2). In contrast to CAIs, which originated in an ^{16}O-rich gaseous reservoir (Clayton 2002; Krot et al. 2002b), chondrule formation occurred in an ^{16}O-poor gas and resulted in oxygen isotope equilibration between the molten silicate droplets and the gas (Boesenberg et al. 2004; Krot et al. 2005a,b) reported several igneous, anorthite-rich (Type C) CAIs from the Allende meteorite that they interpreted as experiencing late-stage melting with addition of chondrule material in the chondrule-forming region. These CAIs are ^{16}O-depleted and have low ($^{26}Al/^{27}Al$) ratio similar to those observed in chondrules. Observations such as these could be extended to other CAI types.

MacPherson & Huss (2003) have found that Al-rich chondrules lie along an apparent mixing line between normal chondrules and anorthite-pyroxene–rich, Type C-like CAIs. They pose a conundrum for melting of simple mixtures of CAI and chondrule material as an explanation for Al-rich chondrules (Krot et al. this volume), in that this pattern shows no evidence for any melilite-rich material in the precursors – yet, melilite-rich CAIs are found in the same meteorites. We suggest that this effect can be explained by the rarity of melilite-rich CAIs in general, and specifically amongst the (small) size range most likely to be incorporated into chondrule precursors (because the smaller CAIs are more easily altered).

Of interest here, is the recent discovery of chimeric, but mainly H-type, chondrites with unusually abundant matrix and refractory inclusions (described by the authors as CO-like; Kimura et al. 2002). Could this have arisen by an unusually late mixing event, which brought together normally matrix-poor H type chondrules, and more primitive matrix-like material?

3.4. Isotopic Properties

Here we emphasize three aspects of this very broad subject: age differences from short- and long-lived isotopes; oxygen isotopes; and the apparent homogeneity of chondritic isotopic compositions.

3.4.1. Isotopic Age Dating

One of the most important advances in recent years is the ability to date various chondritic components with high precision. This capability continues to improve, and we look forward to further progress with great anticipation. Current observations tend to support the legitimacy of the Al-Mg chronometer (Amelin et al. 2002, 2004; Russell et al. 2005; Kita et al. and Goswami et al. this volume). However, it will always be valuable to have independent estimates from different chronometers (Pb-Pb, U-Pb, Rb-Sr, Fe-Ni, Mn-Cr). Even as of this writing it seems clear that the few Myr age difference between most CAIs (and AOAs) and most chondrules is real (Amelin et al. 2002, 2004; Bizzarro et al. 2004). It is also becoming clear that there are discernible age differences between chondrules in different chondrite groups, with chondrules in CB chondrites being the youngest, those in CRs distinctly older, and with CVs apparently containing the oldest chondrules.

What is the age difference between CAIs and chondrules in CV chondrites? Currently CV chondrule ages scatter across about 1.4 Myr after CAIs, with stated measurement errors of maybe a few 10^5 years (Bizzarro et al. 2004). Were any of these forming at the same time as CAIs or not? Nebulae can evolve significantly in temperature and density over the first 10^5 – 10^6 years, and over this time ^{26}Al decays enough for some small planetesimals to form and remain unmelted. The dearth of melted asteroids argues rather strongly against the formation of most parent bodies contemporaneously with most CAIs (section 3.1.3); this argument becomes even more stringent if ^{60}Fe provides a significant heat source. Models of CAI preservation by turbulent mixing (section 2.2) are strongly sensitive to exactly how long the time difference is in this general window.

Currently inferred variances of chondrule ages within a *given* chondrite are comparable to the typical differences between the groups (Kita et al. this volume). Yet, measurement errors remain at the level of these differences. So, while these differences are certainly interesting, much more remains to be done. The variance will tell us how long a given object needs to wander around in the nebula between its formation and its accretion. CAIs can apparently do this for millions of years, so chondrules could also, as long as the same environment persisted. However, nebular lifetimes this extended would have implications for accretion of fine-grained rims, processing in subsequent chondrule-forming events, mixing with other chondrule types and unprocessed matrix material, and so on. Some chondrite groups have such well-defined properties that one imagines accretion must have occurred very quickly, before too much mixing occurred (see section 3.5, and Alexander, this volume). Ultimately, self-consistent primary accretion scenarios will need to reconcile radial mixing and formation of fine (and coarse) grained rims with whatever levels of turbulence might be most consistent with, say, turbulent concentration or other accretionary hypotheses.

Observations could address whether Type I (FeO-poor) and Type II (FeO-rich_ chondrules in a given chondrite have distinguishable ages, and whether Type I chondrules from different chondrite groups have distinguishable ages. This will constrain the nebula lifetime of these objects, and the spatial and temporal scales of the different environments in which these objects form. If their redox differences are explained by oxygen coming from the water evaporation front, its abundance can increase *or* decrease with time, and with distance from the Sun, depending on the epoch.

All these age differences will ultimately provide the constraints for primary accretion models. For instance, turbulent accretion models aim to provide a planetesimal formation rate as a function of time (section 2.5), which can continue for as long as turbulence and chondrule formation continues without depleting all the solids. However, if the nebula ever becomes nonturbulent, growth (incremental or instability) seems to proceed very quickly ($\ll 10^5$ years) converting all available solids to planetesimals large enough to be parent bodies, which soon embark on collisional evolution of their own (stage 3). One would need to introduce new sources of solids and postulate intermittent chondrule formation epochs, one for each chondrite group perhaps, to explain the emerging picture, and still be left without a satisfactory explanation for the remarkable chondrule size distribution – the *shape* of which is the same in all chondrites (except CH and CB) regardless of age, composition, or actual chondrule size (section 3.1.1).

3.4.2. Oxygen Isotopic Composition

On a three-oxygen isotope diagram ($\delta^{17}O$ vs. $\delta^{18}O$), oxygen isotopic compositions of chondrules and refractory inclusions follow a mass-independent fractionation slope of ~1, with refractory inclusions being systematically enriched in ^{16}O relative to chondrules (Clayton et al. 1977; Clayton 1993; McKeegan et al. this volume and references therein).[1]

[1] $\delta^{17}O$ and $\delta^{18}O$ are the deviations in parts per thousand of the $^{17}O/^{16}O$ and $^{18}O/^{16}O$ ratios from those of the standard mean ocean water. Vertical deviations from the terrestrial line on the standard three-isotope plot are characterized by the parameter $\Delta^{17}O = \delta^{17}O - 0.52\ \delta^{18}O$ (e.g., Clayton 1993).

Several hypotheses have been proposed to explain these observations. According to one hypothesis, the protoplanetary nebula was initially ^{16}O-enriched to a level recorded by CAIs and AOAs in primitive chondrites ($\Delta^{17}O \sim -25‰$) (Nuth et al. 1999; Clayton 2002; Yurimoto & Kuramoto 2004; Lyons & Young, this volume). Over some amount of time, irradiated CO in the upper reaches of our nebula, or in the parent molecular cloud, became depleted in $^{17,18}O$, and the heavy O, combined into H_2O, condensed as ice, grew to large particles, which drifted inwards to evaporate and enhance the inner solar system in isotopically heavy oxygen (Yurimoto & Kuramoto 2004; Krot et al. 2005a; Lyons & Young, this volume). Thermal processing, such as evaporation, condensation, and melting of ^{16}O-rich solids in an $^{17,18}O$-enriched inner solar nebula gas resulted in their oxygen isotope exchange and formation of $^{17,18}O$-enriched solids (e.g., Yu et al. 1995; Yurimoto et al. 1998; Jones et al. 2004; Krot et al. 2005a). Thus the later forming silicates (including those that formed the Earth) became more enriched in $^{17,18}O$. An alternate hypothesis envisions that the nebula oxygen was terrestrial-like ($\Delta^{17}O = 0‰$), and that ^{16}O was enhanced in rock-forming vapor in the hot, inner, early solar system where CAIs and AOAs formed – by evaporation of large amounts of ^{16}O-rich primitive solids (Scott & Krot 2001) or by chemical-quantum effects (Thiemens & Heidenreich 1983; Thiemens 1996; Marcus 2004). These hypotheses will soon be better constrained using GENESIS data. Once we understand the evolution of oxygen isotopic compositions of materials in the solar nebula, we might be able to use oxygen isotopes as a chronometer for chondritic ingredients.

Type I and Type II chondrules in the same chondrite appear to have comparable oxygen isotopic compositions, but Types I and II chondrules in carbonaceous chondrites (CCs) are more ^{16}O-rich than those in ordinary chondrites (OCs). One might ask how this can be, if Type I and II chondrules require different redox environments for their formation (section 3.2.2)? We suggest that, under realistic conditions, chondrule redox properties depend to a larger extent on their precursor mineralogy (mainly Fe speciation) than on the redox state of the ambient gas, providing that the O isotopic exchange between a silicate melt and the ambient nebular gas is fast but oxidation kinetics of the Fe,Ni metal is slow. For example, melting a type I chondrule precursor enriched in Fe,Ni metal will lead to Fe-poor silicate melt containing immiscible metal globules - and the metal globules might even get spun out of the silicate melt. Even if the ambient gas were highly oxidizing relative to solar, slow oxidation kinetics of the Fe,Ni metal would prevent formation Fe-O-enriched silicates (e.g., a Type II chondrule), but, the O in the type I silicate melt could equilibrate isotopically with the local gaseous O, leaving the cooled product with a Type I chemistry and redox state, but with isotopic composition more like its co-melted Type II. What is needed to make the Type II FeO rich is precursor silicate minerals which are FeO rich - that is, grains which previously condensed from an O-rich gas. This suggests that the precursors of Type I chondrules in OCs could have been originally made in a different environment from the standpoint of O-isotopes and redox state, but were mixed into the location where Type II chondrule precursors were being made and melted – resetting their isotopes but leaving their mineralogy and redox state (more or less) the same. Relict grains of Type I and Type II material are found in chondrules of the other type; chondrules of intermediate composition might simply be fully molten versions of these.

3.4.3. Isotopic Homogeneity

Finally we mention the perhaps too-obvious fact that more than 99% of all material in chondrites – including fine-grained matrix material - is isotopically "normal"; that is, has only small variance from its mean. Chondrite matrix contains less than 100 ppm of isotopically anomalous material, while interplanetary dust particles (IDPs) contain nearly 1000 ppm (e.g., Messenger et al. 2003; Nagashima, Krot, & Yurimoto 2004). So, there is evidence for only a small amount of verifiably unprocessed, presolar material in both populations, and there is a clear gradient outwards. Perhaps both IDPs and primitive chondrite matrices are telling us the same story: they could all be mixtures of materials from the hot inner nebula (^{16}O-rich dust and refractory inclusions), materials from terrestrial regions (^{16}O-poor crystalline and amorphous dust and chondrule debris), and materials from outer regions (unprocessed presolar grains with isotopic anomalies) – all mixed by turbulence.

3.5. Complementarity

"Complementarity" has been cited as the property of chondrules and matrix in a given meteorite to differ systematically in their chemical properties, but in combination to have a bulk composition more closely resembling CI than either does separately. This property has been used to suggest that chondrules and their associated matrix resided in much the same region and there was not a huge amount of mixing with material from other regions. Most discussion has focused on Fe/Si complementarity of chondrules and matrix (Wood 1985, 1988; Taylor 1992). However, matrix is generally closer to CI than is bulk composition, and even bulk Fe/Si ratios for chondrites show wide variations, so these data do not appear to be conclusive. Furthermore, both chondrules and matrix are generally depleted in moderately volatile elements relative to CI (to different degrees; Bland et al. 2005). Perhaps a stronger argument, although not yet fully documented, is that while Ti/Al ratios in CR matrices are lower than in CI, and those in CR chondrules are higher, the bulk chondrite *has* CI Ti/Al (Klerner & Palme 2000). In addition, we have the K chondrites, admittedly few in number, which have chondrules and matrices that are almost identical in their mineralogy and chemical and oxygen isotopic compositions, which is very unusual (Scott & Krot 2003, 2005). These properties suggest that the matrix in K chondrites is predominantly composed of dust that was processed in the vicinity of the enclosed chondrules. This does not mean that each of the 15 chondrite groups has a unique set of chondrules and associated matrix rims. Chondrule properties suggest that there were at least six quite distinct sets of chondrules, but most carbonaceous chondrites, for example, probably had similar matrices (but not identical; *cf*. Bland et al. 2004).

Clearly complementarity does place some restrictions on how disconnected the formation regions for chondrules and their associated matrix can be. However, turbulent mixing would have affected chondrules, CAIs, matrix, and gas almost exactly the same, so if two different batches of material were each individually complementary in their (different) formation regions, and they were mixed without loss over distances which were not too far, the mixture would also be complementary. Improved measurements of major and trace element abundances and oxygen isotopic compositions in chondrules and matrices of primitive chondrites will ultimately tell the tale. For a more detailed discussion of complementarity see Huss et al. (this volume).

4. Major Current Puzzles and Key Diagnostic Observations

1) Clearly, more high precision age dating of various constituents in all chondrite and achondrite groups will continue to be critical to unraveling the story of nebula processing and parent body accretion (sect 3.4.1). It would be important to date Type I and Type II chondrules within an individual chondrite separately, in order to understand whether their formation was separated in time or in space. While the Al-Mg chronometer is gaining support, it must be continually tested against long-lived isotopic clocks in all samples. Achondrites and iron meteorites might hold important clues to the presence or absence of parent bodies that formed contemporaneously with chondrules or CAIs. Improved thermal evolution models will be an essential aspect of these latter studies (section 3.1.3).

2) Studies of the physical fabric of chondrites still have critical things to tell us. Primary texture is of special interest (section 3.1.1), because it retains material with the highest likelihood of having really accreted together. The size distributions of CAIs, AOAs, and clastic fragments should be measured in different chondrite groups (3.1.1), together with their internal densities, and if possible accounting for their shapes and porosities. Disaggregation or tomographic techniques might be needed. The rim-core relationships should be determined for a variety of different constituents in different chondrite groups, and systematic variation of rim chemical and isotopic properties determined – especially for CAIs, AOAs, and chondrules - as a metric for nebula evolution (3.1.2).

3) The nebula lifetime of chondrules between melting and accretion might be better understood by careful studies of the relationships between fine-grained accretionary rims and coarse-grained igneous rims on different types of chondrules and in different groups. Radial mixing might be constrained by studying the ages of Type I and Type II chondrules in different groups. The discovery of unusual chondrites (such as H types with abundant matrix and CAIs; section 3.3.2) may have implications for turbulent mixing.

4) During its extended temporal evolution, the nebula might have experienced substantial deviations from nominal solar abundance – both radially and temporally (sections 2.3 and 3.2). Essentially all elements would be affected, with variations in carbon, oxygen, silicon, and sulfur being perhaps the most critical. Moderately volatile element variations might be tied to these effects. Observational correlations of the oxygen isotopic content with the redox state of chondrules and other objects would be one useful approach to assessing this effect.

5) Regarding nonequilibrium mineralogy (sect 3.3.1), can we further test the idea that plausible timescales of nebula transport are consistent with various observed cases where alteration to equilibrium mineralogy has been truncated? Which nebula models provide the best opportunities for explaining these effects? Do we need more experimental measurements of diffusion coefficients as functions of temperature for different elements in different minerals? Can any observations of AOAs, hibonite grains, and CAIs as a function of size rule out such a scenario?

6) Are separate formation times/places required for different kinds of CAIs found in different chondrite groups (section 3.3.2)? Can we distinguish different nebula alteration histories from different initial formation environments? Can we use

oxygen isotopes as metrics for nebula thermal alteration? What fraction of CAIs present in chondrule melting regions escaped melting?

7) Better models are needed of the evolution, under solar and self gravity, of the dense particle-rich zones produced by turbulent concentration, including mutual interactions leading to coagulation and dispersal. These are strengthless "objects", but their self-gravity might play a role in stabilizing them, or even in collapsing them into planetesimals as once the Goldreich-Ward particle midplane instability was envisioned to. Mutual encounters might create more stable objects as well as disrupting objects.

8) Improved models are needed for incremental accretion, which allow for large-scale decoupling of solids from the gas, and gradients in chemical or isotopic properties, to be tracked. Specifically, we need models which incorporate "loss" of small, diffusing particles into large objects which evolve in very different ways.

9) Astronomical observations of the relative abundances of water, SiO, CO, and other molecules in the vapor form, relative to each other or to hydrogen, as a function of location (distance from the star) and time (or accretion rate), might be capable of revealing evidence for or against evaporation fronts.

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References

Adachi, I., Hayashi, C., & Nakazawa, K. 1976, Prog. Theor. Phys., 56, 1756
Aléon, J., Krot, A. N., & McKeegan, K. D. 2002, Meteorit. Planet. Sci., 37, 1729
Amelin, Y., Krot, A. N., Hutcheon, I. D., & Ulyanov, A. A. 2002, Science, 297, 1678
Amelin, Y., Krot, A. N., Russell, S. S., & Twelker, E. 2004, Geochim. Cosmochim. Acta, 68, A759
Beckett, J., Stolper, E., Live, D., Tsay, F.-D., & Grossman, L. 1988, Geochim. Cosmochim. Acta, 52, 1479
Bell, K. R., Cassen, P. M., Klahr, H. H., & Henning, Th. 1997, ApJ, 486, 372
Binzel, R. P., Lupishko, D., Di Martino, M., Whitley, R. J., & Hahn, G. J. 2003, in Asteroids III, eds. W. F. Bottke, A. Cellino, P. Paolicchi, & R. P. Binzel, (Tucson: Univ. Arizona Press), 255
Bizzarro, M., Baker, J., & Haack, H. 2004, Nature, 431, 275
Bland, P. A., Prior, D. J., & Hough, R. M. 2003, Meteorit. Planet. Sci., 38, 1587
Bland, P. A., Alard, O., Gounelle, M., Benedix, G., Kearskey, A., & Rogers, N. 2004, Lunar Planet. Sci, 35, 1737
Bland, P. A. et al. 2005, Proc. Nat. Acad. Sci., submitted
Blum, J. 2004, in Astrophysics of Dust, ASP Conference Ser., 309, eds. A. N. Witt, G. C. Clayton, & B. T. Draine, 369

Boesenberg, J. S., Hewins, R. H., & Chaussidon, M. 2004, http://www.lpi.usra.edu/meetings/chondrites2004/pdf/9047.pdf

Bockelee-Morvan, D., Gautier, D., Hersant, F., Hure, J.-M., & Robert, F. 2002, A&A, 384, 1107

Boss, A.P. 2005, ApJ, 616, 1265

Brearley, A. J. 1993, Geochim. Cosmochim. Acta, 57, 1521

Brearley, A. J. 2003, in Meteorites, Comets, and Planets, ed. A. M. Davis, Vol. 1, Treatise on Geochemistry, eds. H. D. Holland, & K. K. Turekian (Oxford: Elsevier-Pergamon), 247

Bunch, T. E., Cassen, P., Podolak, M., Reynolds, R., Chang, S., Schultz, P., Brownlee, D., & Lissauer, J. 1991, Icarus, 91, 76

Burbine, T. H., Meibom, A., & Binzel, R. P. 1996, Meteoritics, 31, 607

Calvet, N., Hartmann, L., & Strom, S. E. 2000, in Protostars and Planets, eds. Mannings, V., Boss, A. P., & Russell, S. S. (Tucson: Univ. Arizona Press), 377

Cassen, P. 1996, Meteoritics, 31, 793

Chambers, J. E. C., 2004, Earth Planet. Sci. Lett., 223, 241

Ciesla, F. J., & Cuzzi, J. N. 2005, Lunar Planet. Sci., 36, 1479

Clayton, R. N. 1993, Ann. Rev. Earth Planet. Sci., 21, 115

Clayton, R. N., & Mayeda, T.K. 1996, Geochim. Cosmochim. Acta, 60, 1999

Clayton, R. N. 2002, Nature, 402, 860

Clayton, R. N., Onuma, N., Grossman, L., & Mayeda, T. K. 1977, Earth Planet. Sci. Lett., 34, 209

Cohen, B. A., & Coker, R. F. 2000, Icarus, 145, 369

Connolly, H. C., Hewins, R. H., Ash, R. D., Zanda, B., Lofgren, G. E. & Bourot-Denise, M. 1994, Nature, 371, 136

Cuzzi, J. N. 2004, Icarus, 168, 484

Cuzzi, J. N., & Hogan, R. C. 2003, Icarus, 164, 127

Cuzzi, J. N., & Zahnle, K. J., 2004, ApJ, 614, 490

Cuzzi, J. N., Dobrovolskis, A. R., & Champney, J. M. 1993, Icarus, 106, 102

Cuzzi, J. N., Davis, A., & Dobrovolskis, A. R. 2003, Icarus, 166, 385

Cuzzi, J. N., Dobrovolskis, A. R., & Hogan, R. C. 1996, in Chondrules and the Protoplanetary Disk, eds. R. H. Hewins, R. H. Jones, & E. R. D. Scott (Cambridge: Cambridge Univ. Press), 35

Cuzzi, J. N., Hogan, R. C., Paque, J. M., & Dobrovolskis, A. R. 2001, ApJ, 546, 496

Cuzzi, J. N., & Weidenschilling, S. J. 2005, in Meteorites and the Early Solar System II, eds. D. Lauretta, & H. Y. Jr. McSween (Tucson: Univ. Arizona Press), in press

Cuzzi, J. N., Petaev, M. I., Ciesla, F. J., Krot, A. N., & Scott, E. R. D. 2005, Lunar Planet. Sci., 36, 2095

Cyr, K., Sears, W., & Lunine, J., 1998, Icarus, 135, 537

Dimanov, A., Jaoul, O., & Sautter, V. 1996, Geochim. Cosmochim. Acta, 60, 4095

Dubrulle, B., Morfill, G., & Sterzik, M. 1995, Icarus, 114, 237

Dullemond, C. P., & Dominik, C. 2005, A&A, 434, 971

Ebel, D. S., & Alexander, C. M. O'D. 2005, Lunar Planet. Sci., 36, 1797

Ebel, D. S., & Grossman, L. 2000, Geochim. Cosmochim. Acta, 64, 339

Fagan, T. J., Krot, A. N., & Keil, K. 1999, Lunar Planet. Sci., 30, 1523

Fedkin, A. V., & Grossman, L. 2004, Lunar Planet. Sci., 35, 1823

Fegley, B., Jr. 1988, in Workshop on The Origins of Solar Systems, ed. J. A. Nuth & P. Sylvester (Houston, TX: LPI technical report 88-04), 51

Fegley, B., Jr. 2000, Space Sci. Rev. 92, 177

Fegley, B., Jr., & Palme, H. 1985, Earth Planet. Sci. Lett., 72, 311

Fegley, B., Jr., & Prinn, R. G. 1989, in The Formation and Evolution of Planetary Systems,

eds. H. Weaver, & L. Danny (Cambridge: Cambridge Univ. Press), 171
Fomenkova, M. N. 1997, in From Stardust to Planetesimals, ASP Conference Series, 122, eds. Pendleton, Y. J., & Tielens, A. G. G. M., 415
Gaffey, M. J., Cloutis, E. A., Kelley, M. S., & Reed, K. L. 2003, in Asteroids III, eds. W. F. Bottke, A. Cellino, P. Paolicchi, & R. P. Binzel (Tucson: Univ. Arizona Press), 183
Gammie, C. F. 1996, ApJ, 462, 725
Ghosh, A., Weidenschilling, S. J., & McSween, H. Y. Jr. 2003, Meteorit. Planet. Sci., 38, 711
Grimm, R. E., & McSween, H. Y. Jr. 1993, Science, 259, 653
Grossman, L. 1972, Geochim. Cosmochim. Acta, 36, 597
Grossman, J. N. 1988, in Meteorites and The Early Solar System, eds. J. F. Kerridge, & M. S. Matthews, (Tucson: Univ. Arizona Press), 680
Hogan, R. C., Cuzzi, J. N. 2001, Phys. Fluids, 13, 2938
Hong, Y. & Fegley, B., Jr. 1998, Meteorit. Planet. Sci. 33, 1101
Hughes, D. M. 1980a, Earth Planet. Sci. Lett., 51, 26
Hughes, D. M. 1980b, Nature, 278, 778
Hutson, M., & Ruzicka, A. 2000, Meteorit. Planet. Sci., 35, 601
Jessberger, E. K. & Kissel, J. 1991, in Comets in the Post-Halley Era, eds. R.Newburn, M. Neugebauer, & J. Rahe (Kluwer Press), 1075
Jones, R. H., Leshin, L. A., Guan, Y., Sharp, Z. D., Durakiewicz, T., & Schilk, A. J. 2004, Geochim. Cosmochim. Acta, 68, 3423
Keil, K. 2000, Planet. Space Sci., 48, 887
Keil, K. 2003, in Asteroids III, eds. W. F. Bottke, A. Cellino, P. Paolicchi, & R. P. Binzel (Tucson: Univ. Arizona Press), 573
Keubler, K., McSween, H. Y., Carlson, W. D., & Hirsch, W. D. 1999, Icarus, 141, 96
Kimura, M., Hiyagon, H., Palme, H., Spettel, B., Wolf, D., Clayton, R. N., Mayeda, T. K., Sato, T., Suzuki, A., & Kojima, H. 2002, Meteorit. Planet. Sci., 37, 1417
Klahr, H. H., & Bodenheimer, P. 2003, ApJ, 582, 869
Kleine, T., Munker, C., Mezger, K., & Palme, H. 2002, Nature, 418, 952
Klerner, S., & Palme, H. 2000, Meteorit. Planet. Sci., 35 Suppl., A89
Kobayashi, S., Imai, H., & Yurimoto, H. 2003, Geochem. J., 37, 663
Kong, P. 1999, Geochim. Cosmochim. Acta, 63, 3673
Krot, A. N. & Wasson, J. T. 1995, Geochim. Cosmochim. Acta, 59, 4951
Krot, A. N., Wasson, J. T., Rubin, A. E., Scott, E. R. D., & Keil, K. 1997, Geochim. Cosmochim. Acta, 61, 463
Krot, A. N., Petaev, M. I., Scott, E. R. D., Choi, B.-G., Zolensky, M. E., & Keil, K. 1998a, Meteorit. Planet. Sci., 33, 1065
Krot, A. N., Zolensky, M. E., Keil, K., Scott, E. R. D., & Nakamura, K. 1998b, Meteorit. Planet. Sci., 33, 623
Krot A. N., Brearley A. J., Petaev M. I., Kallemeyn G. W., Sears D. W. G., Benoit P. H., Hutcheon I. D., Zolensky M. E., & Keil K. 2000a, Meteorit. Planet. Sci. 35, 1365
Krot, A. N., Fegley, B., Palme, H., & Lodders, K. 2000b, in Protostars and Planets IV, ed. A. Boss, V. Manning, & S. S. Russell (Tucson: Univ. Arizona Press), 1019
Krot, A. N., Huss, G. R., & Hutcheon I. D. 2001a, Meteorit. Planet. Sci., 36 Suppl., A105
Krot, A. N., Ulyanov, A. A., Meibom, A., & Keil K. 2001b, Meteorit. Planet. Sci., 36, 611
Krot, A. N., Meibom, A., Weisberg, M., & Keil, K. 2002a, Meteorit. Planet. Sci., 37, 1451
Krot, A. N., McKeegan, K. D., Leshin, L. A., MacPherson, G. J., & Scott, E. R. D. 2002b, Science, 295, 1051
Krot, A. N., Amelin, Y., Russell, S. S., & Twelker, E. 2004a, Meteorit. Planet. Sci., 39 Suppl., A56
Krot, A. N., MacPherson, G. J., Ulyanov, A. A., & Petaev M. I. 2004b, Meteorit. Planet. Sci., 39, 1517

Krot, A. N. et al. 2004c, Chem. Erde, 64, 185
Krot, A. N., Petaev, M. I., & Bland, P. A. 2004d, Antarctic Meteorite Res., 17, 154
Krot, A. N., Keil, K., Goodrich, C. A., Scott, E. R. D., & Weisberg, M. K. 2004e, in Meteorites, Comets, and Planets, ed A. M. Davis, Vol. 1, Treatise on Geochemistry, eds. H. D. Holland, & K. K. Turekian (Oxford: Elsevier-Pergamon), 83
Krot, A. N. et al. 2005a, ApJ, 622, 1333
Krot, A. N., Yurimoto, H., Hutcheon, I. D., & MacPherson, G. J. 2005b, Nature, 434, 998
Krot, A. N., Amelin, Y., Cassen, P. M., & Meibom, A. 2005c, Nature, in press
Kunihiro, T., Rubin, A. E., McKeegan, K. D., & Wasson, J. T. 2004, Geochim. Cosmochim. Acta, 68, 2947
Kurahashi, E., Kita, N. T., Nagahara, H., & Morishita, Y. 2004, http://www.lpi.usra.edu/meetings/chondrites2004/pdf/9039.pdf
Larimer, J. W. 1975, Geochim. Cosmochim. Acta, 39, 389
LaTourrette, T., & Wasserburg, G. J. 1998, Earth Planet. Sci. Lett., 158, 91
Lauretta, D. S., Lodders, K., & Fegley, B., Jr. 1998, Meteorit. Planet. Sci., 33, 821
Lawler, M. E., Brownlee, D. E. 1992, Nature, 359, 810
Lewis, J. S. & Prinn, R. G. 1980, AJ, 238, 357
Lusby, D., Scott, E. R. D., & Keil, K. 1987, Proc. Lunar Planet. Sci. Conf., 17, J. Geophys. Res., 92, E679
Lynden-Bell, D., Pringle, J. E. 1974, Mon. Not. R. Astron. Soc., 168, 603
MacPherson, G. J., A. M. Davis, and E. K. Zinner, 1995, Meteoritics 30, 365
MacPherson, G.J ., Hashimoto, A., & Grossman, L. 1985, Geochim. Cosmochim. Acta 49, 2267
MacPherson, G. J., & Huss, G. R. 2003, Lunar Planet. Sci., 34, 1825
Marcus, R. A. 2004, J. Chem. Phys. 121, 8201
Martin, P. M. & Hughes, D. M. 1980, Earth Planet. Sci. Lett., 49, 175
May, C, Russell, S. S., & Grady, M. M. 1999, Lunar Planet. Sci., 30, 1688
McKeegan, K. D., R. N. Clayon, L. A. Leshin, E. D. Young, & H. Yurimoto (2004) http://www.lpi.usra.edu/meetings/chondrites2004/pdf/9060.pdf
McSween, H. Y. Jr., Ghosh, A., Grimm, R. E., Wilson, L, & Young, E. D. 2003, in Asteroids III, eds. W. F. Bottke, Jr., A. Cellino, P. Paolicchi, & R. P. Binzel (Tucson: Univ. Arizona Press), 559
Meibom, A., & Clark, B. E. 1999, Meteorit. Planet. Sci., 34, 7
Mendybaev, R. A., Beckett, J. R., Grossman, L., Stolper, E., Cooper, R. F., & Bradley, J. P. 2002, Geochim. Cosmochim. Acta, 66, 661
Mendybaev, R. A., Kuyunko, N. S., Lavrukhina, A. K., & Khodakovsky, I. L. 1989, Geokhimiya 4, 467
Mendybaev, R. A., Makalkin, A. B., Dorofeeva, B. A., Khodakovsky, I. L., & Lavrukhina, A. K. 1985, Geokhimiya, 8, 1206
Messenger, S., Keller, L. P., Stadermann, F. J., Walker, R. M., & Zinner, E. 2003, Science, 300, 105
Metzler, K. 2004, Meteorit. Planet. Sci., 39, 1307
Metzler, K., Bischoff, A., & Stöffler, D. 1992, Geochim. Cosmochim. Acta, 56, 2873
Morfill, G. E., & Völk, H. J. 1984, ApJ, 287, 371
Morfill, G. E., & Wood, J. A. 1989, Icarus, 82, 225
Morfill, G. E., Durisen, R. H., Turner, G. W. 1998, Icarus, 134, 180
Muzerolle, J., N. Calvet, L. Hartmann, and P. D'Alessio, 2003, ApJ, 597, L149
Nyquist, L. E., Reese, Y., Wiesmann, H., Shih, C.-Y., & Takeda, H. 2003, Earth Planet. Sci. Lett., 214, 11
Nagashima, K., Krot, A. N., & Yurimoto, H., 2004, Nature, 428, 921
Nakamura, T. M., Sugiura, N., Kimura, M., Miyazaki, A., & Krot, A. N. 2005, Lunar Planet.

Sci., 36, 1249
Nuth, J. A. 1999, Lunar Planet. Sci., 30, 1726
Nuth, J. A., Hallenbeck, S. L., & Rietmeijer, F. J. M. 1999, Earth, Moon, and Planets, 80, 73
Nuth, J. A., Hill, H. G. M., & Kletetschke, G. 2000, Nature, 406, 275
Palme, H., & Jones, A. 2004, in Meteorites, Comets, and Planets, ed. A. M. Davis, Vol. 1. Treatise on Geochemistry, eds. H. D. Holland, & K. K. Turekian (Oxford: Elsevier-Pergamon), 41
Paque, J. M., & Cuzzi, J. N. 1997, Lunar Planet. Sci., 28, 71
Pasek, M. A., Milsom, J. A., Ciesla, F. J., Lauretta, D. S., Sharp, C. M., & Lunine, J. I. 2005, Icarus, in press
Prinn, R. G. & Fegley, B., Jr. 1989, in Origin and Evolution of Planetary and Satellite Atmospheres, eds. S. Attreya, J. Pollack, & M. S. Matthews (Tucson: Univ. Arizona Press), 78
Rubin, A. E., Kallemeyn, G. W., Wasson, J. T., Clayton, R. N., Mayeda, T. K., Grady, M., Verchovsky, A. B., Eugster, O., & Lorenzetti, S. 2003, Geochim. Cosmochim. Acta, 67, 3283
Russell, S. S., Hartmann, L., Cuzzi, J. N., Krot, A. N., Gounelle, M., & Weidenschilling, S. 2005, in Meteorites and the Early Solar System II, eds. D. S. Lauretta, & H. Y. McSween (Tucson: Univ. Arizona Press), in press
Sano, T., Miyama, S. M., Umebayashi, T., & Nakano, T. 2000, ApJ, 543, 486
Sears, D. W. G, & R. T. Dodd 1988, in Meteorites and the Early Solar System, eds. J. F. Kerridge, & M. S. Matthews (Tucson: Univ. Arizona Press), 3
Sekiya, M. 1983, Progress of Theoretical Physics, 69, 1116
Scott, E. R. D. 2003, in Asteroids III, eds. W. F. Bottke, A. Cellino, P. Paolicchi, & R. P. Binzel (Tucson: Univ. Arizona Press), 697
Scott, E. R. D. & Haack, H. 1993, Meteoritics, 28, 434
Scott, E. R. D., & Krot, A. N. 2001, Meteorit. Planet. Sci., 36, 1307
Scott, E. R. D., & Krot, A. N. 2003, in Meteorites, Comets, and Planets, ed. A. M. Davis, Vol. 1, Treatise on Geochemistry, eds. H. D. Holland, & K. K. Turekian (Oxford: Elsevier-Pergamon), 143
Scott, E. R. D., & Krot, A. N. 2005, ApJ, 623, 571
Scott, E. R. D., Love, S. G., & Krot, A. N. 1996, in Chondrules and the Protoplanetary Disk, eds. R. H. Hewins, R. H. Jones, & E. R. D. Scott (Cambridge: Cambridge Univ. Press), 87
Sekiya, M. 1998, Icarus, 133, 298
Shu, F., Shang, H., & Lee, T. 1996, Science, 271, 1545
Shukolyukov, A., & Lugmair, G. W. 2003, in Asteroids III, eds. W. F. Bottke, A. Cellino, P. Paolicchi, & R. P. Binzel (Tucson: Univ. Arizona Press), 687
Simon, S. B., Davis, A. M., Grossman, L., & McKeegan, K. D. 2002, Meteorit. Planet. Sci., 37, 433
Skinner, W. R., & Leenhouts, J. M. 1993, Lunar Planet. Sci., 24, 1315
Srinivasan, G., Goswami, J. N., & Bhandari, N. 1999, Science, 284, 1348
Sreenivasan, K. R., & Stolovitsky, K. 1995, Phys. Rev. E., 52, 3242
Stepinski, T. F., & Valageas, P. 1997, A&A, 319, 1007
Stevenson, D., & Lunine, J., 1998, Icarus, 75, 146
Stone, J. M., Gammie, C. F., Balbus, S. A., & Hawley, J. F. 2000, in Protostars and Planets, eds. Mannings, V., Boss, A. P., & Russell, S. S. (Tucson: Univ. Arizona Press), 389
Sugiura, N., & Hoshino, H. 2003. Meteorit. Planet. Sci., 38, 117
Symes, S. J. K., Sears, D. W. G., Taunton, A., Akridge, D. G., Huang, S., & Benoit, P. H. 1998, Meteorit. Planet. Sci, 33, 13
Taylor, S. R. 1992, in Solar System Evolution: A New Perspective, (Cambridge: Cambridge

Univ. Press), 121
Thiemens, M., & Heidenreich, J., 1983, Science, 219, 1073
Thiemens, M., 1996, in Chondrules and the Protoplanetary Disk, eds. R. H. Hewins, R. H. Jones, & E. R. D. Scott (Cambridge: Cambridge Univ. Press), 107
Trieloff, M., Jessberger, E. K., Herrwerth, I., Hopp, J., Fiéni, C., Ghélis M., Bourot-Denise, M., & Pellas, P. 2003, Nature, 422, 502
Wai, C. M., & Wasson, J. T. 1977, Earth Planet. Sci. Lett., 1, 14
Ward, W. R. 2000, in Origin of the Earth and Moon, eds. R. M. Canup, & K. Righter (Tucson: Univ. Arizona Press), 75
Wark, D. A., 2004, http://www.lpi.usra.edu/meetings/chondrites2004/pdf/9050.pdf
Wark, D. A., Shelley, J. M. G.; O'Neill, H., 2005, Lunar Planet. Sci., 36, 1643
Wasson, J. T., Trigo-Rodriguez, J. M., & Rubin, A. E. 2005, Lunar Planet. Sci., 36, 2314
Wasson, J. T., & Kallemeyn, G. W. 1990, Earth Planet. Sci. Lett. 101, 148
Weidenschilling, S. J., 1977, Mon. Not. Roy. Astr. Soc. 180, 57
Weidenschilling, S. J., 1988 in Meteorites and The Early Solar System, eds. J. F. Kerridge, & M. S. Matthews (Tucson: Univ. Arizona Press), 348
Weidenschilling, S. J., 1997, Icarus, 127, 290
Weidenschilling, S. J., 2000, Space Sci. Rev., 92, 295
Weidenschilling, S. J., & Cuzzi, J.N, 2005, in Meteorites and the Early Solar System II, eds. Ds. Lauretta, & H. Y. McSween (Tucson: Univ. Arizona Press), in press
Whipple, F. 1972, in From Plasma to Planet, ed. A. Elvius (New York: Wiley), 211
Wood, J. A. 1963a, Icarus, 2, 152
Wood, J. A. 1963b, Sci. American, October 1963
Wood, J. A. 1985, in Protostars and Planets II, eds. D. C. Black, & M. S. Matthews (Tucson: Univ. Arizona Press), 687
Wood, J. A. 1988, Ann. Rev. Earth Planet. Science, 16, 53
Wood, J. A. 2004, Geochim. Cosmochim. Acta, 68, 4007
Wood, J. A., & Hashimoto, A. 1993, Geochim. Cosmochim. Acta, 57, 2377
Wood, J. A., & Morfill, G. E. 1988, in Meteorites and The Early Solar System, eds. J. F. Kerridge, & M. S. Matthews (Tucson: Univ. Arizona Press), 329
Wooden, D. 2002, Earth, Moon, & Planets, 89, 247
Woolum, D. S., & Cassen P. 1999, Meteorit. Planet. Sci., 34, 897
Yin, Q., Jacobsen, S. B., Yamashita, K., Blichert-Toft, J., Télouk, P., & Albarède, F. 2002, Nature, 418, 949
Yoshino, T., Walter, M. J., & Katsura, T. 2003, Nature, 422, 154
Youdin, A., & Shu, F. 2002, ApJ, 580, 494
Yu, Y., Hewins, R., H. Clayton, R. N., & Mayeda, T. K. 1995, Geochim. Cosmochim. Acta, 59, 2095
Yurimoto, H., & Kuramoto, K. 2004, Science, 305, 1763
Yurimoto, H., Ito, M., & Nagasawa, H. 1998, Science, 282, 1874