Iron and Stony-iron Meteorites

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1.12 INTRODUCTION

Without iron and stony-iron meteorites, our chances of ever sampling the deep interior of a differentiated planetary object would be next to nil. Although we live on a planet with a very substantial core, we will never be able to sample it. Fortunately, asteroid collisions provide us with a rich sampling of the deep interiors of differentiated asteroids.

Iron and stony-iron meteorites are fragments of a large number of asteroids that underwent significant geological processing in the early solar system. Parent bodies of iron and some stony-iron meteorites completed a geological evolution similar to that continuing on Earth—although on much smaller length- and time-scales—with melting of the metal and silicates, differentiation into core, mantle, and crust, and probably extensive volcanism. Iron and stony-iron meteorites are our only available analogues to materials found in the deep interiors of Earth and other terrestrial planets. This fact has been recognized since the work of Chladni (1794), who argued that stony-iron meteorites must have originated in outer space and fallen during
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fireballs and that they provide our closest analogue to the material that comprises our own planet’s core. This chapter deals with our current knowledge of iron meteorites. How did they form? What can they tell us about the early evolution of the solar system and its solid bodies? How closely do they resemble the material of our own planetary interior? What do we know and don’t we know?

Iron and stony-iron meteorites constitute ~6% of meteorite falls (Grady, 2000). Despite their scarcity among falls, iron meteorites are our only samples of the ~135 asteroids from which meteorites originate (Keil et al., 1994; Scott, 1979; Melhorn and Clark, 1999; see also Chapter 1.05), suggesting that both differentiated asteroids and the geologic processes that produced them were common.

Despite the highly evolved nature of iron and stony-iron meteorites, their chemistry provides important constraints on the processes operating in the solar nebula. Although most of them probably formed through similar mechanisms, their characteristics are diverse in terms of chemistry, mineralogy, and structure. Significant differences in bulk chemistry between iron meteorites from different cores as well as variations in chemistry between meteorites from the same core provide evidence of the complex chemical evolution of these evolved meteorites. Intergroup variations for volatile siderophile elements (e.g., tellurium and germanium) extend more than three orders of magnitude, hinting that iron meteorite parent bodies formed under diverse physical conditions. These differences reflect both the nebular source material and geological processing in the parent bodies.

Can we be sure that the iron meteorites are indeed fragments of cores? Since no differentiated asteroid has yet been visited by a spacecraft, we rely on circumstantial evidence. Some M-type asteroids have spectral characteristics expected from exposed metallic cores (Tholen, 1989), while others exhibit basaltic surfaces, a hallmark of low thermal conductivity of the Earth’s mantle. Highly reduced asteroidal materials may have been exposed to the surface, being divided into hexahedrites, octahedrites (of varying types based on kamacite bandwidth), and ataxites (see Buchwald, 1979, and references therein). Beginning in the 1950s, it was recognized that the chemical composition of iron meteorites varied, often in concert with their physical structure. Today, the chemical classification—based on bulk chemical analysis of the metal—is the standard for iron meteorite groups (Chapter 1.05), with structural and other data (e.g., cosmic-ray exposure ages) providing supporting information about iron meteorite history.

For some asteroids, near-catastrophic impacts could have released triggers to core formation, small asteroid s being divided into hexahedrites, octahedrites to coarsest ataxites. Samples from the small asteroid cores provide us with a rare opportunity to observe core formation—frozen in place.

Iron–nickel alloys are expected in the cores of differentiated asteroids, but what other evidence supports the notion that iron meteorites sample the cores of differentiated asteroids? What suggests that these asteroids were sufficiently heated to trigger core formation, and that iron meteorites sample cores rather than isolated pools of once molten metal? First and foremost, trace-element compositional trends in most groups of iron meteorites are consistent with fractional crystallization of a metallic melt (Scott, 1976; Taylor et al., 1993). In addition, meteorites in different temperature regimes required to form a metallic melt are sufficiently high to cause substantial melting of the associated silicate phases (Taylor et al., 1993). This process is consistent with the associated silicates and the observation of melt inclusions in iron meteorites (Daly, 1963). Core formation from the iron meteorite parent bodies was suggested by the presence of metal pools within meteorites (Wood, 1964), providing evidence of the complex chemical evolution of these evolved meteorites. Intergroup variations for volatile siderophile elements (e.g., tellurium and germanium) extend more than three orders of magnitude, hinting that iron meteorite parent bodies formed under diverse physical conditions. These differences reflect both the nebular source material and geological processing in the parent bodies.

Despite significant overlap in the compositional characteristics of iron meteorites, some of which have fallen on the Earth in the form of meteorites. These fragments provide us with an opportunity to study the evolution of the surviving material. Later impacts on the meteorite parent bodies dispersed fragments in the system, some of which have fallen on the Earth in the form of meteorites. These fragments provide us with an opportunity to study the evolution of the surviving material. Later impacts on the meteorite parent bodies dispersed fragments in the system, some of which have fallen on the Earth in the form of meteorites. These fragments provide us with an opportunity to study the evolution of the surviving material. Later impacts on the meteorite parent bodies dispersed fragments in the system, some of which have fallen on the Earth in the form of meteorites.

1.1.2 CLASSIFICATION AND CHEMICAL COMPOSITION OF IRON METEORITES

To use meteorites as a guide to parent-body evolution, we need groupings that represent individual parent bodies. In this section, we briefly discuss the characteristics of iron and stony-iron meteorites used to discriminate between the different parent bodies. Historically, iron meteorites were classified on the basis of macroscopic features, being divided into hexahedrites, octahedrites, and ataxites (see Buchwald, 1979, and references therein). Beginning in the 1950s, it was recognized that the chemical composition of iron meteorites varied, often in concert with their physical structure. Today, the chemical classification—based on bulk chemical analysis of the metal—is the standard for iron meteorite groups (Chapter 1.05), with structural and other data providing supporting information. The chemical classification of iron meteorites provides us with an unparalleled opportunity to study the evolution of the surviving material. Later impacts on the meteorite parent bodies dispersed fragments in the system, some of which have fallen on the Earth in the form of meteorites.

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Group 1.12.2 Group IAB Iron Meteorites

Group IAB iron meteorites (103 members) have the lowest nickel concentrations of any group of iron meteorites (3.5–6.6 wt% Ni) (see figure 26, Chapter 1.05). A detailed description of the individual iron meteorite groups is found in Scott and Wasson (1975). We include a brief synopsis of some of the important characteristics of the largest groups, particularly those with implications for the evolution of the parent bodies. Differences between these groups serve to illustrate some of the diverse characteristics of iron meteorites. The range of iron meteorite compositions is broad, spanning the volatile and silicate-rich IAB–HICDs to the highly volatile depleted groups IVA and, in particular, HIBs. Differences in the oxygen fugacities of the parent cores are reflected in the difference in mineralogy between the highly reduced IABs and the more oxidized group IIB (two meteorites).

Although the original classification was entirely based on chemistry, additional data on structure, cooling rates, and mineralogy suggest that it is a genetic classification where each group represents one parent core (Scott and Wasson, 1975). The chemical compositions within each group form a distinct structural and mineralogical association, and the chemical crystallization and the members have similar or uniformly varying structure and composition. Despite significant overlap in the compositional clusters, multiple elements, along with structure and mineralogy, distinguish the groups. Most groups have uniform cooling rates, supporting the idea that each group represents an individual asteroid. Cosmic-ray exposure ages further support a genetic classification in groups IAB and IVA, where most members have cosmic-ray exposure ages of 650 Ma and 400 Ma, respectively.

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Iron and Stony-Iron Meteorites

(Buchwald, 1975; Stöffler et al., 1988), and have uniform exposure ages and cooling rates. They have the by far highest abundance of phosphates, suggesting formation from relatively oxidized material (Kerr et al., 1977; Olsen and FREDRIKSSON, 1966; Olsen et al., 1999). Silicate inclusions are documented in only two members (Kracher et al., 1993; Kracher et al., 1996). Olsen et al., 1996). Observed compositional trends can be broadly matched by numerical models of crystallization of a common core (Haack and Scott, 1993; Ulff-Moller, 1998; Chabot and DRAKE, 1999; Wasson, 1999; and DRAKE, 2000). Metallurgical cooling rates of IIAB iron meteorites are fairly uniform with an average of 50 K Myr -1 (Rasmussen, 1966; Kracher Fredriksson, 1966; Olsen et al., 1999). This is a characteristic feature of IIAB iron meteorites and probably the most complete absence of unusual or poorly understood features make group IIAB iron meteorites the most primitive members of the differentiated core.

Although the metal in the latter two crystallized within the solid silicate matrix, it has a normal structure, chemistry, and cooling rate for group IVB (Schaudt et al., 1972; Haack and Scott, 1996a). Metallurgical cooling rates of group IVB iron meteorites are subject to a wide variation in cooling rate, which is an artifact of crystallization of a common core (MORREN and GOLDSTEIN, 1973, 1979; RASMUSSEN, 1982; Rasmussen et al., 1995; Haack et al., 1996a). Haack suggests the Appalachians are a base metal core, indicating that the parent body went through a breakup and reassembly event shortly after core crystallization. The model was disputed by Wasson and Richardson (2001), who argued that the apparent correlation between metallurgical cooling rates and chemistry is a result of error. Like IIAB, they are often heavily shocked and have similar cooling rates. The most primitive members of the group have a cooling rate of 10 -3 K Myr -1 (Voshage, 1967; Voshage and Feldmann, 1979; see Chapter 1.13), suggesting catastrophic dispersal then (Keil et al., 1996a). The large number of IIAB iron meteorites and the almost complete absence of unusual or poorly understood features make group IIAB iron meteorites the best available samples of a crystallized metal core from a differentiated body.

1.12.2.3 Group IVA Iron Meteorites

IVA iron meteorites are fine octahedrites (Buchwald, 1975). With 48 members, IVA is the third-largest group. IVA irons have very low vol.% silicates mixed with metal on a centimeter scale (Haack et al., 1996a; Reid et al., 1974; Scott et al., 1996; Ulff-Moller et al., 1995).

<table>
<thead>
<tr>
<th>Group</th>
<th>Re</th>
<th>Ca</th>
<th>Cr</th>
<th>Fe</th>
<th>Ni</th>
<th>Co</th>
<th>Cu</th>
<th>Au</th>
<th>Go</th>
<th>Ge</th>
</tr>
</thead>
<tbody>
<tr>
<td>IVA</td>
<td>1.780 (250)</td>
<td>12.5 (1.3)</td>
<td>5.6 (4.6)</td>
<td></td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>IIAB</td>
<td>4.39 (500)</td>
<td>32.0 (2.5)</td>
<td>8.5 (1.0)</td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>IVB</td>
<td>2.150 (250)</td>
<td>18.0 (22.0)</td>
<td>17.3 (7.6)</td>
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Data are from the references listed in Section 3.12.2. Numbers in parentheses are single determinations. Data from Chabbal and Drake (1990a, b). Note the significant differences between the estimates of the initial liquid iron and silicate concentrations. High metallographic cooling rates of IVB iron meteorites (Rasmussen et al., 1984) suggest a small parent body.

1.12.2.5 Silicate-bearing Iron Meteorites

Group IAB, IIICD, and IIE iron meteorites differ in a number of properties from other iron groups. The compositional trends in these groups are unlike those in other (e.g., IIA) iron meteorites for which an origin from a common core is inferred. Values for IIIE iron meteorites are lower than in any other group of irons. Except for a few IVB irons, the iron meteorites have very low volatile concentrations consistent with fractional crystallization of a common core, suggesting formation from relatively oxidized material (Kracher Fredriksson, 1966; Olsen et al., 1999). Unlike other groups, irons in IAB and IIICD show considerable ranges in nickel, iron, and germanium concentrations, but very restricted ranges for all other elements. Values for IIICD iron meteorites in these groups vary within a factor of 10, compared to more than three orders of magnitude in IAB iron meteorites. These trends cannot be explained by simple fractional crystallization (Scott, 1972; Scott and Wasson, 1975). IAB irons also have metal compositions that are different from those of other iron meteorites (Scott et al., 1997), suggesting that the metal was molten when mixed with the silicates (Hassanzadeh et al., 1989). Many innovative models have attempted to explain the emplacement of iron meteorites in the core of the Earth. However, the most primitive members of the group are found in Wasson and Kallemeyn (1984). A recent compilation of the chemical compositions of IAB iron meteorites may be found in Wasson and Richardson (2001).

1.12.2.6 Group IVE Iron Meteorites

Despite containing only 13 members, group IVE iron meteorites exhibit unique properties in a special mention. Group IVE iron meteorites have the lowest volatile concentrations of any group (Table 1; see figure 26, Chapter 1.05) and are enriched in refractory elements, with iridium, thallium, and osmium concentrations an order of magnitude higher than in any other iron meteorites (Scott, 1972). These cool -down curves of mesosiderites are almost uniform at a depth of 100 Ma (Voshage, 1967; Voshage and Feldmann, 1979; see Chapter 1.13) suggesting that the metal was molten when mixed with the silicates (Hassanzadeh et al., 1989). Although sometimes termed "nonmagmatic," IAB, IIICD, and IIE iron meteorites closely approached melting during their history, although perhaps of a different type than, for example, group IIA. Unlike other groups, irons in IAB and IIICD show considerable ranges in nickel, iron, and germanium concentrations, but very restricted ranges for all other elements. Values for IIICD iron meteorites in these groups vary within a factor of 10, compared to more than three orders of magnitude in IAB iron meteorites. These trends cannot be explained by simple fractional crystallization (Scott, 1972; Scott and Wasson, 1975). IAB irons also have metal compositions that are different from those of other iron meteorites (Scott et al., 1997), suggesting that the metal was molten when mixed with the silicates (Hassanzadeh et al., 1989). Many innovative models have attempted to explain the emplacement of iron meteorites in the core of the Earth. However, the most primitive members of the group are found in Wasson and Kallemeyn (1984). A recent compilation of the chemical compositions of IAB and IIICD iron meteorites may be found in Wasson and Kallemeyn (2002). Group III is a much smaller group with very diverse characteristics in terms of metal textures and silicate mineralogy. The chemical composition of the metal is very restricted and inconsistent with fractional crystallization (Scott and Wasson, 1975; Wasson and Wake, 1986). The inclusion silicates range from metamorphosed chondrites (e.g., Neuchatel) to highly differentiated silicates (e.g., Kedaikan, Weekeeston Station), with intermediate members present (e.g., Wasson, Techadu) (Bogard et al., 2000). The inclusion silicates of members of the group resemble H chondrites in both mineral chemistry and oxygen isotopic composition (Clayton and Mayeda, 1996; Rigutti et al., 1999), although the match is not perfect (Bogard et al., 2000). The most perplexing feature of this group is the presence of silicate inclusions which give chromatic ages of 2-5 Ma, suggesting formation of 1-4 Gt after the formation of the core. A range of models have been proposed in the past decade, including both impact-induced (Olsen et al., 1994), and incipient parent-body melting (Bogard et al., 2000). Young chromatic ages would virtually require that impact played a role in the formation of some members of this group.

1.12.2.6 Mesosiderites

Mesosiderites are the most enigmatic group of differentiated meteorites. Mesosiderites are breccias composed of roughly equal proportions of Fe-Ni metal and silicates with a bulk composition that is not seen in any stony-iron meteorite. Mesosiderites are made of fine-grained metallic cores, suggesting that the mesosiderite parent bodies were associated with the inner Solar System. The most primitive members of the group are found in Wasson and Kallemeyn (1984). Several relatively new models have attempted to explain the emplacement of iron meteorites in the core of the Earth. These models are the slowest for any natural geological material, suggesting that the mesosiderite parent body must have been large. Mesosiderites have young Ar-Ar ages of 10-20 Myr (Bogard, 1998). The young Ar-Ar ages have been attributed to extended cooling within a large asteroid (Haack et al., 1996a; Bogard and Garrison, 1998) or impact resetting (Bogard et al., 1990, Rubin and Mittlefehldt, 1992).
concentrations of ungrouped irons are not randomly distributed as expected if they represent a large and continuous population of poorly sampled parent bodies (Scott, 1979). Ungrouped irons tend to have gallium and germanium concentrations in the same ranges as those defined by the original Ga–Ge groups I–IV. The origin of ungrouped irons is poorly understood. While many of them sample poorly known asteroidal cores, they include a variety of anomalous types, including highly reduced silicon-bearing irons (e.g., Horse Creek) and at least one that essentially quenched from a modern state (e.g., Niedaiglota).

1.12.3 ACCRETION AND DIFFERENCES IN BULK CHEMISTRY BETWEEN GROUPS OF IRON METEORITES

The most striking differences in chemical composition between groups of iron meteorites are the differences in concentration of volatile elements (several orders of magnitude) and the smaller but important threefold variation in nickel concentration (Table 1). These bulk chemical differences cannot be attributed to fractional crystallization of asteroidal cores (Scott, 1972), but suggest that processes during asteroid accretion and fractionation produced metallic melt bodies with a range of compositions. These processes occurred exclusively in the nebula (condensation–evaporation), in both nebular and parent-body settings (oxidation–reduction) and exclusively on the parent body (metal–silicate segregation, degassing of volatiles and impacts). In some cases, it is possible to relate specific chemical characteristics of an iron meteorite group to a specific process, but in most cases some ambiguity remains. In this section, we discuss the diverse chemical compositions of iron meteorites and its possible origins.

Chondrites are generally considered representative of the material from which asteroids, including iron meteorite parent bodies, and planets formed. The heterogeneity of primitive chondrites shows that solids in the early solar system were, to some extent, chemically and isotopically diverse. Metal abundances range from zero in some carbonaceous chondrites (e.g., CI) to more than 60 wt.% in some CB/CH chondrites (Campbell et al., 2001; see Chapter 1.05). The chemical composition of chondritic metal is also diverse, primarily reflecting oxidation–reduction processes, although in rare cases the metal formed at high temperature condensates (Meibom et al., 1999). These groups are depleted in the volatile elements sulfur, phosphorus, gallium, germanium, phosphorus, antimony, copper, and gold, with the nonmetals sulfur and phosphorus of particular importance, since they have significant effects on core crystallization and Widmanstätten pattern formation. While the depletion of volatile elements in groups II and IIIA results from parent-body degassing (Keil and Wilson, 1993), enrichments in the refractory elements rhenium, osmium, and iridium in group IVB cannot (Kelly and Larimer, 1977; Scott, 1972; Wasson and Wai, 1976). High-temperature condensation for group IVA might also explain the unusually high nickel concentration, although oxidation–reduction processes are likely involved (Figure 2).

The main processes that control the composition of nebular metal—condensation and fractionation from the gas at high temperatures and oxidation–reduction processes (Kelly and Larimer, 1977)—are illustrated in Figure 1. Oxidation of iron from the metal will enrich the remaining metal in elements less easily oxidized than iron. Differences in mineralogy and chemistry between iron meteorite groups show that oxidation–reduction processes were important. The oxygen fugacity of the molten core is reflected in the abundance of the oxygen-bearing phosphates found in some groups of iron meteorites (Olsen and Fredriksson, 1966; Olsen et al., 1999). Phosphate minerals are typical in group IIAB but have never been observed in IIA and IIIAB iron meteorites (Bischelwald, 1975, 1984; Scott and Wasson, 1975). The higher nickel concentration in group IIAB compared to IIAB (Table 1) is consistent with the later forming from more reduced material. Whether this difference is a result of a nebula process or a parent-body process remains an open question.

The other important process in the nebula is condensation and fractionation from the gas. Condensation at high-temperatures results in refractory element enrichment and volatile depletion. High-temperature condensates may be preserved if isolated from the gas before more volatile elements condense (Meibom et al., 2000; Petri et al., 2001). Processes operating within the parent bodies may mask this process. Volatile-element depletions could result from degassing of the parent-body during heating, partial melting, and impact (Rasmussen et al., 1984; Keil and Wilson, 1993). Parent-body degassing will not, however, result in significant refractory element enrichment. Thus, it may be possible to distinguish volatile depletions caused by high-temperature condensation and parent-body degassing.

The best evidence for the importance of condensation and fractionation is the composition of group IVB irons and ungrouped irons of similar composition (Figure 2). The bulk compositions of groups IVA and IVB irons are consistent with high-temperature condensation of the source material (Kelly and Larimer, 1977) (Figure 1). These groups are depleted in the volatile elements sulfur, phosphorus, gallium, germanium, phosphorus, antimony, copper, and gold, with the nonmetals sulfur and phosphorus of particular importance, since they have significant effects on core crystallization and Widmanstätten pattern formation. While the depletion of volatile elements in groups II and IIIA results from parent-body degassing (Keil and Wilson, 1993), enrichments in the refractory elements rhenium, osmium, and iridium in group IVB cannot (Kelly and Larimer, 1977; Scott, 1972; Wasson and Wai, 1976). High-temperature condensation for group IVA might also explain the unusually high nickel concentration, although oxidation–reduction processes are likely involved (Figure 1). The source material for IVB irons is not preserved among known chondrites (Rasmussen et al., 1984), although a few CB chondrites contain metal grains thought to be the first direct evidence that such materials were produced in the nebula. A close relationship...
between type IVA and I VB irons and CB chondrites is also suggested by the elevated $^{87}$Kr values in all three groups (Kerridge, 1985; Bhombro and Clancy, 1985, 1992). Metal from IVA and CB irons would have been isolated from a nebular gas above 1,500 K (Petava et al., 2001), even higher than metal grains in CB chondrites. The postulated chondritic metal corresponding to I VB irons was either entirely incorporated into bodies that differentiated, or these chondrites remained unsampled.

### 1.1.2.4 HEATING AND DIFFERENTIATION

How did asteroidal cores come to exist in the first place? What were the heat sources? In this section, we address these fundamental issues by considering evidence from iron meteorites, primitive chondrites, achondrites, experiments, and numerical calculations. The simplest case for core formation is metal sinking through a silicate matrix that has experienced a high degree of partial melting. Numerous experimental studies (e.g., Takahashi, 1983; Walker and Agee, 1988; McCoy et al., 1998) demonstrate that metal and sulfide tend to form rounded globules, rather than an interconnected network, at moderate degrees of silicate partial melt. These globules then sink through silicate mush. Taylor (1992) calculated that at silicate fractions of 0.5, metal particles $\sim 1,000$ cm sink readily through a crystal mush, although it remains unclear: how millimeter-sized metal particles in primitive chondritic meteorites attain these sizes. Settling would be rapid, with core formation requiring tens to thousands of years, depending on the melt body size and degree of silicate melting. Many large iron meteorites (e.g., Canyon Diablo, Hoba) and shower-producing irons (e.g., Gibeon) exceed 1 - 10 m, supporting rather than dispersed metallic masses.

A more interesting case is porous flow through interconnected networks in a largely solid silicate matrix that has experienced a high degree of partial melting. Rushmer (1992) calculated that at silicate pressures in the range of 10 GPa, Fe$_3$S$_2$ melts at low pressure (anions/cations $\sim 0.8$) would not migrate to form a core under static conditions. However, if dynamic processes play a significant role in the formation of asteroidal cores, then metal-sulfide melt migration may be possible, although the pressure-strain regime may not be applicable to asteroidal-sized bodies. Keil and Wilson (1993) suggested that overpressure during Fe$_3$S$_2$ eutectic melting might create veins and volatiles that might cause these veins to rise in small bodies. Partial melting at low degrees of partial melting may be sulfur-rich or even sulfur-dominated. The small sulfur rich cores would coexist with a mantle that contains essentially chondritic silicate mineralogy and metal abundance. As near-eutectic melting is achieved, metal drains efficiently to the center of the body, forming metal-dominated cores, which may be melted in sulfur relative to the chondritic precursor as a result of explosive removal of the Fe$_{Ni}$-Fe$_3$S$_2$ eutectic melt.

### 1.1.2.5 Melting of asteroids by $^{26}$Al

#### Melting of asteroids by $^{26}$Al

The consequences of an impact during core formation/crystallization of the terrestrial planets—would have produced heating of no more than a few tens of degrees in asteroids. Keil et al. (1997) argued that the maximum energy released during accretion is equal to the gravitational binding energy of the asteroid after accretion. For a 1,000 km body, this equates to a temperature increase of only 6°C. Impacts, however, can also contribute to melting and growth of asteroidal metallic masses.

The most likely heat source for melting of asteroidal parent bodies is in situ heating by the T-Tauri solar wind from the pre-main-sequence Sun (e.g., Sonnet et al., 1970) or short-lived radioactive isotopes. Melting of asteroids by $^{26}$Al has gained broader acceptance since the discovery of excess $^{26}$Al in Pippa Kalan (Srinivasan et al., 1999; Tachibana and Huss, 2003) points out an interesting conundrum. In chondrites, $^{26}$Al is concentrated in plagioclase—whereas in the terrestrial planets—would have produced heating of no more than a few tens of degrees in asteroids. Keil et al. (1997) argued that the maximum energy released during accretion is equal to the gravitational binding energy of the asteroid after accretion. For a 1,000 km body, this equates to a temperature increase of only 6°C. Melting by $^{26}$Al then raised the possibility that $^{26}$Al could cause melting even in the absence of other heat sources. Occurring in both oxidized and reduced forms, $^{56}$Fe could cause melting even in the absence of other heat sources. Melting in oxidized and reduced forms, $^{56}$Fe would be retained throughout the body during differentiation and could provide the heat necessary for global melting and, ultimately, core formation in asteroids.

### 1.1.2.5 Fractional Crystallization of Metal Cores

#### Fractional Crystallization of Metal Cores

Although iron meteorites provide a guide to understanding the ongoing crystallization of the Earth’s core, important differences in the physical
settings exist. First and foremost, the central pressures within the iron meteorite parent bodies did not exceed 0.1 GPa, compared to central pressures in excess of 350 GPa on Earth. The steep pressure gradient in the Earth's core, which causes the core to crystallize from the inside out, was absent in the asteroid cores. Since the core is cooled through the mantle, Haack and Scott (1992) argued that the onset of core crystallization in asteroids was probably from the base of the mantle. Crystallization of the asteroidal cores was likely in the form of kilometer-sized dendrites as light buoyant liquid, enriched in the incompatible element sulfur inhibited crystallization of the outer core.

The fractional crystallization trends preserved in iron meteorites are unparalleled in terrestrial magmatic systems. Trace-element variations span more than three orders of magnitude within group IIIAB iron meteorites (Figure 5). The traditional way to display the bulk compositional data of iron meteorites in a log nickel versus log element diagram. More recent work has used gold as the reference element (Wasson, 1999; Wasson and Richardson, 2001). Gold is a better choice because the distribution coefficient is further from unity giving a natural variation that far exceeds the analytical uncertainty (Haack and Scott, 1993). In a log F versus log Ni diagram, ideal fractional crystallization from a perfectly mixed liquid with constant distribution coefficients will result in straight chemical trends where the slope is given by $D_{\text{Ni}} - 1$, and $D_{\text{Ni}} + 1$, where $D_{\text{Ni}}$ and $D_{\text{As}}$ are the liquid metal/solid metal distribution coefficients for the element and nickel, respectively. Although most trends define almost straight lines, differences in slope from group to group and curved trends for some elements such as gallium and germanium show that the distribution coefficients cannot be constant.

The slopes of the Ni–Ir trends correlate with the volatile concentrations (including sulfur) for the different groups. Experimental work has shown that the distribution coefficients are functions of the phosphorus, sulfur, and carbon concentration in the liquid (Goldstein and Friel, 1978; Naranjy and Goldstein, 1981, 1982; Willis and Goldstein, 1982; Jones and Drake, 1983; Malvin et al., 1986; Jones and Malvin, 1990; Chabot and Drake, 1997, 1999, 2000; Liu and Fleet, 2001; Chabot and Jones, 2003). Using the experimentally determined distribution coefficients it is possible to calculate fractional crystallization trends (Figure 5). Early numerical models assumed that the liquid remained perfectly mixed throughout crystallization (Willis and Goldstein, 1982; Jones and Drake, 1983), whereas later models have included the effects of assimilation (Malvin, 1988), imperfect mixing of the liquid (Chabot and Drake, 1990), liquid immiscibility (Ulf-Möller, 1998; Chabot and Drake, 2000), and trapping of sulfur-rich liquid (Haack and Scott, 1993; Scott et al., 1996; Wasson, 1999; Wasson and Richardson, 2001).

Numerical models of the fractional crystallization process are broadly consistent with the observed trends. There are, however, several features that remain poorly understood.

1.12.1.1 Imperfect Mixing during Crystallization

The compositional data scatter from the average trend for each group are significantly greater than the analytical uncertainty, in particular for compatible elements (Pernicka and Wasson, 1987; Haack and Scott, 1992, 1995; Scott et al., 1996; Wasson, 1999; Wasson and Richardson, 2001). The scatter shows that the assumption of metal crystallizing from a perfectly mixed liquid breaks down during the course of crystallization. An interesting example of this scatter is observed among and within the metallogram fragments of the IIIAB iron meteorite Cape York (Esbensen and Buchwald, 1982; Esbensen et al., 1982). The compositional trend defined by the different Cape York fragments diverges from the general trend defined by IIIAB iron meteorites (Figure 6). The observation that the compositional variation within a single meteorite is almost insoluble in solid metal, and it is therefore not possible to estimate the liquid concentration by analyzing iron meteorites that represent the crystallized metal. Sulfur may only be determined indirectly by treating it as a free parameter in the fractional crystallization models and choose the sulfur composition that provides the best fit to the data (Jones and Drake, 1983; Haack and Scott, 1993; Chabot and Drake, 2000).

The concentration of sulfur in the liquid as crystallization proceeds would lead to changes in the slope of the elemental trends that are not matched by observations. As sulfur and phosphorus concentrations increase, liquid immiscibility may also play an increasingly important role. Shows that at high sulfur and phosphorus concentrations (and high oxygen fugacity), the liquid may be two-phase field (Jones and Drake, 1983; Chabot and Drake, 2000) (Figure 7). Ulf-Möller (1998)
modeled the crystallization of the IIIAB core taking liquid immiscibility into account. He found the end of core crystallization. However, the predicted abundances. For the sulfur- and number of sulfur-rich meteorites that the unsampled volume of eutectic melt is during atmospheric entry (Kracher and Wasson, 1982), it remains a mystery why the number of sulfur-rich meteorites is so low. It is, however, worth noting that although the mantle and crust probably comprised more than 85 vol.% of differentiated parent bodies, achondrites that could represent mantle and crust materials of the iron meteorite parent bodies are astonishingly rare. Clearly, processes that selectively remove more fragile materials en route to the Earth are of significant importance.

11.12.5.3 The Missing Sulfur-rich Meteorites Models predict that a considerable volume of Fe-Cr-FeS eutectic melt should be produced at the end of core crystallization. However, the number of sulfur-rich meteorites falls far short of the predicted abundances. For the sulfur- and phosphorus-rich group IIAB, the models predict that the unsampled volume of eutectic melt is ~70% (Chabot and Drake, 2000). Although arguments may be made that the sulfur-rich material is weaker and, therefore, more easily broken down in space and ablates more rapidly during atmospheric entry (Kracher and Wasson, 1982), it remains a mystery why the number of sulfur-rich meteorites is so low. It is, however, worth noting that although the mantle and crust probably comprised more than 85 vol.% of differentiated parent bodies, achondrites that could represent mantle and crust materials of the iron meteorite parent bodies are astonishingly rare. Clearly, processes that selectively remove more fragile materials en route to the Earth are of significant importance.

1.12.6 COOLING RATES AND SIZES OF PARENT BODIES After the cores of the differentiated asteroids had crystallized, a slow cooling period commenced. During this period the most prominent feature of iron meteorites evolved—the Widmanstätten pattern (Figure 8). Several characteristics of the Widmanstätten pattern may be used to constrain the thermal evolution and the sizes of the iron meteorite parent bodies. In a typical parent-body core with nickel concentrations in the range 7–15 wt% Ni, the low-nickel phase kamacite will grow at the expense of the high-nickel phase taenite between 800°C and 500°C. Several other elements are partitioned between the two phases during growth of the kamacite phase (Rasmussen et al., 1983). The kamacite grows as platelets in four possible orientations relative to the taenite host. Since most of the parent taenite crystals had dimensions larger than typical meteorites, a continuous pattern of four different sets of kamacite plates may be observed on etched surfaces of typical iron meteorites. The limiting factor for the growth of kamacite is the slow diffusion of nickel through taenite. Nickel profiles across taenite lamellae may, therefore, be used to determine the so-called metallographic cooling rate of the parent-body core at ~500°C (Wood, 1964; Rasmussen, 1981; Herpfer et al., 1994; Yang et al., 1997; Hoppe and Goldstein, 2001; Rasmussen et al., 2001). A number of revisions of the metallographic cooling rate method have been implemented since its original formulation (Wood, 1964; Goldstein and Ogilvie, 1965). Revisions have resulted from improvements in ternary (Fe–Ni–P) phase diagrams (Doan and Goldstein, 1970; Romig and Goldstein, 1980; Yang et al., 1996) and, in particular, the discovery that small amounts of phosphorus may increase the diffusion rate of nickel through taenite by an order of magnitude (Rasmussen et al., 1983; Yang et al., 1996). Several other characteristics of the Widmanstätten pattern have provided additional information on the thermal history of the metal. Diffusion controlled growth of schreibersite has been used to determine the cooling rates of hexahedrites (Randich and Goldstein, 1978). Although nickel diffusion through kamacite is much faster than through taenite, the decreasing temperature will eventually result in zoned kamacite as well. Zoned kamacite has been used to infer cooling rates at temperatures below 400°C (Haack et al., 1996b; Rasmussen et al., 2001). Yang et al. (1997) showed that the size of the so-called island phase is inversely correlated with cooling rate and may be used to infer the cooling rate at temperatures ~520°C. The island phase is nickel-rich tetrataenite that forms irregular globules with dimensions up to 470 nm in the nickel-rich rims of taenite.

With the exception of group IVA iron meteorites, the metallographic cooling rates tend to be similar within each group but different from group to group (Table 2). This is consistent with the idea that each iron meteorite group cooled in its own separate metallic core surrounded by an insulating mantle. The cooling rates for the different groups of iron meteorites are generally in the range 10–100°C Myr⁻¹. The metallographic cooling rates combined with numerical models of the thermal evolution of the parent bodies may be used to constrain the sizes of the parent bodies (Wood, 1964; Goldstein and Short, 1967; Haack et al., 1990). The main uncertainties in the numerical models are the states of the mantle, surface, and crust during cooling. Observations of present-day asteroids show that they tend to be covered by a thick, insulating regolith and that they may be heavily brecciated. A brecciated mantle and/or a highly porous regolith cover the surface of an asteroid could potentially slow the cooling rate by a factor of 5–10 (Haack et al., 1990). An approximate relationship between metallographic cooling rates and parent-body radius for regolith covered asteroids was given by Haack et al. (1990): $R = 149 \times CR^{0.63}$, where $R$ is the radius in km and CR the cooling rate in K Myr⁻¹. Table 2 gives a compilation of cooling rates and corresponding parent-body sizes.

<table>
<thead>
<tr>
<th>Group</th>
<th>Cooling rate (K Myr⁻¹)</th>
<th>Parent-body radius (km)</th>
<th>Refs.</th>
</tr>
</thead>
<tbody>
<tr>
<td>I</td>
<td>25</td>
<td>&gt;33</td>
<td>a</td>
</tr>
<tr>
<td>IIAB</td>
<td>6.12</td>
<td>45–65</td>
<td>b</td>
</tr>
<tr>
<td>IIIAB</td>
<td>15–85</td>
<td>20–40</td>
<td>c</td>
</tr>
<tr>
<td>IV</td>
<td>5–15</td>
<td>42–50</td>
<td>d</td>
</tr>
<tr>
<td>IVA</td>
<td>19–3,400</td>
<td>&gt;40</td>
<td>d</td>
</tr>
<tr>
<td>IVB</td>
<td>170–230</td>
<td>12–14</td>
<td>e</td>
</tr>
<tr>
<td>Pallasites</td>
<td>2.5–8</td>
<td>80–100</td>
<td>o</td>
</tr>
<tr>
<td>Meteoroids</td>
<td>0.5</td>
<td>200</td>
<td>a</td>
</tr>
</tbody>
</table>

References: (a) Yang et al. (1997), (b) Salickor and Goldstein (1988), (c) Romig and Goldstein (1978), (d) Rasmussen et al. (1983), and (e) Rasmussen et al. (1995). A radius of the parent body can only be calculated for those groups where the meteorites are believed to have cooled in the core. For other groups a minimum parent-body size may be calculated.
Pallasites are the most abundant group of stony-iron meteorites. Their features are easily reconciled with simple models of asteroid differentiation, yet challenge some of our basic assumptions about core formation and crystallization (Mittlefehldt et al., 1998). The 50 known pallasites are divided into main group pallasites, the Eagle Station group (three members) and the pyroxene pallasite group (two members). These are distinguished based on oxygen isotopic, mineral, and metal compositions. Main group pallasites are comprised of subequal mixtures of forsteritic olivine and iron, nickel metal, often heterogeneous, which suggest that olivine grains reach several centimeters (Uff-Möller et al., 1998), while Bresnan has metal cross-cutting zones with more typical pallasitic texture (Figure 9). Olivine morphology also differs significantly, from angular—subangular in Saars to rounded in Thiel mountains (Buseck; 1977, Scott, 1977b).

Olivines in the main group pallasites cluster around Fo6.8, although some reach Fo50 and exhibit core-to-iron zoning of aluminum, chromium, calcium, and manganese (Zhou and Steele, 1995; Ho et al., 1997). Chromite, low-calcium pyroxene and a variety of phosphates comprise <1 vol.% each (Buseck, 1977; Davis, 1977; Davis and Olsen, 1998). The metal composition is similar to high-nickel IIAB irons, but with a greater scatter in nickel (7-13 wt.%) and some other siderophile elements (Davis, 1977; Scott, 1977b). The pallasite-IIAB iron link is supported by the similarity in oxygen isotopic compositions of their silicates (Clayton and Mayeda, 1996).

The Eagle Station trio (Eagle Station, Cold Bay, Izanagi) is comprised of olivine, iron, nickel metal and olivine with lesser amounts of pyroxene, orthopyroxene, chrome, and phosphates. Olivines in the Eagle Station (FO50, 53) are more ferroan than in the main group and differ substantially in their Fe/Mn ratio (Mittlefehldt et al., 1998). Metal in the Eagle Station pallasites has higher iridium and nickel compared to main group pallasites and is closer to metal in IIAB irons (Kraucher et al., 1980). They also differ dramatically in oxygen isotopic compositions from main group pallasites (Clayton and Mayeda, 1996).

Pyroxene pallasites are represented by only two members: Vermillion and Yamato 8451. The two meteorites share the common feature of core-forming pyroxene, but differ substantially from one another. Yamato 8451 consists of ~60% olivine, 35% metal, 2% pyroxene, and 1% troilite (Hiroi et al., 1993; Yamai and Kojima, 1995). In contrast, Vermillion contains around 14% olivine, and less than 1% each of orthopyroxene, chrome, and phosphates (Boesenberg et al., 2000). The iron concentration in olivine (FO50, 50) is similar to main group pallasites, although pyroxene pallasites have lower Fe/Mn ratios (Mittlefehldt et al., 1998). The two pyroxene pallasites do not share a common metal composition with each other, or in detail, with other pallasites (Wasson et al., 1998).

Pallasites are both intriguing and perplexing. The two lithologies, Fo,Ni metal and olivine, are reasonable assemblages expected at the core—mantle boundary of a differentiated asteroid. The lower mantle should be primarily olivine, either a result from high degrees of partial melting or an early cumulate phase from a global magma ocean. However, the marked density contrast between metal and silicates should lead to rapid separation. Further, some features of pallasitic olivine, such as the angular shapes and marked minor element zoning, seem inconsistent with formation at a relatively quiescent, deep-seated core—mantle boundary. As Mittlefehldt et al. (1998) note, the core—mantle boundary origin remains the most plausible, despite these difficulties.

If we assume a core—mantle boundary origin (see Mittlefehldt et al., 1998, for references suggesting alternative models), how can these disparate features be reconciled and what are the implications for the nature of asteroid cores during cooling and solidification? Scott (1977c) suggested that the main group pallasite metal is a reasonable crystallization product after ~90% crystallization of IIAB metallic melt. However, asteroid cores probably crystallized from the core—mantle boundary inwards and we should expect pallasites to have the signature of early crystallizing, low-nickel melts, not late-crystallizing, high-nickel melts. Haack and Scott (1993) postulated that residual metallic melt migrated to the core—mantle boundary between dendrites and Ulf-Möller et al. (1997) suggested mixing of late-stage metallic melt with earlier solidified metal to produce the range of pallasite compositions. Late-stage intrusions might explain the presence of angular olivines, which were fractured during the intrusion (Scott, 1977b), and the heterogeneous textures seen in pallasites like Bremmel. Ulf-Möller et al. (1997) suggested that high impact pressure injection of metal, perhaps caused by impact, might provide the most plausible mechanism for producing pallasites. Finally, pallasites have experienced a long history of sub solids cooling and annealing. Both minor-element zoning in olivine and rounded olivines might be best explained by sub solidus annealing and diffusion. The small Eagle Station and pyroxene-bearing pallasites are most akin enigmatically, but suggest that similar processes operated on other parent bodies and may have been common among highly differentiated asteroids.

PARENT BODIES OF IRON AND STONY-IRON METEORITES

While iron and stony-iron meteorites provide important snapshots of the origin and evolution of highly differentiated asteroids, they lack geologic context. From which type of asteroid do iron and stony-iron meteorites originate? Have we sampled correspondingly abundant crustal and mantle material in both the meteorite and asteroid populations? These are the questions that we explore in this section.

The major tool that allows us to relate classes of meteorites to asteroids is spectral reflectance (Burbin et al., 2002). Unfortunately, iron meteorites, with their paucity of silicate phases and the likely presence of featureless spectra with red spectral slopes and moderate albedos (e.g., Cloutis et al., 1990) (Figure 10). This appears to be no simple relationship between nickel abundance and spectral redness and, thus, distinguishing different chemical groups (e.g., low-nickel IIAB from high-nickel IVB) is probably not possible. Historically, M-class asteroids, of which ~40 are known, have been linked to iron meteorites. These asteroids exhibit moderate visual albedos, relatively featureless spectra with red spectral slopes, similar to that of iron meteorites. Radar albedos—an indirect measure of near-surface bulk density—have been used as supporting evidence for the link between iron meteorites and M-class asteroids, as M asteroids tend to have higher radar albedos than C or S asteroids (Magni et al., 1999) and the highest asteroid radar albedos are known from the M asteroids 16 Psyche (1986 DA) and 216 Kleopatra (Ostro et al., 1991, 2000). There is reason to believe that not all (perhaps not many) M-class asteroids are either core fragments or largely intact stripped cores of differentiated asteroids. The enstatite chondrites also exhibit nearly featureless spectra (owing to the nearly Fe3+-free composition of the enstatite; Keil, 1968) with red spectral slopes and moderate albedos (Gaffey, 1976) and have been suggested as possible meteoritic analogues to M asteroids (Gaffey and McCord, 1978) (Figure 10). Rivkin et al. (2000) found that more than one-third of observed M-class asteroids have 3 µm absorption features suggestive of hydrated silicates. Although this conclusion remains the subject of significant debate (Gaffey et al., 2002; Rivkin et al., 2002), it would be inconsistent with core material. Finally, recent ground-based measurements of bulk densities for M asteroids 16 Psyche (Vianello, 2000; Brito et al., 2002) and 22 Kalliope (Margot and Brown, 2001) are ~2 g cm3, far below that expected for even highly fragmented core fragments.

Equally puzzling is the lack of olivine-rich asteroids corresponding to the large numbers of
asthenospheric cores sampled by iron meteorites. If we have sampled ~65 asthenospheric cores, we might expect corresponding abundances of olivine-rich meteorites with core-type (olivine-rich) asthenospheres (Chapman, 1986; Bell et al., 1989; Burbine et al., 1996; Gaffey et al., 2002b). The near absence of stony meteorites and a paucity of A-type asteroids has come to be known as "The Great Dunite Shortage." The most likely explanation centers on the nature of differentiation in parent bodies and the durability of metallic meteorites relative to their stony counterparts. If the fragmentation of differentiated bodies occurred relatively early in the history of the solar system, the dunite mantles may have been fragmented to below the current observational limit of ~5 km. Dunite meteorites may well have fallen to Earth as a consequence of this fragmentation of the mantles, but terrrestrial ages for meteorites stretch back only ~2 Myr (Welten et al., 1997). In contrast, core fragments may prove much more rare, and, because of their density, they may still be sampled by impacts today. Indeed, iron meteorites are much more durable, with core-ray exposure ages to more than 2 Gyr (e.g., Deep Springs, Voshage and Feldmann, 1979). Much slower Yarkovsky-driven transfer times from the asteroid belt for metallic objects also suggests that these sample impacts further back in time (Farinella et al., 1998).

A second field ripe for future work is experimental studies designed to increase our understanding of the processes of core crystallization. Although we have provided possible links between iron and stony-iron meteorites and their possible parent as discussed in the previous section, our understanding of the nature of differentiation of M-type asteroids makes it impossible to establish a firm link between iron meteorites and M-type asteroids. The high cosmic-ray exposure ages of iron meteorites show, not surprisingly, that iron--nickel alloys are much more resistant to impacts and space erosion than stony targets. We should, therefore, expect that the surfaces of stripped metal cores or fragments of metal cores display a dramatic crateted terrain unlike anything that we have ever seen before. Although the subtle compositional differences that occur within and between iron meteorite groups could certainly be discerned from orbit, the difference between metal-dominated lithologies, troilite-rich lithologies, and dunite/pallasitic lithologies could be easily distinguished on the basis of both spectral absorption features (e.g., Britz et al., 1992) and, more readily, orbital X-ray or gamma-ray spectroscopic mapping. In short, solving questions about the existence and/or spatial distribution of these important lithologies. The question of whether the distribution of sulfur-rich pockets, for example, can provide evidence on the core crystallization process (Haack and Scott, 1993) and its implications for the nature of late-stage core crystallization. Such a mission might also resolve the role and extent of core-mantle differentiation in the formation of asthenospheric cores. Impact clearly played a role in the formation of both mesosiderites and silicate-bearing meteorites (iron meteorites), but whether those late-stage impacts were global or local remains an open question. The slow cooling rates of mesosiderites suggest that the parent body was very large, probably several hundred kilometers in radius (Haack et al., 1996b). Since large asteroids are more difficult to destroy, this opens the possibility that the parent body may still exist. Davis et al. (1999) suggested that the largest M-type asteroid, 16 Psyche, might have remained as a single fragment of the parent mesosiderite parent body. A mission to Psyche may not only reveal the nature of M-type meteorites, it could possibly provide evidence on one of the biggest impact events documented in our meteorite collections.

Finally, much has been written in recent years about the hazard presented to mankind by asteroid impact (Gehrels, 1994). While considerable effort has been put into understanding possible mitigation strategies, it is important to remember that energy scales linearly with mass. Thus, a metallic asteroid of comparable diameter would release ~2.5 times the energy of a stony asteroid. Further, its physical properties would certainly be quite different from that of a stony asteroid. The aforementioned mission might not only provide substantial insights into the geologic history of asteroids, but also provide essential data for understanding the physical properties (e.g., material distribution, global crags) that would allow us to continue the historical mission.

REFERENCES


Iron and Iron-Meteorites


