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Metal-silicate fractionation in the surface dust layers of accreting planetesimals: Implications for the formation of ordinary chondrites and the nature of asteroid surfaces

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Abstract. Some of the most primitive solar system materials available for study in the laboratory are the ordinary chondrites, the largest meteorite class. The size and distribution of the chondrules (silicate beads) and metal, which leads to the definition of the H, L, and LL classes, suggest sorting before or during aggregation. We suggest that meteorite parent bodies (probably asteroids) had thick dusty surfaces during their early evolution that were easily mobilized by gases evolving from their interiors. Density and size sorting would have occurred in the surface layers as the upward drag forces of the gases (mainly water) acted against the downward force of gravity. The process is analogous to the industrially important process of fluidization and sorting in pyroclastic volcanics. We calculate that gas flow velocities and gas fluxes for the regolith of an asteroid-sized object heated by the impact of accreting objects or by ^{26}Al would have been sufficient for fluidization. It can also explain, quantitatively in some cases, the observed metal-silicate sorting of ordinary chondrites, which has long been ascribed to processes occurring in the primordial solar nebula. Formation of the chondrites in the thick dynamic regolith is consistent with the major properties of chondritic meteorites (i.e., redox state, petrologic type, cooling rate, matrix abundance). These ideas have implications for the nature of asteroid surfaces and the virtual lack of asteroids with ordinary chondrite-like surfaces.

Introduction

The most primitive early solar system materials in our laboratories are the chondritic meteorites, the H, L, and LL chondrites (collectively termed the ordinary chondrites), the EH and EL chondrites (the enstatite chondrites), and the CI, CM, CO, and CV chondrites (the carbonaceous chondrites). A number of small classes, like the CR, CK, and R chondrites and the "primitive achondrites" (the acapulcoites and the brachinites) have also recently been proposed. Each of these classes is characterized by a number of elemental, isotopic, and petrographic properties (Table 1 lists the major classes). The most notable of these properties is the metal to silicate ratio (or the bulk Fe/Si ratio). The present paper is concerned with metal-silicate fractionation in the ordinary chondrites whose metal-silicate ratio (and bulk Fe/Si ratio) decreases along the series H, L, and LL. As will be seen, we think it highly significant that the sizes of chondrules, glassy beads thought to have been produced by flash heating chondritic dust, also increase along this series.

It is widely assumed that all or most of these properties, including the metal-silicate fractionation, are the result of processes occurring in the nebula prior to accretion. The subject was briefly reviewed by *Newsom* [1995]. *Urey* [1961] argued that the separation of metal and silicate did not occur on the meteorite parent body because it would require melting, which chondritic textures clearly preclude. Chondritic textures are essentially sedimentary, or accretionary. They have even been likened to volcanic tuffs [*Ringwood,* 1959; *Urey and Craig,* 1953]. Consequently, a variety of nebula mechanisms for separating metal and silicate have been proposed. *Donn and Sears* [1963] suggested that different accretion efficiencies for compact silicate grains and "fluff balls" of metal would explain metal-silicate fractionations, while *Orowan* [1969] suggested that the greater ductility of metal relative to the silicates would cause the preferential accretion of metal. *Larimer and Anders* [1970] favored a mechanism involving magnetism, but their arguments were based on erroneous thermodynamic calculations [*Sears,* 1978; *Kelly and Larimer,* 1977]. *Larimer and Wasson* [1988] proposed a mechanism involving nebula-wide separation of components falling to the median plane of the early solar nebula following condensation. *Skinner and Leenhouts* [1993b] showed that despite different densities and sizes, metal and silicate chondrules in the CR2 chondrite Acfer 059 had similar terminal velocities and suggested that metal-silicate fractionation occurred by aerodynamic processes in the nebula, along lines discussed by *Weidenschilling* [1977]. *Clayton* [1980] and *Rubin and Keil* [1984] suggested that chondrule sorting may have occurred by aerodynamic processes, and *Haack and Scott* [1993] extended this conclusion to all chondritic components, including the metal and chondrules. All of these mechanisms assume that the metal-silicate fractionation occurred in the nebula, but none has been quantitatively modeled. They all remain speculative and difficult to evaluate.

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Table 1. Particle Sizes and Metal Abundances in Ordinary Chondrites and CI Chondrites

	Ref.	H	L	LL	CI
Physical Property					
Chondrule diameter, mm	1	0.3	0.6-0.8	0.9	Not applicable
Metal grain size, mm	2	0.20	0.18	0.14	Not applicable
Chondrule abundance, vol %	1	65-75	65-75	65-75	None
Metal abundance, wt %	4,5	16	6	2	~0
Matrix abundance, vol %	1	10-15	10-15	10-15	100
Compositional Property					
Carbon, wt %	4,5	0.11	0.12	0.22	2.8
Water, wt %	4,5	0.22	0.46	0.71	16.9
Fe_m/Fe_t, a/a	3	0.58	0.29	0.11	Not applicable
Fe/Si, a/a	3	0.81	0.57	0.52	0.86
Mg/Si, a/a	3	0.96	0.93	0.94	1.05
Ca/Si, a/a	3	0.050	0.046	0.049	0.064
$\delta^{17}O$, ‰	3	2.9	3.5	3.9	~8.8
$\delta^{18}O$, ‰	3	4.1	4.6	4.9	~16.4

References are 1, *Grossman et al.* [1988a]; 2, *Dodd* [1976]; 3, *Sears and Dodd* [1988 and references therein]; 4, *Jarosewich* [1990]; and 5, *Wiik* [1969]. Unit of measure a/a is atom per atom.

Possible Means of Metal-Silicate Fractionation

We argue that melting is not required for large-scale separations of metal and silicate on meteorite parent bodies. We accept that aerodynamic forces played an important part in assembling chondritic meteorites. However, aerodynamic sorting alone fails to explain the relationship between chondrule size and metal abundance. We argue that a weak gravitational field of the sort provided by a 10-100 km asteroid readily explains this very important relationship and that the aerodynamic sorting occurred on the parent body and not in interplanetary space [*Huang et al.*, 1994, 1995].

Our proposal is summarized in Figure 1. We suggest that the body (or bodies) responsible for the ordinary chondrites was originally rich in volatiles, much like the CI and CM chondrites. These meteorites are 10-20 vol% water. Relatively minor amounts of internal heating from ^{26}Al, or external heating by impact during the final stages of accretion, would suffice to dry out the surface layers, cause interior volatiles to evaporate to space, and in doing so, pass through and mobilize the surface dust layers. The process is very difficult to treat theoretically, and laboratory experiments cannot duplicate the low gravitational field. However, a good starting point is to adopt the concept of a "fluidized bed" [*Kunii and Levenspiel,* 1991], which assumes that the dust grains are suspended in a gravitational field by the drag of an upward flowing gas. A continual rain of impacting material, or the stirring caused by the degassing processes, would enable the surface layer to become very thick, perhaps 1-5 km and in this layer aerodynamic sorting by size and density would occur especially as the degassing terminated. We argue that variations in gas flow rate required by differences in the sizes of H, L, and LL chondrules inevitably lead to metal-silicate fractionations characteristic of those classes.

Figures 1b and 1c describe two means by which metal and chondrules could have been separated in a thick dynamic regolith. In Figure 1b, high-density metal falls to the bottom of the fluidized region while chondrules are suspended throughout the layer by the upward flowing gases. The greater the size difference between chondrules and metal, the greater the separation of the two phases. Thus the LL chondrites with their large chondrules experience considerable metal-silicate fractionation, while the H chondrites experience little or none, and the L chondrites experienced intermediate metal-silicate fractionation. In Figure 1c the chondrules are again suspended by the gas flow, but the smaller metal grains rise through the pore space to the surface. In this case the ordinary chondrites would be the material from just below the metal-rich surface layer. Again, the greater the size difference between the metal and chondrules, the greater the metal-silicate fractionation.

Size and density separation as a result of mobilization of particulates by the passage of gases is a common geological process on Earth, where it is associated with pyroclastic volcanism [*Fisher et al.*, 1980; *Wilson*, 1980; *Valentine and Fisher*, 1993]. On Mars, transport of particulates results in the fluidized ejecta blankets surrounding certain craters [*Melosh,* 1989; *Akridge and Sears*, 1996]. On meteorite parent bodies such processes might occur on a regional basis, so that size and density sorting occurs over a significant fraction of the surface, or it might occur through a number of discrete vents. Perfect fluidization is hardly ever achieved in the laboratory or in nature, and a great number of imperfect variations on the theme actually occur. However, the conclusion is inescapable that if mobilization of the dust due to the movement of gases occurs to any degree, then considerable separation of silicates and metal would result.

Some Quantitative Considerations

Need for a Weak Gravitational Field During Aerodynamic Sorting

Whipple [1971, 1972] first discussed a possible role for aerodynamic drag in size-sorting meteorite components in the nebula. He pointed out that for a relatively dense gas,

$$v_1 = 0.36 \mu S / s^2 \rho_s \tag{1}$$

where v_1 is the velocity of a planetesimal of radius S moving through a mixture of gas and dust particles with density ρ_s and radius s and nebular gas of viscosity μ. If the molecular mean free path is larger than the diameter of a particle:

$$v_1 = 0.11 v_m \rho_g S / s \rho_s \tag{2}$$

where v_1 and ρ_g are the mean molecular velocity and density of the gas. *Dodd* [1976] wrote simultaneous equations for a mixture of chondrules and metal and pointed out that one expects the density ratio of chondrules and metal to be equal to their size ratio in the case of a dense gas or the square of their size ratio in the case of a low-density gas:

$$S_m^2 / S_c^2 = \rho_c / \rho_m = \text{const} \tag{3}$$

and

$$S_m / S_c = \rho_c / \rho_m = \text{const} \tag{4}$$

where S_c, ρ_c, S_m, and ρ_m are radius and density of chondrules and metal.

Dodd [1976] showed that the required exponents of S_m/S_c in equations (3) and (4) were about 0.8, 1.0-1.4, and 1.4-4.6 for the LL, L, and H chondrites and suggested that this change in exponent with class reflects a change from the Epstein regime (equation (3)) to the Stokes regime (equation (4)) combined with metamorphic coarsening of the metal grains in H chondrites. This explanation fails because it does not explain values less than 1.0 and because the major factor driving S_m/S_c is the diameter of the chondrules, which at least for the relatively unmetamorphosed meteorites in Dodd's study, is insensitive to metamorphism. Aerodynamic drag alone does not explain the relative size of chondrules and metal in ordinary chondrites, but more important, it does not explain why the amount of metal decreases relative to the amount of silicate along the series H to L to LL. We will show that, given the relative sizes of the metal and silicates, aerodynamic sorting in the presence of a weak gravitational field does explain the metal-silicate fractionation. However, this means that instead of occurring in the nebula, the aerodynamic sorting occurred in the thick dusty surface layers of the accreting planetesimal. The ρ_c/ρ_m terms in equations (3) and (4) are then replaced with complex functions of ρ_c/ρ_m involving the acceleration due to gravity and gas flow rate.

Thick Regoliths on Meteorite Parent Bodies?

Close-up photographs by spacecraft of asteroids Gaspra and Ida and the martian satellite Phobos caused several research

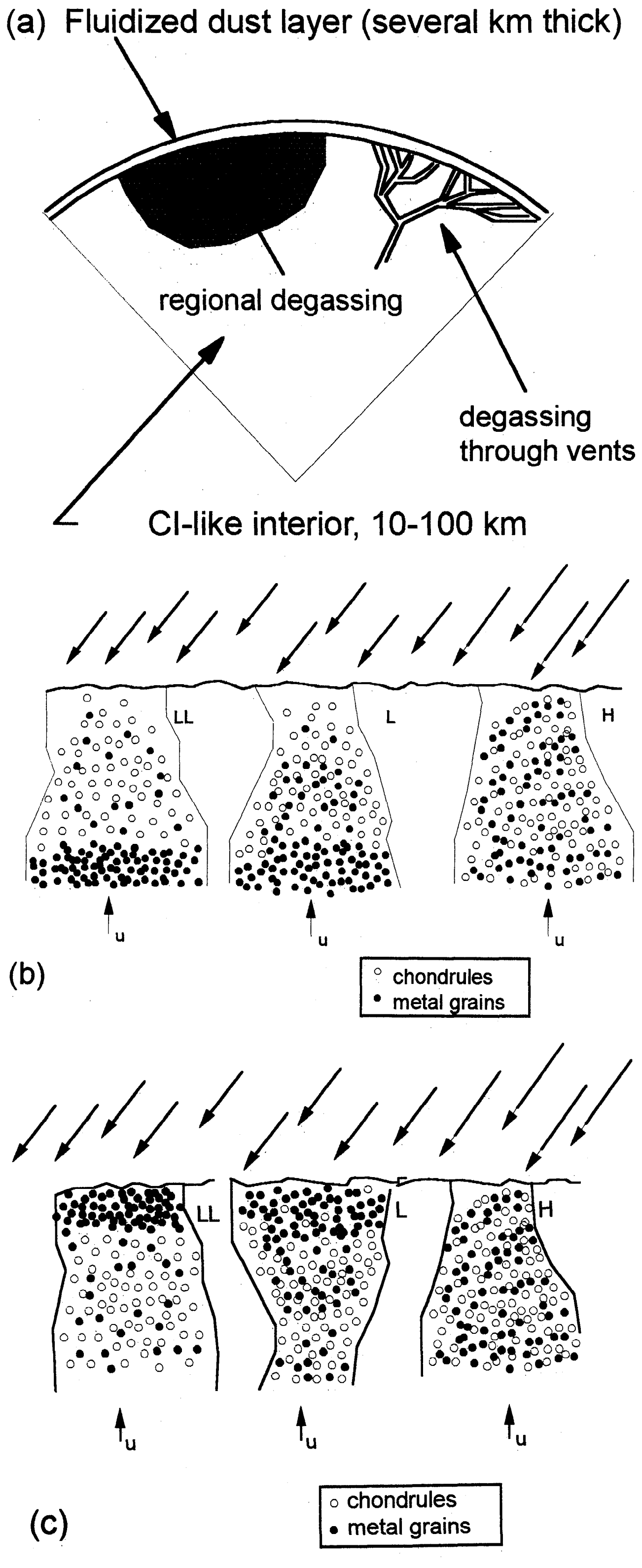

Figure 1. Schematic diagram for the separation of metal and chondrules in the unconsolidated surface layers of an asteroid-sized ordinary chondrite parent body during the final stages of accretion. A volatile-rich (CI chondrite-like) interior body is assumed. (a) Heat from impact or internal radioactivity dries out surface layers, which are then mobilized (or "fluidized") by the upward flow of gases moving from the interior with velocity *u*. The process might be regional or it might be localized in a number of discrete vents. (b) In the small gravitational fields of 10-100 km bodies, chondrules are suspended by the drag forces of the upward flowing gases. Metal grains can then fall to the bottom of the mobilized layer. The degree of metal-silicate separation, and therefore H, L, or LL class, depends on the absolute and relative sizes of the chondrules and metal. (c) However, when the metal is present in relatively small amounts, it can rise to the surface through the pore space between the chondrules to produce a metal-rich surface layer. The ordinary chondrite classes then correspond to the subsurface material, just under the metal-rich surface layer. Various metal-rich chondrite classes may be samples of the metal-rich layer, but it is also possible that such material has not been sampled by Earth, since there is no reason to suppose that the material reaching Earth is representative of any meteorite parent body.

groups to conclude that these objects were covered with substantial regoliths despite their small sizes [*Chapman et al.*, 1992; *Asphaug and Melosh*, 1993; *Carr et al.*, 1994]. Surface features are largely obscured or muted as if covered with regolith and only the largest craters survive unfilled. This has caused considerable revision of theoretical models for ejecta retention and regolith production on asteroid-sized bodies. Early theoretical estimates of regolith thickness by *Langevin and Maurette* [1980] suggested that weak asteroid-sized bodies were capable of retaining little ejecta (<100 m), with relatively little dependence on the size of the asteroid, while strong bodies larger than about 50 km would have regoliths of 100-900 m depending on asteroid size [*Langevin and Maurette,* 1980]. *Housen et al.* [1979] calculated that a 10-km asteroid should have no regolith, but 100-, 300-, and 500-km asteroids should have regoliths of 0.2, 3.5, and 1.2 km, respectively. However, these calculations are strongly dependent on the ejecta velocities, which in turn, depend on target strength. *Housen* [1992] and *Asphaug and Nolan* [1992] have recently presented theoretical treatments that indicate that even asteroids in the 10-km size range are capable of retaining substantial regoliths if they have low strength.

Table 2. Parameters Used for the Calculations

Symbol	Meaning	Value
ϕ	sphericity of chondrules	0.9
ε	void fraction	0.4
ρ_c, ρ_l	density of chondritic silicates	3200
ρ_s, ρ_h	density of metal	7800
d	diameter of chondrules	3×10^{-4} - 9×10^{-4}
g	gravitational acceleration	0.0894
G	gravitational constant	6.67×10^{-11}
R	radius of the asteroid	10^5
M	mass of the asteroid	1.34×10^{19}
C_p	specific heat	1000
T_0	initial surface temperature	200
K_D	constant	6.75×10^{-2}
α_D	constant	0.29
$h(g)$	$g/6$	1.49×10^{-2}
Δs	maximum regolith depth	2200
σ	Stefan-Boltzmann constant	5.67×10^{-8}
φ	radiation exchange factor	0.88
τ	gardening depth	963
K_p	permeability	10^{-10} - 10^{-11}
K	thermal conductivity	1.0
κ	thermal diffusivity	3.3
Q	heat generation	$11.64 \times (^{26}Al/^{27}Al)$
λ	decay constant (^{26}Al)	3.05×10^{-14}
E	emissivity	0.8 ($h = 1.0$ m^{-1})
m	adiabatic flow of water vapor	0.857
μ	viscosity of water vapor (temperature dependent)	1.5 - 2.5×10^{-5}

All values are MKS units.

If one assumes that small asteroids can have regolith depths comparable to their deepest craters, then regolith depths can be estimated from crater sizes. The energy-crater diameter scaling is derived from the "cratering law" [*Housen et al.,* 1979],

$$D = K_D (1/2mv_i^2)^{\alpha} h(g) \tag{5}$$

where D is the diameter of the crater, v_i is the impact velocity, K_D and α are constants, and $h(g)$ is a function of gravitational acceleration. For a weak body the maximum crater diameter is obtained when the impact kinetic energy per unit volume is about 10^4 [*Housen et al.,* 1979]. A maximum crater diameter of 12.6 km is obtained for a weak body of 100 km in radius. For the lunar regolith *Hörz et al.* [1991] determined an empirical equation:

$$\Delta s = aD^b \tag{6}$$

where $a = 1.044$, $b = 0.301$, and D is diameter of crater. In this case the depth is about 2.2 km. Thus, theoretical estimates might still be underestimating regolith depths on small bodies. Regoliths of several kilometers depth might be expected on 10-500 km meteorite parent bodies.

Required Gas Flow Rates

We can estimate theoretical minimum gas flow rates for fluidization of the regolith using the Ergun equation [*Kunii and Levenspiel,* 1991], which equates the upward drag force to the downward gravitational force:

$$\frac{1.75R_e^2}{\varepsilon^3\phi} + \frac{150(1-\varepsilon)R_e}{\varepsilon^3\phi^2} = \frac{d^3\rho_g(\rho_s - \rho_g)g}{\mu^2} \tag{7}$$

where ε is the void fraction under minimum flow conditions, ϕ the sphericity and d the diameter of the particles, ρ_g and ρ_s the densities of the gas (calculated assuming an ideal H_2O gas) and solids, μ the viscosity of the gas, and g the acceleration due to gravity. The Reynold's number, R_e, depends on flow rate of the gases:

$$R_e = dv_{mf}\rho_g / \mu \tag{8}$$

where v_{mf} is the minimum flow rate required for fluidization. Numerical values for the parameters are given in Table 2 and the results are shown in Figure 2. The velocities required to sustain fluidization on these small, asteroid-sized objects are small, namely, ~1, ~10, and ~100 mm/s for 10, 100, and 1000 km radius parent objects, respectively. These very low velocities should be readily achievable, on at least a localized basis, in any number of reasonable scenarios. The main question is whether the flux of volatiles is sufficient, and this will depend on the composition of the planetesimal and the available heat sources.

Required Flux of Volatiles

The most characteristic feature of primitive solar system materials, comets, asteroids and CI chondrites, is the abundance of volatiles, especially water. The CI chondrites contain up to 20 wt% water, while CM chondrites are typically 10 wt% water [*Wiik,* 1969; *Jarosewich,* 1990], and spectroscopic observations of comets indicate perhaps even higher values [*Jessberger et al.*, 1989]. Absorption features in their reflectance spectra near 3 μm suggest the presence of water as hydrated silicates or ice on many asteroids [*Lebofsky et al.,* 1989; *Larson and Veeder,* 1979].

Dehydration of carbonaceous chondrites occurs between 600 and 900 K, and 90-95% by volume of the evolved gas is a mixture of H_2O, CO, and CO_2, the remainder being mainly CH_4 and SO_2 [*Lebofsky et al.,* 1989]. Serpentine will lose its structural water at a rate dependent on fluid pressure and temperature when the shock-induced entropy exceeds a critical value that is lower for porous media than consolidated rock

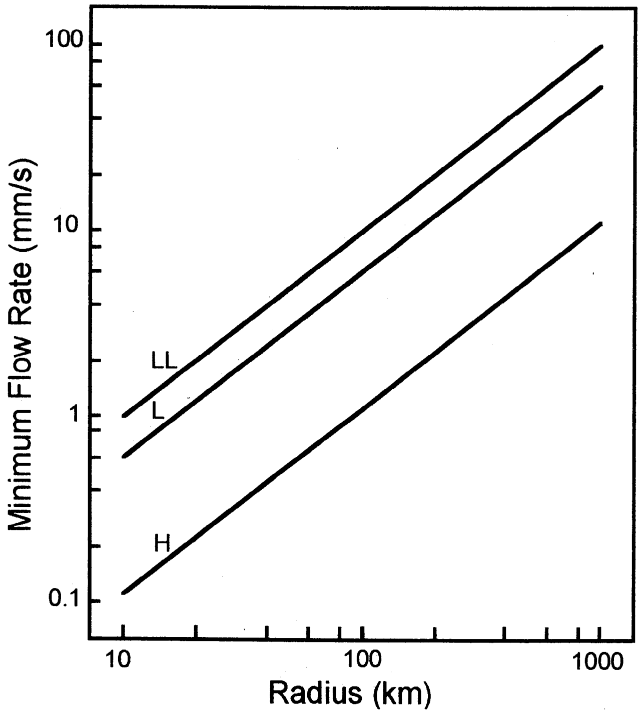

Figure 2. Calculated minimum flow rates required for a gas to fluidize a thick dust layer on an accreting meteorite parent body as a function of the radius of the body. The curves were calculated using the Ergun equation (equation (7)) which equates the upward drag of the gases on the particles to the downward gravitational force. The flow rates required to sustain fluidization of the chondrules, the largest components in the meteorites, are quite small (0.1 to 1.0 mm/s) and increase with chondrule size along the H, L, and LL series.

[*Lange and Ahrens,* 1982]. For the dehydration of serpentine to produce olivine, the equilibrium phase boundary lies at ~630 K for a fluid pressure of 1 kbar [*Johannes,* 1968], and since fluid pressure in asteroids is normally below this [*Grimm and McSween,* 1989], dehydration will occur at temperatures of <630 K. The timescale for impact-induced dehydration of hydrous minerals for kilometer-sized impactors is 10-100 s [*Kieffer and Simonds,* 1980], which we will show below is considerably shorter than the degassing duration.

The flow rate of water vapor in the regolith depends on the pressure gradient generated by differences in temperature and water production rate, the latter depending on the competing processes of dehydration and hydration. One-dimensional flow under steady state conditions is described by Darcy's law

$$v = -\frac{K_p}{\mu}\left[\frac{\partial p}{\partial x} - (\rho_g - \rho_d)g\right] \tag{9}$$

where μ is the viscosity of water vapor and ρ_d the density of the dust particles. The term $\rho_g g$ is negligible compared with the pressure gradient, $\partial p/\partial x$, which is given by [*Muskat,* 1982]

$$\frac{\partial p}{\partial x} = \frac{p_1^{1+m} - p_2^{1+m}}{L(1+m)p_1^m} \tag{10}$$

where p_1 and p_2 are the boundary values of the pressure at $x = 0$ and $x = L$. Combining (9) and (10), the rate of gas flow (m/s) in the regolith is given by

$$v = \frac{K_p}{\mu}\left[\frac{(p_b^{1+m} - p_s^{1+m})}{\Delta s(1+m)p_b^m} - \rho_d g\right] \tag{11}$$

where p_s, p_b are the calculated vapor pressures of water at the surface and bottom of the dust layer. For adiabatic heat flow in water vapor, $m = C_v/C_p = 0.857$. The main variable in equation (11) is permeability (K_p), which has not been determined for asteroid regolith material. The permeability of lunar regolith, as determined from Apollo 11 soils, is about 10^{-12}-10^{-11} m^2 [*Costes and Mitchell,* 1970], in agreement with a value determined in situ by Surveyor III of 1-5 x 10^{-12} m^2 [*Chaote et al.,* 1968]. The permeability of loose beds of sand of similar size are in the range of 10^{-11} to 10^{-10} m^2 [*Kaviany,* 1991]. A fluidized dust layer of the sort envisaged here probably resembles more closely the latter situation so we adopt a permeability of 10^{-11}-10^{-10} m^2 for our calculations. Other parameters are also given in Table 2.

Heat Sources

Both impact heating and decay of ^{26}Al appear to be feasible heat sources for the present process. Our detailed calculations are given in the appendix, and the results appear in Figure 3. Important in the present considerations of impact heating, and in a departure from most earlier work, is (1) that we allow for the presence of a thick insulating regolith and (2) that we assume impact velocities were 1-5 km s^{-1}. Previous studies assumed that most of the impact energy radiates immediately into space [*Sonett et al.,* 1975], so that the surface temperature of the planetesimals was constant throughout accretion. Thermal blanketing by the dust layer will restrict heat loss from the

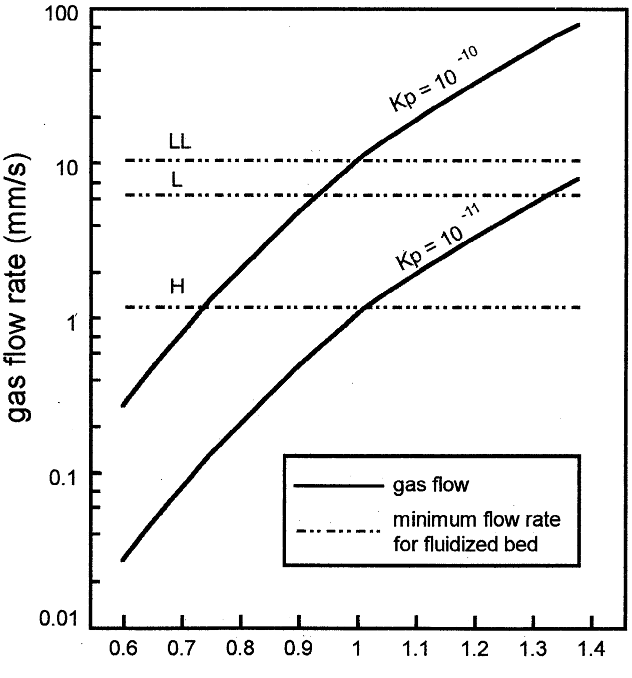

Figure 3a. Calculated gas flow rate in regolith on a meteorite parent body as a function of impact velocity and permeability (K_p, m^2) for an object with radius 100 km whose surface is heated by impact. Surface temperatures, and therefore gas flow rates, increase steadily with increasing impact velocity.

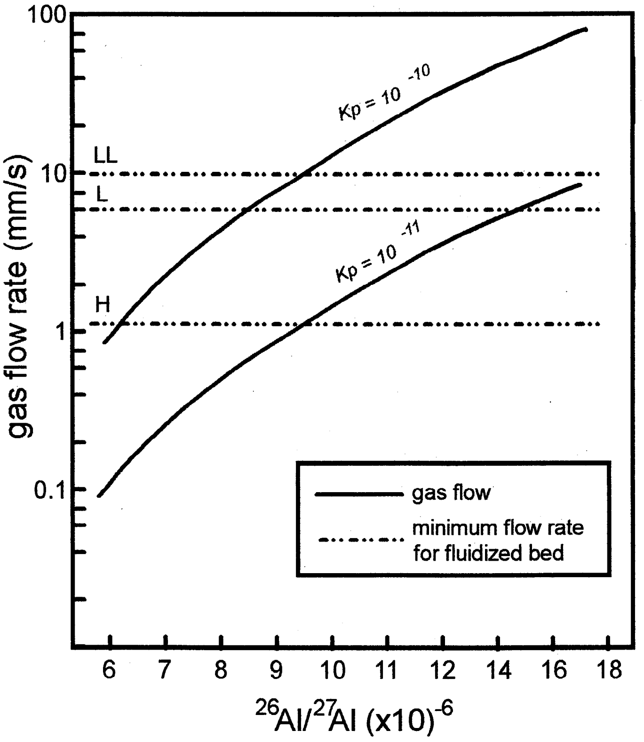

Figure 3b. Calculated gas flow rate in the accreting dust layers or regolith of a meteorite parent body as a function of ^{26}Al abundance and permeability for an object with radius 100 km heated by radioactive decay of ^{26}Al. The regolith depth is assumed to be 2.2 km, and gas flow in the regolith is assumed to be adiabatic and under steady state conditions. The minimum flow rates required to sustain fluidization of H, L, and LL chondrites (taken from Figure 2) are also indicated. Depending on permeability, impact velocities of 0.7 to 1.3 km/s or ^{26}Al with ^{26}Al/^{27}Al ratios of 7×10^{-6} to 1.6×10^{-5} provide sufficient heat to mobilize the surface dust layers.

surface and increase surface temperature, while repeated impact and degassing will stir the regolith and cause thermodynamic equilibrium throughout the layer.

The gravitational potential energy acquired by infalling material is equivalent to a temperature increase of about 9×10^{-4} R^2, assuming a density of 3200 kg m^{-3} and a specific heat of 1000 J kg^{-1} K^{-1}. If impact velocities were comparable to escape velocities during accretion (a few hundred m s^{-1}), as suggested by some investigators [*Safronov,* 1978], then accretion on a 100-km object results in temperature increases of only ~9 K. However, numerical studies by *Davis et al.* [1979] and *Chapman et al.* [1978] suggest that asteroidal encounter velocities were pumped up to the present values of ~5 km s^{-1} within ~10^5 years due to gravitational interactions with Jupiter. Since accretion was occurring for several million years [*Podosek and Cassen,* 1994], impact velocities of several km s^{-1} must have been readily achievable during the late stages of accretion. Surface temperatures increase rapidly with increasing impact velocity, and impact velocities of ~1.0, ~1.2, and ~1.4 km s^{-1} raise the surface temperature to 590, 690, and 785 K, respectively, well above that required to dehydrate silicates.

The effectiveness of ^{26}Al as a heat source depends on the ^{26}Al content at the time of aggregation, which in turn, depends on the time that has elapsed since the end of nucleosynthesis. Inferred initial ^{26}Al/^{27}Al ratios as high as 6×10^{-5} have been determined for refractory inclusions in Allende and other CV chondrites [*Lee et al.,* 1976], but these may not be representative of the entire parent body. The only detected excess ^{26}Mg for material other than the CAIs corresponds to an initial ^{26}Al/^{27}Al of $8 \pm 2 \times 10^{-6}$ [*Hutcheon et al.,* 1989] and was found in a chondrule from Semarkona. Depending on the permeability, ^{26}Al/^{27}Al values of 6×10^{-6} to 8×10^{-5} seem to be capable of providing the required heat output (Figure 3). The calculated gas flow rates are of the order of minimum flow rates required to sustain fluidization of chondritic objects with reasonable ^{26}Al abundances and permeability values.

Duration of Degassing

Density and size separation in a terrestrial fluidized bed occurs very rapidly [*Rowe et al.,* 1972], but it is informative to estimate the duration of degassing for a planetesimal. Assuming vapor transport in a heated porous body is in a state of quasi-static equilibrium, then the degassing zone will be in a state of vapor-condensate equilibrium, and the duration of degassing is given by the simple equation

$$t_d = W_s / (\omega S_s) \qquad (12)$$

where W_s is the mass of water present in the asteroid surface, ω the rate of mass flow per unit cross section and S_s the surface area of the asteroid. For objects of 100-km radius with 10 vol% of water as hydrated silicates and ^{26}Al/^{27}Al of 10^{-5} or impact velocity of ~1 km s^{-1}, the calculated timescale of degassing is of the order of a few months to a few years, depending on permeability.

Metal-Silicate Separation

In a two-component fluidized bed the general rule is that the denser phase falls to the bottom and is referred to as jetsam. Under special, and rather rare circumstances, the denser phase will move to the surface to become flotsam. Both situations suggest interesting explanations for the metal-silicate fractionation in ordinary chondrites. Additionally, the notion that metal might rise to the surface offers new explanations for the formation of metal-rich chondrites and for the metal-rich composition of certain asteroid surfaces which in other respects resemble ordinary chondrites in composition.

The Metal as Jetsam

The behavior of a two-component bed in which the higher density material falls to the bottom of the bed has been experimentally and numerically modeled, and the results are summarized in Figure 4. The degree of mixing can be described by the f parameter [*Rowe et al.,* 1972]:

$$f = x / x_{bulk} \qquad (13)$$

where x is the proportion of jetsam in the upper part of the bed and x_{bulk} is fraction of the jetsam in the entire bed. When the bed is completely segregated, $f = 0$ (Figure 4a), while for perfect mixing $f = 1$ (Figure 4d [from *Rowe et al.,* 1972]).

Segregation is particularly sensitive to density differences, but is also a function of gas flow rate, maximum segregation being

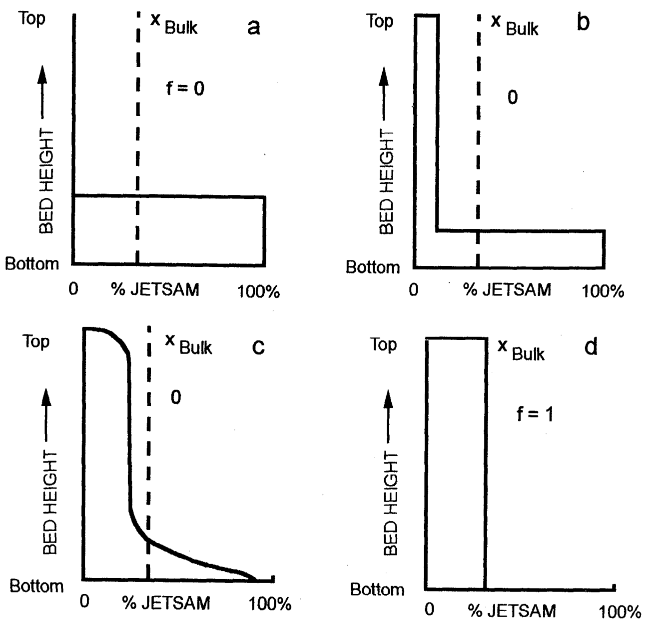

Figure 4. Segregation patterns of particles with different density or size as a function of gas flow velocity in which the dense component sinks to the bottom (i.e. jetsam). Segregation decreases as gas flow rates increase, going from Figure 4a to Figure 4d. Segregation occurs very rapidly in a fluidized regolith, with a relatively pure layer of high-density or large-size particles normally forming at the bottom of the bed and an impure layer of low-density or small-size particles forming at the top, with some dense or large ones being homogeneously dispersed throughout the whole bed. (a) Complete and perfect segregation. (b) Idealized segregation as observed at low flow velocities. (c) Segregation pattern observed at higher flow velocities. (d) Perfect mixing. Figure reproduced from *Rowe et al.* [1972]. When the metal is present in a small amount (<15 vol. %) and chondrule to metal size ratios are high, metal might rise to the surface, and these diagrams are essentially inverted (see Figures 5 and 6).

obtained when the gas flow rate is just above the minimum rate required for fluidization. Higher flow rates result in mixing. *Rowe et al.* [1972] derived empirical relationships for the separation of particles of size d_b and d_s (where the subscripts refer to "big" and "small") and density of ρ_h and ρ_l (where the subscripts refer to "heavy" and "light" using beds of copper, glass, steel, and polystyrene. The resulting expression is

$$x = k(v - v_{mf(f)})(\rho_h / \rho_l)^{-2.5}(d_b / d_s)^{-0.2} \tag{14}$$

where $v_{mf(f)}$ is the minimum flow velocity for the flotsam (the chondrules), which is given by equation (5). In our case, the flotsam is bigger than the jetsam, so we write equation (14) as

$$x = k(v - v_{mf(f)})(\rho_h / \rho_l)^{-2.5}(d_b / d_s)^{0.2} \tag{15}$$

so that we can calculate relative metal-silicate abundances for flotsam and jetsam of H, L, and LL chondrule and metal dimensions components for flow rates just above the minimum required for fluidization. The results are given in Table 3 along with the observed values for the H, L, and LL chondrites. The model readily accounts for the observed metal-silicate fractionations of the ordinary chondrites at gas flow rates just above minimum flow velocity. This is reasonable, since one expects the present-day mixture to reflect metal-silicate fractionation just before flow ceased.

The Metal as Flotsam

The idea that the metal might rise to the top of the layer is prompted by laboratory experiments we have performed and by a study of *Chiba et al.* [1980]. We constructed a 2.5 cm diameter, 100-cm-long column out of Plexiglass with a glass wool diffuser and dry air gas phase. The column was prepared from 800-µm sand and 200-µm iron grains in L chondrite proportions. Air was passed through the column at very high flow rates in order to thoroughly mix the components. The turbulent mixed bed passed through a segregation stage before resulting in a fixed bed when the air flow was stopped. Segregation arises from the large differences in minimum flow rates needed to sustain the sand and metal. The smaller metal grains remained fluidized longer, allowing them to travel upward in the column while the sand sinks (Figure 5). Very similar behavior was reported by *Chiba et al.* [1980] for columns of glass beads and graphite and columns of copper shot and glass beads. Their data are summarized in Figure 6, a plot of the size ratio against volume fraction of the denser component. Two-component beds consisting of large low-density objects and a relatively small fraction of small high-density components can result in the higher-density component rising to the surface. *Chiba et al.* [1980] suggested that the smaller components were rising to the surface through the pore spaces between the larger objects. In beds consisting of >15 vol% of the dense component, the more normal behavior is observed, and the denser component sinks. There is an intermediate range for the amount of the denser phase in which mixing occurs. Data for the H, L, and LL chondrites are superimposed on Figure 6. The L and LL chondrites plot in the region in which metal is expected to be flotsam, while H chondrites plot in the region where metal is expected to be jetsam.

Major Properties of Chondrites and Proposed Model

Our present objectives were limited to offering a new explanation for the metal-silicate fractionation of ordinary chondrites, but clearly our efforts would be futile if our ideas were inconsistent with available information. Without going into great detail, we will point out these ideas are not only consistent with the data in Table 1, but may offer new insights in to some

Table 3. Calculated and Observed Metal Abundances in H, L, and LL Ordinary Chondrite Meteorites

	Metal Weight Ratio LL : L : H
Calculated	
$u = 1.3 \times u_{mf}$	1.0 : 2.1 : 3.9
$u = 1.2 \times u_{mf}$	1.0 : 2.7 : 6.8
$u = 1.1 \times u_{mf}$	1.0 : 4.5 : 7.4
Observed	
	1.0 : 3.0 : 8.0

Here u_{mf} refers to the minimum flow rate of LL chondrules, calculated from the Ergun equation.

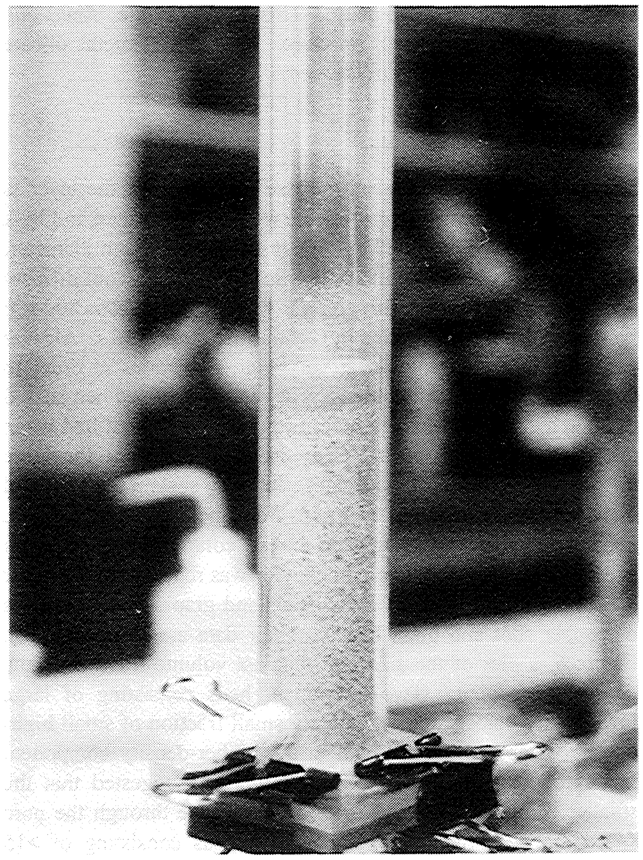

Figure 5. A 2.5-cm-diameter by 20-cm-long fluidized bed of 800-μm sand and 200-μm iron fluidized with air, the metal constituting about 15 volume percent of the bed. The components were first thoroughly mixed by allowing the gas to pass rapidly through the bed. As the gas flow was slowly terminated, the sand settled to the bottom, and the silver-colored metal rose to the surface of the column, forming a metal-sand mixture in the upper half of the bed and a metal-rich surface layer. The results are consistent with the results of *Chiba et al.* [1980], shown in Figure 6.

data. Most significantly, much of the confusion over whether specific chondrite properties were nebular or parent body may be moot if the meteorites spent part of their history in a dynamic regolith where gas-, liquid- and solid-phase reactions were possible.

Metal and Chondrule Size Distributions

We have argued that the bulk Fe/Si of H, L, and LL chondrites is an inevitable result of size and density sorting in a thick regolith. We have not specifically addressed the question of how the chondrules and metal grains acquired their dimensions. It is possible that aerodynamic sorting occurred in a fashion similar to that described here, as discussed by many authors [*Whipple,* 1971, 1972; *Dodd,* 1976; *Rubin and Keil,* 1984; *Leenhouts and Skinner*, 1991; *Skinner and Leenhouts*, 1991, 1993a,b; *Haack and Scott*, 1993; *Scott and Haack*, 1993]. However, there is also the possibility that the present size distributions of the chondrules and the metal grains are a primary property acquired during formation. For example, if the chondrules were melt droplets produced by impact, then ballistic sorting would occur in the manner in which meteorite shower fragments are size-sorted during fall, the largest fragments travelling farthest along the line of travel [*Sears et al.*, 1996; *Nininger,* 1952].

Fate of the Metal-Rich Layer

Whether the metal sank to the bottom or rose to the surface of the regolith layer, there will be a metal-rich layer, and there are several candidates for samples of this metal-rich layer among the recovered meteorites. The IIE iron meteorites have been related to the H chondrites, and the IIIAB iron meteorites have been related to the L chondrites, for instance [*Scott and Wasson,* 1975], but if these represent the metal-rich layer described here, there would have been considerable impact melting in order to explain the trends in trace element chemistry thought to be due to fractional crystallization. A more plausible possibility is that the metal-rich chondrites, Weatherford, Bencubbin, or Allan Hills 85085, or even the CR chondrites are samples of the metal-rich regolith layer [*Mason and Nelen,* 1968; *Weisberg et al.,* 1988, 1990, 1993; *Scott,* 1988; *Grossman et al.,* 1988b]. However, we stress that there is no reason to believe that the metal-rich layer is represented in our meteorite collections. The vagaries of orbit evolution [*Greenberg and Chapman,* 1983; *Greenberg and Nolan,* 1989; *Wetherill and Chapman,* 1988; *Gaffey et al.,* 1993a] and the destructiveness of atmospheric passage [*e.g., Buchwald,* 1975; *Baldwin and Shaeffer,* 1971] ensure that the meteorites falling to Earth are not a representative sample of material in the asteroid belt. At the same time, clustering in the

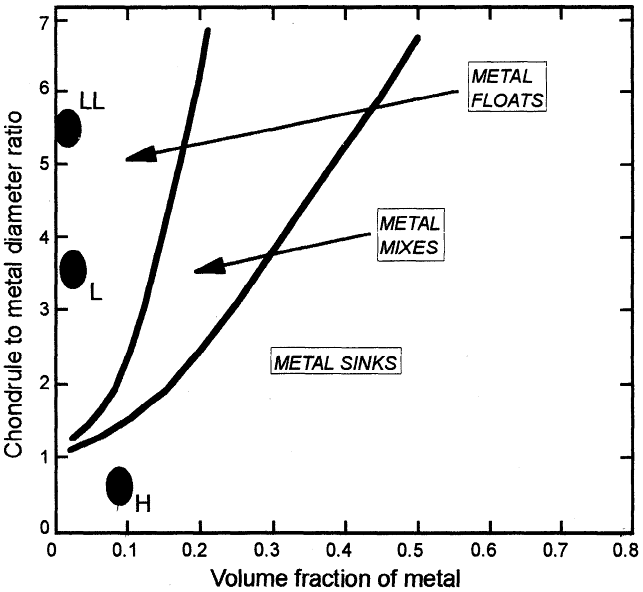

Figure 6. Diagram after *Chiba et al.* [1980] comparing the behavior of the dense component in the two-component fluidized bed as the gas flow is slowly terminated. Chiba et al. used graphite and glass beads for their experiments but mentioned that they found similar results for beds consisting of copper and glass. The denser component sinks in a two-component fluidized bed (e.g., the H chondrites, although the separation is minimal) except when the metal is present in small amounts and the diameter ratio for the low-density and high-density components is large (i.e., the L and LL chondrites).

cosmic ray exposure ages [*Crabb and Schultz,* 1981] indicates that very few bodies are producing the majority of meteorites falling to Earth and that stochastic processes are important in determining the nature of material that comes to Earth [*Benoit and Sears*, 1993].

Silicate Oxidation State and Reactions

The chemical reactions that caused the observed differences in oxidation state of the Fe could have occurred in the nebula prior to accretion, during chondrule formation, or on the parent body during metamorphism. It seems unlikely that significant oxidation occurred during the brief degassing period discussed here. The diffusion time, t, can be estimated from the relationship $t \sim r^2/D(T)$, where r is the grain radius and $D(T)$ is the diffusion coefficient. Diffusion times for a 1-μm grain at 800 K are of the order of 10^5 years. This does not preclude occasional reactions, as for instance, some of the water vapor condensed in cool regions of the regolith. Several of the least metamorphosed ordinary chondrites contain evidence for traces of aqueous alteration [*Hutchison et al.*, 1987; *Alexander et al.*, 1989; *Hutchison,* 1992].

Matrix and Dust Loss

By "matrix" we mean the very fine-grained smoke over which there is considerable controversy as to origin and history [*Scott et al.*, 1988], rather than the relatively coarse grained interchondrule matrix that is clearly comminuted coarse-grained material [e.g., *Sears et al.*, 1991]. Some authors suggest that the fine-grained matrix is interstellar material or nebular condensate [*Allen et al.*, 1980; *Nagahara,* 1984; *MacPherson* et al., 1985], while others have suggested that it is volatiles that have recondensed following evaporation from chondrules [*Alexander,* 1994; *Huang et al.*, 1996] or that it was otherwise derived from chondrules [*Alexander et al.*, 1989]. A variety of theoretical and experiment studies indicate that this fine-grained material should be lost from the regolith by a process known as "elutriation" [*Kunii and Levenspiel,* 1991]. This occurs when the gas exceeds the terminal velocity of the lightest particles and they are carried out of the bed. The terminal velocity for a particle is typically 2 orders of magnitude higher than its minimum flow velocity [*Kunii and Levenspiel,* 1991]. Some of this fine-grained dust reaches the upper atmosphere of Earth as "asteroidal" interplanetary dust [*Bradley et al.*, 1988]. If the degassing occurs through venting, or regional rather than global processes, then the dust might eventually return to other locations on the same asteroid. An important factor in the retention of matrix in the regolith is that much of the matrix is physically attached to the chondrules as a rim [*Metzler et al.*, 1992; *Huang et al.,* 1996]. This would explain why the chondrite classes that have suffered the greatest metal-silicate fractionation, the enstatite chondrites, have the lowest amounts of matrix and why classes that have suffered little or no fractionation of metal and silicate (all but a few of the carbonaceous chondrites) contain considerable amounts of fine-grained matrix.

Oxygen Isotopes

The present model for fractionating metal and silicates is consistent with existing oxygen isotope data, in that it allows for the accreting material to have any isotopic composition, while providing an environment for parent body gas-phase reactions involving oxygen-bearing gases. *Clayton et al.* [1983] suggested that the gas had an oxygen isotope composition similar to that of terrestrial oxygen, while *Clayton et al.* [1991] preferred a reservoir enriched in heavy oxygen. Of course, these arguments are moot if, as suggested by *Thiemens* [1994], the oxygen isotope fractionations occurred during chondrule formation without exchange with the ambient gases being involved.

Thermal History

King et al. [1972] have suggested that chondrites formed in unconsolidated fall-back deposits on the surface layers of a meteorite parent body. Their suggestion, which was prompted by their conclusion that chondrules are of impact origin, is somewhat similar to the present proposal. However, King et al. did not discuss their suggestion quantitatively, particularly the thermal histories of the chondrites. Most ordinary chondrites have suffered peak metamorphic temperatures of 700°C to 950°C, while a few have peak temperatures of ~350°C to 700°C (see *McSween et al.* [1988] for a brief review). Cooling rates for the ordinary chondrites are 1-100°C/Myr [*Wood,* 1979]. These cooling rate estimates suggest burial depths of ~100 km, assuming that the heat source was ^{26}Al [*Minster and Allégre,* 1979; *Miyamoto et al.*, 1981; *Grimm,* 1985; *Grimm and McSween,* 1989; *Bennett and McSween,* 1996] or ~300 km, assuming that the heat source was the decay of U, Th, and K [*Fricker et al.*, 1970; *Herndon and Rowe,* 1973]. It should be stressed that all these estimates are for burial depths, and thereby minimum radii. In both cases, compact parent bodies with thermal properties equal to those of the present lithified rocks were assumed. However, thermal history is highly sensitive to the physical form of the body and, in particular, the presence or otherwise of a regolith [*Haack et al.*, 1990]. An unconsolidated regolith of 1-10 km dimensions is also capable of producing the observed chondrite cooling rates. In a homogeneous regolith the cooling time t at a depth Δs is given by $t \sim \Delta s^2/\kappa_r$, where κ_r is the thermal diffusivity of regolith ($K_r/\rho_r C_{p,r}$, where K_r is the thermal conductivity, 0.01W m^{-1} K^{-1} [*Wood,* 1979], ρ_r the density (1500 kg m^{-3} [*Hörz et al.*, 1991]), and $C_{p,r}$ the specific heat (1000 J kg^{-1} K^{-1})). Cooling rates of the order of 2 to 200 K Myr^{-1} correspond to depths of 1-10 km in an unconsolidated regolith. Wood came to a similar conclusion when he showed that the observed cooling rates could be produced by 20-km dust balls being sintered to 2-km objects. The observed metallographic cooling rates and peak temperatures (600-1000°C) are clearly consistent with the formation of chondrites in the unconsolidated regolith several kilometers thick in which we think the metal-silicate fractionation occurred.

There is considerable controversy over meteorite parent body structure and whether cooling rates correlate with metamorphic temperature. A correlation would imply a simple onion-shell structure for the parent objects or, in the present scenario, that the regolith was thermally zoned. We would then assume that the most intensely metamorphosed material was from the center of the parent body or bottom of the regolith, while the essentially unmetamorphosed material was close to or at the surface. A lack of correlation between cooling rate and peak metamorphic temperature would imply a more complex history, perhaps involving "rubble stone" structures [*Miyamoto et al.*, 1981; *Scott and Rajan,* 1981] or considerable brecciation.

Lithification

It has always been unclear as to how unconsolidated chondritic material, namely, the matrix, chondrules, metal, sulfides, and other primary components, came to be converted to compact stones. The comet simulation experiments of *Grün et al.* [1992] involve a degassing body containing particulates. They find that when water and carbon monoxide ice evaporate under conditions resembling those of the inner solar system, silicate grains accumulate on the surface and form a crust. One might well imagine that body with much higher amounts of dust maintained at black body temperatures appropriate for the asteroid belt, say, 200 K, would form thick dusty layers that would crust over as degassing came to an end. Petrographic observations of regolithic breccias suggest that the lithification process involved surficial melting of feldspar grains [*Ashworth and Barber,* 1976; *Bischoff et al.*, 1983], but this would have to await the end of degassing when grains can come into close physical contact. *Ashworth and Barber* [1976] and *Bischoff et al.* [1983] were considering the lithification of gas-rich regolith breccias and assumed that continued impact would provide the minor amounts of heating required for lithification. We assume that similar mechanisms might have lithified all chondrites.

Implications for the Nature of Asteroid Surfaces

Except for recent estimates of the density of Ida, which have large uncertainties [*Belton et al.,* 1995], and geochemical arguments for Vesta [*Pieters and Binzel,* 1994; *Delaney,* 1995], our knowledge of asteroidal compositions is inferred almost entirely from the radiation reflected from their surfaces [*Gaffey et al.,* 1993a]. However, the ordinary chondrites have remarkably few asteroid matches. *Gaffey et al.* [1993b] argue that some members of the SIV type are ordinary chondrites, but even then their surfaces appear to be richer in metal than the ordinary chondrites. It might be that ordinary chondrites derive from particularly small asteroids [*Bell et al.,* 1989], but recent surveys make this unlikely [*Gaffey et al.,* 1993a]. Alternatively, maybe S asteroids are the parent bodies of ordinary chondrites but their surfaces have been altered by impact [*Britt and Pieters,* 1991] or preferential loss silicates, as suggested by the inverse relation between absorption band and asteroid diameter [*Gaffey et al.,* 1993b]. The present work provides a mechanism by which silicates can be lost from their surfaces by elutriation but also provides a means by which metal grains might be enriched on the surface of the parent body.

Appendix: Impact and Radiogenic Heating of the Thick Dusty Layer on an Asteroid-Sized Body

Impact Heating

The equilibrium between accretional energy and radiative energy is then given by

$$\frac{1}{2}M_s u^2 + \frac{GM}{R}M_s = 4\pi R^2 F\Delta t + C_p M_s(T - T_0) \qquad \text{(A1)}$$

where u is the velocity of accreting material, M and M_s the masses of the body and mass of the surface layer, respectively, G the gravitational constant, Δt the time interval for the parent body to grow from (R-Δs) to R (Δs being the thickness of the surface layer), C_p the specific heat of the body, T_0 and T the temperature of the surface dust layer and the dust at depth Δs, respectively, and F the energy flux escaping from the surface into space through the dust layer. We have ignored the heat evaporation and other chemical transformations, since they are relatively small and we are interested in first-order calculations. For simplicity, Δt is calculated by assuming constant mass flux of 10^{-6} kg m^{-2} s^{-1}, which means that the accretion time of a 100-km-radius body is ~3.5 x 10^6 years, but this is not a critical assumption. Heat conduction to the interior from the hot surface regions is negligible. Energy lost with escaping material is also ignored, since it is minor compared with the internal energy [*O'Keefe,* 1977]. Numerical values for the parameters are given in Table 2.

If a significant proportion of the mass accreted was as sizable planetesimals then some of the heat generated by their infall would have been buried deep enough not to have been lost by the asteroid [*Kaula,* 1979]. Thus the surface layer is likely to be in thermal and dynamic equilibrium. The radiative heat transfer into space through the porous regolith under steady state conditions is given by [*Vortmeyer,* 1978]

$$F = \frac{\varphi\sigma}{\frac{2-E}{E} + \frac{\tau}{d}}(T^4 - T_0^4) \qquad \text{(A2)}$$

where ϕ is the radiation exchange factor (assumed to be wavelength-independent), σ the Stefan-Boltzmann constant, and E and d the surface emissivity and diameter of particles (we assume ~900 μm chondrules) in the surface layer, respectively. Also, τ is typical gardening depth (~960 m, evaluated from the maximum crater depth [*Housen et al.,* 1979, equations 6 and 6a]. Numerical values of the parameters are also given in Table 2.

We also find from equations (A1) and (A2) that impact heating in such a dusty environment is insensitive to the asteroidal growth time. Rapid formation is not required, and assumed mass fluxes between 10^{-7} and 10^{-5} kg m^{-2} s^{-1} yield similar temperatures.

Heating by ^{26}Al

Since the discovery of its decay product (i.e., excess ^{26}Mg) in Allende [*Lee et al.,* 1976], ^{26}Al has been considered a potential heat source for parent body thermal evolution because of its high decay energy, expected high abundance, and reasonably long half-life. For a body with an initial uniform temperature T_0 of 200 K in a medium at temperature T_0, the temperature as a function of time and depth is given by [*Miyamoto et al.,* 1981]

$$T = T_0 + \frac{2hR^2A}{rK}\sum_{n=1}^{\infty}\frac{1}{1-\frac{\lambda}{\kappa\alpha_n^2}}\frac{[\exp(-\lambda t) - \exp(-\kappa\alpha_n^2 t)]\sin(r\alpha_n)}{\alpha_n^2[R^2\alpha_n^2 + Rh(Rh-1)]\sin(R\alpha_n)} \qquad \text{(A3)}$$

where a_n, n = 1, 2 ..., are the roots of $R\,\alpha\cot(R\,\alpha) + R\,h - 1 = 0$. R and r are the radius and radial distance from the center of the parent body, respectively, and Q the heat generation by ^{26}Al [*Herndon and Herndon,* 1977]. We fairly arbitrarily chose R = 100 km for our calculations, but other values produce very similar results. Other parameters are given in Table 2, which are evaluated from *Matsui and Osaka* [1979]. The gas flow rates in the surface layer calculated using equation (A3) as a function of $^{26}Al/^{27}Al$ and permeability are shown in Figure 3.

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