# Differentiated Planetesimals and the Parent Bodies of Chondrites

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## Abstract

Meteorites are samples of dozens of small planetary bodies that formed in the early Solar System. They exhibit great petrologic diversity, ranging from primordial accretional aggregates (chondrites), to partially melted residues (primitive achondrites), to once fully molten magmas (achondrites). It has long been thought that no single parent body could be the source of more than one of these three meteorite lithologies. This view is now being challenged by a variety of new measurements and theoretical models, including the discovery of primitive achondrites, paleomagnetic analyses of chondrites, thermal modeling of planetesimals, the discoveries of new metamorphosed chondrites and achondrites with affinities to some chondrite groups, and the possible identification of extant partially differentiated asteroids. These developments collectively suggest that some chondrites could in fact be samples of the outer, unmelted crusts of otherwise differentiated planetesimals with silicate mantles and metallic cores. This may have major implications for the origin of meteorite groups, the meaning of meteorite paleomagnetism, the rates and onset times of accretion, and the interior structures and histories of asteroids.

# 1. DEVELOPMENT OF THE PARENT BODY PARADIGM

Modern classification schemes divide meteorites into two broad lithologic types (Weisberg et al. 2006). Chondrites are composed of diverse silicate and metallic minerals that formed in the solar nebula. Their aggregational textures, near-solar composition, and petrologic and isotopic evidence for chemical disequilibrium indicate that they have never experienced melting since they accreted on their parent planetesimals. In contrast, achondrites are igneous rocks that formed as melts or partial melt residues on their parent bodies. Since the advent of the modern era of meteoritics in the mid-1960s, this clear textural difference has been interpreted almost exclusively as evidence that chondrites and achondrites formed on different parent planetesimals, such that chondrite parent bodies never experienced large-scale melting or differentiation (e.g., Weisberg et al. 2006).

However, many early meteoriticists seriously considered the possibility that all chondrite and achondrite groups originated from a single or a small number of Moon-sized primordial bodies (Wood 1958) (**Figure 1**). Although the structures and origins imagined for these bodies varied

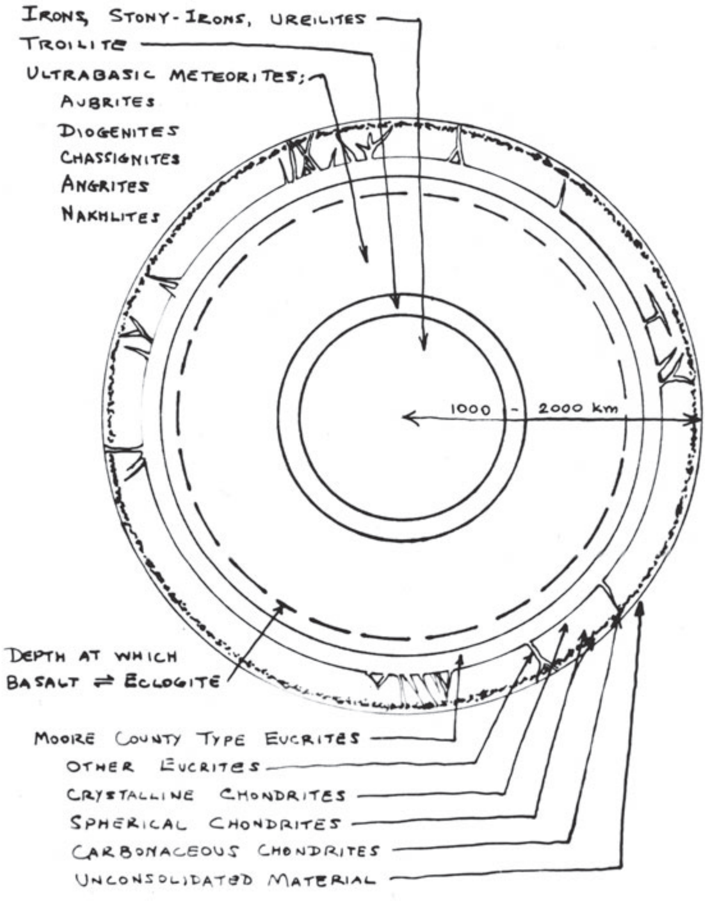


**Figure 1**

Sketch of a Moon-sized meteorite parent body with a radially layered igneous interior overlain by an unmelted chondritic crust. Reproduced from Wood (1958). Various achondrite groups are samples from different stratigraphic depths in the interior, whereas variably metamorphosed ordinary and carbonaceous chondrite groups are samples of the unmelted crust. Wood and most other meteoriticists later discarded this model as it became clear that meteorites originated from multiple, asteroid-sized parent bodies.

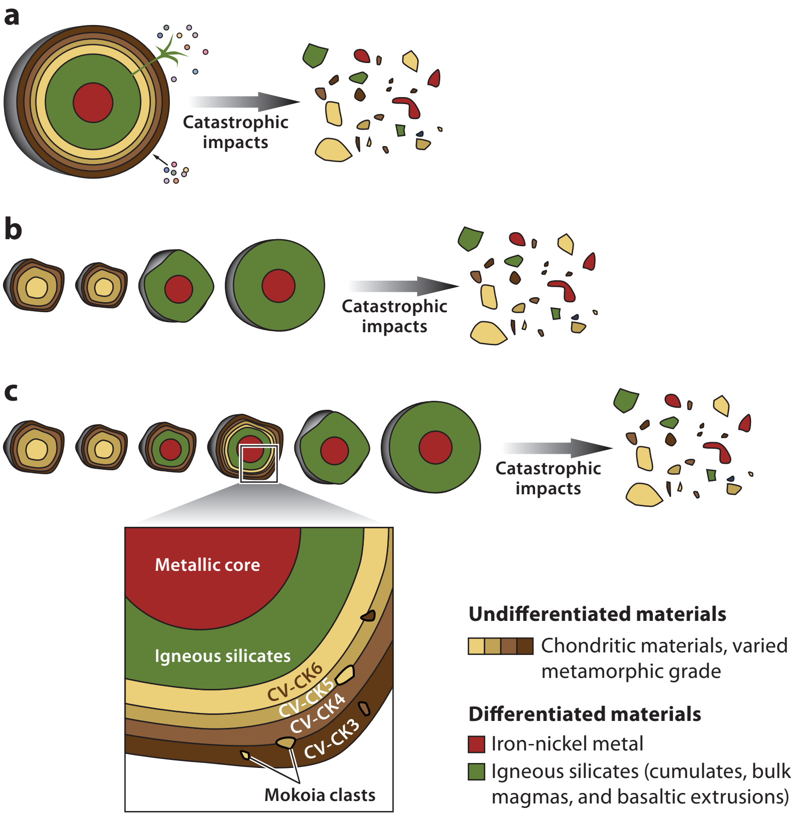


**Figure 2**

Meteorite parent body models. (*a*) All meteorites originated from one or a few Moon-sized partially differentiated bodies with chondritic surfaces formed by tuffaceous volcanism (Ringwood 1961), impact-induced melting, and/or deposition of exogenous material (Wood 1963). (*b*) Meteorites originated from multiple asteroid-sized bodies. Individual bodies were fully differentiated or fully undifferentiated (Mason 1967). (*c*) Meteorites originated from multiple, asteroid-sized bodies (Anders & Goles 1961). Individual bodies were fully differentiated, fully undifferentiated, or partially differentiated with an unmelted chondritic crust. (*Inset*) Schematic asteroid showing a possible structure of a partially differentiated CV-CK carbonaceous chondrite parent planetesimal (see Section 3.3). Metamorphosed and/or partially melted materials like clasts found in the CV chondrite Mokoia may be samples excavated from the deep interior.



greatly, they were commonly envisioned to have radially layered, melted interiors composed of iron meteorites and basaltic achondrites overlain by accretional crusts composed of chondrites (Anders & Goles 1961, Lovering 1962, Ringwood 1961, Wood 1963) (**Figures 1** and **2*a*,*c***). The chondritic surfaces were typically thought to have formed from volcanic fire fountaining, impact processing of the deep interior materials, or possibly even deposition of exogenous nebular solids (Wood 1963). The existence of polymict breccias containing both chondritic and achondritic clasts was taken as important evidence for the coexistence of the various meteorite groups on a single body (Lovering 1962, Wood 1963), and the observation of natural remanent magnetization in some chondrites was attributed to dynamo generation by an interior convecting metallic core (Anders 1964, Lovering 1962, Ringwood 1961, Stacey et al. 1961).

Beginning around the landmark review of meteoritics by Anders (1964), the single parent body hypothesis for meteorites was questioned and then eventually discarded as an accumulation of discoveries demonstrated that multiple parent bodies are represented by the world's meteorite inventory. These discoveries included orbital, dynamical, and geochemical observations demonstrating that most meteorites are samples of asteroids rather than of a disaggregated Moon-sized primordial body (Anders 1964) (**Figure 2**). Arguably, the development that propelled the multiple parent body paradigm to general acceptance was the observation that elemental and isotopic variations observed among meteorite groups were unlikely to have been produced solely by planetary igneous processes. In particular, meteorite groups were recognized to exhibit striking mass-independent variations in their bulk oxygen isotopic compositions (Clayton et al. 1976). These data also showed that chondritic clasts in many achondritic breccias were clearly foreign impactors, derived from different parent bodies rather than from an overlying unmelted crust (Clayton & Mayeda 1978). Over the past two decades, it has become clear that different meteorite groups also have distinct bulk isotopic compositions for a diversity of elements, including Cr, Ti, Mo, Ni, and Cu (e.g., Burkhardt et al. 2011). These differences cannot be easily explained except by formation on different parent bodies with isotopically distinct bulk compositions inherited during accretion.

In the new paradigm, achondrites and chondrites were assumed to have originated from separate bodies (Mason 1967) (**Figure 2*b***). However, over the past three decades, this simple picture has been muddled by the discovery of a broad class of meteorites that reached only subliquidus melting temperatures. Known as primitive achondrites (Weisberg et al. 2006), these partial melt residues provide prima facie evidence for the existence of planetesimals on which igneous differentiation did not go to completion. Because the lowest-temperature melts of chondrites typically have dense, iron-sulfide compositions that could sink to the planetary center, such bodies may have formed iron sulfide–iron metal cores (or at least localized metal-rich zones) overlain by primitive achondritic crusts that may even have included localized metamorphosed chondritic materials. A widely recognized example of such an object is the common parent body proposed for the winonaite primitive achondrites and the IAB iron meteorites (Benedix et al. 2000, Ruzicka & Hutson 2010). Candidate parent bodies for primitive achondrites have even been tentatively identified in the asteroid belt (Gaffey et al. 1993).

Although the existence of primitive achondrite parent bodies with partially melted mantles and incipient iron sulfide–iron metal cores seems broadly established, it has major implications for meteoritics and asteroid evolution that are still being explored. In this review, we examine the possibility that some planetesimals formed large interior regions that experienced total silicate melting but were overlain by a substantial unmelted chondritic crust. This would permit the possibility that known or as-yet-undiscovered chondrite and achondrite groups could in fact have originated from a single partially differentiated asteroid-sized body (**Figure 2*c***). This classic idea (Anders & Goles 1961) deserves serious renewed consideration because of several important recent developments in meteoritics. First, as mentioned above, the discovery of primitive achondrites indicates that in fact early planetesimals reached a continuum of differentiation end-states. Second, over the past decade, short-lived radioisotopic chronometers have established that the parent bodies of chondrites, despite containing the oldest known solids, accreted contemporaneously with or after many achondrite parent bodies (Kleine et al. 2009). This has motivated renewed proposals that some chondrules, and therefore some chondrites, formed from the remnants of already-differentiated planetesimals (Asphaug et al. 2011, Libourel & Chaussidon 2011, Libourel & Krot 2007). Third, the origin of remanent magnetization in most chondrites is largely unexplained by the existing paradigm. In fact, new paleomagnetic measurements have provided strong evidence that the recorded magnetic fields originated from the parent bodies themselves (Carporzen et al. 2011b). Fourth, asteroid spectroscopy studies and spacecraft flybys have now provided circumstantial evidence for the existence of partially differentiated bodies with chondritic crusts extant in the asteroid belt (Mothé-Diniz et al. 2008, Nathues 2010, Weiss et al. 2012). Fifth, there is growing geochemical evidence for direct genetic relationships between some achondritic and chondritic groups. Finally, virtually all modern models of planetesimal thermal evolution now predict the formation of an unmelted, primordial crust as an inevitable outcome of melting on early parent bodies from the decay of short-lived radionuclides (Elkins-Tanton et al. 2011, Ghosh & McSween 1998, Hevey & Sanders 2006, Sahijpal & Gupta 2011, Šrámek et al. 2012). In this review, we discuss how these and other recent developments in meteoritics and asteroid science suggest that it is time again to seriously consider the idea of partially differentiated parent bodies for some chondrites. We evaluate the evidence for and challenges to this hypothesis and consider its broad implications for the accretion and evolution of planetesimals, the interior structures of asteroids, and the origins of meteorite groups.



## 2. PLANETESIMAL THERMAL EVOLUTION

### 2.1. The Onion Shell Structure

Partially differentiated chondrite parent bodies should have formed naturally as a consequence of radiogenic heating within early planetesimals whose surfaces remained cold due to exposure to space. This heating and the attendant ice melting are the favored agents for the thermal metamorphism and aqueous alteration of chondrites and the melting processes that produced achondrites. The primary source of the heat for these processes is now thought to be the decay of the short-lived radionuclide $^{26}$Al (Fish et al. 1960, Huss et al. 2006, LaTourrette & Wasserburg 1998, Lee et al. 1976, Urey 1955). Impact heating may also have played an important role in chondrite metamorphism, but subcatastrophic impacts are inefficient at producing large-scale partial melting (Davison et al. 2010). Internal radiogenic heating coupled with radiation from the planetary surface means that a planetesimal would initially form an "onion shell" structure, with the highest temperatures reached in its center, surrounded by concentric zones of materials exposed to progressively lower peak temperatures (see McSween et al. 2002 and references therein).

Until recently, all planetesimal thermal models assumed the body accreted to its final size instantaneously, which leads to the prediction that the entire body will remain unmelted if it accretes $>\sim$1.5 Ma after the formation of calcium-aluminum-rich inclusions (CAIs) (due to the exponential decay of $^{26}$Al) or if it never grows to $>\sim$7 km in radius (due to conductive heat loss from the surface) (Hevey & Sanders 2006). A key finding of such models is that because the outer surface of a planetesimal is exposed to space, it will inevitably resist melting even when the interior of the asteroid is completely molten (e.g., Ghosh & McSween 1998). However, these early models did not consider convective heat transport within the molten interior of a body, which otherwise should lead to extremely efficient outward heat transport and a very thin unmelted crust (e.g., only $\sim$2 km thick for a 50-km-radius body that instantaneously accreted 0.75 Ma after CAI formation) (Hevey & Sanders 2006).

### 2.2. Incremental Accretion Models

A major recent step forward has been the development of thermal models that permit accretion to occur incrementally rather than instantaneously (Ghosh et al. 2003, Merk et al. 2002). The incremental accretion models, which model the thermal evolution of a growing body while incorporating heating from impacts, demonstrate clearly that the volumes of material in the parent body that reach different metamorphic grades differ significantly from those predicted by instantaneous models. In particular, incremental accretion strongly influences cooling in at least three ways. First, a small body loses heat from its surface far more rapidly than does a large body because of its proportionally larger surface-area-to-volume ratio; thus, the small body cools rapidly between accretionary impacts as it grows. Second, a thickening, undifferentiated, porous crust added to an initially melted planetesimal slows the planetesimal's heat flux into space. Third, because accretion could in principle begin when $^{26}$Al is abundant and continue well after it has largely decayed, incremental accretion permits an internally molten body to form an unmelted crust that is much thicker (Elkins-Tanton et al. 2011, Sahijpal & Gupta 2011, Šrámek et al. 2012) than that achievable in instantaneous accretion models (e.g., Hevey & Sanders 2006). In particular, numerical modeling suggests molten planetesimals can build up substantial (kilometers to tens of kilometers thick) crusts if they accrete to radii of at least a few tens of kilometers by 1.5 Ma after CAI formation and continue to accrete over a minimum period of perhaps one to several million years (with the precise time constraints depending on the rate of accretion, the time of onset of accretion, the size of the parent body, and its bulk composition and structure) (Sahijpal & Gupta 2011, Šrámek et al. 2012) (**Figure 3**). However, these numbers are uncertain, given that convection was only roughly parameterized in these models, that the temporal evolution of planetesimal porosity is poorly understood, and that accretion codes have not yet been developed to model this problem.



## 2.3. Survival of the Unmelted Crust

We have discussed how recent incremental accretion models indicate it is highly likely that partially differentiated planetesimals with unmelted crusts formed in the early Solar System. However, whether an unmelted crust could survive is less certain. This is because the crust is threatened from above by impacts and from below by eruption, foundering, and convective thinning. We discuss each of these processes in turn.

Impacts may melt, mix, or remove the crust. The question of whether a relict crust could survive is highly complex and depends upon poorly understood factors such as its thickness, the physical effects of impacts, the stochastic nature and duration of accretion, and the timelines for solidification and collisional evolution in the early Solar System. However, the preservation of most of asteroid 4 Vesta's basaltic crust (De Sanctis et al. 2012), much of which crystallized within the first 10 Ma following CAI formation (Ghosh & McSween 1998, Kleine et al. 2009), clearly demonstrates that early-formed planetary crusts can in principle be preserved for billions of years.

Eruption and intrusion of underlying magmas may also disrupt the crust. Wilson & Keil (1997) predict fire-fountaining lava eruptions on Vesta driven by volatiles in magmas. Ghosh & McSween (1998) describe an end-member model for Vesta in which all melt from the interior erupts onto the surface and another end-member model in which no melt extrudes; their careful efforts demonstrate the difficulty of arguing completely for one or another eruptive scenario. On Earth and other large terrestrial planets, eruption is driven primarily by two factors: density and volatile release. Hot, molten-silicate magmas can have lower density than crustal rocks in some cases. However, in the case of a planetesimal with a chondritic crust, melts from the interior are less likely to be positively buoyant because of the potentially high porosity of chondritic materials. An unsintered crust of CV carbonaceous chondrite composition may have a density between ~2,700 and 2,900 kg m$^{-3}$, whereas the density of molten CV chondrite over a range of temperatures and pressures is between ~2,900 and 3,000 kg m$^{-3}$ (Elkins-Tanton et al. 2011). Therefore, there may be little gravitational driving force for eruption (and internal convection), and furthermore, there may not be sufficiently large pressure range or competency in the crust to produce a pressure head on any magma body lying beneath. As on the Moon but to an even greater extreme, magmas on a planetesimal may require a significant impact basin to erupt onto the surface through the more buoyant crust.

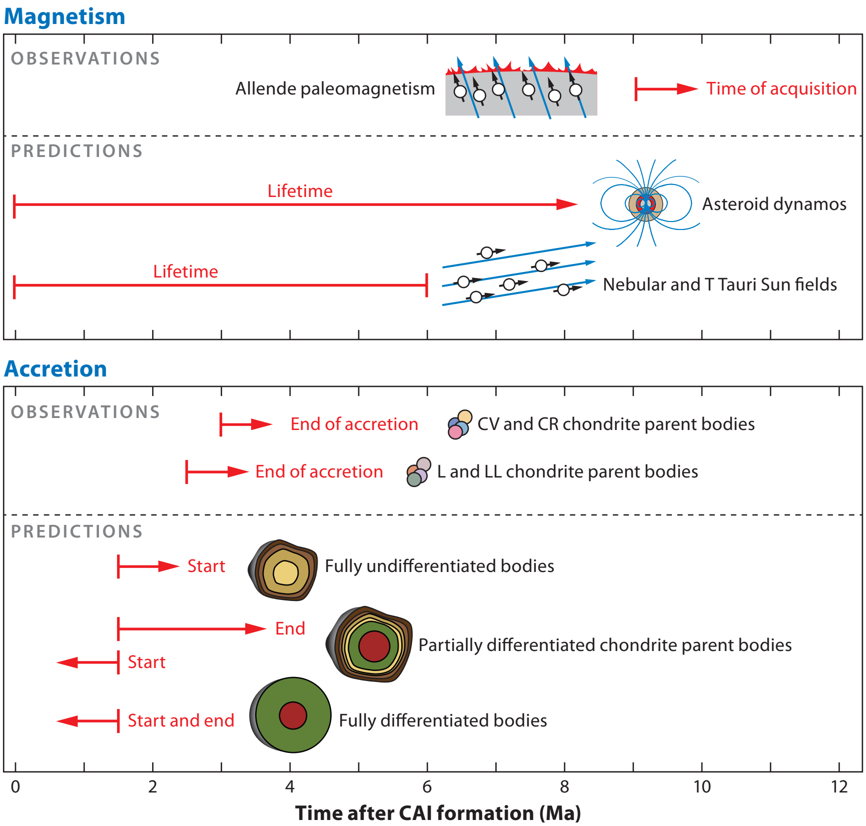


**Figure 3**

Predictions and observational constraints on the formation of partially differentiated chondrite parent bodies. (*Bottom*) Thermal modeling predicts that large (>∼5–7-km-radius) fully undifferentiated bodies must accrete most of their masses after ∼1.5 Ma after CAI formation, fully differentiated bodies must complete most of their accretion before ∼1.5 Ma after CAIs, and partially differentiated bodies with substantial chondritic crusts must begin accreting before ∼1.5 Ma after CAIs and finish accreting after ∼2.5 Ma after CAIs. The youngest chondrule ages for a given chondrite group provide a maximum age for the end of accretion. Constraints on crystallization ages are shown for CV, CR, L, and LL chondrules. (*Top*) Nebular and T Tauri Sun magnetic fields were present during the lifetime of the protoplanetary nebula (<∼6 Ma after CAI formation), whereas asteroid dynamos could have lasted ten to perhaps hundreds of Ma after CAI formation. Geochronometry suggests that Allende's unidirectional magnetization was acquired after the lifetime of nebular and T Tauri fields. Abbreviation: CAI, calcium-aluminum-rich inclusion.

Volatile release from solution into bubbles is one of the primary drivers of eruptions on Earth and may also have driven fire-fountaining eruptions on the Moon. On Earth, the highest volatile contents of magmas are introduced in subduction zones. In contrast, on early-forming planetesimals, gradual radiogenic heating would drive off volatiles before silicate melting begins. For peridotitic compositions, Ohtani et al. (2004) showed that at the very low pressures relevant for small bodies, talc and chlorite are the hydrous phases stable below approximately 650°C. Between 650 and 800°C, talc is replaced by amphibole; between 800 and approximately 1,100°C, water or hydroxyl can be held in amphibole alone; and above 1,100°C there is no stable hydrous silicate phase. Schmidt & Poli (1998) studied a wet basaltic composition and reported that above 500°C, no hydrous phase is present. Therefore, as temperatures rise above some temperature between

500 and 1,100°C, depending upon bulk composition, water will be released as a free fluid. If the free fluid migrates away, as is highly probable due to its buoyancy (Young et al. 1999), the resulting dry chondritic material will not melt until it heats to approximately 1,200°C (Agee et al. 1995). Silicate magmas on planetesimals therefore would often have very low volatile contents, consistent with measurements of achondrites (Jarosewich 1990).

However, if the magmas are often unable to erupt, some may still intrude and freeze in the lower crust. It has been suggested that a dense frozen magma lower crust overlying less-dense chondritic material is liable to founder through gravitational instability. This foundering could also drag down the overlying unmelted chondritic crust or at least thin the solid outer planetesimal shell sufficiently that it could be breached later by impacts or volcanism. The Rayleigh-Taylor instability timescale for overturn of an unstably stratified layer of viscous fluid of thickness $d$ and viscosity $\eta$ with stress-free top and bottom boundaries is given by

$$t_{\text{overturn}} = \frac{4\pi^2\eta}{\gamma g d},$$

where $\gamma$ is the compositional density difference (in kg m$^{-3}$) and $g$ is gravity (Hess & Parmentier 1995). Therefore, a frozen magma layer 5 km thick with a viscosity of $10^{18}$ Pa s (similar to Earth's mantle at solidus temperatures) and a density of 3,200 kg m$^{-3}$ that is perched above 5 km of solid bulk lid material of density 3,000 kg m$^{-3}$ on a body with gravitational acceleration of 0.3 m s$^{-2}$ will begin to sink in about 2 Ma. This would be after the time of peak heating for many bodies. Even if this first dense layer succeeds in sinking before cooling raises its viscosity, the development of a second dense layer from new rising magmas is unlikely. Thus, some thinning of the lithosphere through foundering is possible, but it is unlikely to be a continuous process.

Finally, portions of the lower crust may be subject to removal by convective traction. However, convective traction in a planetesimal is probably modest, primarily because of low gravity but also because of the near absence of viscous traction. Earth's sublithospheric mantle has a viscosity between $10^{18}$ and $10^{20}$ Pa s (Ghosh & Holt 2012), and the viscosity of the lithosphere is even higher. Traction is produced on the bottom of the lithosphere as a result of the relative velocity of the sublithospheric mantle and lithosphere. In contrast, the viscosity of molten mafic silicate may be 0.1 Pa s or less (Rubie et al. 2003). Thus, unlike in Earth's mantle, where solid mantle material is partially coupled to the bottom of the lithosphere, on planetesimals there is little traction between the liquid "mantle" and the solid conductive lid. Furthermore, convective velocity will be lower than that on larger planets because of low gravity; Vesta's gravity field, for example, is only ~0.32 m s$^{-2}$ (Zuber et al. 2011). The small density changes produced by cooling under the lid will drive convection under this small gravity field that is far less vigorous than convection on Earth. If all other parameters are equal, Rayleigh number is directly proportional to gravity. Heat flux from the convecting magma ocean is proportional to Rayleigh number to the one-third power, and in the soft convective regime, convective velocity is proportional to heat flux to the one-third power (Solomatov 2000). Thus, by changing gravity from 9.8 m s$^{-2}$ on Earth to 0.32 m s$^{-2}$ on Vesta, convective velocity is lowered by approximately 30%. The lower velocity will further limit erosion of the lid. Thus, traction from the underlying magma ocean is an unlikely mechanism for thinning the conductive solid lid.

# 3. IMPLICATIONS FOR AND EVIDENCE FROM METEORITES

## 3.1. Overview and Predictions

The chondrite record should provide critical tests of the partial differentiation hypothesis. However, a key limitation is that an unmelted chondritic crust is predicted to have little petrologic communication with the melted interior, with the important exception that the crust would be variably metamorphosed and metasomatized from outward-migrating heat and fluids. The consequent lack of an igneous signature of the differentiated interior within surface rocks is both a prediction of the chondrite differentiation hypothesis and a key impediment to testing it. Nevertheless, we suggest that there are at least four possible consequences of a melted interior that can be tested with these meteorites. In this section, we provide a broad overview of three of these predictions and examine the evidence from selected chondrite groups; we discuss the fourth prediction in Section 5.

The first prediction, which is virtually a tautology of the hypothesis, is that partially differentiated chondrite parent bodies should have generated a suite of rock types extending from moderately metamorphosed chondrites to highly metamorphosed chondrites and, critically, to achondrites. The latter almost certainly should have included primitive achondrites and possibly also basaltic achondrites and cumulates as well as stony-iron and iron meteorites. Such achondrites might exist in the world's meteorite collections, but because of the biases associated with meteorite ejection from asteroids and their transfer to Earth, this is certainly not a requirement. Recognizing these affiliated achondrites is challenging, but it would probably involve demonstrating that they share a similar isotopic composition, paleomagnetic record, cosmic ray exposure age distribution, and/or radioisotopic impact age distribution with a particular chondrite group. Their chemical compositions should also be consistent with fractionations expected from igneous differentiation of their chondrite protoliths. As was recognized early on (see Section 1), particularly strong evidence would come from the identification of chondritic and achondritic materials that share these properties and are colocated within a single brecciated meteorite. For example, the existence of howardites, breccias containing diogenitic and eucritic clasts, provides strong evidence for a common parent body for most howardites, eucrites, and diogenites (Weisberg et al. 2006). The strongest geochemical tests are provided by a comparison of the isotopic compositions of nonradiogenic elements in bulk meteorite parent bodies that exhibit mass-independent heterogeneities typically inherited from the protoplanetary nebula. Parent body origins can also be constrained using isotopes exhibiting only mass-dependent variations or elements with only two naturally occurring stable isotopes, although caution must be exercised because a parent body could have a composition of these isotopes that is spatially heterogeneous as a result of postaccretional igneous, metamorphic, or aqueous alteration processes.

A second prediction is that remanent magnetization in chondrites could provide a record of an interior advecting metallic core dynamo. Recent paleomagnetic studies of basaltic achondrites, coupled with theoretical constraints on the physics of dynamo generation, have shown that many early planetesimals with convecting metallic cores were likely capable of generating magnetic fields beginning within $\sim$3 Ma of CAI formation and lasting for at least $\sim$10 to possibly $>$100 Ma (Fu et al. 2012, Sterenborg & Crowley 2013, Tarduno et al. 2012, Weiss et al. 2010b). Therefore, if chondrites originated on bodies that formed metallic cores, they could have become magnetized by an early dynamo magnetic field at the time they were metamorphosed or aqueously altered. In this sense, paleomagnetism is a remote core detector, making it one of the few properties of meteorites that can provide a record of conditions in the deep planetesimal interior.

This possibility was not fully appreciated until recently because of the widespread assumption that chondrite paleomagnetism is a record of fields produced externally to the parent bodies. The T Tauri Sun generated a long-lived, strong ($\sim$0.1-T surface field) dipolar dynamo magnetic field as well as transient (hours-long) flares (Vallée 2011). The protoplanetary nebula itself may have contained large-scale fields (Sugiura & Strangway 1988, Weiss et al. 2010b). Both T Tauri and nebular field sources should have dissipated when the Sun entered its main sequence phase and the nebula was dispersed. Astronomical observations of dust emission from young stellar objects of solar mass indicate that ~50% lose their disks within 2 ± 1 Ma of their formation and >~90% lose their disks within 6 Ma. (Evans et al. 2009, Haisch et al. 2001) (**Figure 3**). The only way an asteroid could have been magnetized by these external fields would be if its magnetization were acquired earlier than this limiting time. This means that an asteroid would need to have experienced its final major episode of thermal metamorphism and/or aqueous alteration within just a few million years of CAI formation, which is unlikely for at least some chondrite parent bodies (see Section 3.3 and Kleine et al. 2009). Even if this condition were met, the expected field intensities available to magnetize an asteroid were not great. In particular, the weak intensities of the T Tauri fields in the asteroid belt region (~1–10 nT for both dynamo and flare fields, assuming dipolar field geometry and a stellar radius three times that of the Sun today) make them an unfavorable source of magnetization in asteroids. Magnetohydrodynamical simulations predict nebular fields of order 0.01–0.1 mT in the midplane of the asteroid belt region but find that these fields were dominated by toroidal and radial components that periodically reversed every few hundred years (Turner & Sano 2008). Such reversing fields would have produced large-scale net magnetization in an asteroid only if the time period over which the magnetization is acquired were shorter than this reversal timescale. In summary, it is difficult for externally generated magnetic fields to strongly magnetize asteroids because of the short lifetimes and predicted weak intensities of the temporally steady components of these fields.

A third prediction motivated by thermal evolution models (see Section 4) is that partially differentiated parent bodies must have been large. An absolute minimum radius of ~5–7 km is set by the requirement that they reach melting temperatures as a result of $^{26}$Al heating (Hevey & Sanders 2006, Merk et al. 2002). Bodies that undergo >50 vol% melting must have radii greater than ~20 km (Hevey & Sanders 2006). Those bodies that went on to form substantial metallic cores and dynamos lasting for millions of years would probably have to have been even larger: Dynamo scaling laws suggest a minimum metallic core radius (which is a lower limit on parent body radius) of ~100–200 km (Weiss et al. 2010b). Parent body sizes can be inferred from petrographic and thermochronometric measurements of meteorite cooling rates (Kleine et al. 2009).

The fourth prediction is that to build up a substantial unmelted crust overlying a molten interior, accretion must begin prior to 1.5 Ma after CAI formation and continue incrementally over perhaps 1 Ma or longer. This prediction is discussed in Section 5.

## 3.2. Ordinary Chondrites

**3.2.1. Overview.** Ordinary chondrites, which make up 85% of meteorite falls, are thought to be derived from at least three separate parent bodies (H, L, and LL) (Weisberg et al. 2006). Each of these ordinary chondrite groups forms a complete sequence from petrologic types 3 to 6, with the equilibrated members (types 4–6) representing >90% of known meteorites from each group. A variety of data have long suggested an association between ordinary chondrites and two particular groups of iron meteorites. In particular, geochemical data indicate a possible common parent body for L or LL chondrites and IVA iron meteorites, the third-largest group of iron meteorites (**Supplemental Table 1**; see the Supplemental Material link in the online version of this article or at **http://www.annualreviews.org/**). Here we focus on the even more intriguing IIE iron meteorites (IIE irons), which contain a diversity of silicate inclusions with geochemical affinities to H ordinary chondrites.

**3.2.2. Connection with IIE iron meteorites.** IIE irons are thought to be impactites assembled from a mixture of molten iron and petrologically diverse silicate-rich inclusions ranging from unmelted chondrites with relic chondrules (**Figure 4**), to metamorphosed chondrites, to partial melts and melt residues (Goldstein et al. 2009). As such, IIE irons exhibit almost the full range of

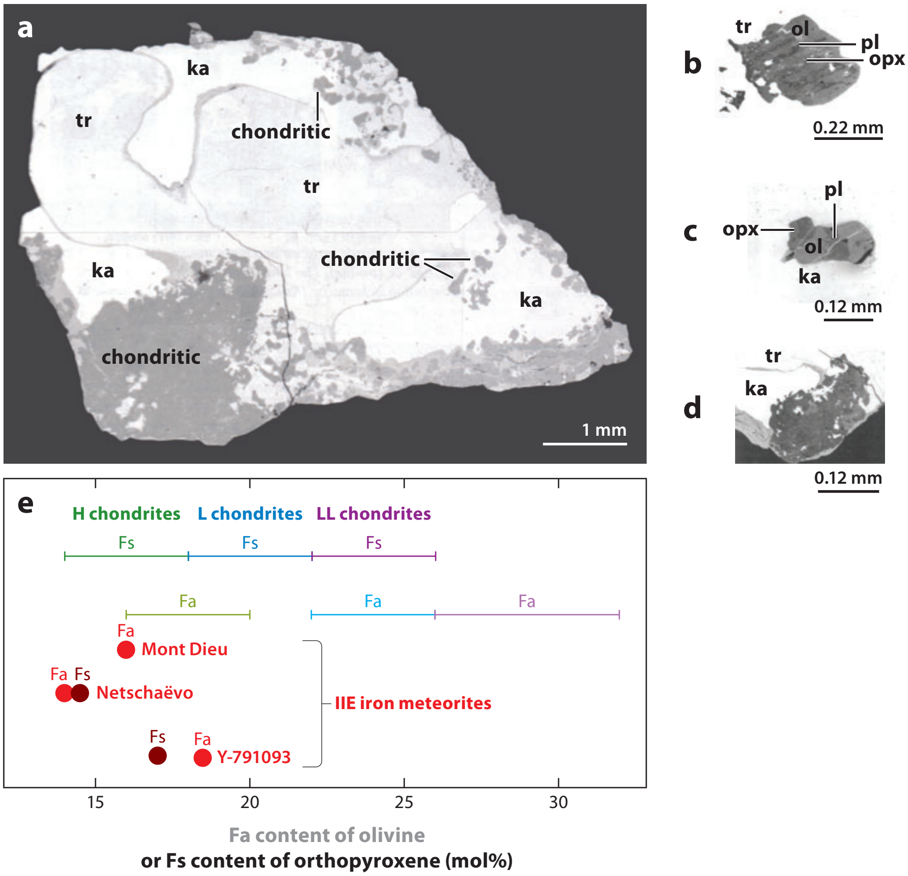


**Figure 4**

Backscattered scanning electron microscopy images of the IIE iron meteorite Y-791093 and chemical compositions of chondritic inclusions in IIE iron meteorites. (*a*) Thin section of meteorite showing kamacite (ka) and troilite (tr) with chondritic inclusions. (*b*) Barred-olivine chondrule containing olivine (ol), plagioclase (pl), and orthopyroxene (opx). (*c*,*d*) Chondritic silicate inclusions. (*e*) Iron composition of olivine (fayalite content) and orthopyroxene (ferrosilite content) in chondritic inclusions in three different iron meteorites (Y-791093, Mont Dieu, and Netschaëvo) compared with the three different ordinary chondrite groups. Images in panels *a*–*d* from Ikeda et al. (1997). Data in panel *e* from Ikeda et al. (1997), Olsen & Jarosewich (1971), and Van Roosbroek et al. (2012). Abbreviations: Fa, fayalite; Fs, ferrosilite.

igneous differentiates—iron core, basaltic and primitive achondritic mantle, and a relict chondritic crust—expected for a partially differentiated planetesimal!

Unlike in most iron meteorite groups, siderophile elements in IIE irons do not exhibit strong trends that can be explained by fractional crystallization during solidification of a molten iron core (Goldstein et al. 2009). IIE irons are also unusual in that their metals exhibit a wide range of $^{182}$W isotopic compositions, including the largest $^{182}$W/$^{184}$W value known for iron meteorites (Kleine et al. 2009). This has motivated two general classes of models for the origin of IIE irons (Ruzicka & Hutson 2010). The first set of models posits that the IIE irons formed in small melt pools produced by the selective melting and mobilization of metallic and silicate liquids by localized impact-induced heating (**Figure 5*a***). A second set of hybrid models proposes that like the IAB iron-winonaite parent body (Benedix et al. 2000), the IIE parent body was partially differentiated

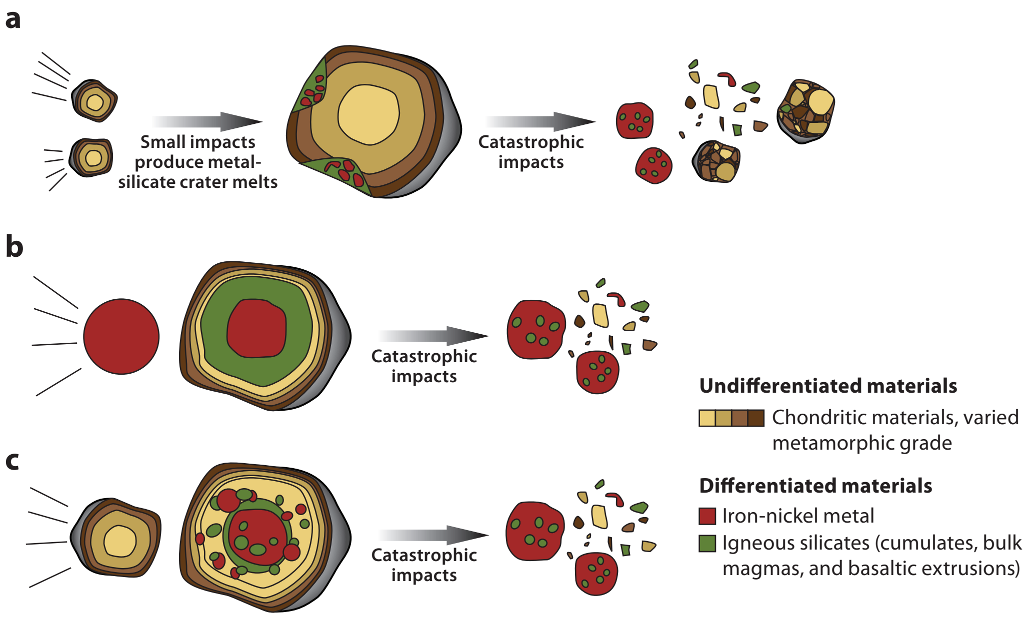




**Figure 5**

IIE iron meteorite parent body models. (*a*) IIE iron meteorites formed in localized, impact-generated pools of metallic and silicate liquids on an H chondrite–like body. (*b*) A partially differentiated asteroid (melted from $^{26}$Al decay) with an H chondrite–like crust, achondritic silicate mantle, and possibly a metallic core is impacted by a molten metal–rich impactor. The IIE iron meteorites form from mixtures of molten metal from the impactor and silicates from the target. (*c*) A partially differentiated asteroid with an H chondrite–like crust, achondritic mantle, and incipient metallic core is structurally scrambled by a large impact.

(due to $^{26}$Al-induced heating) and was then structurally scrambled as a result of a catastrophic impact. In this scenario, the metal in IIE irons was derived from a molten metal impactor or from an incipient metal core inside the partially differentiated target (**Figure 5*b*,*c***). The single parent body variant of the latter group of models is essentially a partially differentiated planetesimal with a relict chondritic crust (see Section 2).

Although impact mixing of metal and silicates undoubtedly led to the assembly of IIE irons, impacts almost certainly were not the sole heat source responsible for the formation of the igneous silicate inclusions (as proposed by the first set of models above) because it is difficult for impacts on asteroid-sized bodies to produce large quantities of silicate partial melts (Davison et al. 2010, Ruzicka & Hutson 2010). Furthermore, it is challenging for impacts to produce the two-stage cooling histories (initial slow cooling followed by rapid melt solidification) observed for fractionated IIE irons (Ruzicka & Hutson 2010). These observations favor the second set of models invoking impact scrambling of a previously partially differentiated object. These observations are also consistent with the proposal that the broad range of W isotopic compositions of IIE irons formed via impact-induced isotopic exchange between silicates and metal (Kleine et al. 2009). Furthermore, the similar isotopic compositions of IIE iron and H chondrite metals and the apparent feasibility of deriving the siderophile element patterns of IIE irons by reduction and partial melting of H chondrites (Teplyakova et al. 2012) (**Supplemental Table 2**) favor the hypothesis that IIE iron metal was extracted from a protolith very similar to the chondritic silicate inclusions. This, in turn, favors formation of IIE irons by impact-induced mixing of silicates with the interior core from a single partially differentiated body rather than with the core of a foreign impactor.

Moreover, it suggests that this parent body is the H chondrite parent body itself! This would be consistent with its estimated ~100-km radius (Ghosh et al. 2003, Kleine et al. 2009), which well exceeds the minimum size required for $^{26}$Al-induced differentiation (see Section 2).

An H chondrite parent body is supported by the similar elemental and isotopic compositions of chondritic silicate inclusions in IIE irons and H chondrites (**Supplemental Table 2**). An important exception is that the high Fe/FeO and high ratio of siderophile elements to Si in apparently unmelted chondritic inclusions in the IIE iron Netschaëvo suggest that it instead may sample a different "HH" parent body (Bild & Wasson 1977, Olsen & Jarosewich 1971). However, recent analyses of chondritic silicate inclusions in two other IIE irons have found that they have compositions closer to or even indistinguishable from that of H chondrites (Ikeda et al. 1997, Van Roosbroek et al. 2012) (**Figure 4**).

**3.2.3. Paleomagnetism of ordinary chondrites.** The endogenic heating models for IIE irons imply that some ordinary chondrites originated from a body with a liquid metallic core. This core could have generated a dynamo whose surface field could have been imprinted as remanent magnetization in metamorphosed ordinary chondrites. In fact, the first detailed paleomagnetic studies of meteorites, which were conducted on ordinary chondrites (see Section 1), identified natural remanent magnetization interpreted to have formed in ~10–100-μT fields originating from an interior core dynamo. Subsequently, remanent magnetization was identified in dozens of other chondrites and inferred to have formed in similarly strong fields (Weiss et al. 2010b). However, the emergent paradigm of undifferentiated chondrite parent bodies led most workers to reinterpret this magnetization as the record of a nebular or solar field rather than a core dynamo (Sugiura & Strangway 1988).

Thermoremanent magnetization acquired during cooling in the presence of a magnetic field should be unidirectionally oriented throughout an unbrecciated rock. However, the natural remanent magnetization observed within the extensively studied L and LL ordinary chondrite groups has usually been found to be heterogeneously oriented down to at least the millimeter scale (**Figure 6**). Depending on the meteorite, this heterogeneity may reflect ancient magnetization acquired by chondrules and inclusions in the nebula prior to accretion, an initially unidirectional postaccretional magnetization that was subsequently jumbled during cold brecciation, or magnetization formed in a null magnetic field (**Figure 7*b*,*d*,*e***). With respect to the latter, the early-formed ferromagnetic iron-nickel mineral taenite transforms to the ferromagnetic mineral tetrataenite during slow cooling below 320°C, during which it may not preserve any preexisting natural remanent magnetization (Weiss et al. 2010b). Thermochronometry indicates that this temperature was reached between ~10 and 150 Ma after CAI formation for type 4–6 H chondrites (Kleine et al. 2009). Given the ≤6-Ma lifetime of nebular fields (see Section 3.1) and the ~10–100-Ma lifetime of planetesimal dynamos, this indicates that tetrataenite in a chondrite may have formed in a null-field environment and therefore not acquired new magnetization. Even more importantly, most ordinary chondrites have $^{40}$Ar/$^{39}$Ar ages <4 billion years (Ga) (Bogard 2011), indicating that regardless of the presence of tetrataenite, they were completely thermally remagnetized or demagnetized long after any putative planetesimal dynamo or nebular field would have decayed.

Interestingly, in stark contrast to L and LL chondrites, eight out of the nine H chondrites analyzed using mutually oriented subsamples have approximately unidirectional natural remanent magnetization (Funaki et al. 1999, Hamano & Matsui 1983, Westphal & Whitechurch 1983). Furthermore, $^{40}$Ar/$^{39}$Ar analyses have found that several percent of H chondrites were not significantly shock heated following ~4.4 Ga ago, suggesting that they could have recorded an early planetesimal dynamo (Bogard 2011). However, previous paleomagnetic studies have not ruled out the possibility that these meteorites were remagnetized after landing on Earth by weathering or

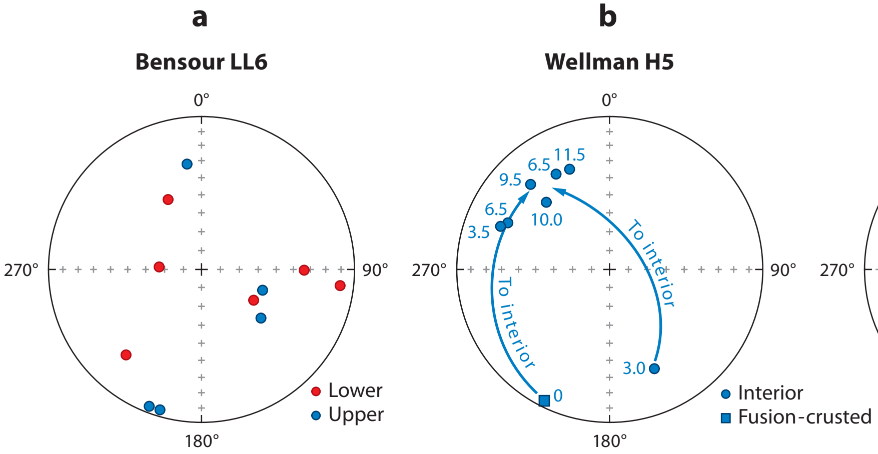


**Figure 6**

Magnetization directions of mutually oriented subsamples of ordinary and carbonaceous chondrites. Each symbol represents a single subsample; squares denote fusion-crusted (exterior) subsamples and circles denote interior subsamples. Shown are equal-area stereographic projections, with blue symbols denoting upper hemisphere and red symbols denoting lower hemisphere. (*a*) Nonunidirectional magnetization in the LL6 chondrite Bensour (Gattacceca et al. 2003), possibly produced by tetrataenite formation in a zero field (**Figure 7*d***) and/or by cold brecciation (**Figure 7*e***). (*b*) Magnetization in the Wellman H5 chondrite (Hamano & Matsui 1983). The distance of each subsample from the closest fusion crust is listed in millimeters. There is a progressive rotation of magnetization directions from the interior to the exterior. The magnetizations of subsamples from within ~3 cm of the exterior are ~100 times stronger than those of interior subsamples, even though all subsamples have similar quantities of ferromagnetic material. This depth is several times deeper than the expected thermal remagnetization zone from atmospheric heating (**Figure 7*f***) and is characteristic of the application of a strong magnet following collection on Earth (**Figure 7*g***). (*c*) Unidirectional magnetization in the Allende CV3 chondrite (Carporzen et al. 2011b) likely produced in a parent body dynamo field during thermal metamorphism and aqueous alteration (**Figure 7*c***).



collectors' magnets (Weiss et al. 2010b) (**Figure 7*g***). Unfortunately, this applies to one of the ordinary chondrites best characterized using paleomagnetic methods, the H5 chondrite Wellman (Hamano & Matsui 1983) (**Figure 6**). A new generation of paleomagnetic measurements, just getting underway at the time of this writing, is now required to resolve these ambiguities before the full implications of ordinary chondrite paleomagnetism can be appreciated.

## 3.3. Carbonaceous Chondrites

**3.3.1. Overview.** Several carbonaceous chondrite groups contain abundant CAIs and chondrules that provide a direct rock record from the preaccretional phase of the early Solar System. Until recently, no major carbonaceous chondrite group other than CK chondrites was known to have members of type higher than 3.9, with various petrologic and geochemical data indicating peak temperatures of ~300–700°C for the most-metamorphosed CV and CO chondrites (Huss et al. 2006, Weisberg et al. 2006). In comparison, CK chondrites form a complete metamorphic sequence from types 3 to 6, with peak temperatures ranging up to 650–1,000°C (Righter & Neff 2007). The seeming lack of metamorphosed members among non-CK carbonaceous chondrite groups has been ubiquitously interpreted as evidence that their parent bodies did not experience high metamorphic temperatures, let alone melting.

However, as discussed in Section 2, accretion beginning before ~1.5 Ma after CAI formation and lasting for perhaps ≥1 Ma can produce differentiated bodies of carbonaceous chondrite

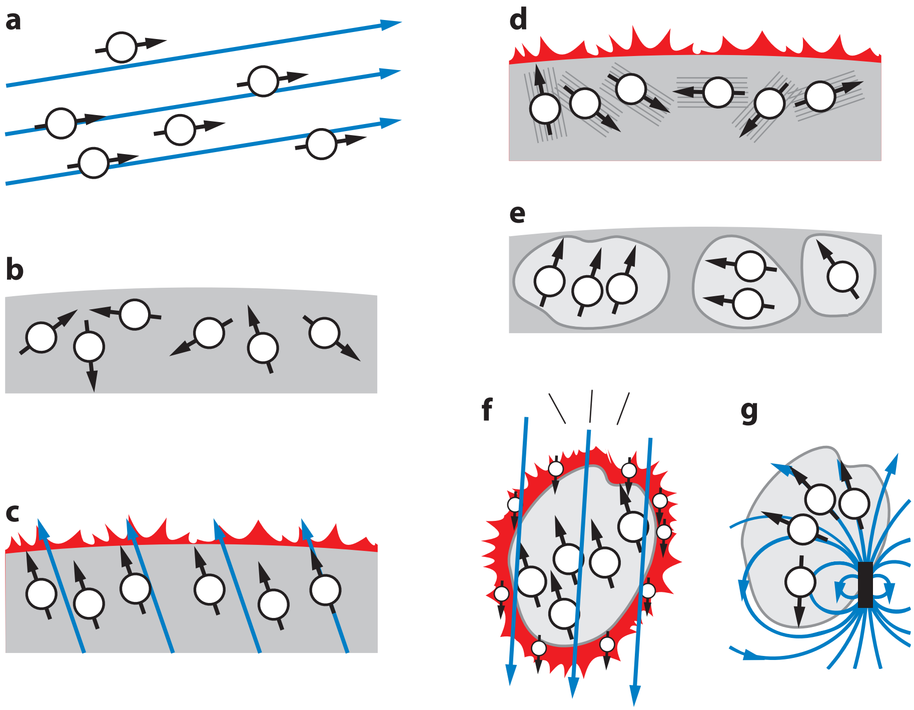


**Figure 7**

Mechanisms for magnetizing and remagnetizing chondrites. (*a*) Prior to accretion, lithic objects (chondrules, CAIs, and other rock fragments) (*circles*) can become magnetized in nebular or solar magnetic fields. Arrows show the direction of magnetization in each object. (*b*) Subsequent accretion in a weak field will produce nonunidirectional magnetization at the scale of individual lithic objects. (*c*) Subsequent thermal metamorphism and cooling in the presence of a magnetic field (e.g., a parent body core dynamo) will produce a unidirectional overprinting magnetization. (*d*) Following the events in panel *c*, thermal metamorphism and subsequent cooling in the absence of a magnetic field will produce fine-scale nonunidirectional magnetization controlled by local crystallographic axes or mineral fabrics (*parallel lines*). (*e*) Alternatively, following the events in panel *c*, cold brecciation will produce magnetization that is nonunidirectional at large scales but unidirectional within clasts. (*f*) During atmospheric passage, the outer several millimeters of stony meteorites will thermally remagnetize in the direction of Earth's field. (*g*) After landing, a chondrite can be remagnetized by hand magnets and weathering. Abbreviation: CAI, calcium-aluminum-rich inclusion.

composition with metamorphosed crusts. Furthermore, over the past two decades, there has been a growing realization that much of the thermal and aqueous alteration of carbonaceous chondrites, once thought to have occurred mainly in the nebula, actually occurred following accretion on the parent body (see Krot et al. 1995 and references therein). This metamorphism and alteration in turn provides evidence for heating by $^{26}$Al decay, opening the possibility that deeper portions of some carbonaceous chondrite parent bodies reached melting temperatures.

An interesting possible example of partial differentiation is provided by the CR carbonaceous chondrites, a metal-rich group with heavily hydrated matrices once thought to contain only type 2 members (Krot et al. 2002, Weisberg et al. 2006). However, recent discoveries of type ≥6 CR chondrites, along with silicate-bearing iron meteorites and a basaltic achondrite (NWA 011) with geochemical affinities to CR chondrites, indicate that the CR parent body also experienced substantial thermal metamorphism and possibly even internal igneous differentiation

(Wittke et al. 2011) (**Supplemental Table 3**). This suggests that the CR parent body was large and partially differentiated. However, there have been virtually no paleomagnetic studies of CR chondrites that could further test this idea.

In comparison, numerous paleomagnetic measurements of CV chondrites have provided robust paleomagnetic evidence for ancient magnetic fields on the CV chondrite parent body (see references in Weiss et al. 2010b). Furthermore, highly metamorphosed chondrites and achondrites with affinities to CV chondrites have recently been recognized. These and other data provide evidence not only that some carbonaceous chondrite-like bodies experienced extensive melting but also that the CV parent body itself may be partially differentiated. Furthermore, as discussed in Section 4, this parent body itself may even have been identified in the asteroid belt.

**3.3.2. Paleomagnetism of CV carbonaceous chondrites.** The Allende CV carbonaceous chondrite is the meteorite that has been most extensively studied using paleomagnetic methods. More than two dozen studies have now demonstrated that Allende contains natural remanent magnetization (see references in table 2 of Weiss et al. 2010b). Unlike nearly all analyzed ordinary chondrites (**Figure 6*a***), more than 90% of the natural remanent magnetization of matrix-rich subsamples of Allende is unidirectionally oriented throughout the interior of the meteorite and is divergent in direction from the magnetization in fusion-crusted (i.e., exterior) subsamples. Given that the fusion crust should be thermally remagnetized during atmospheric passage in Earth's field (**Figure 7*f***), these two observations, now demonstrated by four different laboratories (Carporzen et al. 2011b, Weiss et al. 2010b) (**Figure 6*c***), unambiguously demonstrate that this magnetization component formed after accretion of the CV parent planetesimal but before arrival at Earth. The paleointensity of the field that produced this magnetization was substantial (approximately that of the present-day Earth) (Carporzen et al. 2011b). Similar paleointensities have been obtained for the CV chondrite Leoville (Nagata et al. 1991). Furthermore, the two other CV chondrites studied using mutually oriented subsamples, ALH 84028 and Kaba, also exhibit highly stable, unidirectional magnetization that therefore must also postdate accretion (Gattacceca et al. 2012, Weiss et al. 2010a).

The origin of the magnetization in Allende has largely been a mystery. Part of the confusion has been that in addition to the strong unidirectional magnetization discussed above, individual chondrules also contain weak magnetization components with scattered directions, previously interpreted as preaccretional (e.g., **Figure 7*b***). However, because most of the ferromagnetic minerals in Allende are parent body alteration products and because recent analyses have shown that the magnetization within individual chondrules is nonunidirectional, this cannot be nebular paleomagnetism but rather is probably a near-zero-field magnetization (Carporzen et al. 2011a) (e.g., **Figure 7*d***). Regardless, nearly all investigators since the mid-1970s have suggested that the field that produced the unidirectional component was generated externally to the CV parent body, either by the T Tauri Sun or the nebula (Sugiura & Strangway 1988). With the exception of one study (Funaki 2005), the possibility of an internally generated dynamo field has been universally discounted because chondrites were assumed to be samples of undifferentiated bodies.

The key to distinguishing between these field sources is a determination of when and over what time period the magnetization was acquired. I/Xe thermochronometry indicates that the magnetization in Allende was acquired over several million years during metasomatism on the CV parent planetesimal ∼9–10 Ma after CAI formation (Carporzen et al. 2011b). This means that the magnetization in Allende is too young and was acquired over too long a time period to have been produced by early nebular or solar magnetic fields (see Section 3.1). The authors concluded that the only magnetic field source consistent with the age and period of magnetization acquisition and the paleointensity constraints is an internal core dynamo. Thermal modeling indicates a

>200-km-radius CV parent planetesimal can produce a dynamo magnetic field lasting for 10 Ma after CAI formation if, as discussed above, the body accreted gradually over a $\geq\sim$1 Ma period beginning before $\sim$1.5 Ma after CAI formation (Elkins-Tanton et al. 2011, Sahijpal & Gupta 2011, Šrámek et al. 2012). Such an early accretion time is broadly consistent with the fact that among chondrite groups, CV chondrites contain chondrules with some of the oldest crystallization ages (Bizzarro et al. 2004, Connelly et al. 2012, Kleine et al. 2009). Because gas drag rapidly depletes the largest objects in this size range, an early accretion age is also consistent with the observation that CV chondrites contain the largest chondrules and CAIs as well as the most abundant CAIs (Cuzzi & Weidenschilling 2006).

**3.3.3. Metamorphosed chondrites and achondrites with affinities to CV chondrites.** If the CV parent planetesimal evolved into a radially layered body with a central metallic core, silicate magma ocean mantle, and metamorphosed chondritic crust, then it is possible that we have iron, stony-iron, basaltic achondrite, and/or heavily metamorphosed meteorites from the deep interior of this body. However, because this body should have been large (see Section 3.3.2), we expect that its collisional lifetime (the mean time required for a collision to produce fragments the largest of which has a mass less than half that of the original asteroid) should exceed the age of the Solar System (Bottke et al. 2005), which may limit sampling of the deep interior (but see Section 4.3.2).

Until very recently, there were no known type $\geq$4 chondrites in the CV group. However, renewed studies of CV and CK meteorites indicate that they may form a continuum in mineralogy and composition and therefore a complete metamorphic sequence from types 3–6, representing a stratigraphic cross section of an onion shell asteroid (Greenwood et al. 2010, Isa et al. 2012; but see Davidson et al. 2012) (**Figure 2*d***). In support of this possibility, CK and CV meteorites overlap in their oxygen (Greenwood et al. 2010) and possibly their $^{54}$Cr/$^{52}$Cr (Qin et al. 2010, Trinquier et al. 2007) isotopic compositions. However, the $^{48}$Ti/$^{47}$Ti composition of the only CK chondrite for which it has been measured differs from that of the CV chondrite Allende (Trinquier et al. 2009). Although the CV-CK metamorphic sequence was recently proposed to have been produced by radiative heating from close approaches to the Sun rather than by internal radiogenic heating (Chaumard et al. 2012), the petrologic grade and the observed amount of diffusive cosmogenic $^3$He loss are not clearly correlated as would be expected for that model (Scherer & Schultz 2000).

Another important recent development is the discovery of heavily metamorphosed CV chondritic materials. In particular, a type $\geq$6 chondrite with bulk elemental and oxygen isotopic compositions indistinguishable from those of CV chondrites, NWA 3133, has recently been identified (**Figure 8**) (Schoenbeck et al. 2006), although its Cr isotopic composition appears distinct from that of CV chondrites (Shukolyukov et al. 2011). Even more importantly, clasts that are thought to have experienced peak metamorphic temperatures of $\geq$830°C and possibly even partial melting and that have bulk elemental and oxygen isotopic compositions similar to those of CV chondrites have been identified within two brecciated CV chondrites (Jogo et al. 2013) (**Figure 8**). The direct association of these clasts with weakly metamorphosed CV3 materials in a single breccia provides very strong evidence that the clasts originated on the CV parent body and that the interior of this body experienced high temperatures that may even have exceeded the solidus (see Section 3.1).

CV-like achondritic clasts in CV chondrites have even been reported. Kennedy & Hutcheon (1992) described a plagioclase olivine inclusion in Allende with subophitic texture and a composition interpreted as reflecting the incorporation of a chemically fractionated component produced by planetary melting processes. More recently, Libourel & Chaussidon (2011) argued that relict Mg-rich olivine aggregates found in CV chondrules and as isolated matrix grains are the products of early planetary differentiation. The oxygen isotopic compositions of these olivine aggregates

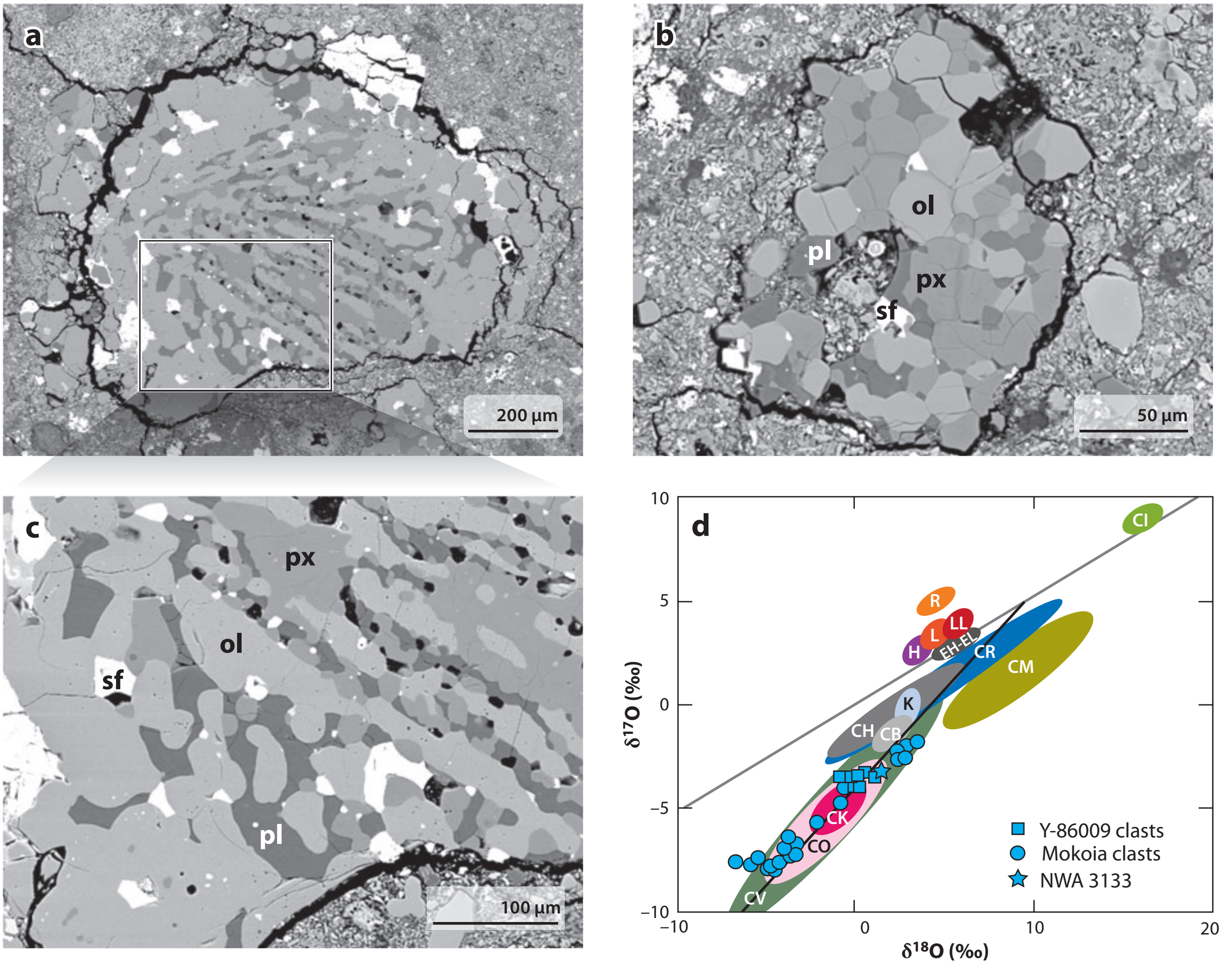


**Figure 8**

Metamorphosed chondritic clasts in CV3 chondrites. (*a*,*b*) Backscattered electron images of metamorphosed and/or partially melted chondritic clasts in the Mokoia breccia. (*c*) Close-up of the boxed region in panel *a*. Olivine (ol), diopside (px), anorthitic plagioclase (pl), and pentlandite (sf) grains are labeled. (*d*) Oxygen isotopic compositions of metamorphosed chondrite NWA 3133 (*blue star*) and metamorphosed clasts in the CV chondrites Mokoia (*blue circles*) and Yamato 86009 (Y-86009) (*blue squares*) compared with the ranges of values measured for various chondrite groups (*colored ovals*). The compositions of the clasts lie in the CV, CK, and CO carbonaceous fields, distinct from CB, CH, CM, CI, CR, K, R, H, L, LL, EH, and EL chondrites. The gray line is the terrestrial fractionation line, and the black line is the carbonaceous chondrite anhydrous mixing line. After Jogo et al. (2013), Krot et al. (1998), and Weisberg et al. (2006).

lie in the CV bulk chondrite field, consistent with the possibility that they are derived from the interior of the CV parent planetesimal itself.

There are also several achondrites that may have genetic relationships to CV chondrites (Irving et al. 2004). Although a common parent body has long been suspected between one of the known carbonaceous chondrite groups and ureilites, the second-largest known group of achondrites, recent isotopic measurements strongly disfavor this idea (Warren 2011) (**Supplemental Table 4**). Similarly, a relationship between CV chondrites and the three Eagle Station pallasites has long been suspected on the basis of their similar oxygen isotopic and siderophile element compositions, but this link, too, has been weakened by recent isotopic measurements (**Supplemental Table 5**). Two ungrouped silicate-bearing irons that contain clasts with chondritic bulk compositions, Bocaiuva and NWA 176, also have geochemical affinities to Eagle Station pallasites (Liu et al. 2001, Malvin et al. 1985).

In summary, there are tantalizing but unconfirmed hints that some achondritic materials originated from the CV parent body, some of which may even be found as clasts in CV chondrites. Furthermore, the existence of ureilite and the Eagle Station pallasites makes it certain that carbonaceous chondrite-like parent bodies experienced igneous differentiation and should therefore have been capable of forming metallic cores and dynamo magnetic fields.

## 3.4. Enstatite Chondrites

**3.4.1. Overview.** Enstatite chondrites are extremely reduced rocks composed dominantly of essentially Fe free enstatite, Si-bearing FeNi metal, sulfides, and a diversity of other unusual minerals that originated from at least two (EH and EL) parent bodies (Weisberg et al. 2006). The highly reduced state of enstatite chondrites has long been interpreted as evidence that they were the protolith for aubrites, a class of highly reduced achondrites with which they share similar mineralogies and compositions (Anders 1964, Ringwood 1961). This interpretation has led numerous authors to propose the existence of a partially differentiated parent body composed of aubrites and enstatite chondrites.

**3.4.2. Relationship with aubrites.** A genetic connection between aubrites and enstatite chondrites was strengthened by the discovery that they share identical oxygen isotopic compositions (Clayton et al. 1976). This possibility of partially differentiated enstatite chondrite parent bodies has been further strengthened by the discoveries of primitive enstatite achondrites inferred to be partial melt residues from an enstatite chondrite precursor (Keil 2010) and of a basaltic enstatite achondrite vitrophyre (Keil et al. 2011), which indicate that the full suite of silicate lithologies (unmelted crust, basaltic melts, melt residues, and cumulates) expected for a putative partially differentiated enstatite chondrite body are known.

Although there is general agreement that most aubrites and primitive enstatite achondrites formed as a result of melting of enstatite chondrite–like precursors, Keil (2010) has marshaled a series of arguments against the hypothesis that aubrites are derived from the EH or EL parent body. First, as discussed above, one might expect breccias that contain both lithologies to have formed. Although candidate aubrite clasts formed from partial melting of an enstatite chondrite protolith have indeed been identified in both an EH and an EL chondrite, these clasts may instead be impact melts (Keil 2010). A second possible problem is that no simple way has been identified to produce the high-Ti ($\sim$6 wt%) troilite and low ratios of metal to troilite ($\sim$0.45) found in aubrites as a result of the melting of enstatite chondrites, which have low-Ti ($\sim$0.5 wt%) troilite and high ratios of metal to troilite ($\sim$2.6). However, this seeming inconsistency may reflect silicate melting under reducing conditions and/or cocrystallization of troilite and oldhamite (McCoy 1998). Third, it is difficult to derive high-petrologic-type aubrites, which contain abundant diopside, as partial melts from high-petrologic-type enstatite chondrites (none of which were known to contain diopside) (Keil 2010). However, a diopside-bearing EL6 chondrite has now been identified (Floss et al. 2003). There are numerous other possible tests for a connection between aubrites and enstatite chondrites, including a diversity of new elemental and isotopic measurements (**Supplemental Table 6**). Overall, the new isotopic data, with the possible exception of Ca, continue to be consistent with the proposal for a common parent body for aubrites and enstatite chondrites.

**3.4.3. Paleomagnetism of enstatite chondrites and aubrites.** Exploratory paleomagnetic studies were conducted on both EH and EL chondrites in the 1970s and 1980s (see references in Weiss et al. 2010b). The only enstatite chondrites studied in any detail are the EH4 chondrites Abee and, to a much lesser extent, Indarch (Sugiura & Strangway 1983). Both meteorites appear to have been magnetized by extremely strong fields (0.1 to >1 mT). If these fields were really present on the parent body, they would be the highest paleointensities known for any Solar System object. Given that Abee was metamorphosed to at least 700–900°C (Rubin & Scott 1997), virtually all of its remanent magnetization must have been acquired following accretion. $^{40}Ar/^{39}Ar$ dating suggests that these thermal events occurred ∼4.4–4.5 Ga ago with no major thermal remagnetization events since (Bogard et al. 2010). These ages are too imprecise to distinguish early nebular and solar magnetic fields from a longer-lived planetesimal dynamo as the field source (Elkins-Tanton et al. 2011, Weiss et al. 2008). Nevertheless, if the high paleointensities found by Sugiura & Strangway (1983) are accurate, they would favor a dynamo in the EH parent body given that they are out of the expected range of external field sources (unless the EH parent body happened to form within ∼0.1 AU of the T Tauri Sun, where the Sun's dipolar field could have reached 0.1 mT).

However, an alternative explanation for these extremely high paleointensities is that these meteorites were partially remagnetized by strong fields during sample handling or by collectors' magnets, a common problem for meteorite paleomagnetic studies (Weiss et al. 2010b). Further evidence for this possibility is provided by the pattern of magnetization observed in Abee: Subsamples with fusion crust are magnetized in the same direction as many interior subsamples (unlike what is expected for a meteorite not remagnetized following landing; see **Figure 7*f***), and there is a characteristic large-scale, coherent rotation of magnetization directions away from the fusion crust magnetization direction with depth from the fusion crust (see figures 8 and 9 of Sugiura & Strangway 1983 and compare with **Figures 6*b*** and **7*g***). Additional paleomagnetic studies are therefore necessary before the origin of enstatite chondrite paleomagnetism can be understood.

Of the only five aubrites studied using paleomagnetic techniques (Weiss et al. 2010b), two (Norton County and Pesyanoe) apparently possess stable natural remanent magnetization. If these magnetizations were acquired during cooling below 700°C on the aubrite parent body, they would indicate a stable paleofield about 20% of the intensity of the present-day Earth's magnetic field. However, the complex brecciation and shock history of aubrites make the timing and origin of this magnetization uncertain.

## 4. IMPLICATIONS FOR AND EVIDENCE FROM ASTEROIDS

### 4.1. Partially Differentiated Asteroids with Chondritic Crusts

If some chondrite groups originate from the surface layers of partially differentiated parent bodies, then such objects or their fragments may be extant in the asteroid belt. The large (perhaps of hundreds of kilometers) sizes expected for many of these bodies (see Section 3) imply that many should have survived collisional disruption to the present day (Bottke et al. 2005). Their discovery would not only validate the hypothesis of partially differentiated chondrite bodies but might also explain a longstanding conundrum: ∼80% of known meteorite parent bodies melted and formed metallic cores (see table 1 of Burbine et al. 2002), yet only a small fraction of known asteroids have been positively identified as differentiated (i.e., having a basaltic, primitive achondritic, or metallic surface composition). Even if we excluded iron meteorites from consideration because of their anomalously long collisional lifetimes, this would still leave approximately equal numbers of known chondrite and achondrite parent bodies. Three possible explanations for this discrepancy

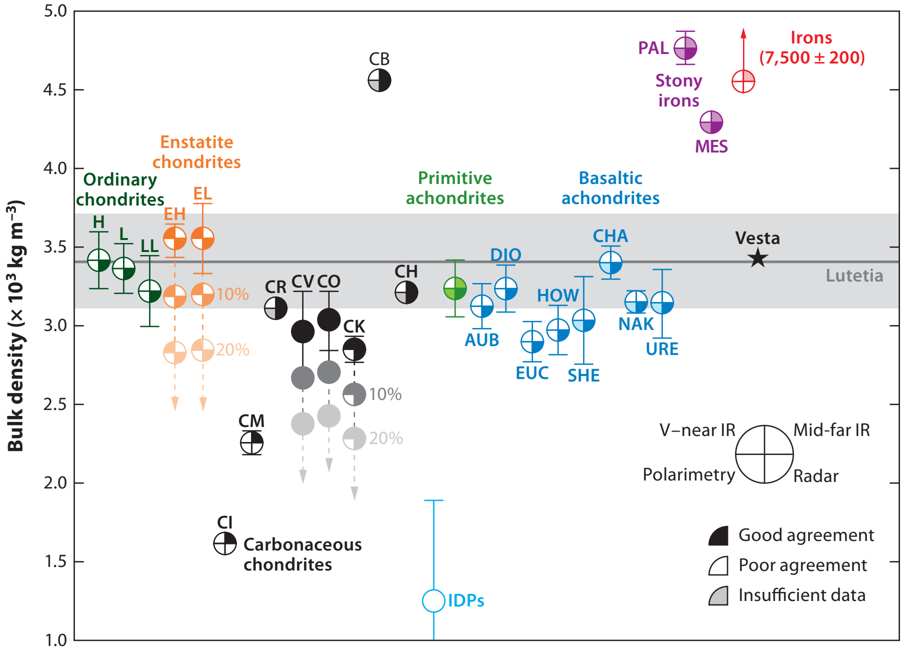


**Figure 9**

Measured bulk densities and observational properties of various meteorite groups compared with asteroid 21 Lutetia. The vertical position of each circular symbol gives bulk density. Quadrants in each symbol denote agreement with four different surface compositional constraints for Lutetia: visible to near-infrared reflectance spectra (0.5–3.0 μm) (*top left*; best-studied constraint); mid-far-infrared reflectance spectra (>3 μm) (*top right*); visible polarimetry (*bottom left*); and OC radar albedo (*bottom right*; second-best-studied constraint). The effect of adding 10% and 20% macroporosity to the bulk densities of bodies composed of select chondrite groups is shown with lighter colored symbols and arrows. The densities of asteroid Lutetia (mean value given by the gray horizontal line and uncertainty range shown by the gray field) (Sierks et al. 2011) and 4 Vesta (*black star*) (Russell et al. 2012) are also shown. Error bars give one standard deviation of meteorites within each group when known. Adapted from Weiss et al. (2012). Abbreviations: AUB, aubrites; CHA, chassignites; DIO, diogenites; EUC, eucrites; HOW, howardites; IDPs, interplanetary dust particles (unmelted chondritic); MES, mesosiderites; NAK, nakhlites; PAL, pallasites; SHE, shergottites; URE, ureilites; IR, infrared; V, visible.

are: (*a*) The meteorite suite is not representative of the present-day asteroid belt (Burbine et al. 2002); (*b*) few large, differentiated asteroids have survived to the present day (Bottke et al. 2006); or (*c*) some asteroid spectral classes typically associated with chondritic bodies in fact contain differentiated members (Gaffey et al. 1993). A fourth possibility is that there are large numbers of asteroids with chondritic surfaces and differentiated interiors (Carporzen et al. 2011b, Elkins-Tanton et al. 2011).

We next discuss our current understanding of the compositions and large-scale structures of asteroids. The ultimate evidence for partially differentiated bodies with chondritic crusts would come from the identification of surface exposures of compositional terranes corresponding to chondritic and achondritic materials. Such variations could be produced by volcanism, outgassing, convective traction, or impacts that breached the chondritic crust and exposed the melted interior. A metallic core could also be identified from geophysical data. As discussed in Section 1, the objects we are seeking are distinct from the primitive achondrite parent bodies of Gaffey et al. (1993), which were instead envisioned to be partially or fully melted throughout essentially all of their interiors and therefore to lack a chondritic crust.

## 4.2. Extant Partially Differentiated Asteroids with Relict Crusts?

**4.2.1. Lutetia.** One of the best studied candidates for a partially differentiated body with a relict crust is the Xc-type asteroid 21 Lutetia, encountered in July, 2010, by the Rosetta spacecraft. With a mean diameter of ~100 km (Sierks et al. 2011), Lutetia is the first asteroid that is unambiguously in the size regime capable of large-scale melting to have been visited by a spacecraft. Its long collisional lifetime and ancient surface ages indicate that it is also sufficiently large to have retained a mostly intact record of any early metamorphic and melting processes.

Weiss et al. (2012) discuss how the surface properties of Lutetia, as measured by Rosetta and ground-based observatories, indicate a primitive surface composition like that of CV, CO, and CK carbonaceous chondrites or possibly enstatite chondrites (**Figure 9**). This is perplexing given that Lutetia's bulk density is 3,400 ± 300 kg $m^{-3}$ (Sierks et al. 2011), among the largest known density for any asteroid and within error of 4 Vesta's density (Russell et al. 2012). Given the uncertainty range for Lutetia's bulk density, if the asteroid has >0% or >13% macroporosity (due, for example, to void space from macroscopic fractures, jointing, and gaps between boulders), then the bulk densities of its rocky constituents would exceed the mean densities of CV/CO/CK or enstatite chondrite groups, respectively. Given that such macroporosities are low for asteroids in Lutetia's size range, Lutetia may be partially differentiated, with a relict chondritic crust and metallic core (**Figure 9**). Although there is no evidence of igneous rocks exposed on the surface, Lutetia's density permits a bulk composition like that of non-CB chondrites and an unmelted crust that is thicker than the deepest unambiguous crater (Weiss et al. 2012), indicating that the absence of clear exposures of igneous rocks is not inconsistent with partial differentiation.

## 4.3. Dynamical Families

**4.3.1. Probing asteroid interiors using collisional fragments.** The ambiguity about Lutetia's interior composition illustrates the key difficulty with searching for partially differentiated asteroids: A relict crust by definition covers the exterior of the body and therefore should conceal the melted interior if it has not been subsequently disrupted. This motivates the alternative approach of examining members of asteroid dynamical families—fragments of single large bodies, many of which were in the size regime capable of igneous differentiation, that have been disaggregated by catastrophic impacts (Cellino & dell'Oro 2010). Such families could provide a stratigraphic cross section across the interior of the parent asteroid. However, a major limitation is that nearly all family members larger than 1 km in size are expected to be reaccumulated rubble piles (Michel et al. 2004), which may make it difficult to identify spectral signatures characteristic of individual achondrite or chondrite types.

Most dynamical families have been found to have spectrally homogeneous members (Mothé-Diniz et al. 2005). This may indicate either that most parent bodies were not differentiated or that reaccumulation processes have masked differentiation by mixing materials from compositionally distinct layers. As a result, the assumption of spectral homogeneity is now routinely used to identify family members from a background of compositionally distinct objects with similar orbital elements (Parker et al. 2008). The latter interlopers are then typically interpreted as genetically unrelated bodies that did not originate as fragments from the parent asteroid. Although this is undoubtedly a powerful approach (e.g., Binzel & Xu 1993), it will inevitably overlook differentiated family members if it is applied rigidly. Here we focus on two asteroid dynamical families that have exceptional spectral diversity and are compositionally distinct from background objects in the surrounding region of the asteroid belt.

**4.3.2. Eos family.** With an estimated 4,400 members (Vokrouhlický et al. 2006), the Eos dynamical family is one of the most numerous and earliest recognized asteroid families. The estimated ~220-km-diameter parent body (well within the size range capable of differentiation and even dynamo generation) was catastrophically destroyed (Cellino & dell'Oro 2010) such that existing family members should contain fragments of the deep interior. Interestingly, the Eos family has the highest diversity of taxonomic classes of any known family (Mothé-Diniz et al. 2008). Many members are of K-complex spectral type, which is uncommon outside the family and is similar to the spectra of CV, CK, CO, and CR carbonaceous chondrites (e.g., Mothé-Diniz et al. 2005). Other family members have spectra resembling stony-iron meteorites, dunites, and primitive achondrites (Mothé-Diniz et al. 2008). This diversity has led to the suggestion that the Eos parent body was partially differentiated (Burbine et al. 2002, Doressoundiram et al. 1998, Mothé-Diniz et al. 2008).

It is predicted that family members are being actively delivered to the 9:4 and 7:3 Jovian mean motion resonances that divide and truncate the family, respectively (Vokrouhlický et al. 2006). Indeed, numerous bodies are presently located in both resonances, many of which have spectra indicative of Eos family membership (Tsiganis et al. 2003, Zappalà et al. 2000). Objects entering these two resonances are expected to become Earth crossers on average after 90 and 6 Ma, respectively (Gladman et al. 1997), which is broadly consistent with the exposure ages of CK, CO, and CV meteorites (Eugster et al. 2006). There is even a possible K-complex near-Earth asteroid that might be one of these transferred bodies (Clark et al. 2010). Therefore, the Eos family may not only be a partially differentiated carbonaceous chondrite parent body, but it could actually be the source of the CV-CK meteorite clan! As described in Section 3.3, a variety of paleomagnetic and petrologic data independently suggest that these meteorites originated from a partially differentiated planetesimal.

**4.3.3. Eunomia family.** The Eunomia family is the largest family in the intermediate part of the main belt, with an estimated >4,500 members (Leliwa-Kopystński et al. 2009). Unlike the Eos parent asteroid, the inferred ~300-km-diameter Eunomia family parent asteroid is thought to have remained mostly intact as the present-day 264-km-diameter S-type asteroid 15 Eunomia (Leliwa-Kopystński et al. 2009). 15 Eunomia appears to have a compositionally variegated surface, with one face resembling stony-iron meteorites (olivine- and possibly also metal-rich) and the opposite face more basaltic (pyroxene-rich) (Nathues 2010). Most family members are also of S type and span the spectral diversity observed across 15 Eunomia [Gaffey subclasses S(I) to at least S(IV)], with a few others scattered among K, L, C, T, and X classes (Mothé-Diniz et al. 2005, Nathues 2010). Recently, at least one confirmed and five additional candidate V types have also been identified (Moskovitz et al. 2008, Roig et al. 2008). These findings suggest that the mineralogies of members might range from dunitic achondrites, to basaltic achondrites, to ordinary chondrites and/or primitive achondrites.

## 5. IMPLICATIONS FOR PLANETESIMAL ACCRETION

As discussed in Section 2, essentially all recent thermal models of planetesimal growth and differentiation indicate that objects will melt substantially while retaining chondritic crusts if at least three major conditions are met: (*a*) They began to accrete within ~1.5 Ma of CAI formation (required to produce substantial melting within the lifetime of $^{26}$Al); (*b*) they continued to accrete, perhaps sporadically, until at least ~2.0–2.5 Ma after CAI formation (sufficiently long to permit the construction of a substantial unmelted crust); and (*c*) they grew to final radii of $>\sim$5–7 km, possibly reaching hundreds of kilometers or larger (such that they were able to efficiently retain radiogenic heat in their interiors against radiation to space). These conditions, particularly the second, are somewhat uncertain given that planetesimal thermal models have yet to employ detailed simulations of convection and accretion. Despite these uncertainties, it is useful to consider whether these conditions are consistent with meteoritic data and accretion models.

The existence of numerous iron meteorite groups with parent body Hf/W core formation ages within ∼1.5 Ma of CAI formation (Kleine et al. 2009) indicates that many bodies clearly fulfilled the first criterion. The cooling rates of many of these meteorites indicate that many bodies formed with radii far exceeding 7 km, consistent with the third criterion (Goldstein et al. 2009). A diversity of thermal models for chondrite parent bodies, with outcomes ranging from metamorphism to melting, also indicate final radii that can usually exceed this criterion (Elkins-Tanton et al. 2011, Kleine et al. 2009, Merk et al. 2002, Sahijpal & Gupta 2011). The formation of large bodies is also supported by the present size distribution of the asteroid belt, which contains more than 200 bodies of radius >100 km (Bottke et al. 2005, Morbidelli et al. 2009).

Fulfillment of the second criterion is less certain. Given that known chondrite parent bodies must have finished accreting no earlier than the formation ages of their youngest chondrules, the known spread of Al/Mg and Pb/Pb ages for chondrules from CV, CR, and CB chondrites (ranging from 0–4 Ma after CAI formation) (Bizzarro et al. 2004, Connelly et al. 2012, Kleine et al. 2009) confirms that some chondrite parent bodies finished accreting after the major early epoch of $^{26}$Al heating. Although this is consistent with the second criterion, it does not establish that a given chondrite parent body accreted continuously over a period of several million years beginning before ∼1.5 Ma after CAI formation. In fact, the distinct sizes and compositions of chondrules for each chondrite group have been suggested to indicate that the accretion period of a given parent body was extremely short (taking place over <1,000 years) in order to avoid turbulent mixing of different chondrule types (Alexander 2005). However, recent Al/Mg chronometry indicates that chondrule populations in individual ordinary and carbonaceous chondrite parent bodies formed over periods lasting ∼1–1.5 Ma, with the youngest chondrules originating 2.4 Ma or more after CAI formation (Hutcheon et al. 2009, Kita & Ushikubo 2011). Furthermore, recent measurements of highly siderophile elements in eucrites and angrites (Dale et al. 2012, Riches et al. 2012) indicate that differentiated planetesimals acquired a late veneer of chondritic material following core formation but prior to solidification of crustal igneous rocks (i.e., between ∼2 and ∼10 Ma after CAI formation). The observed intrasample variability of highly siderophile element abundances means that it is currently not possible to accurately estimate the mass of this veneer, but it could perhaps range up to 1 wt% of the planetesimal mantle mass (Brenan 2012). Finally, new high-precision analyses indicate that unequilibrated H chondrites have oxygen isotopic compositions distinct from those of equilibrated H chondrites, hinting at the possibility of radial stratification in an onion shell parent body (McDermott et al. 2012). Therefore, the available meteoritic evidence permits but does not require that the second criterion has been met by some chondrite parent bodies.

Recent accretion models describe two broad mechanisms for forming planetesimal-sized ($10^1$–$10^2$-km-radius) parent bodies. In incremental accretion scenarios, mutual interactions among planetesimals in the asteroid belt region with initial sizes ranging from <1 to 10 km led to the hierarchical growth of larger bodies (Weidenschilling 2011). Recent simulations (Weidenschilling & Cuzzi 2006) show that 2 Ma after the onset of runaway growth (during which gravitational focusing leads to rapid formation of the largest planetesimals), accretion produces planetesimals with radii ranging from less than ten to several thousand kilometers. The continued accretion of small planetesimals can provide a source of unmelted chondritic material to the surfaces of larger, differentiated bodies. Therefore, these simulations appear consistent with the growth of numerous bodies to sizes sufficient for large-scale melting (i.e., >7 km in radius) and over a time period lasting sufficiently long for the construction of an unmelted crust (i.e., >∼1 Ma). However, a major limitation of the incremental models is that they do not naturally provide a mechanism for planetesimals to grow from meters to kilometers in size in a turbulent nebula.

This problem has motivated accretion models that involve agglomeration of clumps of centimeter- to meter-sized objects concentrated by streaming instabilities or by turbulent eddies into 10- to >100-km-sized solid bodies on timescales of just tens to hundreds of years in the asteroid belt region (Cuzzi et al. 2008, Johansen et al. 2007, Morbidelli et al. 2009). Although such short timescales would seem to be at odds with the second criterion above, for many bodies this early collapse phase may have been followed by a subsequent period of incremental growth. This is because turbulent concentration is an inefficient process, owing to disruption of clumps by rotational breakup and ram pressure from the surrounding gas, such that bodies were likely produced sporadically over millions of years (Chambers 2010), leaving the remaining solid material potentially available to gradually accrete onto the surfaces of these bodies. For example, Ormel & Klahr (2010) found that the accretion of chondrule-sized (∼0.1–1-mm) objects onto 50–100-km planetesimals at 5 AU will double the size of the planetesimals over a period of $10^5$ to $>10^7$ years. However, taking into account the interaction of particle streams with the nebular gas, Johansen & Lacerda (2010) observed that such bodies can accrete 10–50% of their masses within just a few tens of years.

There are numerous uncertainties associated with all of these models, including the presence and nature of turbulence, the possible effects of gravitational perturbations from a growing Jupiter, the timing of mass depletion in the asteroid belt, the lifetime of nebular gas, the relative abundances of gas and dust in planetesimal-formation regions (in particular, the extent to which dust settles to the midplane), the unknown formation mechanism for chondrules (including the possible role of precursor planetesimals), and the uncertainties in the sticking behavior of accreting solids. Therefore, the detection and characterization of partially differentiated planetesimals with relict chondritic crusts could be used to place potentially powerful constraints on the timing and nature of accretional processes.

## SUMMARY POINTS

1. The world's meteorite inventory represents dozens of parent bodies. These bodies were mostly asteroidal in size and ranged from unmelted primordial aggregates, to partially melted objects, to fully melted and differentiated bodies.
2. It is widely assumed that known chondrites did not originate from the same parent bodies as achondrites or primitive achondrites. However, the retention of a chondritic crust on a partially molten planetesimal is a natural outcome of the condition that the surface could have initially been maintained at subsolidus temperatures by radiative equilibrium. This suggests that some chondrites could have been derived from internally differentiated bodies.
3. Planetesimal thermal models find that partially differentiated bodies with molten, differentiated interiors overlain by relict, unmelted chondritic crusts will form if at least three criteria are met: (*a*) Accretion initiates within ∼1.5 Ma of CAI formation, (*b*) accretion extends (perhaps sporadically) over a period lasting approximately an additional one to several million years (with the precise requirement somewhat uncertain), and (*c*) the final body is at least ∼5–7 km in radius.

4. Although these primitive crusts could be subsequently destroyed by explosive volcanism, impacts, or foundering into the molten interior, it seems improbable that all such crusts would be destroyed by these processes.
5. Known meteorite groups might have originated from partially differentiated parent bodies with relict unmelted crusts. Remanent magnetization in CV chondrites is consistent with a partially differentiated parent body with an internal convecting metallic core and dynamo. IIE irons may be composed of materials drawn from a partially differentiated body with an H chondrite crust, basaltic mantle, and metallic core.
6. The partial differentiation hypothesis predicts that there may be hidden igneous diversity in the asteroid belt. It is difficult to identify intact partially differentiated asteroids with chondritic crusts because the melted interior is by definition hidden. For example, 21 Lutetia has a density indistinguishable from that of 4 Vesta but has a surface composition like that of some carbonaceous or enstatite groups.
7. Asteroid collisional families provide a natural stratigraphic cross section of their parent asteroids. The spectral uniformity observed for most families suggests that most of the parent bodies were undifferentiated. Nevertheless, compositional interlopers should be reexamined as potential indicators of a differentiated interior.
8. The existence of partially differentiated planetesimals with chondritic crusts would have several broad implications for planetary science: (*a*) It would indicate that planetary processes produced a continuous range of differentiation end-states in small bodies, (*b*) it could provide a natural explanation for postaccretional magnetization in some chondrites, (*c*) it may constrain the initiation time and duration of accretion, and (*d*) it suggests there may be hidden igneous diversity in the asteroid belt.

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# Contents

# Supplemental Tables for B. P. Weiss and L. T. Elkins-Tanton (2012) Differentiated Planetesimals and the Parent Bodies of Chondrites, *Annu. Rev. Earth Planet. Sci.*

**Table S1.** Characteristics of IVA iron meteorites and their relationship to L and LL chondrites.

| Characteristic | Consistent | Inconsistent | References |
|---|---|---|---|
| Isotopic composition of silicates | O | | a |
| Isotopic composition of metal | N[1], Ni | Cu | b, c, d, e |
| Elemental composition of metal | Low Ni/Fe, Ge/Ni, and Ga/Ni of IVA parental core consistent with origin by melting and reduction of L/LL protolith[2]<br><br>Highly siderophile elements broadly consistent[3] | | f, g |
| Elemental composition of pyroxenes | Low FeO content of IVA irons consistent with origin by melting and reduction of L/LL protolith | | f, h |

*Notes:* The first column gives the characteristic being compared between IVA iron meteorites and L and LL chondrites; the second and third columns list how they are similar and/or dissimilar; the fourth column lists the references for the associated measurements. Key: a = Clayton & Mayeda (1996), b = Mathew et al (2000), c = Mathew et al (2005), d = Bishop et al (2012), e = Luck et al (2003), f = Wasson et al (2006), g = McCoy et al (2011), h = Ruzicka & Hutson (2006).

[1]N isotopic composition of the light IVA subgroup is similar to L, LL, and H chondrites but does not distinguish amongst them.
[2]The Ni content, Ge/Ni, and Ga/Ni of IVA iron meteorites are much lower than for L and LL chondrites, but Wasson et al (2006) argue that these differences are explainable by planetary fractionation processes.
[3]Pd and Au modestly enriched in IVA parental melt.

**Table S2.** Characteristics of IIE iron meteorites and their relationship to H chondrites.

| Characteristic | Consistent | Inconsistent | References |
|---|---|---|---|
| Isotopic composition of silicates and chromite | $O^{Y,O}$, $^{54}Cr^{O}$? | | a, b, c |
| Isotopic composition of metal | $N^{Y,1}$, $Ni^{Y}$, $Cu^{2}$ | $N^{O}$ | d, e, f, g, h |
| Cosmic ray exposure ages | 4-15 $Ma^{Y}$ | 50-600 $Ma^{O}$ | i, j, k |
| Impact ages | Similar groupings at ~3.7 and ~4.5 Ga | | l, m |
| Elemental composition of chondritic silicate inclusions | Fa and Fs compositions of Y-791093 and Mont Dieu and Fs content of Netschaëvo similar to H chondrites | Si/siderophiles and Fa composition of Netschaëvo lower than for H chondrites | n, o, p, q |
| Siderophile element composition of matrix metal | Consistent with fractionally crystallized partial melt of H chondrites?[3] | | r, s, t |

*Notes:* The first column gives the characteristic being compared between IIE iron meteorites and H chondrites; the second and third columns list how they are similar and/or dissimilar; the fourth column lists the references for the associated measurements. The Cr isotopic composition of IIE iron meteorites is uncertain due to cosmogenic effects and the fact that only one sample has been measured. IIE irons fall into two young and old age subgroups (clustered at ~3.7 and ~4.5 Ga, respectively) which are denoted above with superscripts $^{Y}$ and $^{O}$, respectively (Bogard 2011, Bogard et al 2000, Goldstein et al 2009).

Key: a = McDermott et al (2012), b = Qin et al (2010a), c = Trinquier et al (2007), d = Mathew et al (2000), e = Mathew et al (2005), f = Moynier et al (2007), g = Luck et al (2003), h = Bishop et al (2012), i = Casanova et al (1995), j = Bogard et al (2000), k = Eugster et al (2006), l = Swindle et al (2009), m = Bogard (2011), n = Bild & Wasson (1977), o = Ikeda et al (1997), p = Van Roosbroek et al (2012), q = McDermott et al (2011), r = Ebihara et al (1997), s = Teplyakova et al (2012), t = Wasson & Scott (2011).

[1]N isotopic composition of the young IIE subgroup is similar to L, LL, and H chondrites but does not distinguish amongst them.
[2]Mean Ni isotopic composition of one young and one old IIE iron meteorite resembles that of H chondrites.
[3]The implications of siderophile element compositions of IIE iron matrix metal are not yet resolved.

**Table S3.** Characteristics of the NWA 011 basaltic achondrite and its relationship to the CR chondrite clan.

| **Characteristic** | **Consistent** | **Inconsistent** | **References** |
|---|---|---|---|
| Isotopic composition of whole rock and leachates | O, Ti[1], $^{54}$Cr[1] | | a, b, c, d |
| Elemental composition of whole rock | High Fe/Mn and enrichment in refractory and normal siderophile elements similar to CH and CB groups | Inconsistent with partial melting of CR group? | a, e |
| Cosmic ray exposure ages | ~20 Ma, not inconsistent with ~1-25 Ma range of CR clan | | f, g |

*Notes:* The first column gives the characteristic being compared between NWA 011 and CR chondrites; the second and third columns list whether and how they are similar or dissimilar; the fourth column lists the references for the associated measurements. Key: a = Floss et al (2005), b = Trinquier et al (2009), c = Qin et al (2010a) d = Trinquier et al (2007), e = Boesenberg (2003), f = Patzer et al (2003), g = Eugster et al (2006).

[1]Measured on basaltic achondrite NWA 2976, paired with NWA 011.

**Table S4.** Characteristics of ureilites and their relationship to CV carbonaceous chondrites.

| Characteristic | Consistent | Inconsistent | References |
|---|---|---|---|
| Isotopic composition of whole rock and silicates | Os, Si | O (mean value), Ti, $^{54}Cr$, Ni, C | a, b, c, d, e, f, g, h, i, j |
| Carbon abundance | | More carbon-rich compared to CV chondrites | i, j |
| Noble gas isotopic and elemental composition | "planetary" pattern | | k |
| Cosmic ray exposure ages | ~1-40 Ma | | k, l |

*Note:* The first column gives the characteristic being compared between ureilites and CV chondrites; the second and third columns list whether and how they are similar or dissimilar; the fourth column lists the references for the associated measurements. Key: a = Rankenburg et al (2007), b = Armytage et al (2011), c = Clayton & Mayeda (1996), d = Leya et al (2008), e = Trinquier et al (2009), f = Qin et al (2010b), g = Yamakawa et al (2010), h = Warren (2011), i = Grady et al (1985), j = Kerridge (1985), k = Scherer & Schultz (2000), l = Eugster et al (2006).

**Table S5.** Characteristics of Eagle Station pallasites and their relationship to CV chondrites.

| Characteristic | Consistent | Inconsistent | References |
| --- | --- | --- | --- |
| Isotopic composition of bulk rock, silicates, metal, and/or chromite | O | Ni, $^{54}$Cr?[1], Mo?[2] | a, b, c, d, e, f |
| Siderophile elemental composition of metal | Consistent with partial melting of CV chondrites | | g |

*Notes:* The first column gives the characteristic being compared between Eagle Station pallasites meteorites and CV chondrites; the second and third columns list whether and how they are similar or dissimilar; the fourth column lists the references for the associated measurements. Key: a = Clayton & Mayeda (1996), b = Moynier et al (2007), c = Papanastassiou & Chen (2011), d = Shukolyukov & Lugmair (2006), e = Dauphas et al (2002), f = Burkhardt et al (2011), g = Humayun & Weiss (2011).

[1]Eagle Station chromite and olivine have a heterogeneous Cr isotopic composition.
[2]Mo isotopic measurements on Eagle Station are in disagreement for unknown reasons.

**Table S6.** Characteristics of aubrites and their relationship to enstatite chondrites.

| Characteristic | Consistent | Inconsistent | References |
|---|---|---|---|
| Isotopic composition of whole rock and leachates | O, $^{54}$Cr, Ti[1], Si, Zn?[2] | Ca?[2], $^{88}$Sr/$^{86}$Sr?[2] | a, b, c, d, e, f, g |
| Cosmic ray exposure ages | | 12-120 Ma (aubrites) vs. <1 to 70 Ma (enstatite chondrites)[3] | h, i |
| Impact ages | >4.3 Ga | | j |
| Elemental and mineralogical composition | Overall broadly consistent with fractionally crystallized partial melt of EH or EL chondrites<br><br>Siderophile element compositions of metals and sulfides<br><br>Fe, Ni, Co, Ge and Ga content of bulk metal | High Ti content in troilite?[4]<br><br>Abundant pigeonite in Norton County<br><br>Diopside abundances?[4]<br><br>High olivine content of LAP 03719<br><br>Rare earth element abundances?[5] | i, k, l, m |

*Notes:* The first column gives the characteristic being compared between aubrites and enstatite chondrites; the second and third columns list how they are similar and/or dissimilar; the fourth column lists the references for the associated measurements. Key: a = Newton et al (2000), b = Trinquier et al (2007), c = Trinquier et al (2009), d = Armytage et al (2011), e = Moynier et al (2011b), f = Simon & DePaolo (2010), g = Moynier et al (2010), h = Eugster et al (2006), i = Keil (2010), j = Bogard et al (2010), k = Okada et al (1988), l = van Acken et al (2012), m = Easton (1986).

[1]Measured on the Itqiy primitive enstatite achondrite
[2]Aubrites have isotopically lighter Zn, $^{88}$Sr/$^{86}$Sr, and Ca isotopic compositions relative to enstatite chondrites. For at least Zn and Ca [and probably also Sr (Moynier et al 2011a)], the aubrites and enstatite chondrites lie on the same solar system-wide mass-dependent fractionation line such that their isotopic differences may not be an indicator of nebular heterogeneity. Therefore, is possible that the fractionation between aubrites and enstatite chondrites was produced by parent body differentiation processes as described for Zn (Moynier et al 2011b).
[3]Many aubrites and enstatite chondrites had complex exposure histories, such that a comparison of nominal cosmic ray exposure ages has ambiguous meaning (Lorenzetti et al 2003).
[4]See main text for further discussion on these points.
[5]Athough the rare earth element abundances in aubrites and enstatite chondrites differ, the origin of these differences are poorly understood and therefore do not presently rule out a common parent body.

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