Asteroid Differentiation

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Differentiation of silicate bodies was widespread in the inner solar system, producing an array of igneous meteorites. While the precursor material was certainly chondritic, direct links between chondrites and achondrites are few and tenuous. Heating by short-lived radionuclides (^{26}Al and ^{60}Fe) initiated melting. Formation of a basaltic crust would occur soon after the onset of melting, as would migration of the cotectic Fe,Ni-FeS component, perhaps forming a S-rich core. While some asteroids appear to be arrested at this point, many continued to heat up, with efficient segregation of metal to form large cores. Extensive melting of the silicate mantle occurred, perhaps reaching 40–75%. In the largest asteroids, the original, basaltic crust may have been resorbed into the mantle, producing a magma ocean. The crystallization of that magma ocean, whether by equilibrium or fractional crystallization, could have produced the layered structure typical of large differentiated asteroids like 4 Vesta. Spacecraft exploration of 4 Vesta promises to further extend our knowledge of asteroid differentiation.

1. INTRODUCTION

The transformation of rocky bodies from an agglomerate of nebular particles to fully differentiated, layered worlds consisting of core, mantle, and crust is, without question, one of the most fundamental processes to have occurred during the birth of our solar system. The products of this transformation abound. The basaltic eucrites, iron meteorites, and pallasites all originated on asteroidal bodies that experienced extensive differentiation. Since the publication of *Meteorites and the Early Solar System* (*Kerridge and Matthews,* 1988), primitive achondrites — those meteorites that experienced only partial differentiation — have also been widely recognized, marking one of the major advances in our understanding of asteroidal differentiation. While primitive chondrites may dominate the flux of material encountering Earth, the majority of distinct asteroids that have delivered meteorites to Earth are differentiated (*Burbine et al.,* 2002). All the terrestrial planets and our own Moon experienced igneous differentiation. In short, differentiation appears to have been widespread in the inner solar system.

While planetary differentiation in general appears to have been a dominant process in the inner solar system, the differentiation of asteroids in particular forms an end member in understanding large-scale igneous processes. In comparison to Earth, asteroidal differentiation occurred at much lower gravity (radii ≤ 500 km; g < 0.03 terrestrial); was driven by early, intense heat sources; and operated over a much wider range of f_{O_2} conditions [~IW+2 for angrites (*Jurewicz et al.,* 1993) to IW-5 for aubrites (*Keil,* 1989)] and over a broader range of states of hydration. Thus, studies of asteroidal differentiation provide a natural complement to studies of igneous systems under terrestrial conditions.

Despite the obvious interest in asteroid differentiation, a complete understanding of this process has been and remains elusive, in large part owing to the remoteness in time and space of the locations in which this event occurred. While Earth experienced planetwide melting and differentiation, subsequent geologic processes have largely obliterated the early record of its differentiation. Large parts of the differentiated Earth remain inaccessible, and thus analogous materials to asteroidal cores are not available. Finally, all meteorites are samples without geologic context and enormous effort has been expended to link meteorites from common asteroidal parents. Even now, the first detailed examination of a differentiated asteroid and its components remains years in the future.

2. APPROACH

In this chapter, we address the differentiation of an asteroidal body from its primitive, precursor material through heating; the onset of differentiation; the separation of a core, mantle, and crust; to the solidification of the silicate portion of the asteroid. We briefly discuss possible heat sources for differentiation. Core crystallization is addressed by *Chabot and Haack* (2006). As we demonstrate, this is an enormously complex process both physically and chemically. Furthermore, myriad variations on the theme of asteroid differentiation must have occurred to produce the variety of meteorites observed in our collections. Most notable among these variations are differences in the chemical composition of the starting material, the style of melting that might be engendered by different heat sources, the role of f_{O_2} during igneous processes, and the random influence of impacts occurring during igneous differentiation [as opposed to those occurring later and forming breccias; these are addressed by *Bischoff et al.* (2006)]. For our purposes, we outline differentiation of a common precursor material and attempt to trace differentiation of an asteroid throughout that process from the onset of melting through solidification. We discuss excursions from this path only as they are warranted, particularly where differences in condition (e.g., f_{O_2}) produce dramatically different results. Finally, we emphasize the *process* of, rather than the *products* of, differentiation. Specific meteorite types are presented as evidence supporting the broad conceptual framework we outline here, but we do not discuss specific meteorite types and their origin in a systematic manner. The reader is referred to the works of *Mittlefehldt et al.* (1998), *Mittlefehldt* (2004), and *Haack and McCoy* (2004) for treatments of individual meteorite types.

3. STARTING MATERIAL

The nature of the precursor material for a differentiated asteroid can have a profound influence on the process of differentiation. It is therefore disconcerting that, for most differentiated asteroids, solid links to a precursor assemblage simply do not exist. It is widely assumed that differentiated asteroids began their existence as chondritic materials. This in large part is owed to the complete lack of any chemically and petrologically primitive materials that fall outside the broad scope of the term "chondritic" (*Brearley and Jones,* 1998; *Scott and Krot,* 2003) and, to a lesser extent, limited evidence linking chondrites and achondrites. Among the latter is the presence of relict chondrules in some acapulcoites (*McCoy et al.,* 1996), winonaites (*Benedix et al.,* 2003), and silicate-bearing IIE irons (*Bild and Wasson,* 1977), and similarities in O-isotopic composition, mineralogy, and mineral compositions between silicate-bearing IIE irons and H chondrites (*Bogard et al.,* 2000) and enstatite chondrites and aubrites (*Keil,* 1989). Even within these groups, it remains highly disputed whether we have any achondrite groups that can be linked directly to any specific group of chondrites. *Meibom and Clark* (1999) addressed the issue of the abundance of ordinary chondritic composition parent bodies within the subset of asteroids sampled as meteorites. In their assessment of likely precursors for various groups of achondrites, they found that carbonaceous chondrites appeared to be the dominant precursor material. There are, of course, important exceptions, such as the silicate-bearing IIE and IVA irons and the aubrites.

For the purposes of this chapter, we will trace the evolution of an ordinary chondrite protolith through melting and differentiation. There are several reasons for following this course. First, ordinary chondrites are well characterized both chemically and mineralogically. Normative mineralogies provide a reasonable approximation of the modal mineralogy for ordinary chondrites (e.g., *McSween et al.,* 1991) and allow comparison across groups with differing oxidation states. [The normative calculation for ordinary chondrites utilizes wet-chemistry analyses of bulk samples. A standard CIPW norm is calculated for the metal- and sulfide-free portion of the analysis, and then adding the metal and sulfide components and renormalizing to 100%; see *McSween et al.* (1991) for full description of the method.] Ordinary chondrites can be described as subequal mixtures of olivine and pyroxene, with lesser amounts of feldspar, metal, troilite, phosphates, and chromite. Carbonaceous chondrites are more olivine normative, while enstatite chondrites lack olivine in their normative compositions. Second, the melting of ordinary chondrites can be reasonably well understood both theoretically [through application of the Fe-FeS and olivine-anorthite-silica phase diagrams (e.g., *Stolper,* 1977)] and experimentally [through partial melting of ordinary chondrites (e.g., *Kushiro and Mysen,* 1979; *Takahashi,* 1983; *Jurewicz et al.,* 1995; *Feldstein et al.,* 2001)]. This approach allows reasonably robust conclusions about the temperatures for onset of melting, the composition of the melts, and, by extension, the physical properties (e.g., density, viscosity) of those melts. Together, this information proves invaluable in tracing the differentiation of an asteroid-sized body.

4. HEAT SOURCES

The differentiation of an asteroid requires a potent heat source, capable of producing high temperatures [in some cases, in excess of 1773 K (*Taylor et al.,* 1993)] in bodies with high surface-to-volume ratios. The heat source must have also operated quite early in the history of the solar system. Hafnium-182/tungsten-182 and ^{187}Re-^{187}Os systematics in iron meteorites (e.g., *Horan et al.,* 1998) suggest core formation within 5 m.y. of the formation of the first solids in the solar system. Early differentiation is also supported by the presence of excess ^{26}Mg from the decay of extinct ^{26}Al (half-life of 0.73 Ma) in the eucrites Piplia Kalan (*Srinivasan et al.,* 1999) and Asuka 881394 (*Nyquist et al.,* 2001). The identity of the heat source (or sources) remains unresolved. Several heat sources can be ruled out. Heating was not the result of the decay of long-lived radionuclides (e.g., U, K, Th) that contributed to the melting of Earth and other terrestrial planets. Similarly, accretional heating was probably minimal. *Haack and McCoy* (2004) argue that accretional heating would raise the temperature only 6 K on a body 100 km in diameter.

Impacts may have contributed to the heating of asteroids, either through thermal metamorphism or limited melting. Some authors have championed complete melting of asteroids by impact (e.g., *Lodders et al.,* 1993). However, *Keil et al.* (1997) effectively marshal the case against impacts as a global heat source and this view is now widely accepted. Perhaps most compelling among their arguments is that impacts large enough to produce postshock temperature increases of hundreds of degrees would shatter the asteroidal target. There is no question that impact played a major, perhaps even dominant, role in shaping individual asteroids and the asteroid belt as it exists today. It is also possible that these impacts produced local heating. *Rubin* (2004) documented postshock annealing in ordinary chondrites that occurred hundreds of millions of years after the birth of the solar system and must have resulted from impact heating. He argued that localized heating of the walls and floors of craters in rubble-pile asteroids might produce sufficient heat for thermal metamorphism. Impact as a heat source for local melting and differentiation has been championed in the case of the silicate-bearing IAB irons by *Wasson et al.* (1987) and *Choi et al.* (1995). While this idea continues to stimulate debate, it is clear that impact did not produce complete differentiation of asteroids and the wide range of achondrites observed in our collections.

Electrical conduction heating by the T Tauri solar wind from the pre-main-sequence Sun (e.g., *Sonett et al.,* 1970) and short-lived radioactive isotopes remain the most likely heat sources for melting of an asteroidal body. The latter were discussed extensively by *McSween et al.* (2002) in their review of asteroid thermal histories. Among the short-lived radioactive isotopes, ^{26}Al has gained increasing acceptance since the discovery of excess ^{26}Mg in Piplia Kalan (*Srinivasan et al.,* 1999). In feldspar-bearing chondritic materials, Al is evenly distributed and could act as an effective heat source. However, once melting begins, Al is sequestered into early partial melts and could effectively migrate out of the core and mantle, leaving these parts of the parent body with essentially no heat source. *Shukolyukov and Lugmair* (1992, 1996) have argued that ^{60}Fe might play a role as a heat source, although *McSween et al.* (2002) suggest that its concentration would be too low to effectively heat the mantle, producing a reverse thermal gradient. However, more recent analyses of chondritic troilite imply a higher $^{60}Fe/^{56}Fe$ ratio for the early solar system than previously thought (*Mostefaoui et al.,* 2003; *Tachibana and Huss,* 2003), suggesting that metallic cores might contain a potent heat source. It seems likely that a range of short-lived radioactive isotopes might play a role in heating asteroids and their interplay is likely complex.

5. ONSET OF MELTING

Heating and, ultimately, melting of asteroids is inherently a high-temperature, low-pressure process. Even within the centers of the largest asteroids, static pressures do not exceed 2 kbar. If the protolith is an ordinary chondrite, melting also occurs under very dry conditions. Thus, the first melts generated do not occur until ~1223 K and the first silicate melts at ~1323 K. This is in sharp contrast to Earth, where melting under even shallow crustal conditions can occur at ~1073 K for mafic rocks in the presence of water.

The earliest melt produced from an ordinary chondrite protolith occurs at the Fe,Ni-FeS eutectic at ~1223 K and contains ~85 wt% FeS (*Kullerud,* 1963). There are numerous experimental, theoretical, and petrologic data to constrain the processes that occur once this temperature is reached and melting begins. A considerable number of experimental studies (e.g., *Herpfer and Larimer,* 1993; *Ballhaus and Ellis,* 1996; *Minarik et al.,* 1996; *Shannon and Agee,* 1996, 1998; *Gaetani and Grove,* 1999; *Rushmer et al.,* 2000; *Bruhn et al.,* 2000) discuss melt migration and core formation in olivine-dominated systems. Although most of these studies were conducted at pressures considerably greater than those occurring during asteroidal melting, taken as a whole, they suggest that only anion-dominated metallic melts (mostly S in the experiments) exhibit dihedral angles less than 60° and form interconnected networks (Fig. 1). Thus, even the FeS-rich eutectic melt (anion/cation ~0.8) would not migrate to form a core under static conditions.

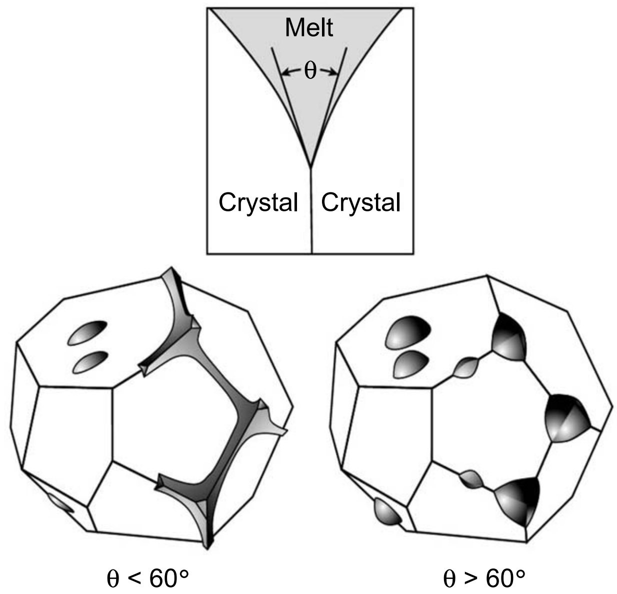

Fig. 1. Illustration of the dihedral angle θ and a depiction of two endmember microstructures for static systems. At θ < 60°, an interconnected network will form and melt migration can occur. If θ > 60°, melt forms isolated pockets. In experimental systems, interconnectedness occurs in static systems only if they are anion-rich.

During Fe,Ni-FeS eutectic melting, dynamic conditions may actually dominate in the mobility of these melts. *Keil and Wilson* (1993) suggested that overpressure created by melting at the eutectic might create veins and these veins might actually rise within the parent body due to their incorporating bubbles of preexisting gases, thus overcoming the gravitational influence on these high-density melts. The effect of this melting and melt migration is clearly seen both petrographically and geochemically in several groups of meteorites, most notably the acapulcoites and lodranites. *McCoy et al.* (1996) documented micrometer-sized veinlets occurring ubiquitously in all acapulcoites and rarer, centimeter-sized veins in the Monument Draw acapulcoite (Fig. 2).

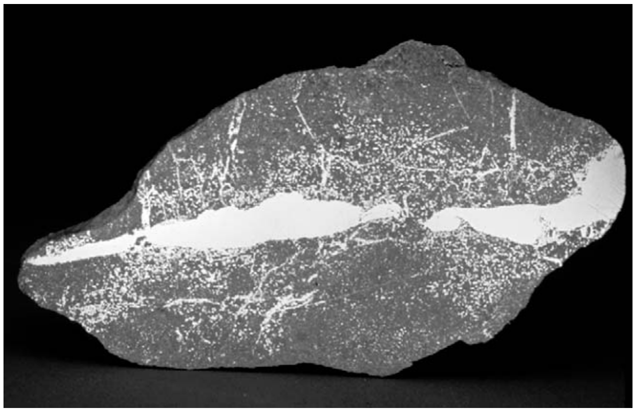

Fig. 2. Centimeter-scale veins of Fe,Ni metal and troilite in the acapulcoite Monument Draw formed during the first melting of an asteroid. In the absence of silicate melting, these veins were unable to migrate substantial distances and thus would not have contributed to core formation. Length of specimen is ~7.5 cm. (Smithsonian specimen USNM 7050; photo by C. Clark.)

These veins cross-cut silicates, suggesting that the silicates were solid at the time of Fe,Ni-FeS melt intrusion and remained solid throughout the history of the rocks. The veins migrated reasonably short distances (hundreds of micrometers to centimeters), leaving essentially unfractionated compositions. Similar veins and veinlets are observed in winonaites (*Benedix et al.,* 1998). Selenium/cobalt ratios (a proxy for sulfide/metal) in acapulcoites are essentially unfractionated relative to H-group ordinary chondrites, suggesting that the sulfide-rich early partial melt was not separated from the metal-rich residue (*Mittlefehldt et al.,* 1996) (Fig. 3).

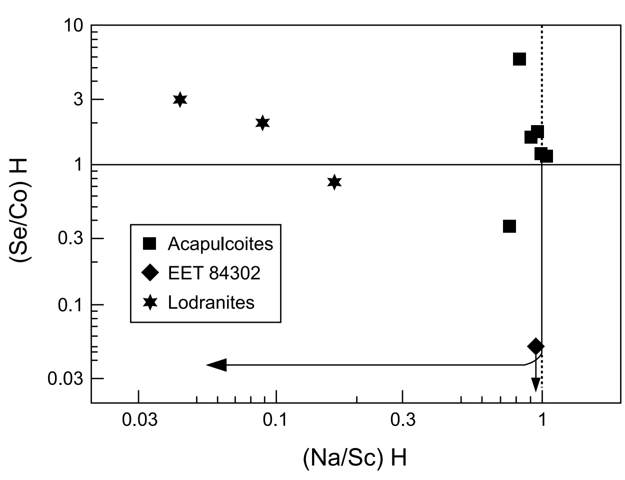

Fig. 3. H-chondrite normalized Se/Co vs. Na/Sc for acapulcoites, lodranites, and the intermediate EET 84302 after *Mittlefehldt et al.* (1996). The ratios serve as proxies for troilite/metal and plagioclase/pyroxene respectively. Partial melting should follow the arrowed curve. While lodranites have clearly lost the plagioclase-rich silicate partial melt, redistribution of the S-rich Fe,Ni-FeS cotectic melt was uneven, with melt probably being both removed from and introduced into the lodranites.

Silicate partial melting of an ordinary chondrite begins at between ~1323 and 1373 K (*Kushiro and Mysen,* 1979; *Takahashi,* 1983). This first melt is broadly basaltic in composition (dominated by pyroxene and plagioclase). In the Fo-An-SiO_2 system, the first melt is at the peritectic and contains ~55% plagioclase (*Morse,* 1980), although the phase relations in an iron-bearing system with albitic feldspar are incompletely understood. Despite these uncertainties, removal of a plagioclase-rich melt (basalt) from an ordinary chondrite protolith that contains ~10 wt% plagioclase (*McSween et al.,* 1991) would produce a residue substantially depleted in feldspar, provided that the melt separates efficiently. This must have occurred in the case of the lodranites, as evidenced by Fig. 3. Ratios of Na/Sc (a proxy for plagioclase/pyroxene) are substantially lower in lodranites relative to acapulcoites or ordinary chondrites, reflecting efficient removal of feldspar (in the form of basaltic melt). It is interesting to note that basalt-depleted lodranites exhibit chondritic to slightly enriched Se/Co ratios, suggesting that some melt of the Fe-Ni-S system might have even been introduced from another region of the parent body. Some intermediate members of the group, such as EET 84302, exhibit petrologic evidence for silicate melting, but have chondritic Na/Sc ratios and highly depleted Se/Co ratios (*Mittlefehldt et al.,* 1996). Thus, it appears that the onset of silicate partial melting first provided the conduit for efficient melt migration and set the stage for the formation of a basaltic crust and metallic core.

6. CRUST FORMATION

There are two linked requirements for the formation of a compositionally distinct crust on an asteroid: a sufficiently large degree of partial melting of the region that becomes the mantle, and efficient spatial separation of melt from the residual mantle. The relevant issues are (1) the nature of the mineral population, which determines which grain-grain contacts are the sites of first melting and also the typical separation between such sites; (2) the contact angles between melt and unmelted solid for each mineral species present, which determine the minimum degree of melting that must be attained before a widespread continuous fluid network exists; (3) the difference between the density of the fluid network and the bulk density of the unmelted matrix, which determines the buoyancy force (which could be positive or negative) acting on the melt; (4) the volume fraction of empty pore space between mineral grains; (5) the nature of, and initial pressure in, any gas species occupying the pore space; and (6) the uniformity of the ambient stress regime in the region of melt formation.

Numerical simulations imply that a few percent partial melting must occur before interconnected networks are formed (*Maaloe,* 2003). Initial basaltic melts have a somewhat smaller density and hence larger volume than the solids from which they form and so will invade available pore space, compressing whatever gas species may be present and causing a pressure increase. Buoyant melt will percolate upward along grain boundaries due to gravity alone, especially if the presence of a significant shear stress due to inhomogeneity in the asteroid structure encourages grain compaction. However, at low pore space volume fractions the large pressure rise in trapped gas may become a significant factor in fracturing the contacts between mineral grains and facilitating formation of buoyant liquid-filled veins that are longer than the typical mineral grain size of the matrix (*Muenow et al.,* 1992). As soon as veins become able to expand by fracturing their tips they will grow and may intersect other veins. A vein network will form, and larger veins will preferentially drain melt from smaller veins (*Sleep,* 1988) and grow at their expense (*Keil and Wilson,* 1993), eventually forming structures large enough to qualify as dikes.

The speed at which melt can move through a fracture, whether it be classified as a vein or a dike, depends on the inherent driving forces of buoyancy and excess pressure and on the shortest dimension of the fracture. Once a sufficiently wide dike forms, it will be able to propagate toward the surface and very efficiently drain much of the melt from the vein network feeding it. Thus two competing mechanisms operate in the mantles of asteroids (and larger bodies): a slow, steady percolation of melt along grain boundaries, and an intermittent but fast drainage of most of the available melt from a laterally and vertically extensive region into a dike that can penetrate a large fraction of the way from the melt source zone to the asteroid surface. There is a very large difference between these two mechanisms in regard to the time that a given batch of melt spends in contact with rocks at depths shallower than its source (*Kelemen et al.,* 1997; *Niu,* 1997), and so the relative importance of these two mechanisms is beginning to be understood from studies of the partitioning of rare Earth elements in, for example, the residual asteroid mantle rocks represented by the ureilites (*Goodrich and Lugmair,* 1995; *Warren and Kallemeyn,* 1992).

The timescale τ for the heating of asteroids due to the decay of ^{26}Al is on the order of 1 m.y. The distance λ that a thermal wave can travel in silicate rocks with thermal diffusivities κ typically on the order of 10^{-6} m^2 s^{-1} is $\lambda = (\kappa\tau)^{1/2} = 5.5$ km. Thus, the ability of heat to be conducted radially within an asteroid during the time for which the most plausible heat source is available is limited to a few kilometers, and so in asteroids with radii larger than a few tens of kilometers and ^{26}Al abundances sufficient to reach the solidus temperature, the melting process must occur simultaneously over almost the entire range of depths, as is found in detailed numerical models of asteroid evolution (*Grimm and McSween,* 1989; *Ghosh and McSween,* 1998). By the same token, there will be a region occupying the outermost ~5 km of all asteroids that is completely unmelted because it can always lose heat to the surface fast enough to compensate for heat flow from the interior (*Ghosh and McSween,* 1998).

The initial structure of this near-surface region at the end of asteroid accretion is critical. If it has a high porosity and low bulk density, melts rising from the mantle will not be able to penetrate it due to buoyancy alone unless they are very volatile-rich and contain a sufficiently large proportion of gas bubbles. Instead they will stall to form intrusions at a depth where they are neutrally buoyant. On large bodies like Earth, Mars, and Venus, tectonic control of major locations of mantle melting ensures that magma injection events are repeated frequently in essentially the same place. Previous magma batches then remain molten until subsequent ones arrive and intersect them, leading to development of shallow magma reservoirs and the growth of volcanic edifices such as shield volcanos on the surface above them. It is by no means clear that this pattern of activity will have dominated on any asteroids, and volcanism may have been characterized by the random distribution in space and time of monogenetic events, each of which produced either an eruption or the intrusion of a dike or sill, the outcome depending on the previous history of the eruption site and the volatile content of the magma. Bearing in mind the fact that initial crustal growth involves eruptions onto and intrusions into the primitive parent material of the asteroid, and that random impacts will be taking place on the surface of the growing crust, it is clear that predicting the pattern of crustal growth is a complex problem. There is probably still much to be learned by studying the proportions of, and detailed morphologies and thermal histories of, groups of meteorites from the outer layers of a common parent body, such as the eucrites, howardites, and diogenites (*Taylor et al.,* 1993; *Wilson and Keil,* 1996a).

A major issue related to the growth of asteroid crusts is the efficiency of volcanic advection of heat. There are two aspects of this. The first relates to the release of heat at the asteroid surface. Clearly even shallow intrusions were less efficient than surface eruptions at releasing heat. When surface eruptions occurred, explosive activity was probably more efficient than lava flow formation, but the situation is not clear cut. The negligible atmospheric pressure caused maximal expansion of released gas and hence high velocities of ejected magma droplets, and the relatively low gravity allowed wide dispersal of pyroclasts (*Wilson and Keil,* 1991). However, the high degree of gas expansion would also have fragmented the erupted magma into very small (submillimeter to tens of micrometers to judge by lunar samples) droplets and there is then a wide range of eruption conditions (*Wilson and Keil,* 1996b) that could have produced optically dense fire fountains in which all droplets except those in the outer envelope of the fountain retained their heat and formed uncooled ponds feeding lava flows.

7. CORE FORMATION

Formation of a metal-sulfide core would occur essentially contemporaneously with the early stages of crust formation. As discussed earlier in this chapter, cotectic Fe,Ni-FeS melts are unlikely to have dihedral angles exceeding 60° and migrate under static conditions. However, dynamic conditions may produce very different results. *Keil and Wil-*

son (1993) explored the loss of the Fe,Ni-FeS melt by explosive volcanism as a mechanism for producing the S depletions (relative to S within the Fe,Ni-FeS component of ordinary chondrites) inferred for several groups of iron meteorites (most notably the IIIAB, IVA, and IVB groups). These authors argued that even the dense cotectic Fe,Ni-FeS melts, which would normally be negatively buoyant, can be driven to the surface and lost by expansion of trapped gases on small asteroidal bodies (radii < 100 km). While this mechanism may set the stage for S loss on small asteroidal bodies, it has equal implications for larger bodies. On those asteroids with radii in excess of 100 km, such melts would likely migrate to the center of the body and form a small, S-rich core. If partial melting were essentially arrested at this step (such as appears to have happened on the parent bodies of many primitive achondrites), this small S-rich core might crystallize, overlain by a shell of unfractionated or weakly fractionated silicates. Several authors (e.g., *Kracher,* 1982; *Benedix et al.,* 2000) have argued that this is the mechanism by which IAB and IIICD irons formed, although they recognize that impact must have played a role in mixing silicate inclusions into these meteorites. An alternative hypothesis, championed most recently by *Wasson and Kallemeyn* (2002), argues that these groups are formed in localized impact-melt pools near the surface of their parent body.

In many asteroids, much more extensive melting must have occurred, leading to formation of a core that incorporated essentially the entire metal-sulfide component present in the asteroid (recognizing that small droplets could have remained entrained in an overlying, convecting silicate shell). *Taylor* (1992) examined the physics of core formation, both at low and high degrees of partial melting. While the conclusions of this work at low degrees of partial melting have been challenged by subsequent experimental work, the calculations for separation of metallic melt blobs at high degrees of partial melting appear sound. *Taylor* (1992) argued that metal droplets might coarsen from the millimeter-size range seen in chondrites to 10 cm by a variety of largely unknown processes (e.g., Ostwald ripening) during early heating and melting of an asteroid. For these droplets to overcome the yield strength of the melt depends on the

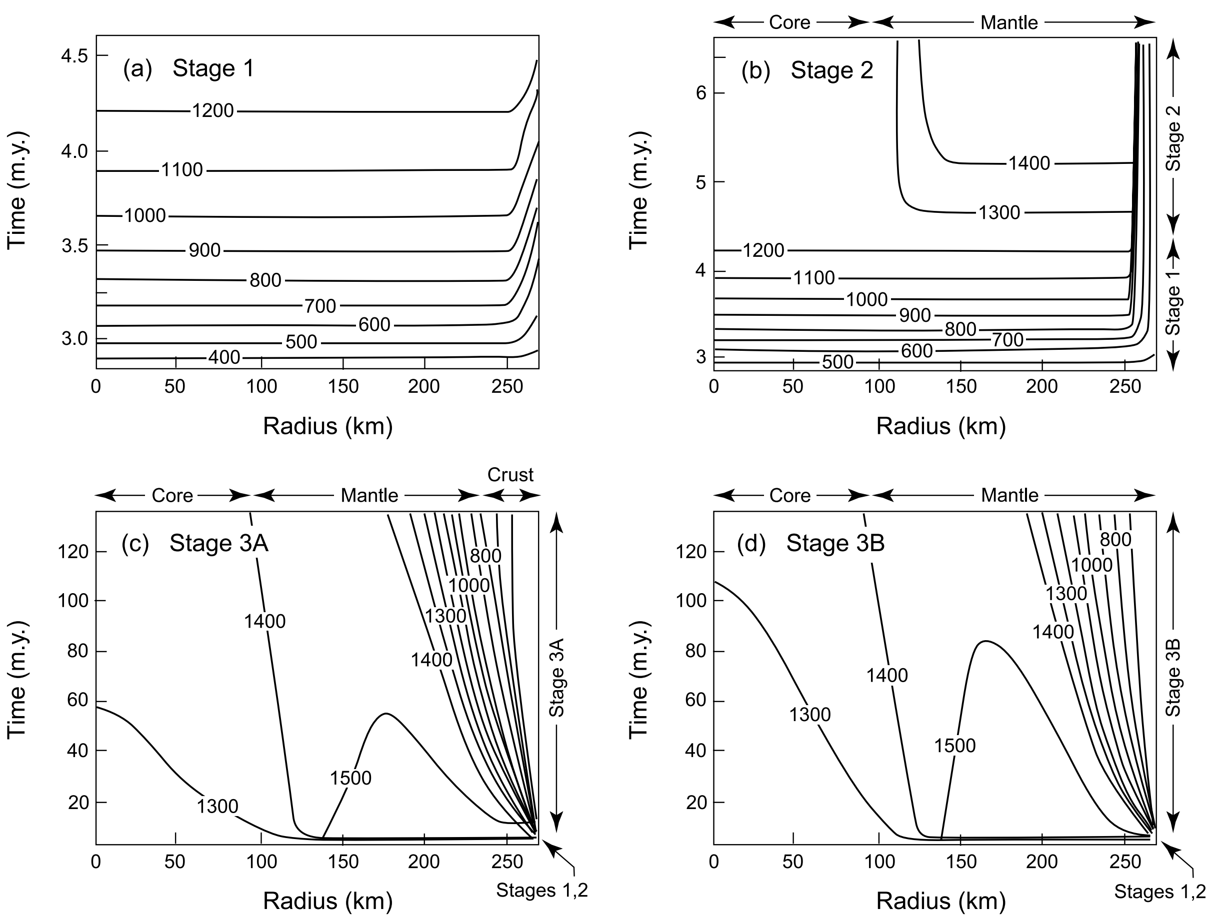

Fig. 4. Temperature contours for 4 Vesta on plots of time elapsed since CAI formation and radial distance from asteroid center after *Ghosh and McSween* (1998). **(a)** Stage 1 is the interval from accretion to core formation. **(b)** In stage 2, core formation redistributes ^{26}Al into the mantle, where temperatures continue to increase. Stages 3A **(c)** and 3B **(d)** compare heating in the cases of complete melt extraction from the mantle and retention of the melt in the mantle respectively. Peak temperatures reach 1500 K in the mantle, corresponding to 25% melting of their model mantle.

gravity and, hence, on the radius of the parent body. The percentage of melting required exceeded 50% in each case, reaching as much as 90% melting for efficient core formation on a 10-km-radius asteroid. Uncertainties in the yield strength propagate to uncertainties of ±10% in the degree of melting. Given that iron meteorite parent bodies were in the range of 10–100 km in radius (*Haack and McCoy,* 2004), high degrees of partial melting must have been reasonably widespread, given the abundance of iron meteorite groups and ungrouped meteorites (see *Weisberg et al.,* 2006).

8. COMPLETE MELTING?

While the presence of iron meteorites implies high degrees of partial melting (≥50%), did differentiated asteroids melt completely? We can examine this issue from a number of theoretical standpoints, as well as by looking at the meteorite record.

Several authors, most notably *Righter and Drake* (1997), have argued that the parent body of the howardite-eucrite-diogenite (HED) meteorites — likely 4 Vesta — had a magma ocean formed by extensive partial melting. They modeled partitioning and concentrations for several siderophile elements for a variety of likely chondritic precursors (H, L, LL, CI, CM, CO, CV) and found their best matches with peak temperatures of 1773–1803 K and degrees of silicate melting between 65% and 77%.

Ghosh and McSween (1998) did modeling of the thermal history of an asteroid of chondritic composition the size of Vesta heated mostly by ^{26}Al (Fig. 4). The modeling was simplified in that instantaneous cold accretion was assumed, and core formation was assumed to occur instantaneously when the "Fe-FeS liquidus temperature" was reached. (The temperature used in the model was 1223 K, too low to explain the low-Ni, high-Ir irons of classic magmatic iron meteorite groups such as IIAB and IIIAB.) Because ^{60}Fe produced only a small fraction of the heat, the core acted as a heat sink in the early history of the asteroid (*Ghosh and McSween,* 1998). More recent estimates of the nebular abundance of ^{60}Fe are 30–40× higher than the value used in the thermal models (*McKeegan and Davis,* 2003). In their models, peak temperatures within the mantle reach ~1500 K, although this peak temperature appears to be constrained by the model parameter of reaching only 25% melting in the mantle.

Somewhat higher temperatures and degrees of partial melting were inferred by *Taylor et al.* (1993) in examining the liquidus temperatures for the Fe,Ni-FeS component of LL (1630 K) and H (1700 K) ordinary chondrites and the S-depleted IVB irons (1770 K). Comparing these temperatures with melting relationships for peridotites (Fig. 5), these authors inferred 35% and 60% melting.

In each case, the inference is that high degrees of partial melting were achieved for many asteroidal parent bodies, but that a substantial fraction of unmelted crystals (25–50%) might remain. *Taylor* (1992) argued that these high degrees of melting were consistent with the olivine-only silicate residues at partial melting in excess of 45% and the absence

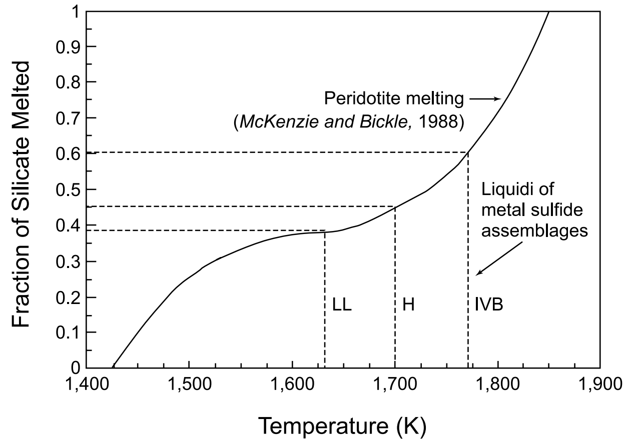

Fig. 5. Fraction of silicate melted vs. temperature for peridotite melting (*McKenzie and Bickel,* 1988). Liquidi are shown for H and LL chondrites and the S-poor IVB irons. Degrees of partial melting between 39% and 60% are indicated, consistent with the presence of magma oceans on the parent asteroids of most iron meteorites and a residual mantle with 40–60% residual crystals. The residual crystals would be dominantly olivine, although a small pyroxene component might be present at the lower ranges of degrees of partial melting, consistent with the discovery of pyroxene pallasites. Figure from *Taylor et al.* (1993).

of significant pyroxene in pallasites. Since the publication of that work, the recognition of pyroxene pallasites might suggest that the degree of partial melting achieved on some asteroidal parent bodies that formed cores was lower than that inferred by *Taylor* (1992).

The enigma that remains is that of whether differentiated asteroids achieved a truly homogeneous magma ocean prior to solidification or whether an early-formed crust overlay a peridotite mantle. As *Ghosh and McSween* (1998) note, the realistic case is probably intermediate between early and complete extrusion of basaltic melts from the mantle and their complete retention in the mantle. In addition, these authors point out that the outer 5 km of an asteroid dissipates heat as quickly as it enters from below, and thus might remain largely unmelted. If an early basaltic crust forms or a relict chondritic layer remains, the two most likely scenarios for incorporation of this material back into the mantle are recycling by impacts and remelting of this crust, primarily at the base.

It is difficult to model the former. *Ghosh and McSween* (1998) cited earlier workers that suggested either minimal heating by impact or localized heating. However, we envision remelting not due to heating as a consequence of impact, but by recycling cold outer crust into the deeper, hotter parts of the mantle.

It is easier to estimate the ability to melt the crust from below and this appears to be particularly dependent on parent-body radius. For example, on a 20-km-radius parent body, the volume of melt in a 10-km-thick mantle at 50% melting would equal only 35% of the volume of the outer 5-km shell. Thus, the ability of this relatively small amount of melt to recycle the entire outer shell would be very limited. In contrast, a 250-km-radius asteroid (e.g., 4 Vesta) with a 50-km-diameter core and 5-km outer shell at 50% melting of the mantle would contain nearly 32× as much melt in the mantle as the volume of cold, outer crust. Thus, it seems quite probable that large, differentiated asteroids may have in fact reincorporated any early-formed crust and achieved a true magma ocean, while smaller asteroids retained the early-formed basaltic crust and perhaps even highly metamorphosed traces of their chondritic precursor.

9. SOLIDIFICATION

In this section, we address the solidification of an initially chemically uniform, ultramafic magma ocean. As a starting point in this discussion, we will assume that the rate of heating is much more rapid than the rate of heat dissipation. Thus, for example, if ^{26}Al is the heat source, Al-rich melt does not migrate out of the interior rapidly enough to shut down heating of the deep interior. Under this assumption, immiscible metal-sulfide melts and refractory metallic solids will segregate downward to form a core, which will be overlain by a molten silicate shell of ultramafic composition.

At present, little detailed modeling has been done to try to understand the physical behavior of such a silicate melt shell, and thus chemical modeling of solidification is somewhat ungrounded. If the heat source is a short-lived lithophile nuclide such as ^{26}Al, then the silicate melt shell will be approximately evenly heated internally, with thermal boundaries at the top and bottom (see *Ghosh and McSween,* 1998). This could act to subdue convection, although calculations using the canonical solar nebula value for $^{26}Al/^{27}Al$ indicate that molten asteroids would be within the convection realm (*Righter and Drake,* 1997; *Taylor et al.,* 1993). On the other hand, if ^{60}Fe is a major heat source, the silicate melt shell would be heated from the base and would contain an evenly distributed heat source, and strong convection would certainly occur. The style of solidification — equilibrium vs. fractional crystallization — could be very different in these two cases.

A compositionally uniform, heated silicate shell maintains a high temperature for an extended period of time. For example, at a radius of 175 km (depth of 100 km) the temperature is above 1400 K for 140 m.y. [For comparison, the silicate solidus for ordinary chondrites is probably <1375 K; *Jurewicz et al.* (1995) found ~13% melt in ordinary chondrite experimental charges run at 1445 K.] The "hot zone" in the asteroid thermal model is located in the mid regions of the silicate shell, roughly at a radius of 200 km at the start, lowering to radius of roughly 150 km as the body cools (*Ghosh and McSween,* 1998). (The core-mantle boundary is at radius 100 km.) If this thermal model is correct, solidification would be going on simultaneously at the base and top of the molten silicate shell.

Tonks and Melosh (1990) discussed the nature of convection in planetary-scale magma oceans. In a terrestrial or lunar magma ocean, the Rayleigh number would be very high, and convection would be in the realm of inertial flow, where turbulent eddies overcome settling forces to keep growing crystals in suspension, although this conclusion has been challenged by subsequent workers (see *Solomatov,* 2000). *Taylor et al.* (1993) and *Righter and Drake* (1997) have estimated the Rayleigh numbers for a molten silicate shell over a metallic core on 100-km-radius and Vesta-sized asteroids. Both studies found that an asteroidal magma ocean would convect in the inertial realm, keeping crystals suspended in a well-mixed melt. Under these conditions, equilibrium crystallization would occur for a substantial fraction of the crystallization sequence (*Righter and Drake,* 1997).

Taylor et al. (1993) calculated that crystals would not grow large enough for settling to occur. However, they based this conclusion on calculations that the magma ocean would cool and solidify on a very short timescale for the 100-km-radius asteroid they considered. Their model did not include continued heating of the asteroid through ^{26}Al decay even while heat is being lost to space. The thermal models of *Ghosh and McSween* (1998) for a Vesta-sized asteroid have a prolonged cooling history, and crystals would reach much larger sizes, possibly allowing settling to occur. The settling velocity of a crystal is proportional to the square of the crystal radius (*Walker et al.,* 1978), so a growing crystal rapidly increases its settling velocity.

All models of the physical setting of asteroid solidification assume uniform low gravity, but this is a simplification. Figure 6a shows the gravity as a function of fractional radius for a Vesta-sized asteroid, assuming core masses of 5% and 15% the mass of the asteroid (cf. *Righter and Drake,* 1997). The core and molten silicate shell are each assumed to have uniform density, with the silicate melt density that of *Righter and Drake* (1997). Gravity reaches a local minimum in the lower portion of the magma shell before increasing to the maximum at the surface. Figure 6b shows calculated settling velocity following *Walker et al.* (1978) for 1-cm-radius crystals in the silicate magma shell. Because of the linear dependence on gravity, the settling velocity similarly reaches minimum in the lower portion of the magma shell. The mean convection velocity scales as the cube-root of gravity (*Tonks and Melosh,* 1990, their equation (17)), so the grain-settling velocity (linear function of gravity) is a stronger function of depth within the magma shell than is convection velocity. There is thus a potential that grains growing in the upper region of the magma sphere could settle, only to become stalled in convecting magma at an intermediate depth.

As outlined above, we have only a rudimentary understanding of the physics of a cooling asteroid magma shell. For this reason, models for the petrologic evolution of the magma are crude approximations. Only two end members have been modeled in detail in the literature: equilibrium crystallization (*Righter and Drake,* 1997) and fractional crystallization (e.g., *Ruzicka et al.,* 1997). Both of these have been done to model the evolution of the howardite-

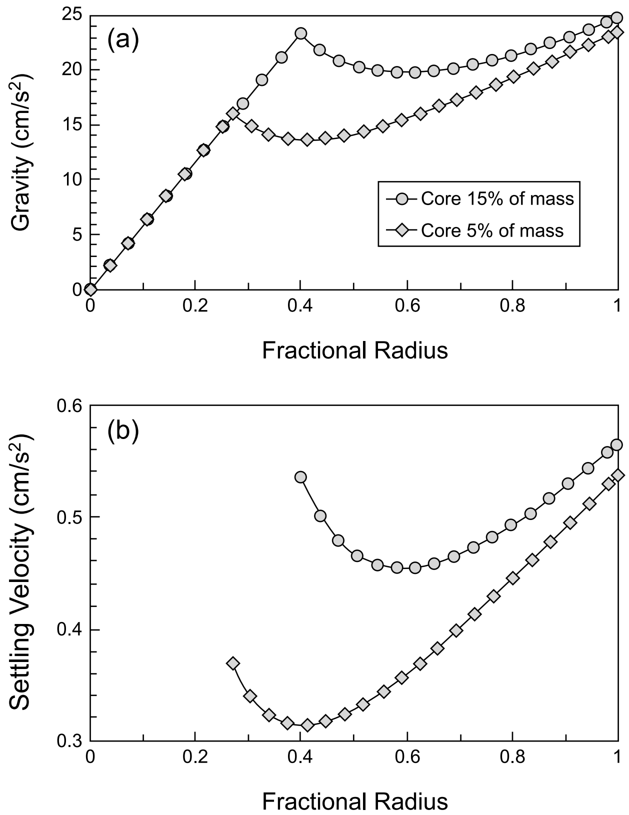

Fig. 6. Calculated **(a)** gravity and **(b)** grain-settling velocity vs. fractional radius for a Vesta-sized molten asteroid. Core density calculated for a mixture of 90% Fe,Ni and 10% FeS; silicate melt density is 2900 kg m^{-3}, following *Righter and Drake* (1997). Settling velocity calculated following *Walker et al.* (1978) assuming 1-cm grains with a density of 3200 kg m^{-3}.

eucrite-diogenite parent body, presumed to be 4 Vesta. The compositions investigated are roughly chondritic, but that modeled by *Ruzicka et al.* (1997) is depleted in Na.

9.1. Equilibrium Crystallization

A magma with roughly chondritic relative proportions of lithophile elements will be highly olivine normative (Fig. 7). The maximum temperature reached in the Vesta thermal models is ~1540 K (*Ghosh and McSween,* 1998). *Jurewicz et al.* (1995) found that LL-chondrite charges run at 1600 K contained ~55% residual olivine, while H-chondrite charges run at 1550 K contained ~70% residual olivine. Thus an asteroid magma shell might consist of turbulently convecting melt entraining refractory olivine grains in equilibrium with the melt.

Righter and Drake (1997) modeled the equilibrium crystallization of a Vestan magma shell assuming a composition equivalent to 30% CV–70% L chondrite (chosen to approximate the HED O-isotopic composition) and an initial temperature of 1875 K. They calculated the Rayleigh number for their model magma, and concluded that convec-

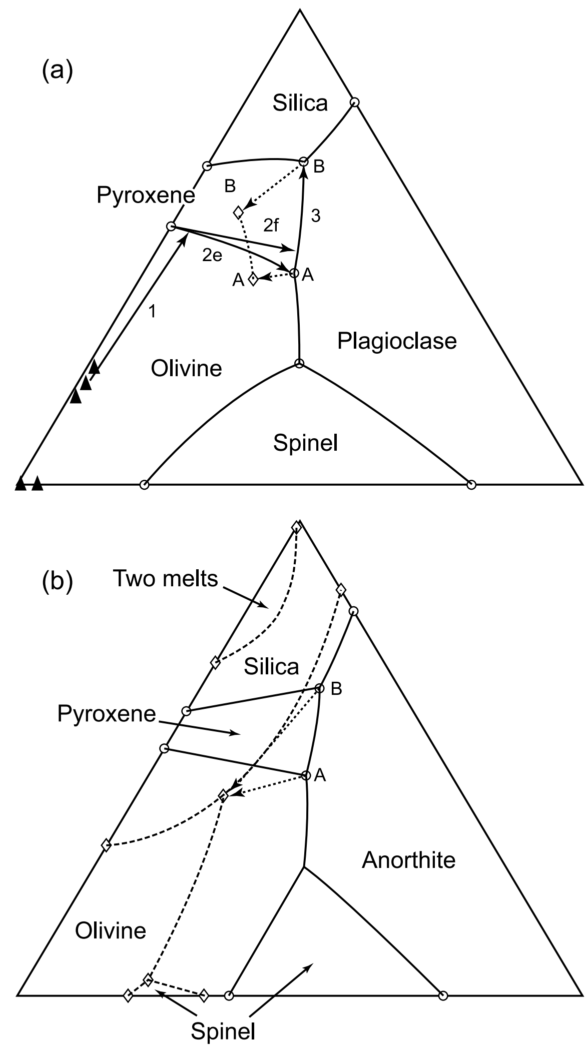

Fig. 7. **(a)** Schematic olivine-plagioclase-silica pseudoternary diagrams showing idealized paths for crystallization of molten chondritic asteroids. Phase boundaries are taken from *Walker et al.* (1973), devised for modeling lunar compositions; mg# of 70 and Na-depleted relative to chondrites. Dashed arrows show the effect of decreasing mg# on the peritectic (A) and eutectic (B) points based on experiments on eucritic compositions (connected by dotted line), with mg# ~40 (taken from *Stolper,* 1977). A schematic equilibrium crystallization path is given by arrows 1 and 2e; crystallization will terminate at the peritectic point (A). A schematic fractional crystallization path is given by arrows 1, 2f, and 3; crystallization will terminate at the eutectic (B). Phase boundaries shift with changes in composition, pressure, and f_{O_2}, and no single diagram can truly represent full crystallization path of a magma. Solid triangles are the locations of the silicate portions of ordinary chondrites (upper three) and CM and CV chondrites, using average fall data from *Jarosewich* (1990). **(b)** The shifting phase boundaries with Fe content are illustrated by comparing the Fo-An-Si (solid lines) and Fa-An-Si (dashed lines) phase diagrams. With increasing Fe content, the anorthite and silica stability fields expand, the peritectic (A) and eutectic (B) points merge, and the stability field of pyroxene disappears. Substituting albite for anorthite (not shown) in the Fo-Ab-Si system greatly decreases the stability field of plagioclase, and points A and B shift toward the Si-Ab join.

tion would keep the crystal and melt well-stirred up to about 80% crystallization. Convection lockup would occur at this point, and the remaining 20% solidification would follow a fractional crystallization path. An idealized equilibrium crystallization path is shown in Fig. 7a. Olivine is the sole crystallizing phase for about the first 35% solidification (Fig. 7a, arrow 1). At that point, orthopyroxene and minor Cr-rich spinel join the crystallization sequence at nearly the same time (*Righter and Drake,* 1997) (Fig. 6). These three phases co-crystallize until about 1515 K (Fig. 7a, arrow 2e). [The olivine-orthopyroxene boundary is normally a reaction boundary at low pressure. In Fe-rich mafic melts, it becomes a cotectic boundary. The modeling programs used by *Righter and Drake* (1997) treated the boundary as a cotectic.] Pigeonite replaces orthopyroxene below 1515 K, when the silicate system is ~85% crystallized. Plagioclase joins the crystallization sequence at the end, and some olivine will react with the residual, basaltic melt to form pigeonite. Crystallization ends at the pseudoternary peritectic point, A.

If equilibrium is maintained throughout the crystallization sequence, the asteroid would end up as a metal-sulfide core surrounded by a homogeneous shell of coarse-grained olivine, orthopyroxene, pigeonite, plagioclase, and chromite. This seems unlikely. At some point the convective motion in the magma will cease to entrain the growing crystals, and as discussed by *Righter and Drake* (1997) the residual melt will begin to follow a fractional crystallization path. Should this occur before the melt composition reaches the peritectic point in the pseudoternary system, orthopyroxene will fractionate from the melt, followed by fractionation of pigeonite plus plagioclase. In this case, the asteroid would consist of a metal-sulfide core, a thick shell of harzburgite, overlain by a thin shell of orthopyroxenite, capped by a gabbroic/basaltic crust. *Righter and Drake* (1997) estimate a thickness for the mafic crust on a Vesta-sized asteroid of 10–15 km. The thickness of the orthopyroxenite shell would depend on the point at which the transition from equilibrium to fractional crystallization occurred. If the transition to fractional crystallization occurs only after the peritectic point is reached, than there will be no orthopyroxenite layer. The asteroid mantle would consist of harzburgite.

9.2. Fractional Crystallization

Ruzicka et al. (1997) modeled fractional crystallization of molten asteroids of several chondritic compositions; all Na-poor because the intent was to model the eucrite-diogenite suite. Olivine was the sole crystallizing phase for ~60 vol% crystallization (Fig. 7a, arrow 1). [*Ruzicka et al.* (1997) tabulated model results by volume, rather than by mass.] The olivine-orthopyroxene boundary was treated as a reaction boundary. When the melt composition reached this boundary, orthopyroxene became the sole crystallizing phase, followed by orthopyroxene + chromite, followed by pigeonite (Fig. 7a, arrow 2f). The melt composition reached the pyroxene-plagioclase cotectic, and formation of gabbroic cumulates (plagioclase + pigeonite) began (Fig. 7a, arrow 3). *Ruzicka et al.* (1997) noted that melt viscosity would be substantially higher by this stage, and suggested the further crystallization would approach equilibrium crystallization. Finally, the plagioclase-pyroxene-silica ternary eutectic was reached (point B) and silica plus ferroan olivine joined the crystallization sequence. The resulting asteroid is slightly different from the model produced by equilibrium crystallization: a metal-sulfide core, overlain by a 150-km-thick dunite shell, overlain by roughly 13 km of orthopyroxenite, capped by a 26-km-thick mafic crust (*Ruzicka et al.,* 1997).

The discussions of equilibrium and fractional crystallization paths given above in relation to phase relations (Fig. 7a) are highly schematic because phase boundaries shift as the system evolves. This is qualitatively illustrated in Fig. 7b where the phase boundaries in the forsterite-anorthite-silica (Fo-An-Si) and fayalite-anorthite-silica (Fa-An-Si) are compared. Adding Fe to the system almost completely changes the topology of the diagram. In real silicate systems, the mg# [molar 100*MgO/(MgO + FeO)] decreases as crystallization proceeds, and the position of the pyroxene-plagioclase cotectic shifts toward the evolving melt composition, foreshortening the pyroxene crystallization interval (Fig. 7a, arrow 2f). The olivine-pyroxene boundary also changes from a reaction boundary to a cotectic as mg# decreases, which will alter the crystallization sequence. Finally, the peritectic point crosses the plagioclase-pyroxene join as pressure is increased, and the olivine-pyroxene boundary will become a cotectic by ~100 MPa (*Bartels and Grove,* 1991). This corresponds to a depth of ~190 km in a Vesta-sized asteroid. These issues have been discussed by *Bartels and Grove* (1991), *Boesenberg and Delaney* (1997), and *Warren* (1985), for example.

Figure 7 can be thought of as an asteroidal equivalent to the tholeiitic basalt sequence on Earth — crystallization results in low-Ca pyroxene and silica-normative residua. *Jurewicz et al.* (1993) and *Longhi* (1999) have shown that increasing the f_{O_2} of chondritic systems can change the phase relations, resulting in minimum melts that are critically silica unsaturated, and magmas with angritic affinities are produced. The solidification of such an asteroid would be quite different from that outlined above for HED analog asteroids, but no detailed modeling has been done. Olivine would still be the liquidus phase, and form the bulk of the solidified asteroid. Plagioclase ± spinel would next join olivine in the crystallization sequence, followed by augite. Finally, kirschsteinite ($CaFeSiO_4$) joins the crystallization sequence at the very end.

The modeling by *Ruzicka et al.* (1997) considered Na-poor starting compositions because the object was to model the HED parent body. Although *Righter and Drake* (1997) were also attempting to explain HED petrogenesis, they used true chondritic Na contents. The phase relations will be different at higher, more chondrite-like, Na contents (see *Stolper et al.,* 1979). The stability field of plagioclase greatly decreases, while those of olivine and pyroxene increase as the Na content of the system increases. Pyroxene also becomes more calcic as Na replaces Ca in plagioclase (*Stolper et al.,* 1979). Some of the difference in model results between *Ruzicka et al.* (1997) and *Righter and Drake* (1997) is probably a reflection of the different assumed magma compositions. There have been no attempts to model the solidification of true chondritic composition asteroids (e.g., CV, H, EH, etc.) using a common methodology. Because of this, detailed predictions of the solidification and structure of chondritic composition asteroids are not known.

10. SYNOPSIS

Despite the broad array of chondritic precursors known to exist, the wide range of geologic processes acting during melting of these bodies (e.g., heating, differentiation, impact), and the complex interplay between these parameters, we can define — in the broadest sense — the melting of a typical asteroid. As heating exceeded the solidus, melting would begin first at the Fe,Ni-FeS cotectic and be followed by melting of silicates. These low-degree partial melts would likely not form networks, but would migrate primarily through the formation of veins and dikes driven in part by expansion of the melts and their included gases. Formation of a basaltic crust would occur soon after the onset of melting, perhaps in as little as 1 m.y. At the same time, the Fe,Ni-FeS melts would begin to migrate and, depending on parent-body size, might be erupted either from the asteroid or the sink to form a small S-rich core. The melting of the parent bodies of some meteorites (e.g., ureilites, acapulcoites-lodranites, perhaps IAB irons) were arrested at this stage. These parent bodies produced achondrites variably depleted in basaltic components and melts of the Fe-Ni-S system (ureilites, lodranites). Most differentiated asteroids continued to higher degrees of melting, eventually reaching percentages of partial melting where efficient metal segregation occurred by sinking of metallic melt droplets and metal-dominated cores were formed. The mantles of these asteroids likely did not melt completely, with percentages of melting ranging from ~40% to 75%. In the largest asteroids, impacts and basal melting may have recycled the early-formed basaltic crust (as well as any remaining chondritic material at the surface) back into the mantle, producing a homogeneous magma ocean. In contrast, smaller asteroids might have retained this primary basaltic crust. The crystallization of that magma ocean, whether by equilibrium or fractional crystallization, would produce a core mantled with a thick layer of olivine. Depending on the style of crystallization and the nature of the olivine-pyroxene phase boundary, the dunite layer will be overlain by harzburgite, orthopyroxenite, or some combination of the two. The ultramafic shell will be capped by a 10–25-km-thick basaltic to gabbroic crust. This gross structure should be typical of large differentiated asteroids like 4 Vesta.

11. FUTURE DIRECTIONS

Our understanding of asteroid differentiation is likely to continue unabated for the foreseeable future. Recovery of new meteorites and attendant geochemical and experimental studies of the processes that formed them proceed and shed new light on these processes. In many respects, the most exciting new avenue for research into the next decade will be the first spacecraft exploration of a differentiated asteroid. The DAWN mission is scheduled to orbit 4 Vesta beginning in 2010, collecting a variety of spectral, geochemical, shape, and magnetic data. Existing groundbased rotational spectroscopy (*Binzel et al.,* 1997; *Gaffey,* 1997) suggests geologic units consistent with dunite, basalt, and orthopyroxenite interpreted as the mantle, eucrite, and diogenite components of the HED suite. Given the importance that the study of this suite has in shaping our views of asteroidal differentiation (perhaps even to the exclusion of considering alternative scenarios for smaller and/or less-melted asteroids), these new data could prove revolutionary in shaping our understanding of the process by which primitive, chondritic asteroids ultimately become the layered worlds we know from our meteorite collections.

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