

Meteoritic Parent Bodies: Their Number and Identification

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Extensive collection efforts in Antarctica and the Sahara in the past 10 years have greatly increased the number of known meteorites. Groupings of meteorites according to petrologic, mineralogical, bulk-chemical, and isotopic properties suggest the existence of 100–150 distinct parent bodies. Dynamical studies imply that most meteorites have their source bodies in the main belt and not among the near-Earth asteroids. Spectral observations of asteroids are currently the primary way of determining asteroid mineralogies. Linkages between ordinary chondrites and S asteroids, CM chondrites and C-type asteroids, the HEDs and 4 Vesta, and iron meteorites, enstatite chondrites, and M asteroids are discussed. However, it is difficult to conclusively link most asteroids with particular meteorite groups due to the number of asteroids with similar spectral properties and the uncertainties in the optical, chemical, and physical properties of the asteroid regolith.

1. INTRODUCTION

Since the publication of *Asteroids II*, our knowledge concerning the composition, orbital dynamics, and geologic histories of meteorites and asteroids has increased substantially. These advances in meteoritics have been primarily due to the thousands of new meteorites discovered in Antarctica (e.g., *Zolensky, 1998*) and the Sahara (e.g., *Bischoff, 2001*) coupled with more precise and sensitive analytical equipment to characterize them. The near-exponential increase in computational power and improvements in integration algorithms have allowed for vastly improved studies of the transport of meteorites from the asteroid belt. Our knowledge of the diversity of asteroid mineralogies has increased considerably from the use of charge-coupled device (CCD) detectors to obtain visible and near-infrared reflectance spectra of smaller and smaller objects.

With these advances, it may be possible to determine parent bodies for particular meteorites. Geochemical and petrological evidence (e.g., *Lipschutz et al., 1989*) implies that almost all meteorites originated on subplanetary-sized objects that formed ca. 4.56 Ga. The exceptions are more than 50 unpaired meteorites that appear to originate from the Moon and Mars. Meteorites from other sources such as Mercury (*Love and Keil, 1995*), Venus (*Melosh and Tonks,*

1993), comets (e.g., *Campins and Swindle, 1998*), and even Earth (*Melosh and Tonks, 1993*) also appear possible (*Gladman et al., 1996*). Interplanetary dust particles (IDPs) (e.g., *Rietmeijer, 1998; Dermott et al., 2002*) sample comets and also asteroids. Meteorites from planets in other stellar systems appear to be very unlikely (*Melosh, 2001*).

These linkages between asteroids and meteorites are important for a variety of reasons. Scientifically, they allow for an understanding of compositional and thermal gradients in the solar nebula by allowing the orbital locations of objects with different chemical and isotopic compositions to be pinpointed. Economically, many Fe-rich asteroids could be important resources for elements relatively rare on Earth's surface (*Kargel, 1994*). For the preservation of the human race, an asteroid on a collision course with Earth may have to be diverted or destroyed, and it would be vital to know the object's composition when formulating scenarios for keeping the approaching body from impacting Earth.

The goal of this chapter is to determine the number and identity of meteoritic parent and/or source bodies. A parent body is the body from which the meteorite acquired its current chemical and mineralogical characteristics. A source body is a fragment of the parent body by which the meteorite was completely shielded from cosmic rays for most of

TABLE 1. Meteorite groups and their compositional characteristics.

Groups	Composition*	$\Delta^{17}\text{O}^\dagger$	Compositional Linkages with Other Groups
<i>Carbonaceous Chondrites</i>			
CI	phy, mag	0	
CM	phy, toch, ol,	-3	
CO	ol, px, CAIs, met	-4	
CR	phy, px, ol, met	-1.5	
CH	px, met, ol,	-1.5	
CV	ol, px, CAIs	-4	
CK	ol, CAIs	-4	
<i>Enstatite Chondrites</i>			
EH	enst, met, sul, plag, \pm ol	0	
EL	enst, met, sul, plag	0	
<i>Ordinary Chondrites</i>			
H	ol, px, met, plag, sul	0.73	IIE (<i>Olsen et al.</i> , 1994)
L	ol, px, plag, met, sul	1.07	
LL	ol, px, plag, met, sul	1.26	
<i>R Chondrites</i>			
R	ol, px, plag, sul	2.7	
<i>Primitive Achondrites</i>			
Acapulcoites	px, ol, plag, met, sul	-1.04	Iodranites (<i>Nagahara and Ozawa</i> , 1986)
Lodranites	px, ol, met, \pm plag, \pm sul	-1.18	acapulcoites
Winonaites	ol, px, plag, met	-0.50	IAB, IIIICD (<i>Bild</i> , 1977)
<i>Differentiated Achondrites</i>			
Angrites	TiO ₂ -rich aug, ol, plag	-0.15	
Aubrites	enst, sul	0.02	
Brachinites	ol, cpx, \pm plag	-0.26	
Diogenites	opx	-0.27	euclrites, howardites (<i>Mason</i> , 1962)
Euclrites	pig, plag	-0.24	howardites, diogenites
Howardites	euclritic-diogenitic breccia	-0.26	euclrites, diogenites
Ureilites	ol, px, graph	-1.20	
<i>Stony-Irons</i>			
Mesosiderites	basalt-met breccia	-0.24	
Main group pallasites	ol, met	-0.28	
<i>Irons</i>			
IAB	met, sul, ol-px,-plag incl	-0.48	IIIICD, winonaites
IC	met, sul		
IIAB	met, sul, schreib		
IIC	met, sul		
IID	met, sul		
IIE	met, sul, ol-px-plag incl	0.59	H chondrites
IIF	met, sul		
IIIAB	met, sul	-0.21	
IIIICD	met, sul, ol-px-plag incl	-0.43	IAB, winonaites
IIIE	met, sul		
IIIF	met, sul		
IIVA	met, SiO ₂ -px incl	1.17	
IVB	met		

*Minerals or components are listed in decreasing order of average abundance. Abbreviations: ol = olivine, px = pyroxene, opx = orthopyroxene, pig = pigeonite, enst = enstatite, aug = augite, cpx = clinopyroxene, plag = plagioclase, mag = magnetite, met = metallic iron, sul = sulfides, phy = phyllosilicates, toch = tochilinite, graph = graphite, CAIs = Ca-Al-rich refractory inclusions, schreib = schreibersite, incl = inclusions, \pm = may be present.

† Average values (e.g., *Clayton et al.*, 1991; *Clayton and Mayeda*, 1996, 1999) for $\Delta^{17}\text{O}$, where $\Delta^{17}\text{O} = \delta^{17}\text{O} - 0.52 \times \delta^{18}\text{O}$.

the solar system's history and from which the meteorite was recently liberated at a time measured by its recent cosmic-ray-exposure age.

We first detail the number of postulated distinct parent bodies. This is followed by a short discussion on the dynamical issues for delivering meteorites to Earth. We then discuss the telescopic and spacecraft data used to identify meteoritic parent bodies and some proposed asteroid-meteorite linkages. We conclude the chapter with a discussion of the work that needs to be done to better identify meteoritic parent bodies.

2. THE DIVERSITY OF METEORITES AND THEIR PARENT BODIES

Meteorites provide the most tangible evidence of the chemical and physical makeup of asteroids, the processes by which asteroids were formed and modified during the history of the solar system, and the number of distinct asteroidal bodies for which we currently have samples. Classification of the more than 22,500 known meteorites (Grady, 2000) provides a means for grouping samples formed from common constituents and processes and, more importantly for the purposes of this discussion, from a likely common parent body.

The bulk composition, mineralogy, and petrology of a meteorite are functions of the original bulk composition of its parent body and the amount of heating and melting it has experienced. The precursor assemblages ranged from highly reduced to highly oxidized material. Meteorites can be broken into two types: those that experienced heating but not melting (chondrites) and those that experienced melting and differentiation (achondrites, primitive achondrites, stony-irons, irons). A more detailed discussion of chondritic meteorites can be found in *Brearley and Jones (1998)* and a more complete discussion of differentiated meteorites can be found in *Mittlefehldt et al. (1998)*.

Chondrites are the most primitive material in the solar system. Mild thermal metamorphism and aqueous alteration has not destroyed the nebular components they contain or erased the record of their formation in the solar nebula ca. 4.56 Ga. On the other hand, differentiated meteorites have experienced significant degrees of melting, leading to an erasing of most of the evidence of their chondritic precursors.

2.1. Classification

Meteorites that are similar in terms of petrologic, mineralogical, bulk-chemical, and isotopic properties are separated into groups (Table 1). Particularly important parameters for the classification of silicate-bearing meteorites include refractory lithophile ("oxygen-loving") elements (e.g., Ca, Al, Ti), the FeO concentration in olivine ($(\text{Fe,Mg})_2\text{SiO}_4$) and orthopyroxene ($(\text{Fe,Mg})_2\text{Si}_2\text{O}_6$), and the whole-rock O-isotopic composition (Figs. 1 and 2). Iron meteorites are classified according to siderophile ("iron-loving") element (Ga, Ge, Ir, Ni) concentrations. It is noteworthy that the proper-

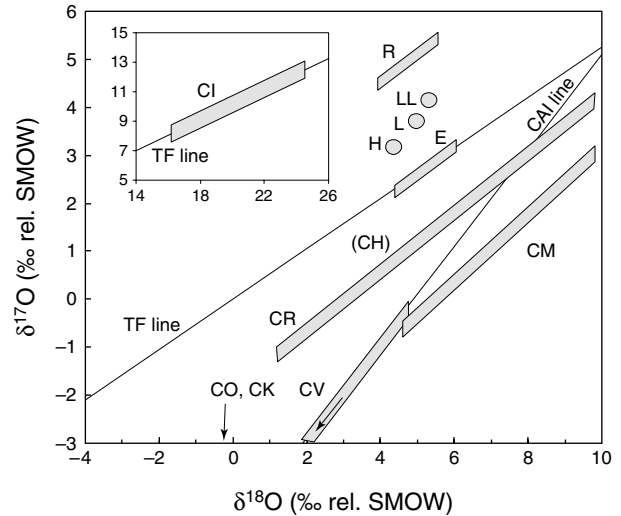


Fig. 1. Oxygen-isotopic compositions relative to standard mean ocean water (SMOW) for the chondritic meteorite groups. The terrestrial fractionation (TF) line and the refractory inclusion (CAI) mixing line are also plotted. CI chondrites are plotted in the inset in the upper left part of the diagram. The CO and CK chondrites plot off the diagram along the CAI line. The label for the CH chondrites is in parentheses because the CH chondrites occupy only part of the CR chondrite region. The region labeled E contains both the EH and EL chondrites. Figure revised from *Meibom and Clark (1999)*.

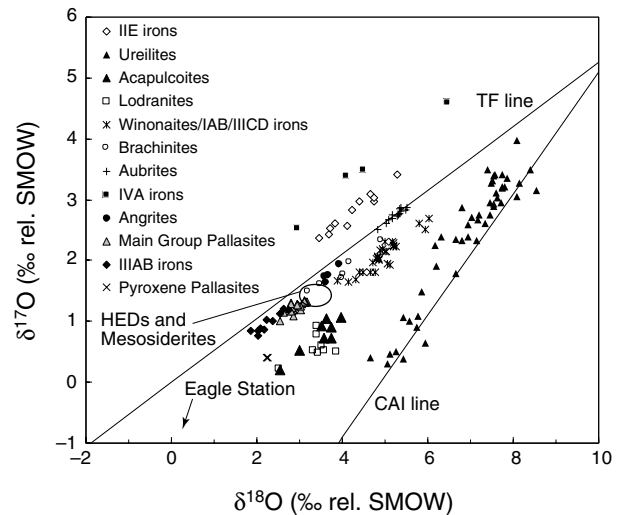


Fig. 2. Oxygen-isotopic compositions relative to standard mean ocean water (SMOW) for the differentiated meteorite groups plus the members of the pyroxene-pallasite grouplet. The terrestrial fractionation (TF) line and the refractory inclusion (CAI) mixing line are also plotted. Members of the Eagle Station pallasite grouplet ($\delta^{18}\text{O} \approx -2.80\text{‰}$, $\delta^{17}\text{O} \approx -6.15\text{‰}$) plot off the diagram. The standard deviations for $\delta^{18}\text{O}$ and $\delta^{17}\text{O}$ are $\sim 0.1\text{‰}$ for individual points. Figure revised from *Meibom and Clark (1999)*.

TABLE 2. Meteorite groups and their postulated parent or source bodies.

Group	Fall Percentage (%) [*]	Postulated Parent or Source Bodies [†]
L	38.0	S(IV) asteroids (<i>Gaffey et al.</i> , 1993)
H	34.1	6 Hebe [S(IV)] (<i>Gaffey and Gilbert</i> , 1998)
LL	7.9	S(IV) asteroids (<i>Gaffey et al.</i> , 1993)
Irons	4.2	M asteroids (<i>Cloutis et al.</i> , 1990; <i>Magri et al.</i> , 1999)
Eucrites	2.7	4 Vesta (V) (<i>Consolmagno and Drake</i> , 1977; <i>Drake</i> , 2001)
Howardites	2.1	4 Vesta (V) (<i>Consolmagno and Drake</i> , 1977; <i>Drake</i> , 2001)
CM	1.7	19 Fortuna (G, Ch) (<i>Burbine</i> , 1998)
Diogenites	1.2	4 Vesta (V) (<i>Consolmagno and Drake</i> , 1977; <i>Drake</i> , 2001)
Aubrites	1.0	3103 Eger (E) (<i>Gaffey et al.</i> , 1992)
EH	0.8	M asteroids (<i>Gaffey and McCord</i> , 1978)
EL	0.7	M asteroids (<i>Gaffey and McCord</i> , 1978)
Mesosiderites	0.7	M asteroids (<i>Gaffey et al.</i> , 1993)
CV	0.6	K asteroids (<i>Bell</i> , 1988)
CI	0.5	C asteroids (<i>Gaffey and McCord</i> , 1978)
CO	0.5	221 Eos (K) (<i>Bell</i> , 1988)
Pallasites	0.5	A asteroids (<i>Cruikshank and Hartmann</i> , 1984; <i>Lucey et al.</i> , 1998)
Ureilites	0.5	S asteroids (<i>Gaffey et al.</i> , 1993)
“Martian”	0.4	Mars (<i>McSween</i> , 1994)
CR	0.3	C asteroids (<i>Hiroi et al.</i> , 1996)
CK	0.3	C asteroids (<i>Gaffey and McCord</i> , 1978)
Acapulcoites	0.1	S asteroids (<i>McCoy et al.</i> , 2000)
Angrites	0.1	S asteroids (<i>Burbine et al.</i> , 2001a)
Lodranites	0.1	S asteroids (<i>Gaffey et al.</i> , 1993; <i>McCoy et al.</i> , 2000)
R	0.1	A or S asteroids
Winonaites	0.1	S asteroids (<i>Gaffey et al.</i> , 1993)
(Tagish Lake) [‡]	0.1	D asteroids (<i>Hiroi et al.</i> , 2001)
Brachinites	Only finds	A asteroids (<i>Cruikshank and Hartmann</i> , 1984; <i>Sunshine et al.</i> , 1998)
CH	Only finds	C or M asteroids
“Lunar”	Only finds	Moon (<i>Warren</i> , 1994)

^{*}Fall percentages are calculated from the 942 classified falls that are listed in *Grady* (2000), *Grossman* (2000), and *Grossman and Zipfel* (2001).

[†]Asteroid classes are a combination of those of *Tholen* (1984), *Gaffey et al.* (1993), and *Bus* (1999).

[‡]Tagish Lake is a newly discovered type of carbonaceous chondrite (*Brown et al.*, 2000) and is listed in the table because of its spectral similarity to D asteroids.

ties most useful for classifying meteorites are not readily measured by Earth-based remote-sensing techniques.

Oxygen-isotopic data (Figs. 1 and 2) are very useful for distinguishing different meteorite groups because samples from the same parent body will cluster together or tend to fall on a mass-fractionation line with a slope of 0.52 (e.g., *Clayton*, 1993). The exceptions are the CK, CO, and CV chondrites, which fall on a line with a slope of ~1 due to the mixing of two distinct nebular components (*Clayton*, 1993). We note that although a common O-isotopic composition is often taken as evidence of a common parent-body origin, it only requires formation in a similar region and does not preclude separate parent bodies.

In general, meteorite groups contain five or more members. Grouplets contain two to four members. Meteorites that do not fit in any well-defined group are termed anomalous or ungrouped. Each group is generally thought to represent a distinct parent body, although some cases exist where multiple meteorite groups apparently originated from

a common body (Table 1). It is more difficult to interpret the origin of anomalous or ungrouped meteorites, but many of these are thought to be the only samples of their respective parent bodies in our meteorite collections.

2.2. Chondrites

Currently, 13 groups (Table 1) of chondritic meteorites have been defined, largely based on refractory lithophile element abundances and O-isotopic compositions. Chondrites, particularly ordinary chondrites (~80% of all falls), dominate the current flux of meteorite landings (Table 2). The classification of ordinary and enstatite chondrites has changed little in the past 10 years, with ordinary chondrites divided into the classic H, L, and LL groups, and enstatite (FeO-poor pyroxene) chondrites divided into the EH and EL groups on the basis of total Fe content.

In contrast, several new groups of carbonaceous chondrites have been defined. In addition to the traditional CI,

CM, CO, and CV groups, researchers now recognize the CK group of thermally metamorphosed carbonaceous chondrites and the metal-rich CH and CR groups. Also recently recognized are the highly oxidized R chondrites (e.g., Schulze et al., 1994). Chondritic groups are subdivided according to petrologic type (1–6), with 1 being the most aqueously altered, 3.0 being the least altered, and 6 being the most thermally altered. Some CI1 and CM2 chondrites have also undergone some late-stage thermal metamorphism where heating to 500°–700°C (Lipschutz et al., 1999) has dehydrated their phyllosilicates.

In addition to the 13 well-defined groups, ~14 chondritic grouplets or unique meteorites (Meibom and Clark, 1999; Brown et al., 2000; Weisberg et al., 2001) have been recognized. Thus, the full range of chondritic meteorites probably requires at least 27 distinct parent bodies.

2.3. Differentiated Meteorites

In contrast to chondrites, the differentiated meteorites (Table 1) represent a much larger number of meteorite parent bodies. The differentiated meteorites range from those that experienced only limited differentiation (primitive achondrites) to those (achondrites, stony-irons, irons) that were produced by extensive melting, melt migration, and fractional crystallization. Although classification of differentiated meteorites is relatively straightforward, igneous differentiation by its very nature can produce radically different lithologies of a common parent body.

2.3.1. Primitive achondrites. Among the primitive achondrites, the acapulcoites and lodranites appear to have originated on a single parent body, as evidenced by the transitional nature of some members of these groups. The winonaites, which tend to be similar in mineralogy and chemistry to acapulcoites and lodranites, appear to sample another parent body based on distinct O-isotopic compositions.

2.3.2. Achondrites. Among the fully differentiated achondrites, the basaltic angrites (containing a TiO₂-rich augite), the ultrareduced, pyroxenitic aubrites, the olivine-dominated brachinites, and the C-rich ureilites each require a unique parent body. The basaltic eucrites and orthopyroxenitic diogenites appear to sample a common parent body, as evidenced by the occurrence of polymict breccias known as howardites, which contain both eucritic and diogenitic material. These meteorites are referred to as the HEDs. However, the recent discovery (Yamaguchi et al., 2002) of a eucrite with an O-isotopic value very different (near the CR region) from the HEDs argues for the formation of another HED-like body in the belt.

2.3.3. Stony-iron meteorites. Stony-iron meteorites include the pallasites and mesosiderites. All pallasites are composed primarily of centimeter-sized olivine grains imbedded in metallic Fe. They differ in terms of O-isotopic composition, olivine composition, and pyroxene abundance, and these features have been used to delineate the main group, the Eagle Station grouplet, and the pyroxene-pallasite grouplet, each of which requires a separate parent body.

Pallasites are generally thought to be fragments of the core-mantle boundary.

Mesosiderites are breccias composed of HED-like clasts of basaltic to orthopyroxenitic material mixed with metallic clasts. They likely formed by impact mixing of a core fragment on the surface of a basaltic asteroid. Their O-isotopic compositions are indistinguishable from the HEDs. Whether they originated on the same parent body as the HEDs or sample yet another basaltic asteroid remains uncertain.

2.3.4. Iron meteorites. The number of types of differentiated meteorites is roughly doubled by 13 groups of iron meteorites defined by siderophile-element (Ga, Ge, Ir, Ni) compositions. The large differences in the most volatile siderophile elements (Ga and Ge) between groups suggest that each iron group formed in a separate parent body. Evidence that these irons existed in cores include fractional crystallization trends suggestive of prolonged cooling in 10 of the groups (excluding IAB, IIE, and IIICD) and the subsolidus exsolution of Ni-rich and Ni-poor phases (the familiar Widmanstätten pattern) requiring cooling of 1–100 K/m.y. at low temperatures.

The IAB, IIE, and IIICD irons all contain abundant silicate inclusions and do not display well-developed fractional crystallization trends. The IAB irons and primitive achondritic winonaites appear to sample a common parent body and also have been linked with the IIICD irons (e.g., Mittlefehldt et al., 1998). H chondrites and silicate inclusions in IIE irons share similar bulk and mineral compositions, textures, and O-isotopic compositions (Olsen et al., 1994; Casanova et al., 1995) and appear to be related.

The largest number of distinct parent bodies is probably sampled by the ungrouped irons, which numbered 95 in Grady (2000) and account for ~10% of known irons. Wasson (1995) argues that the ungrouped irons required ~70 distinct parent bodies. However, some of these ungrouped irons are probably related to the 13 major groups, perhaps representing an extreme composition produced by fractional crystallization from a poorly represented parent body. We suggest that these ungrouped irons more likely represent ~50 distinct parent bodies.

The large number of ungrouped irons appears to be due to the strength of metallic Fe and its resistance to terrestrial weathering compared to silicates. These two factors allow for more parent bodies rich in metallic Fe to be sampled in our meteorite collections. Irons have cosmic-ray exposure ages ranging from hundreds of millions to a few billion years (e.g., Voshage and Feldman, 1979) while stony meteorites tend to have much shorter exposure ages of <100 m.y. (e.g., Marti and Graf, 1992; Scherer and Schultz, 2000). (Cosmic-ray exposure ages record the time an object has spent as a meter-sized (or less) body in space or within a few meters of the surface.) The greater physical strength of irons allows them to survive as meter-scale objects in space much longer than silicate bodies. The relative resistance of irons to terrestrial weathering increases the probability that iron meteorites that fall to Earth will survive to the present day.

2.4. Number of Parent Bodies

In total, our meteorite collections could represent as few as ~100 distinct asteroidal parent bodies (~27 chondritic, ~2 primitive achondritic, ~6 differentiated achondritic, ~4 stony-iron, ~10 iron groups, ~50 ungrouped irons) or perhaps as many as 150, if assumed relationships between ungrouped iron meteorites prove untrue. This number is evolving because of thousands of new meteorites collected each year, primarily from Antarctica and the Sahara, resulting in a continuous supply of new samples that expand our perception of the diversity of the asteroid belt.

It is interesting to note the enormous disparity between the 100–150 meteorite parent bodies sampled on Earth and the approximately 1,000,000 asteroids (e.g., *Ivezić et al.*, 2001) in the main belt with diameters greater than 1 km. Although we cannot expect to have sampled the vast number of asteroids in their entirety, two explanations for this discrepancy are worth noting. First, meteorite researchers focus on “parent bodies,” the primordial asteroids as they existed in the first tens of millions of years of solar system history. In contrast, asteroid researchers study fragments produced by 4.5 b.y. of impact and fragmentation. A single “parent body” may have produced tens, hundreds, or thousands of current asteroids. Secondly, although we commonly refer to a single parent body (e.g., the H-chondrite parent body), we have no direct evidence that rules out multiple asteroids composed of essentially identical material. Thus, our meteorite collections may well sample a large percentage of the types of materials present in the asteroid belt, despite the apparent mismatch.

We also do not understand how biased our meteorite collection is compared to the asteroid belt as a whole. Many types of carbonaceous chondritic material may be too weak (e.g., *Sears*, 1998) to be able to make it through the atmosphere to Earth’s surface as meteorites and may only be sampled as IDPs. It is also unclear, as discussed in the next section, how well the dynamical mechanisms that deliver meteorites to Earth sample the asteroid belt.

3. ASTEROIDS TO EARTH: THE DYNAMIC CONNECTION

Since the review of meteorite transport in *Asteroids II* (*Greenberg and Nolan*, 1989), advances in computer speed and numerical algorithms have produced several surprising revelations. *Wetherill* (1985) showed that the expected flux from continual interasteroid collisions in the main belt could be sufficient to push the necessary mass of ejected meteorite-sized bodies into the main orbital resonances. These resonances would then be responsible for increasing the orbital eccentricity to planet-crossing values, allowing some fraction of the material to impact Earth. However, modern computer power has allowed *Farinella et al.* (1994) to show that the v_6 resonance (Fig. 3) could rapidly raise orbital eccentricities to 1, leading to collision with the Sun. *Gladman et al.* (1997) show this result to be generic for all main orbital resonances in the inner belt out to the 3:1 resonance

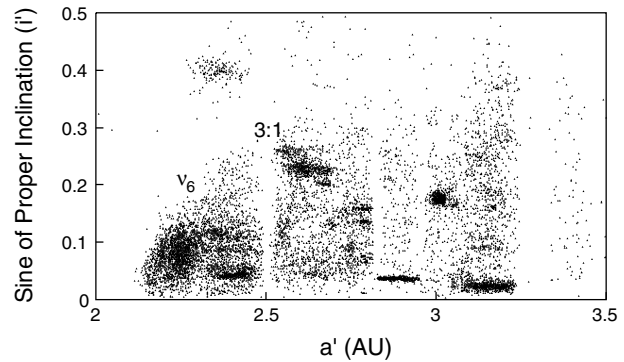


Fig. 3. Proper semimajor axis (a') (AU) vs. sine of proper inclination (i') for the first ~10,000 numbered main-belt asteroids. The 3:1 and v_6 resonances are labeled.

at 2.5 AU (Fig. 3). Outside 2.5 AU, most emerging meteoroids fall under the control of Jupiter and are ejected from the solar system, strongly reducing the fractional contribution from the outer belt to the meteorite flux at Earth.

Dynamical studies of meteorite delivery are most easily constrained by the cosmic-ray-exposure ages measured in falls. *Morbidelli and Gladman* (1998) extend the previous dynamical studies by calculating the entry geometries, speeds, and delivery times from the main inner-belt resonances and show that the distribution of radiants and entry speeds is in good agreement with that determined by the fireball camera networks. However, the delivery timescales calculated were typically 2–4 m.y., which is 3–10× shorter than that recorded in most ordinary chondrites. This disagreement is serious and robust; the dynamical calculations are now direct approximation-free N-body integrations, and similar studies on the transport of lunar and martian meteorites (*Gladman et al.*, 1996) match the spectrum of transfer ages to Earth from these two source bodies extremely well. Therefore, the inescapable conclusion is that the “classic” scenario of meteorite delivery is incorrect: Meteorites (especially ordinary chondrites) are not delivered to Earth by being liberated by large collisions in the main belt and directly injected into the belt’s orbital resonances for transport to the inner solar system. So how do meteorites arrive at Earth? Where are their source bodies, and can we learn about their parent bodies from dynamical studies? One possibility is that most source bodies are near-Earth asteroids [NEAs; see *Morbidelli et al.* (2002)] distributed throughout terrestrial space. However, it is unlikely that the meteoroid mass flux off NEAs (which are in a less-collisional environment) can rival that of the entire main belt, and *Morbidelli and Gladman* (1998) showed that the orbital distribution of fireballs does not match an NEA source. Therefore, the major chondrite groups almost certainly have their source bodies in the main belt.

A main-belt source for meteorites (especially ordinary chondrites) requires that the bulk of their cosmic-ray exposure occurs after collisional liberation of the meteoroids but before they reach the main resonant “escape hatches.” The

most-developed models for doing this use the Yarkovsky effect, which causes a slow drift in semimajor axis due to the anisotropic emittance of thermal radiation of a meteoroid (see *Bottke et al.*, 2002). In this context, meteoroid orbits spiral in or out by a distance determined by their collisional lifetimes, producing a sampling zone around the resonances, which evacuate them from the belt. Note that this process allows the ejection velocities during collisional events that liberate meteorites to be effectively zero (in contrast to the classic picture that required meteoroids to be thrown out at hundreds of meters per second to reach the resonances). This allows even small (kilometer-scale) asteroids to be source bodies (*Vokrouhlický and Farinella*, 2000).

The number of source bodies for the various meteorite classes is a problem perhaps best illustrated by using the HEDs as an example. If the widely accepted link between HEDs and Vesta is correct, given that Vesta is almost as far as possible from any orbital resonance (being midway between the ν_6 and 3:1), it would seem that identifying “actual” source bodies is a hopeless quest because all main-belt bodies can be sampled. But perhaps Vesta is not the source body for the HEDs and is rather just the parent body of the “vestoids,” which are closer to the resonances and easier to sample. However, distinct peaks in the cosmic-ray-exposure age distribution among the howardites, eucrites, and diogenites (*Welten et al.*, 1997) argue for major impacts on one body (4 Vesta) ejecting all three types of material.

The contribution to the meteoroid flux from big and small parent bodies depends on the size-frequency distribution of objects in the main belt down to subkilometer sizes, the physics of collisional disruption, and the relative importance of slow transport mechanisms like the Yarkovsky effect or orbital diffusion. If the most sophisticated size-frequency distribution models (e.g., *Durda et al.*, 1998) are correct, *Bottke et al.* (2000) found that the flux of meteorite-sized ejecta produced by the largest asteroids (with the largest collisional cross sections) should dominate the flux produced by the smaller asteroids (which lose nearly all their ejecta due to low escape velocities). Hence, many of the chondrites and HEDs falling on Earth today may ultimately be derived from a few large source objects (e.g., 4 Vesta and 6 Hebe), despite that fact that nearly all main-belt asteroids can potentially provide meteorite samples to Earth. If true, the peaks seen in histograms of cosmic-ray-exposure ages for many meteorite groups would represent individual impact events on large asteroids, while the background continuum would represent meteorites produced by numerous smaller impacts occurring across the main belt. If not, then the peaks are due to the large relative mass contribution from a recent major disruption of a “medium-sized” asteroid that was well situated for efficient delivery.

4. DETERMINING METEORITE PARENT BODIES

From analyzing a meteorite in the laboratory, we can learn a variety of details on its composition, the history of its parent body, and its passage through space. However,

until samples are returned to Earth, all compositional measurements of asteroids will need to be done through observations using telescopes and spacecrafts.

4.1. Telescopic Data

Compositional data on asteroids are usually derived from the analysis of sunlight reflected from their surfaces. Many minerals (e.g., olivine, pyroxene) have diagnostic absorption bands in the visible and NIR. Spectral surveys (e.g., *Zellner et al.*, 1985; *Bus*, 1999) are primarily done in the visible (~ 0.4 – $1.1 \mu\text{m}$) due to the peaking of the illuminating solar flux in the visible and the relative transparency of the atmosphere at these wavelengths. More than 2000 asteroids have been observed in the visible. CCD detectors now allow for objects as small as a few hundred meters in near-Earth orbit and a few kilometers in the main belt to be observed by Earth-based telescopes.

Asteroids are generally grouped into classes based on their visible spectra (~ 0.4 to ~ 0.9 – $1.1 \mu\text{m}$) and visual albedo (when available). The most widely used taxonomy (*Tholen*, 1984) classifies objects observed in the eight-color asteroid survey (ECAS) (*Zellner et al.*, 1985). *Bus* (1999) develops an expanded taxonomy with many more classes and subclasses to represent the diversity of spectral properties found in his CCD spectra of more than 1300 objects. However, NIR observations (e.g., *Gaffey et al.*, 1993; *Rivkin et al.*, 1995, 2000) of a few hundred objects show that most asteroid classes contain a variety of surface assemblages.

Reflectance spectra of an asteroid can be directly compared to spectral data obtained on tens-of-milligram-sized to gram-sized samples of meteorites. A few hundred meteorites have had their spectra measured. *Gaffey* (1976) shows that different meteorite types tend to have distinctive spectra from 0.3 to 2.5 μm . Besides mineralogy, the shapes and depths of absorption bands and the spectral slope are a function of many other surface parameters including particle size (e.g., *Johnson and Fanale*, 1973) and temperature (e.g., *Singer and Roush*, 1985; *Hinrichs et al.*, 1999). It has also been proposed (e.g., *Chapman*, 1996; *Sasaki et al.*, 2001; *Hapke*, 2001) that the optical properties of an asteroidal surface will change over time due to processes such as micrometeorite impacts and sputtering due to solar wind.

4.2. Spacecraft Data

The best way to identify the parent body of a meteorite is through a spacecraft mission. Meteorites from the Moon have been identified (e.g., *Warren*, 1994) from lunar samples retrieved by Apollo astronauts, since these samples have distinctive petrologic and chemical (e.g., bulk Mn:Fe) properties. The most conclusive evidence that meteorites originate from Mars is the finding (e.g., *Bogard and Johnson*, 1983) that the measured abundances and isotopic ratios of trapped noble gases in glasses in these meteorites are similar to those measured for the martian atmosphere by the *Viking* landers. Other arguments for linking these meteorites to Mars can be found in *McSween* (1994).

Spacecraft have visited a number of S-type asteroids (e.g., 243 Ida, 433 Eros, 951 Gaspra) and one C-type asteroid (253 Mathilde). Spacecraft are able to obtain data that is difficult to impossible to obtain from Earth, including high-resolution images, reflectance spectra of different lithologic units, bulk densities, magnetic field measurements, and bulk-elemental compositions. For determining meteoritic parent bodies, bulk-elemental compositions are probably the most important pieces of information that can be obtained because the data can be directly compared to meteorite bulk compositions (e.g., *Jarosewich, 1990*). Only one previous mission (*NEAR Shoemaker*) has determined elemental ratios (*Nittler et al., 2001; Evans et al., 2001*) of the surface of an asteroid (433 Eros) using measurements of discrete-line X-ray and γ -ray emissions.

5. METEORITES AND POSSIBLE PARENT BODIES

The following sections will discuss a number of postulated asteroid-meteorite linkages for some of the most common meteorites to fall on Earth. A more complete list of postulated meteorite parent bodies are in Table 2.

5.1. Ordinary Chondrites and S Asteroids

Ordinary chondrites are composed (e.g., *Gomes and Keil, 1980; Brearley and Jones, 1998*) of abundant chondrules containing olivine, pyroxene, plagioclase feldspar, and glass with lesser amounts of an olivine-rich matrix, metal, and sulfides. Although grouped under the heading “ordinary chondrites,” they actually comprise three separate chemical groups (H, L, and LL) and a range of petrologic types. They range substantially in mineralogy (ol:px ratio of ~54:46 in H chondrites to ~66:34 in LL), metal concentration (~18 vol% in H chondrites to ~4% in LL), mineral composition (particularly the FeO concentrations in olivine and pyroxene),

and degree of thermal metamorphism (from virtually none in type 3 to extensive in type 6).

Spectrally, ordinary chondrites have features due to olivine and pyroxene (Fig. 4). LL chondrites, because of their higher olivine contents, have more distinctive olivine bands while H chondrites have more distinctive pyroxene bands. Even though ordinary chondrites contain significant amounts of metallic Fe, their spectra are not appreciably reddened (reflectances tending to increase with increasing wavelength) as found in other metallic Fe-rich meteorites.

S asteroids are the most abundant type of asteroid observed in the inner main belt. S asteroids have long been considered possible parent bodies of the ordinary chondrites, because a large number of these objects have spectral features due to both olivine and pyroxene. However, S asteroids are spectrally redder than ordinary chondrites (Fig. 4) and tend to have weaker absorption bands. It has long been unclear (e.g., *Wetherill and Chapman, 1988*) if these spectral difference are due to a compositional difference or an alteration process that can redden ordinary chondrite material. Some researchers (e.g., *Chapman, 1996*) argue that some percentage of the S asteroids have ordinary chondrite compositions. Others (*Bell et al., 1989*) believe that the ordinary chondrite parent bodies are found among small (diameters <10 km) objects and not among the S asteroids, which are believed to be primarily differentiated or partially differentiated bodies.

One asteroid (3628 Božněmcová) was originally announced (*Binzel et al., 1993*) as the first main-belt object with a visible spectrum similar to ordinary chondrites. However, NIR spectra (*Burbine, 2000; Binzel et al., 2001*) of Božněmcová (Fig. 4) show an unusual bowl-shaped 1- μ m feature unlike any measured meteorite spectrum from ~1.2 to 1.5 μ m.

Because of the diversity of assemblages found in the S-asteroid population, *Gaffey et al. (1993)* devised a classification system based on analyses of NIR spectral data (*Bell et al., 1988*). Using the ratio of the areas of Band II (2- μ m feature) to Band I (1- μ m feature) and the Band I centers, the S asteroids were broken into seven subtypes, S(I) through S(VII). S(I) objects have surfaces mineralogies dominated by olivine, and S(VII) objects have surfaces dominated by pyroxene. The pyroxene abundance tends to increase with increasing number of the subtype.

Only a subset (~25%) of the measured S asteroids, designated as S(IV) objects, fall within the region defined by the ordinary chondrites. However, a few other meteorite types will also plot within this region. Primitive achondrites such as lodranites and acapulcoites tend to overlap the most pyroxene-rich part of the ordinary chondrite area (*Burbine et al., 2001b*). Some ureilites have compositions (e.g., *Mittlefehldt et al., 1998*) consistent with falling in this region.

S(IV) asteroids include many of the largest asteroids in the belt including 3 Juno, 6 Hebe, 7 Iris, and 11 Parthenope. Hebe is an often-discussed candidate due to its location near the 3:1 and v_6 resonances; if ejecta speed distribution favors production from large bodies, then Hebe could be a major contributor to the terrestrial meteorite flux (*Farinella et al.,*

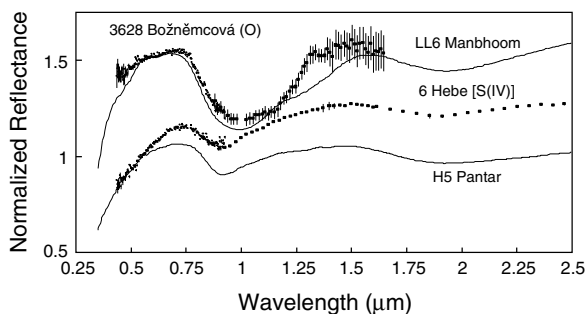


Fig. 4. Normalized reflectance vs. wavelength (μ m) for S-type 6 Hebe vs. H5 chondrite Pantar and O-type 3628 Božněmcová vs. LL6 chondrite Manbhoom. All meteorite spectra are from *Gaffey (1976)*. All spectra are normalized to unity at 0.55 μ m. The asteroid spectra are offset by 0.5 in reflectance. Visible asteroid data (points) are from *Bus (1999)*. Near-infrared data (dark squares) for Hebe are from *Bell et al. (1988)* and Božněmcová are from *Burbine (2000)* and *Binzel et al. (2001)*.

1993). Mineralogies derived from visible and NIR spectra of Hebe (Gaffey and Gilbert, 1998) appear consistent with H chondrites.

However, Hebe is spectrally redder than measured ordinary chondrites (Fig. 4). Gaffey and Gilbert (1998) argue that the reddening on Hebe is due to “red” metallic Fe similar to that found in iron meteorites. Assuming a very coarse metallic Fe component, they propose that a surface mixture of ~60% H-chondrite material and ~40% metallic Fe would duplicate Hebe’s spectral characteristics. They argue that the existence of H6 chondrite Portales Valley, which contains numerous metallic veins with a distinctive Widmanstätten pattern (Kring et al., 1999), and H-chondrite silicate inclusions in IIE irons are evidence that the H-chondrite parent body contains numerous metallic regions. Other researchers (e.g., Sasaki et al., 2001; Hapke, 2001) propose that the spectral reddening is due to some type of surface alteration processes (e.g., micrometeorite impacts or solar wind sputtering) believed to produce vapor-deposited coatings of nanophase Fe to redden the spectra. Another chapter (Clark et al., 2002) discusses “space weathering” mechanisms on asteroids.

However, the best spectral matches to ordinary chondrites are in the near-Earth population (Binzel et al., 2002). For example, 1862 Apollo has a visible spectrum (McFadden et al., 1985) similar to an LL chondrite and was classified as a Q type by Tholen (1984). In their spectral survey of NEAs, Binzel et al. (2001) discovered that one-third of their observed objects resembled ordinary chondrites.

Binzel et al. (2001) also found that there is a continuum of spectral properties between the S asteroids and the ordinary chondrites among the Q- and S-type NEAs in the visible and NIR. This spectral continuum may be related to surface gravity and/or surface age, because the NEAs are much smaller and should have much younger surfaces (on average) than main-belt objects. These observations argue that only “fresher” asteroidal surfaces would resemble ordinary chondrites.

It was hoped that many of these questions on the composition of S asteroids would be answered by the NEAR Shoemaker mission to S(IV) asteroid 433 Eros. The average olivine-to-pyroxene composition derived from band area ratios (McFadden et al., 2001) and elemental ratios (Mg:Si, Fe:Si, Al:Si, and Ca:Si) derived from X-ray measurements (Nittler et al., 2001) of Eros are consistent (McCoy et al., 2001) with ordinary chondrites. However, the S:Si ratio derived from X-ray data (Nittler et al., 2001) and the Fe:O and Fe:Si ratios derived from γ -ray data (Evans et al., 2001) are significantly depleted relative to ordinary chondrites. McCoy et al. (2001) found that the best meteoritic analogs to Eros are an ordinary chondrite, whose surface chemistry has been altered by the depletion of metallic Fe and sulfides, or a primitive achondrite, derived from a precursor assemblage of the same mineralogy as one of the ordinary chondrite groups. They note that the biggest obstacle in identifying a meteoritic analog for Eros is the lack of understanding (e.g., McKay et al., 1989) of the nature of the chemical and physical processes that affect asteroid regoliths.

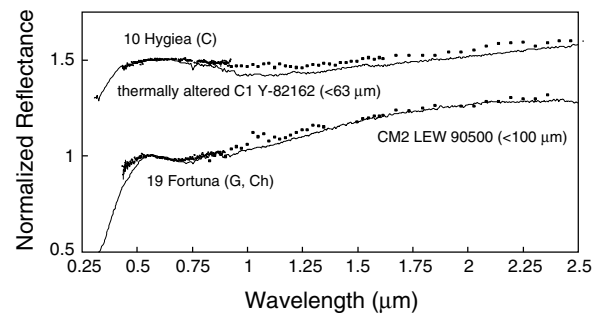


Fig. 5. Normalized reflectance vs. wavelength (μm) for G-type (also classified as a Ch-type) 19 Fortuna vs. CM chondrite LEW 90500 (Burbine, 1998) and C-type 10 Hygiea vs. thermally altered CI chondrite Y-82162 (Hiroi et al., 1993). All spectra are normalized to unity at $0.55 \mu\text{m}$. The asteroid spectra are offset by 0.5 in reflectance. Visible asteroid data (points) are from Bus (1999). Near-infrared asteroid data (dark squares) are from Bell et al. (1988).

Returned samples are needed to answer these questions concerning the properties of asteroidal regoliths.

5.2. CM Chondrites and C-type Asteroids

CM chondrites are typically composed of small chondrules set in an aqueously altered matrix (e.g., Buseck and Hua, 1993; Brearley and Jones, 1998) of Fe^{3+} -bearing phyllosilicates (e.g., cronstedtite, greenalite) and tochilinite (alternating layers of a sulfide and a hydroxide). The chondrules themselves tend to be FeO-poor, while the phyllosilicates tend to be FeO-rich. In some cases, alteration has been considerably more extensive, leading to aqueous alteration of the chondrules and the replacement of tochilinite but preserving the chondritic structure (Zolensky et al., 1997). CM chondrites typically contain 6–12 wt% water (Jarosewich, 1990).

Spectrally, CM chondrites (Fig. 5) have low visible albedos, (~ 0.04), relatively strong UV features, a number of weak features between 0.6 and $0.9 \mu\text{m}$, and relatively featureless spectra from 0.9 to $2.5 \mu\text{m}$ (Johnson and Fanale, 1973; Gaffey, 1976; Vilas and Gaffey, 1989). CM chondrites have $3\text{-}\mu\text{m}$ features with average band strengths of $\sim 45\%$ (Jones, 1988).

The strongest ($\sim 5\%$ band strength) of the weak features between 0.6 and $0.9 \mu\text{m}$ is a band centered at $\sim 0.7 \mu\text{m}$, attributed to an absorption due to ferric Fe (Fe^{3+}) in the phyllosilicates. This $0.7\text{-}\mu\text{m}$ band has been found in a number of CM chondrite spectra (Vilas and Gaffey, 1989; Burbine, 1998), but not in CI- or CR-chondrite spectra.

CM chondrites have been generally linked to the C-type asteroids because they also have low albedos and relatively featureless spectra (Fig. 5) from low-resolution photometric data. Burbine (1998) identified two asteroids (13 Egeria and 19 Fortuna) as possible CM-chondrite parent bodies. These asteroids have a $0.7\text{-}\mu\text{m}$ band, similar spectral slopes to CM chondrites out to $2.5 \mu\text{m}$, and relatively strong $3\text{-}\mu\text{m}$ features (strengths of $40 \pm 15\%$ for Egeria and $25 \pm 6\%$ for

Fortuna) (Jones *et al.*, 1990). Egeria and Fortuna are classified as G asteroids by Tholen (1984) but as Ch objects by Bus (1999) on the basis of having a 0.7- μm band in his higher-resolution spectra.

However, Bus (1999) finds that almost half of the C-type asteroids observed in the main belt have this 0.7- μm feature. What makes Egeria and Fortuna likely candidates as CM-chondrite parent bodies is that they are the two largest objects (diameters >200 km) with this feature, and both are located relatively near the 3:1 resonance at ~ 2.5 AU. For samples to survive passage to Earth, the parent body or bodies of the CM chondrites would have to be relatively near a meteorite-supplying resonance due to their very fragile nature (Scherer and Schultz, 2000). CM chondrites have relatively low cosmic-ray-exposure ages (<7 m.y.) (e.g., Eugster *et al.*, 1998) with approximately half the CMs having ages <1 m.y., consistent with the expectation that they could not survive long in space.

Hiroi *et al.* (1993, 1996) measured the spectra of a number of CI and CM chondrites that have undergone late-stage thermal metamorphism. These meteorites and laboratory-heated CM chondrite material tend to have UV features that are weaker than “typical” carbonaceous chondrites and no ~ 0.7 - μm band, which disappears at temperatures of $\sim 400^\circ\text{C}$. Heating also tends to weaken the strength of the 3- μm feature.

Hiroi *et al.* (1993, 1996) find that the weaker UV and 3- μm features were similar to those found for a number of large C-type asteroids such as 10 Hygiea (Fig. 5) and 511 Davida. They suggest that these bodies might have been heated sufficiently to destroy some of their aqueous alteration products. The rarity of thermally altered carbonaceous chondrite material in our meteorite collections could be due to these bodies being located too far from meteorite-supplying resonances.

5.3. HEDs and 4 Vesta

The V-type asteroid 4 Vesta, the vestoids, and Vesta’s relationship to the basaltic achondrites (HEDs) are discussed in detail in Keil (2002). The one question that will be discussed here is whether other bodies like Vesta have existed in the belt.

According to arguments put forth by Consolmagno and Drake (1977), Vesta appears to be the parent body of the HEDs because it is the only large (few-hundred-kilometer) body surviving with an intact “basaltic” crust. Spectrally, Vesta is most similar to a howardite (Hiroi *et al.*, 1994), consistent with a surface mixture of eucritic and diogenitic material. The much smaller (~ 10 -km-sized) vestoids (e.g., Binzel and Xu, 1993) have been found in the Vesta family and between Vesta and the 3:1 and the v_6 resonances, consistent with derivation from Vesta.

However, the grouped and ungrouped irons imply the formation and later disruption of at least 50 differentiated parent bodies. Recent observational and meteoritical evidence now unequivocally supports the existence of at least one other “Vesta.” An object (1459 Magnya) with a “Vesta-like” spectrum (Lazzaro *et al.*, 2000) has been identified at

3.15 AU, too far to be easily related to Vesta at 2.36 AU. Magnya appears spectrally indistinguishable from inner-main-belt vestoids. The discovery (Yamaguchi *et al.*, 2002) of the eucrite with a very different O-isotopic value from the HEDs also confirms the formation of other “Vesta-like” bodies. Vesta appears to be the parent body of most HEDs and vestoids, but not all of them.

5.4. Iron Meteorites, Enstatite Chondrites, and M Asteroids

Iron meteorites are composed (e.g., Buchwald, 1975; Mittlefehldt *et al.*, 1998), primarily of metallic Fe with usually 5–20 wt% Ni. The variations in bulk Ni content for most meteorite groups is a few weight percent except for the IAB and IIICD irons, which have variations over 15 wt%. Accessory phases include troilite (FeS), schreibersite ((Fe,Ni)₃P), and graphite. IAB, IIICD, and IIE irons contain abundant inclusions of olivine, pyroxene, and plagioclase feldspar, while IVA irons contain abundant inclusions of trydymite (SiO₂) and pyroxene. The Widmanstätten pattern found in most irons is an oriented intergrowth of body-centered cubic α -Fe,Ni (<6 wt% Ni) (kamacite) and high-Ni (30–50 wt%) regions composed of a variety of phases.

Spectrally, iron meteorites (e.g., Cloutis *et al.*, 1990) have relatively featureless spectra (Fig. 6) with red spectral slopes and moderate albedos (~ 0.10 – 0.30). The reflectance properties of metal are functions of composition and grain size/surface roughness of the sample. Contrary to the conclusion of Gaffey (1976), Cloutis *et al.* (1990) find no simple correlation between Ni abundance and spectral redness in measured metallic Fe samples. This finding argues that it may not be possible to differentiate between different iron meteorite groups by their spectral properties. Cloutis *et al.* (1990) also found that the spectra of iron meteorites become redder with increasing grain size.

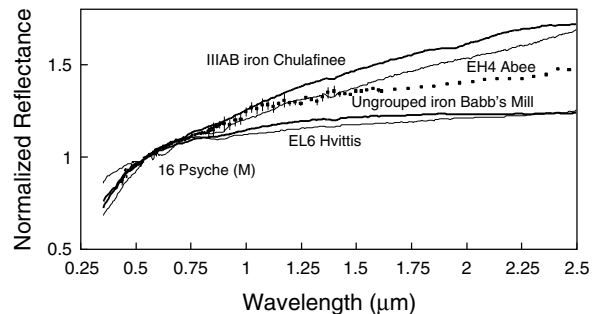


Fig. 6. Normalized reflectance vs. wavelength (μm) for M-type 16 Psyche vs. iron and enstatite chondrite meteorites (lines). All meteorite spectra are from Gaffey (1976). In order of increasing reflectance at 2.0 μm , the meteorites are EL6 chondrite Hvittis, ungrouped iron Babb’s Mill, EH4 chondrite Abee, and IIIAB iron Chulafinee. All spectra are normalized to unity at 0.55 μm . Visible asteroid data (points) are from Bus (1999). Near-infrared asteroid data (dark squares) are from Bell *et al.* (1988).

Enstatite chondrites also have relatively featureless spectra (Fig. 6) with red spectral slopes and moderate albedos (Gaffey, 1976). Besides enstatite, they are composed of FeO-poor chondrules, metal, and a range of sulfides. Like aubrites, they are extremely reduced, with virtually all iron occurring in the metallic form (~13–28 vol%) (Keil, 1968). They are divided into two groups (EH and EL) and exhibit a range of metamorphic types (3–6). These meteorites have no distinctive absorption features from 0.3 to 2.5 μm because of the absence of Fe^{2+} in their silicates.

M asteroids have moderate visual albedos (~0.10 to ~0.30), relatively featureless spectra, and red spectral slopes. Tholen (1989) identifies approximately 40 M asteroids. Because metallic Fe has the same reflectance characteristics in this wavelength region (Fig. 6), the M asteroids have been historically identified as the disrupted cores of differentiated objects that have had their silicates removed. However, enstatite chondrites have similar reflectance characteristics (Fig. 6) to metallic Fe and have also been proposed (e.g., Gaffey and McCord, 1978) as meteoritic analogs to the M asteroids. Complicating possible interpretations, Rivkin et al. (2000) find that more than one-third of observed M asteroids have 3- μm absorption features, implying hydrated silicates on the surfaces of the objects with weight percents of a few tenths of a percent. Further discussion on whether these absorption features on M asteroids really indicate hydrated assemblages can be found in Rivkin et al. (2002) and Gaffey et al. (2002).

Radar observations have been used to estimate the metal contents of asteroids. Radar albedos are functions of the near-surface bulk density, which is related to both the solid-rock density and the surface porosity of the object. M asteroids (e.g., 16 Psyche, 216 Kleopatra) tend to have higher radar albedos than C or S asteroids (Magri et al., 1999). M asteroids 6178 1986 DA (Ostro et al., 1991) and 216 Kleopatra (Ostro et al., 2000) have the highest observed radar albedos (~0.6–0.7), implying surfaces of metallic Fe with “lunarlike” porosities (35–55%) or solid enstatite chondritic material with little to no porosity. Without knowing the porosity of the surface, it is impossible to conclusively differentiate between these two types of assemblages. However, recent groundbased measurements of bulk densities, determined from astrometric observations, for M asteroids 16 Psyche (Viateau, 2000; Britt et al., 2002) and 22 Kaliope (Margot and Brown, 2001) are ~2 g/cm³. Their bulk densities are much lower than expected and may imply that these objects are not metallic in composition. If these objects are exposed Fe cores, they must be extremely porous with substantially more internal empty space than solid material (Britt et al., 2002).

6. CONCLUSIONS AND FUTURE WORK

In our meteorite collections, there is evidence for 100–150 asteroidal parent bodies. The “true” number is dependent on how well we can discern the chemical, mineralogical, petrologic, and isotopic characteristics of asteroidal-sized parent bodies from gram- to kilogram-sized meteorite samples.

It is very difficult to “conclusively” identify the parent bodies of even the most well-studied meteorite groups with current telescopic and spacecraft data. Asteroids and meteorites with similar spectral and mineralogical characteristics can be identified, and theoretical models for delivering fragments to Earth from these objects can be formulated. However, most postulated parent bodies are not unique spectrally so it is very difficult to rule out all other possible asteroids as parent bodies. Likely parent bodies can be easily identified, but it is very difficult to identify the “true” parent body in most cases. The obvious exception is 4 Vesta, which appears to be the parent body of almost all HEDs.

Making this problem even harder is our limited knowledge of the chemical and physical properties of asteroid regoliths. We do not understand the significance of spectral and chemical differences between meteorites measured in the laboratory and asteroids observed from Earth and by spacecraft.

How can we better answer these questions in the next decade?

1. Samples need to be returned to Earth from the upper (top few millimeters) and lower (tens of centimeters) surface layers of a variety of asteroid types to understand the optical, chemical, and physical properties of asteroid regoliths. *MUSES-C*, a Japanese sample return mission (Zolensky, 2000), is currently being prepared for launch to a near-Earth object and it is hoped that it will return ~1 g of material back to Earth.

2. Spacecraft orbiters and landers should be sent to taxonomic types (e.g., A, E, M, V) not yet visited by spacecrafts. These missions should focus on determining chemical compositions of the asteroids.

3. Near-infrared spectra from ~0.9 to 3.5 μm are needed to complement visible spectral surveys. Only with this extended wavelength coverage can we more accurately determine asteroid mineralogies.

4. Reflectance spectra of rare meteorite types need to be obtained.

5. More research needs to be done on the petrology of ungrouped irons to understand possible relationships with other meteorite groups.

6. Processes by which small meteoroids reach the resonances need to be explored in a model that considers all the meteorite types in a self-consistent scenario. In particular, how can fragile carbonaceous chondrites reach the resonances in sufficient numbers to provide the observed flux?

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