Chondrites and the Protoplanetary Disk

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Key Words

chondrules, solar nebula, planetesimal accretion, meteorite ages

Abstract

Major advances in deciphering the record of nebula processes in chondrites can be attributed to analytical improvements that allow coordinated isotopic and mineralogical studies of components in chondrites and to a wealth of new meteorites from hot and cold deserts. These studies have identified a few rare pristine chondrites that largely escaped heating and alteration in asteroids, which have matrices composed of submicrometer-sized grains of enstatite and forsterite and amorphous silicates, as found in comets. Isotopic analyses of components in pristine chondrites using short-lived nuclide chronometers, Pb-Pb dating, and oxygen isotopes aided by laboratory and theoretical studies of chondrites and differentiated meteorites have provided key constraints on the processes that shaped the early solar system. These processes were once thought to have followed one another sequentially over a period of several million years: chondrule formation; planetesimal accretion; alteration, metamorphism, and melting in planetesimals; and finally, high-velocity collisions between asteroids. Radiometric dating shows, however, that these processes overlapped so that chondrules were still forming in the nebula several million years after early-formed planetesimals had melted and collided. Chondrites are extraordinary mixtures of presolar and solar nebula materials and asteroidal debris.

INTRODUCTION

IDP: interplanetary dust particle

Chondrites are sedimentary rocks principally composed of chondrules, which are roughly millimeter-sized particles that were once wholly or partly molten in the nebula and were deposited with several other kinds of particles at the midplane of the solar nebula. Chondrules are largely composed of olivine, $(Mg_xFe_{1-x})_2SiO_4$, and low-Ca pyroxene, $Mg_xFe_{1-x}SiO_3$, where x is the Mg/(Mg+Fe) ratio, which crystallized in hours or minutes between ~1800 and ~1300 K. These silicates are also major minerals in the chondrite matrix—the fine-grained silicate material that coats chondrules and other coarse chondritic ingredients and in some cases fills the interstices between them (e.g., Scott & Krot 2003). The other two important ingredients of chondrites are the refractory inclusions, which are composed almost entirely of crystalline silicates and oxides rich in Ca, Al, and Ti and formed above 1300 K (MacPherson 2003, MacPherson et al. 2005), and the metallic Fe,Ni grains, which appear to be closely associated with chondrules and coeval with them (Campbell et al. 2005). Three features suggest that the chondritic ingredients come from the solar nebula disk from which the planets formed. First, their bulk chemical compositions match that of the Sun's photosphere (neglecting the incompletely condensed elements H, He, C, N, O, and the inert gases). Second, chondrite matrices contain small amounts of interstellar and circumstellar grains (Clayton & Nittler 2004, Lodders & Amari 2005). Finally, the mineralogy, chemical, and oxygen isotopic compositions of the chondritic ingredients can generally be understood in terms of thermal processing over diverse temperatures in the solar nebula.

The second kind of primitive materials that strike Earth are dust particles ~5–100 μm in size that spiral inward toward the sun under Poynting-Robertson and solar-wind drag and are also broadly chondritic in composition (Brownlee 1985, Bradley 2003). The particles collected in the stratosphere, which are called interplanetary dust particles (IDPs), and those that are recovered from deep-sea sediments or polar ice, which are called micrometeorites (Maurette et al. 1991, Genge et al. 2005), both contain presolar grains like those found in chondrites (Messenger et al. 2003, Yada et al. 2006). Chondrites come from asteroids, whereas IDPs and micrometeorites come from both asteroids and comets. However, friable, C-rich carbonaceous chondrites, such as Orgueil, may be cometary in origin (Gounelle et al. 2006b). IDPs that are very porous, largely anhydrous, and tend to have high atmospheric entry speeds are thought to come from comets, whereas compact hydrated particles, which appear to have lower entry speeds, probably come from asteroids (Bradley 2003, Joswiak et al. 2005). However, there is some overlap in the properties of comets and asteroids, for example, some asteroids contain ice, like comets (Hsieh & Jewitt 2006), and the Trojan asteroids are probably captured cometary bodies (Morbidelli et al. 2005). In addition, as noted below, matrix material in the least-altered, matrix-rich carbonaceous chondrites appears to be related to the cometary IDPs.

Most silicates in chondrites and many grains in cometary IDPs and cometary nuclei are crystals that formed at high temperatures (>1200 K), unlike the interstellar silicates, which are very largely (>99.8%) amorphous (Bradley 2003, Kemper et al. 2004, Brownlee et al. 2006, Wooden et al. 2006). A major goal of research on chondrites and cometary IDPs is to understand how amorphous silicate dust was thermally processed in the solar nebula to form the refractory inclusions, chondrules, in chondrites and the submicrometer silicate crystals present in chondrite matrices and comets. How were these materials accreted into asteroids, comets, and planets?

A wealth of review articles on chondrites and IDPs have been published in volumes edited by Davis (2003), Krot et al. (2005a), and Lauretta & McSween (2006). In addition, the book edited by Reipurth et al. (2007) contains articles about chondrites and protoplanetary disks, and the monograph by Sears (2004) reviews 200 years of chondrite studies on chondrites. Nearly all these papers focus on specific aspects of chondrite studies: overviews by Alexander (2005) and Scott & Krot (2003, 2005a) are exceptions. The textbook by Hutchison (2004) covers the whole field of meteoritics.

In this review, I focus on evidence in chondrites that illuminates the early Solar System processes responsible for the formation of the ingredients in chondrites. Major advances since Wood's (1988) review can be attributed to remarkable improvements in the techniques used for coordinated isotopic and mineralogical studies of components in chondrites, and to a wealth of new meteorites from hot and cold deserts. These studies have identified a few rare pristine chondrites that largely escaped heating and alteration in asteroids. Studies of these rocks by analysis of short-lived nuclides and oxygen isotopes and Pb-Pb dating aided by an amazing wealth of other experimental and theoretical studies of chondrites and differentiated meteorites have provided key constraints on the processes that shaped the early Solar System. These processes were once thought to have followed one another sequentially, each one starting when the previous one ended: chondrule formation; planetesimal accretion; alteration, metamorphism, and melting in planetesimals; and finally, high-velocity collisions between asteroids. We now recognize that these processes overlapped so that chondrites are actually mixtures of presolar, solar nebula, and asteroidal materials.

CHONDRITES

Chondrite Groups

Chondrites are divided into 16 groups based largely on their bulk chemical composition, mineralogical properties, and oxygen isotopic composition (**Table 1**). All but two (the K and R groups) can be assigned to the ordinary, enstatite, or carbonaceous chondrite classes. In addition, there are 12 unclassified carbonaceous chondrites that probably represent another 8 poorly sampled groups (see Krot et al. 2003). Cosmic-ray exposure ages show that most meteorites were exposed to space in a few major impacts and many groups have characteristic shock features suggesting that each group comes from one, or possibly a few asteroids. Note that carbonaceous chondrites have diverse concentrations of carbon: CI and CM chondrites have 1–6 wt.% C, but CO chondrites have <1%. They are now defined on the basis of their abundance of refractory lithophile abundances (**Table 1**) and oxygen isotopic compositions. The proportions

Table 1 Chondrite groups and their properties (Scott & Krot 2003)

Group	Type	Refractory inclusions (vol. %)	Chond. (vol.%)[a]	Chond. ave. diam. (mm)	Fe,Ni metal (vol. %)	Matrix (vol.%)[b]	Fall freq.[c] (%)	Refract. lith./Mg rel. CI[d]	Examples
Carbonaceous									
CI	1	<0.01	<5	–	<0.01	95	0.5	1.00	Ivuna, Orgueil
CM	1–2	5	20	0.3	0.1	70	1.6	1.15	Murchison
CO	3	13	40	0.15	1–5	30	0.5	1.13	Ornans
CV	2–3	10	45	1.0	0–5	40	0.6	1.35	Vigarano, Allende
CK	3–6	4	15	0.8	<0.01	75	0.2	1.21	Karoonda
CR	1–2	0.5	50–60	0.7	5–8	30–50	0.3	1.03	Renazzo
CH	3	0.1	~70	0.05	20	5	0	1.00	ALH 85085
CB_a	3	<0.1	40	~5	60	<5	0	1.0	Bencubbin
CB_b	3	<0.1	30	~0.5	70	<5	0	1.4	QUE 94411
Ordinary									
H	3–6	0.01–0.2	60–80	0.3	8	10–15	34.4	0.93	Dhajala
L	3–6	<0.1	60–80	0.5	3	10–15	38.1	0.94	Khohar
LL	3–6	<0.1	60–80	0.6	1.5	10–15	7.8	0.90	Semarkona
Enstatite									
EH	3–6	<0.1	60–80	0.2	8	<0.1–10	0.9	0.87	Qingzhen, Abee
EL	3–6	<0.1	60–80	0.6	15	<0.1–10	0.8	0.83	Hvittis
Other									
K	3	<0.1	20–30	0.6	6–9	70	0.1	0.9	Kakangari
R	3–6	<0.1	>40	0.4	<0.1	35	0.1	0.95	Rumuruti

[a]Includes chondrule fragments and silicates inferred to be fragments of chondrites.
[b]Includes matrix-rich rock fragments, which account for all the matrix in CH and CB chondrites.
[c]Fall frequencies based on 918 falls of differentiated meteorites and classified chondrites.
[d]Mean ratio of refractory lithophiles relative to Mg, normalized to CI chondrites.

and the nature of the chondrules, refractory inclusions, matrix, and metal grains vary widely among the groups. The chondrules and other components in a group tend to have characteristic size ranges, reflecting size-dependent concentration in the nebula (Cuzzi et al. 2001, Liffman 2005). Thus each chondrite group appears to be composed of a unique mixture of materials that formed under diverse conditions and were mixed together and accreted from the solar nebula at a specific time and place (e.g., Wood 2005).

CAI: calcium-aluminum inclusion

Carbonaceous chondrites, which are the largest and fastest growing class, are extraordinarily diverse. CI chondrites, which are closest to solar composition and lack chondrules, and CM chondrites are rich in carbon and probably formed further out in the nebula than enstatite and ordinary chondrites. However, the CB and CH chondrites are very poor in carbon and matrix material and rich in metallic Fe,Ni, although they share isotopic characteristics with the matrix-rich CR chondrites (Weisberg et al. 1990, Krot et al. 2002). CB chondrites have so many unique features—very high metal concentrations, extreme volatile depletions, unique chondrule types and metal grains—that they clearly formed under very unusual circumstances. The CB_b subgroup have chemically zoned metal grains and chondrules with unusually diverse refractory/Ni elemental ratios (0.03–10 $\times$ those in CI chondrites), suggesting early formation by condensation following complete vaporization in a dust-enriched nebula (Krot et al. 2001). However, the CB_a subgroup contains S-rich metal grains that are very different and may have formed in an impact plume (Rubin et al. 2003). The mystery deepens because Krot et al. (2005b) discovered that chondrules in both subgroups are younger than chondrules in other groups, and they suggest the chondrules all formed in a collision between planetary embryos. However, CB chondrites also contain calcium-aluminum inclusions (CAIs) and altered matrix-rich clasts, whereas CH chondrites contain a mixture of normal and CB-like chondrules. This suggests that even if they are impact products, the CB chondrules and metal grains seem to have accreted from a turbulent nebula that mixed materials from diverse locations, just like other chondrites.

Alteration and Metamorphism

To understand how the components of chondrites formed in the nebula, we must first distinguish the mineralogical changes caused by metamorphism, aqueous alteration, and impacts on asteroids. Systematic studies to identify metamorphic effects were begun by Van Schmus & Wood (1967), who assigned chondrites to six petrologic types on the basis of simple petrological and chemical criteria. Types 3–6 represent a metamorphic sequence of rocks that were heated in asteroids from $\sim$500 to 900°C (Huss et al. 2006), whereas types 1 and 2 are products of aqueous alteration of type 3 material at temperatures generally below $\sim$100°C (e.g., Brearley 2003, 2006). Rare phyllosilicates may have formed in the nebula during passage of shock waves (Ciesla et al. 2003) or by alteration of micrometer-sized amorphous silicate grains. However, the bulk of the alteration probably occurred in the meteorite parent bodies after ice that accreted with rocky materials was melted. Alteration in smaller planetesimals (Bischoff 1998) cannot be excluded.

The relatively rare type 3 chondrites in the ordinary, CO, and CV groups have been subdivided into 3.0 (least metamorphosed) to 3.9 subtypes (e.g., Chizmadia et al. 2002). Subtypes 3.0 and 3.1 have been divided further into 3.00 to 3.15 on the basis of Cr concentrations in olivine with >2 wt.% FeO (Grossman & Brearley 2005). Mineralogical changes in the chondritic components during alteration and metamorphism are described for all chondrite groups in a comprehensive treatise by Brearley & Jones (1998). The identification of the least altered and most pristine chondrites is a complex process, in large part because of the difficulty in distinguishing nebula and asteroidal alteration effects, the remarkable range in alteration products—from hydrated phyllosilicates to anhydrous minerals such as fayalite and magneite—and the additional complexity caused by shock and brecciation (e.g., Scott 2002). Many chondrites are actually breccias composed of clasts with different metamorphic and alteration histories.

Identifying and distinguishing nebula and asteroidal alteration has been especially controversial for the CV chondrites (Krot et al. 1997, 1998, 2006c; Brearley 2006). The ferrous olivine that coats chondrules and is a major matrix mineral in many CV3s was initially thought to have formed in an oxidizing nebula gas (Kurat et al. 1989, Weisberg & Prinz 1998). However, transmission electron microscopy studies of matrix (Brearley 1999); oxygen isotopic analysis of secondary fayalite, magnetite, and Ca-rich pyroxene (e.g., Choi et al. 2000); and thermodynamic analysis (Krot et al. 2004a, Fedkin & Grossman 2006) all indicate that asteroidal fluid-assisted metamorphism at temperatures above $\sim$200°C was the cause of alteration, and not nebula processing. As a result of these studies, the Allende meteorite, which for many years was the sole source of CAIs and the most studied chondrite, is no longer considered to be one of the most pristine chondrites. Studies of the crystallinity of carbonaceous material and the abundance of presolar grains and gas components show that Allende is a type 3.6 chondrite that was heated to 550–600°C and is one of the two most metamorphosed CV3 chondrites (Bonal et al. 2006, Huss et al. 2006). This recognition has greatly improved our understanding of the properties and origin of pristine CAIs (MacPherson 2003) and matrix material (Scott & Krot 2003). MacPherson (2003) notes that, "the fall of so much Allende material at such a fortuitous time (in 1969) unquestionably had a profound positive influence on our understanding of the early solar system, but the path to progress would have been straighter and quicker if several tons of Vigarano or Efremovka had fallen instead!"

The nature of the heat source that caused melting, metamorphism, and alteration in asteroids has proved controversial (Huss et al. 2006). Rubin (2004, 2005) has argued that for ordinary chondrites, shock heating played a significant role in metamorphism. However, for asteroid-sized bodies, it is doubtful that impacts were a major source of heat (Keil et al. 1997). Electromagnetic induction heating has lost favor, as it requires an unrealistically high solar wind flux. This leaves short-lived isotopes as the most plausible heat sources: primarily ^{26}Al with a minor contribution from ^{60}Fe. Because of their potential importance in dating chondritic components and the insights that their genesis offer into early Solar System processes, the short-lived nuclides have assumed major importance in chondrite studies.

Parent Bodies

This section is necessarily brief as few asteroids have been visited by spacecraft and ground-based spectral studies are not able to link specific asteroids and meteorites (except for Vesta and the so-called HED achondrites). Ordinary chondrites probably come from asteroids with radii of ~100 km (Trieloff et al. 2003) with the spectral characteristics of S-type asteroids, which are concentrated in the inner belt (Burbine et al. 2002, Chapman 2004). But we do not know whether most S-type asteroids contain ordinary chondrite material or if only a few do. Matrix-rich carbonaceous chondrites probably come from C-type and D-type asteroids, which are concentrated around 3.0 AU and beyond 4 AU, respectively. E chondrites may come from M-type asteroids that are concentrated at 2–3 AU, although formation inside 2 AU cannot be excluded.

Our ignorance about the current location of the chondrite parent bodies and their formation locations will hopefully be rectified by new missions and innovative studies of asteroid families, which probably dominate the meteorite sources.

Oxygen Isotopes

Oxygen is a unique element in the solar nebula because it has three or more isotopes and exists in several forms—both gaseous (CO and H_2O) and solid (ice, silicates, and oxides)—over a wide range of temperatures. As a result, mass-independent chemical fractionation of oxygen isotopes in gaseous molecules in the nebula or molecular cloud can be preserved in solids. Mass-independent oxygen effects are so pervasive among chondrules and chondrites that rocks from nearly all meteorite parent bodies, including, for example, the Moon, Mars and Vesta, can be distinguished from one another using oxygen isotopes (Clayton 2007). Thus oxygen isotopes provide an excellent way of characterizing the formation environments of diverse cosmic samples. But without a good understanding of the mechanism that produced these anomalies, we cannot draw firm conclusions about the formation location or age of any chondritic component. Nevertheless, there are several promising models that explain the origin of the oxygen isotopic anomalies in terms of chemically produced mass-independent fractionation (Thiemens 2006, Yurimoto et al. 2006).

On a standard three-isotope plot of $^{17}O/^{16}O$ versus $^{18}O/^{16}O$ with units of deviations in parts per thousand from standard mean ocean water, most chemical reactions fractionate oxygen isotopes according to their masses along a line of slope 0.52 so that rocks from Earth, Mars, and Vesta plot on three separate parallel lines. The exceptions that lie off these lines include molecular atmospheric species and minerals formed by reaction with these species. On this three-isotope plot, refractory inclusions, chondrules, and matrix samples plot on or near a slope-1 line, with the chondrules and matrix samples closest to the terrestrial line and the refractory inclusions furthest away, with $\Delta^{17}O$ values approximately –25‰ (**Figure 1*a***). The vertical deviations from the terrestrial line are given by $\Delta^{17}O = \delta^{17}O - 0.52 \times \delta^{18}O$ (Clayton 1993).

Understanding the origin of the mass-independent oxygen isotopic effects in chondritic components is one of the most important outstanding problems in cosmochemistry. The oxygen isotopic variations among CAIs were initially attributed to

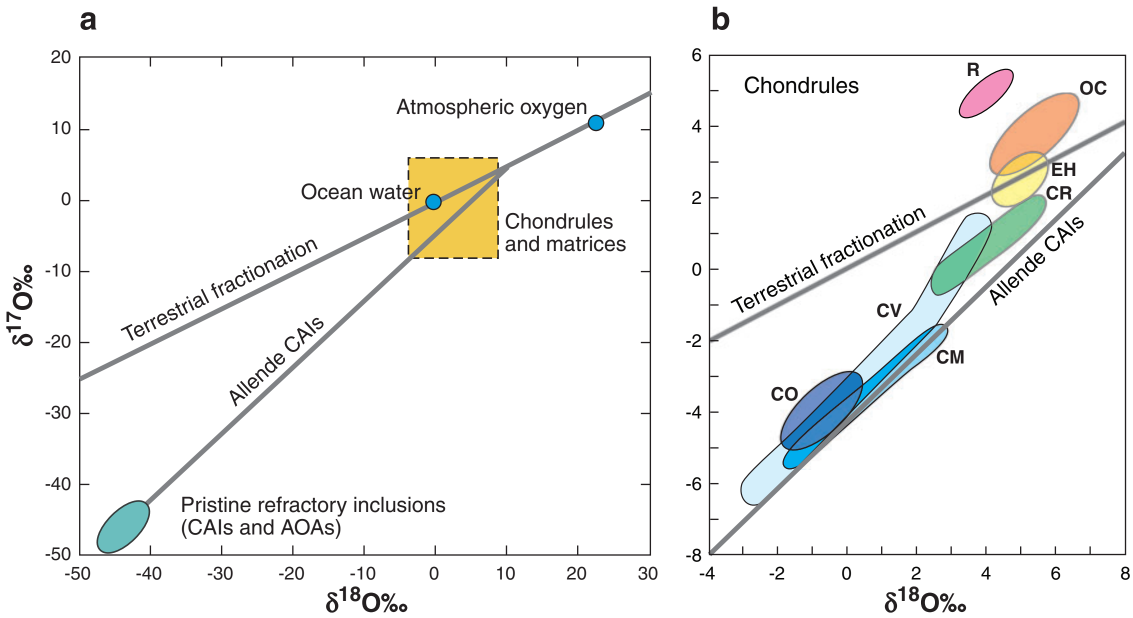

Figure 1

Oxygen isotopic plots of $^{17}O/^{16}O$ versus $^{18}O/^{16}O$ with isotopic compositions plotted as deviations from standard ocean water in parts per 10^3 (δ-units). (*a*) Terrestrial samples show mass-dependent isotopic variations and plot on the line labeled terrestrial fractionation, whereas refractory inclusions in most pristine chondrites plot in the lower left of the diagram. Ferromagnesian chondrules and matrices plot closer to the terrestrial line within the box. CAIs in the Allende chondrite were modified after formation and scatter along the slope-1 line. (*b*) Enlarged view of boxed region in (*a*) showing ranges of oxygen isotopic compositions of large chondrules in carbonaceous chondrites, enstatite chondrites, ordinary chondrites (OCs), and R chondrites. There is little overlap in composition between the chondrules in these four kinds of chondrites. (After Rubin 2000.)

stardust bearing the signature of nucleosynthetic processes in stars (Clayton et al. 1973). However, this explanation is no longer favored as stardust is largely ^{17}O-rich and CAIs do not show comparable anomalies in other elements (Clayton & Nittler 2004). Thiemens & Heidenreich (1983), who first demonstrated mass-independent fractionation using ozone, suggested that the CAI effects were due to chemical effects and invoked self-shielding. They noted that large variations in isotopic abundances of CO had been observed in molecular clouds, presumably from isotopic self-shielding. UV photons that dissociate $C^{16}O$ cannot penetrate beyond the surface of the cloud, whereas inside the cloud, $C^{17}O$ and $C^{18}O$ can be preferentially dissociated. Yurimoto & Kuramoto (2004) suggested that if the atomic ^{17}O and ^{18}O from this process reacted with H to form water ice, then mass-independent oxygen effects could be transferred to Solar System materials as a result of preferential concentration of water ice enriched in ^{17}O and ^{18}O (see Ciesla & Cuzzi 2006). Although many unresolved issues with self-shielding remain and experimental studies are needed (Thiemens 2006), there are several plausible sites for the photodissociation of CO: at the innermost edge of the disk (Clayton 2002), in the protosolar molecular cloud (Yurimoto & Kuramoto 2004), and in the outer surface of the solar nebula disk (Lyons & Young 2005). These models assume that the bulk solar composition has a $\Delta^{17}O$ value around –25‰, like most CAIs.

AOA: amoeboid-olivine aggregate

The bulk oxygen isotopic composition of the Sun has been investigated using collectors on the *Genesis* spacecraft as well as lunar metal grains from the lunar regolith. Oxygen isotopic analyses of solar wind implanted in the outermost surface layers of the lunar metal grains by Hashizume & Chaussidon (2005) revealed the presence of a component with $\Delta^{17}O < -20$‰ that was interpreted as solar wind oxygen. This apparent confirmation of the self-shielding model was, however, contradicted by Ireland et al. (2006) who inferred the presence of a solar wind component on lunar metal grains with $\Delta^{17}O$ around +26‰. Even though the two studies measured solar particles with different energies from samples of very different ages, the conflicting data are difficult to reconcile (Huss 2006).

Despite our ignorance about the bulk solar oxygen composition and the cause of the mass-independent oxygen isotopic variations in the Solar System, oxygen isotopic data are invaluable in helping distinguish diverse processes that established and then modified the composition of the chondritic components. Most CAIs and amoeboid-olivine aggregates (AOAs) in pristine chondrites have $\Delta^{17}O$ –25‰, implying that the gaseous reservoir from which AOAs and CAIs condensed was ^{16}O-rich. However, some CAIs appear to have been formed or altered in an $^{17,18}O$-rich reservoir. CAIs in metamorphosed CO chondrites show large fractionation effects owing to exchange with an $^{17,18}O$-rich aqueous fluid in their parent body (Wasson et al. 2001, Itoh et al. 2004). However, in CR, CH, and CB chondrites, the $^{17,18}O$-rich nature of igneous CAIs suggests that oxygen isotopic exchange occurred during melting either in the chondrule- or CAI-formation region. Oxygen isotopic variations in melilite in several coarse-grained CAIs in CV3 chondrites were interpreted by Yurimoto et al. (1998), Ito et al. (2004), and Yoshitake et al. (2005) in terms of a nebula gas with fluctuating ^{16}O-rich and $^{17,18}O$-rich composition in the CAI formation region.

The oxygen isotopic compositions of chondrules suggest they formed from many different nebula reservoirs, which were richer in $^{17,18}O$ than the CAI source (**Figure 1*b***). Except for refractory-rich, ^{16}O-rich chondrules, chondrules in ordinary chondrites generally plot above the terrestrial fractionation line, carbonaceous chondrite chondrules lie below the line, and chondrules in enstatite chondrites plot near the line. However, within each chondrite, there are nearly always significant $\Delta^{17}O$ variations. Ion probe analyses of minerals in ferromagnesian chondrules in pristine chondrites show that each chondrule is generally rather homogeneous, with $\Delta^{17}O$ variations of <3‰. However, Al-rich and other refractory chondrules are much more heterogeneous because of ^{16}O-rich relict grains (Pack et al. 2004, Krot et al. 2006a). Some type I chondrules also show large ^{16}O variations (e.g., Jones et al. 2004): olivine, low-Ca pyroxene, and glass show sequential increases in $\Delta^{17}O$ that reflect relict grains and gas-liquid exchange (Chaussidon et al. 2006a). We can infer that the oxygen isotopic compositions of chondrules, and presumably their chemical compositions also, reflect the diverse compositions of the materials that were assembled and melted as well as exchange between chondrule melt and $^{17,18}O$-rich nebula gas.

The chondrules in CH chondrites, which are small and fine-grained, show extraordinarily large ranges of mass-independent oxygen isotopic variations, although each is relatively homogeneous. Kobayashi et al. (2003) and Yoshitake & Yurimoto (2004) found that CH chondrules plotted close to the Allende CAI line, with $\Delta^{17}O$ ranging from –15 to +5‰. One cryptocrystalline chondrule was found to be even richer in ^{16}O than CAIs, with a $\Delta^{17}O$ value of –35‰. Thus, aside from this chondrule, oxygen isotopic data indicate that type I and possibly other chondrules appear to have formed by mixing of refractory ^{16}O-rich solids with $\Delta^{17}O$ as low as −25‰ and $^{17,18}O$-rich solids with $\Delta^{17}O$ –3 to +3‰ followed by exchange between melt and $^{17,18}O$-rich gas.

CHONDRITIC COMPONENTS

The primary tool for understanding the mineralogy of chondritic components is the thermodynamic equilibrium diagram for the solar nebula that shows what minerals are stable in a nebula of solar bulk composition (Davis & Richter 2003). **Figure 2** shows such a diagram calculated for a pressure of 10^{-3} bar between 900 and 1800 K. Although very few objects actually represent equilibrium assemblages, the thermodynamic calculations provide a very useful framework for interpreting the mineralogy and bulk compositions of refractory inclusions, chondrules, metal grains, and matrix silicates. Equilibrium condensation calculations have also been performed for systems enriched in various components, such as ice, carbonaceous matter, and chondritic dust (Wood & Hashimoto 1993). Liquids become stable if chondritic dust is enriched by a factor of 10 or more, and under these more oxidizing conditions, significant fractions of the iron condense into silicates rather than in metal (Ebel 2006, Fedkin & Grossman 2006). Petaev & Wood (2005) have also explored the effects of fractional condensation—the continuous removal of small fractions of solids during cooling—as chondritic components typically resemble condensates formed over a narrow temperature interval. Condensation calculations for trace elements are summarized by Lodders (2003). Changes in the gas composition may result from the preferential concentration and evaporation of solids: Ciesla & Cuzzi (2006) have developed models showing this effect for water ice.

Refractory Inclusions

Refractory inclusions are composed of minerals that are stable above 1400 K (**Figure 2**). There are two basic types: (*a*) CAIs that are composed of refractory Ca-Al-Ti minerals, such as corundum, hibonite, grossite, perovskite, spinel, melilite (gehlenite-Åkermanite solid solution), Al-diopside, and anorthite, that are stable above 1400 K and (*b*) the amoeboid olivine aggregates that are composed of various proportions of forsterite and metallic Fe,Ni, which are both stable below ~1450 K, and nuggets of spinel, anorthite, and Al,Ti-pyroxene (MacPherson 2003, MacPherson et al. 2005). CAIs are the oldest known objects that formed in the Solar System and

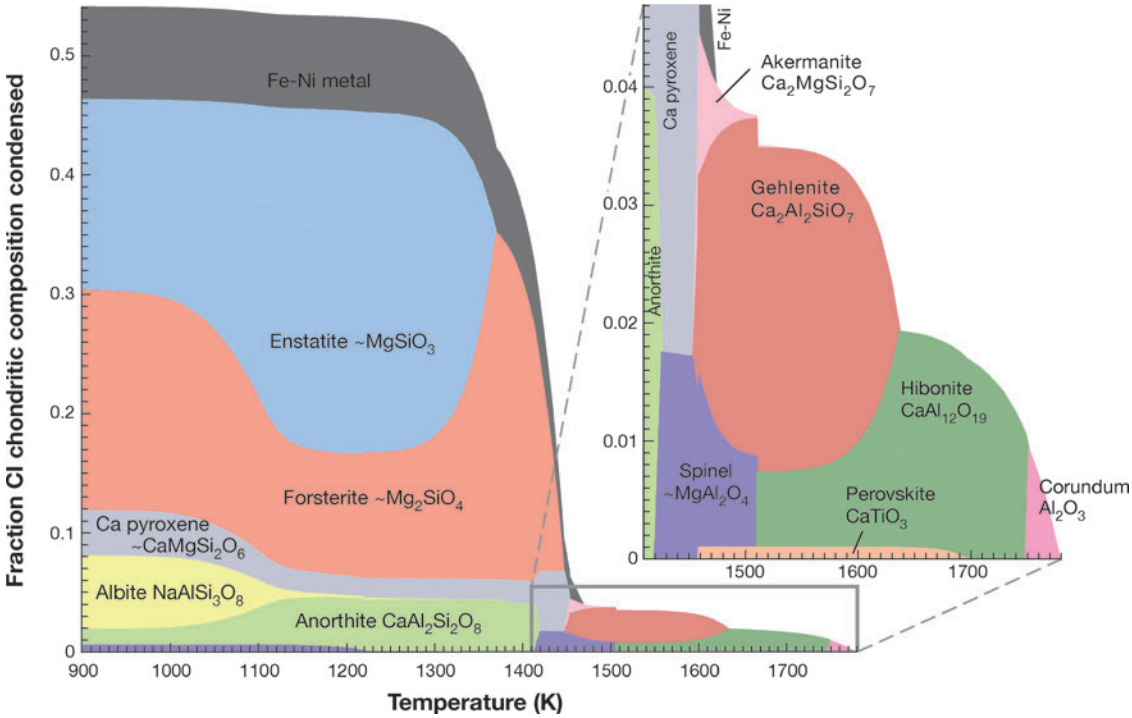

Figure 2

Equilibrium diagram showing which minerals are stable between 900 and 1800 K in a nebula of solar composition at 10^{-3} bar (after Davis & Richter 2003). At 900 K, half the atoms (0.55) in a CI chondrite are in minerals; S and other volatile elements are in the gas. Minerals stable above 1400 K are found in refractory inclusions; minerals stable below 1400 K predominate in chondrules and matrix material. Only three minerals condense entirely from the gas on cooling—corundum (Al_2O_3), forsterite (Mg_2SiO_4), and Fe,Ni metal—the remainder form by reaction between solids and gas. Liquids are unstable unless the total pressure or the dust/gas ratio is increased 10–100 ×.

probably formed in under a few hundred thousand years (see below) in a uniquely ^{16}O-rich environment during the most energetic phase of disk evolution (Wood 2004). They commonly have outermost rims of forsterite with minor refractory minerals that closely match the mineralogy of AOAs, suggesting that AOAs and forsterite rims on CAIs are coeval and postdate CAIs. Preliminary Al-Mg dates suggest that AOAs may be ~0.5 Myr younger (Itoh et al. 2002). Both types of refractory inclusions formed in a highly reducing (solar), ^{16}O-rich environment before chondrules. Roughly equal proportions of AOAs and CAIs are present in nearly all chondrite groups, but their total volume varies enormously from 0.01 to 10% (**Table 1**). CAI sizes in each chondrite group are very roughly correlated with chondrule size, although the range in a given chondrite may be very large. CAIs show mass-independent isotopic anomalies that are larger than those in chondrules; however, they are much smaller than those in presolar grains, suggesting that CAIs formed in the Solar System, not around evolved stars (Nittler 2005).

REE: rare-earth element

Some fine-grained CAIs are thought to be condensates, as they have irregular shapes, fluffy textures, and a mineralogy that is consistent with the condensation sequence (e.g., Simon et al. 2002). The strongest evidence that some CAIs are condensates comes from their group II rare-earth element (REE) patterns, which show much lower abundances of the heavy REE, and requires the prior removal of ultra-refractory elements (Palme & Boynton 1993, Ireland & Fegley 2000). Other CAIs with larger isotopic anomalies may be evaporative residues.

CAIs that are coarse-grained with spheroidal shapes and igneous textures probably formed from fine-grained inclusions by melting and crystallization on timescales of days. The best studied, though atypical, CAIs are the centimeter-sized inclusions (type B) that were first studied in Allende and are rare outside CV3 chondrites. These show mineralogical, chemical, and isotopic evidence for evaporative loss of Mg, Si, and O during crystallization (Davis & Richter 2003). The most common CAIs are spinel-pyroxene and melilite-rich (Lin et al. 2006). Although certain chondrite groups contain characteristic CAIs, e.g., grossite-rich CAIs in CH chondrites, these differences may reflect size-dependent processing in the nebula [e.g., more rapid loss of large CAIs (Cuzzi et al. 2003)], resistance to alteration, diverse sampling techniques, and other factors (Lin et al. 2006). AOAs in different chondrites appear to be identical.

AOAs are irregularly shaped objects with grain sizes of 5–20 μm that occupy up to a few vol.% of chondrites (Chizmadia et al. 2002; Krot et al. 2004b,c; Weisberg et al. 2004). In the least altered chondrites, AOAs are porous and their mineralogy matches that expected for high-temperature condensates, e.g., olivine is forsterite (mostly $Fa_{<1}$), whereas in metamorphosed chondrites, such as Allende where they were first described, the olivine is more Fe-rich (Fa_{5-30}) with the fayalite concentration correlated with the degree of metamorphism (Chizmadia et al. 2002). (Because AOAs are readily altered, they provide an exquisite tool for identifying pristine chondrites.) Some AOAs have recrystallized textures indicative of high-temperature annealing with only very minor melting, and about 10% contain low-Ca pyroxene replacing forsterite at grain boundaries in the AOA periphery or rimming the AOA (Krot et al. 2004c, 2005c). They appear to preserve a record of nebula alteration in ^{16}O-rich and ^{16}O-poor environments.

Chondrules

This section focuses on the properties of ferromagnesian chondrules that help to constrain their origin (Connolly & Desch 2004, Zanda 2004, Hewins et al. 2005, Jones et al. 2005, Lauretta et al. 2006). Al-rich chondrules are not discussed below as they appear to be formed by incorporation of CAIs into ferromagnesian chondrules (e.g., Russell et al. 2005). Unlike CAIs, chondrules formed in $^{17,18}O$-rich environments that were more oxidizing than the CAI region, at much lower ambient temperatures (<1000 K) and at higher total pressures, or higher partial pressures of rock-forming elements (Russell et al. 2005).

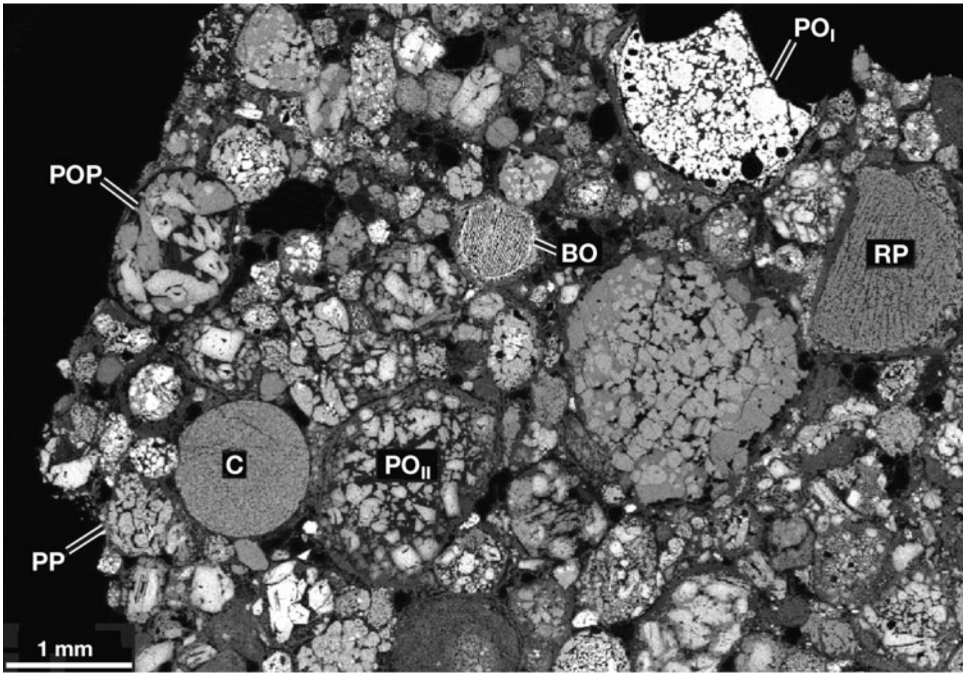

Figure 3

Magnesium concentration map showing chondrules with diverse textures in the Tieschitz ordinary chondrite derived from Mg Kα X-rays generated in an electron microprobe. Most chondrules have porphyritic textures and are rich in olivine (PO) or pyroxene (PP), or less commonly both (POP). The Mg-rich ones are called type I (marked PO$_I$); type II chondrules have lower Mg/(Mg+Fe) ratios (<0.9) (PO$_{II}$). Chondrules without porphyritic textures include those with cryptocrystalline (C), radial pyroxene (RP), and barred-olivine (BO) textures. (From Scott & Krot 2003.)

Chondrule compositions are close to solar except for depletions of elements more volatile than Fe, which are typically more abundant in the matrix. The major minerals in chondrules are olivine and low-Ca pyroxene, with minor glassy mesostasis rich in Ca, Al, Na, and K. Metallic Fe,Ni and FeS are also present in some chondrules. Those with more reduced silicates called type I, in which ferromagnesian silicates have Fe/(Fe+Mg) <0.05–0.1, tend to contain metallic Fe,Ni droplets, which are less abundant in the more oxidized type II chondrules. Chondrules show many different textures that reflect their diverse chemical compositions, peak temperatures (1400–1850°C) and cooling histories (**Figure 3**). Those heated above their liquidus temperatures show fine-grained, cryptocrystalline textures, whereas those heated below preserve their nuclei and develop porphyritic textures. Synthetic chondrules that have the appropriate textures and chemical zoning in olivine have cooling rates of 10–1000°C/hour (Hewins et al. 2005). Chondrules have very diverse proportions of olivine and pyroxene: olivine abundances typically vary from 80 to 20 vol.%.

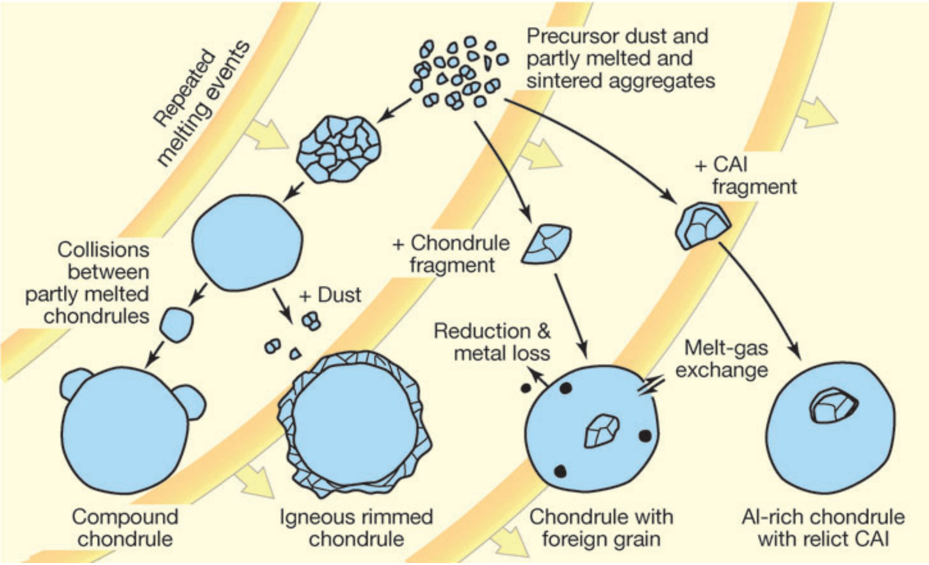

Figure 4

Sketches illustrating the processes involved in chondrule formation: heating and melting of dust, growth by collisions between solid particles and melted or partly melted objects, and exchange between gas and melt. These processes account for the existence of fragments of chondrules and CAIs within chondrules, igneous rims, and adhering chondrules on the exterior, and the growth of pyroxene around type I chondrules.

Much can be learned about chondrule formation from studies of igneous rims, relict grains, and compound chondrules, which all formed as a result of collisions involving chondrules that were molten or partly molten (**Figure 4**). About 15% of chondrules have relict grains that did not crystallize in situ and are fragments or grains from other chondrules, or refractory material. For example, type I chondrules may contain fragments of type II chondrules and vice versa (Jones et al. 2005). Because types I and II chondrules formed in different environments, chondrule formation was clearly a repetitive process that involved mixing of materials from diverse sources. From the abundance of compound chondrules and plausible estimates of the duration of melting and chondrule speeds, a density of 1–10 chondrules per meter cubed can be inferred in the region where chondrules were molten, 10^{2-3} times higher than predicted for canonical nebula conditions (Gooding & Keil 1981, Cuzzi & Alexander 2006). A few rare chondrules contain fragments of CAIs, which were melted to various degrees consistent with the radiometric ages that show CAIs formed before chondrules. Relict chondrules in igneous CAIs appear to represent remelting of CAIs in an ^{16}O-poor chondrule-forming region (Krot et al. 2005d, Russell et al. 2005), although other interpretations are possible (Itoh & Yurimoto 2003). Finally, many chondrules—half those in CV chondrites and 10% of ordinary chondrite chondrules—have igneous fine-grained rims, which formed by collisions between dust and partly molten chondrules or remelting of accretionary rims (Krot & Wasson 1995). These rims show that the dust in the type I regions was partly made of magnesian olivine and pyroxene, whereas in the type II regions, silicates were more ferroan.

Chondrule formation was once considered as a two-stage process in which dustballs of various compositions were first formed and then melted by some flash heating mechanism. However, we now recognize that many other processes were involved. The abundance of relict grains, rims, and compound chondrules suggests that chondrules grew by collisions and remelting. Thus, melting and sintering effectively promoted particle growth from micrometer-sized dust, at least in the region where carbonaceous material and ice were scarce. The melting process was also instrumental in modifying the chemical, mineralogical, and oxygen isotopic compositions as a result of gas-melt reactions, at least for type I chondrules. The pyroxene-rich igneous rims on type I chondrules were reproduced by Tissandier et al. (2002) by reacting gaseous SiO with chondrule melts. These experiments and the discovery of Si-rich igneous rims on CR type I chondrules show clearly that Si and volatile elements like Na condensed on cooling during the formation of these chondrules and provide an explanation for the formation of pyroxene-rich chondrules from olivine-rich chondrules (Krot et al. 2004d). Olivine-rich chondrules may be older than pyroxene-rich chondrules (Mostefaoui et al. 2002), but it is surprising that current Al-Mg dating techniques would detect such a difference.

An important constraint on chondrule formation comes from the lack of evidence for loss of volatiles such as Na, K, and S from molten chondrules. If type II porphyritic chondrules cooled at 100 K/hour as inferred from the olivine zoning, Na and S should have been readily lost by evaporation from the melt (Hewins et al. 2005, Nagahara et al. 2005). Only if the chondrules were immersed in a dense atmosphere with dust enriched by at least 1000 times could the volatiles be retained. Lack of evidence for isotopic mass fractionation of K and Si in diverse chondrules also suggests that chondrule melts were stable (see Davis et al. 2005). Wasson & Rubin (2003) attributed the preservation of volatiles in chondrules to numerous brief partial melting events followed by cooling in seconds. However, this suggestion fails to account for the chemical zoning of olivine and is inconsistent with the appreciable K isotopic fractionation observed in cosmic spherules that were melted for seconds on atmospheric entry (Davis et al. 2005).

According to Cuzzi & Alexander (2006), the lack of isotopic fractionation in chondrules cannot be attributed to high gas pressures—chondrules must have been densely packed when molten. They inferred that the partial pressures of volatiles in the nebula were very close to their saturation pressures for chondrule melts, and that the chondrule density was therefore at least $\sim$10 per m^3, 10^3 times higher than in a standard nebula. These conditions also help to explain why chondrule silicates are much richer in FeO than predicted for canonical nebula conditions (**Figure 2**). To ensure that the dense vapor did not diffuse away from a significant fraction of the chondrules before they crystallized, the chondrule-forming regions would have to have been at least $\sim$150 km in size, and probably close to the nebula midplane so that solids could have been concentrated by large factors (see also Desch 2006).

Matrix Material

The fine-grained, silicate-rich material with grain sizes of 10 nm to 5 μm that coats chondrules, refractory inclusions, and, when present, large grains of metallic Fe,Ni is called matrix. It may also fill the interstices between the chondrules and other ingredients. Matrix is a mixture of materials that formed in diverse locations in the Solar System. In addition, there are small amounts (<100 ppm) of presolar grains that are recognized by their large isotopic anomalies: circumstellar oxides and silicates and organic material that formed in the interstellar medium or the cool exterior of the solar nebula (Clayton & Nittler 2004, Lodders & Amari 2005, Busemann et al. 2006). The rims on chondrules and CAIs, which have thicknesses of ~10–100 μm, are mineralogically indistinguishable (Chizmadia et al. 2005), and rim thickness appears to be correlated with chondrule size (see Cuzzi 2004). Most authors infer that chondrule rims were acquired in the nebula (e.g., Cuzzi 2004), but Trigo-Rodriguez et al. (2006) argue that rims formed in asteroidal regoliths. The truth probably lies between these extremes: Transient fluffy rims may have been acquired in the nebula, but permanent rims were probably formed during the earliest stages of planetesimal growth from chondritic materials.

Because of its fine-grained nature, matrix was especially susceptible to alteration during the heating and aqueous alteration of asteroids. Detailed studies of these processes showed that the abundant phyllosilicates in types 1 and 2 chondrites and fayalite-rich olivines in many type 3 chondrites were the products of aqueous alteration and fluid-assisted metamorphism, respectively, in asteroids (Buseck & Hua 1993; Brearley 2003, 2006; Krot et al. 2004a). This realization allowed the unusual matrix mineralogy of ALHA77307 (Brearley 1993), Acfer 094 (Greshake 1997), and a few other carbonaceous chondrites to be identified as pristine nebula materials that are closely related to the constituents of cometary IDPs and the dust in comets (Scott & Krot 2005b). Although there are many diverse views about the origins of chondrite matrix materials (e.g., Huss et al. 2005), detailed studies of minerals in pristine matrices have provided significant insights.

Unaltered carbonaceous chondrite matrices are composed of crystalline Mg-rich silicates and amorphous Fe-rich silicates, with smaller amounts of metallic Fe,Ni, sulfides, refractory oxides, and carbonaceous material. Only 10–100 ppm of the matrix consists of presolar silicates and other minerals (Nagashima et al. 2004, Nguyen & Zinner 2004). Most of the crystalline silicates formed in the solar nebula as olivine or low-Ca pyroxene single crystals, which are 100–1000 nm in size and have Fe/(Fe + Mg) ratios of 0.01–0.05. Many low-Ca pyroxene crystals in the matrix show intergrowths of ortho- and clino-pyroxene, like those in chondrules, showing that they cooled at ~1000 K/hour from the stability field of protopyroxene above 1300 K. Forsterite and enstatite may both show high concentrations of MnO (0.6–2% wt.% MnO). Similar MnO concentrations are observed in AOAs, suggesting that the matrix silicates are also nebula condensates. The amorphous silicate material in the matrices of pristine carbonaceous chondrites is more Fe-rich, and in ALHA77307 it forms 1–5 μm units, which are very heterogeneous (Brearley 1993). Some crystalline silicates may have formed by annealing of amorphous particles, but unless Fe is sequestered in metal or sulfide, one would expect Fe-Mg silicates to have formed by annealing. Condensation under disequilibrium conditions appears a more plausible origin for the silicate crystals (Nagahara et al. 1988, Klöck et al. 1989, Greshake 1997).

Very little is known about the original mineralogy of matrices in ordinary chondrites, as Semarkona, the most primitive type 3 ordinary chondrite, experienced significant alteration. In the Kakangari K3 chondrite, the abundant matrix is composed largely of 200–1500 nm crystals and crystal aggregates of enstatite Fs_{2-5} and forsterite Fa_{0-3} that are similar in composition to its chondrule silicates, which are slightly more reduced—Fs_{4-8} and Fa_{2-4} (Brearley 1989). Similarly, in E chondrites, the matrix minerals closely resemble those in chondrules. In their successful search for presolar silicates in EH3 chondrites, Ebata et al. (2006) found that the matrix, like the chondrules, consists largely of enstatite, metallic Fe,Ni, and sulfides.

Primitive chondrite matrices contain ~100 ppm of micrometer-sized grains of corundum, spinel, and hibonite that were found as a by-product of the search for presolar oxides (Nittler et al. 1994, Choi et al. 1999). Those grains with Solar System isotopic compositions probably formed as dust grains in the CAI-forming region. Possibly related to these grains are the rare micrometer-sized ^{16}O-rich forsterites revealed by oxygen isotopic mapping of primitive matrices (Kunihiro et al. 2005). ^{16}O-rich refractory micrograins are also found in chondritic IDPs (Greshake et al. 1996).

The organic matter in chondrite matrices is largely insoluble and may be partly or wholly interstellar based on large D/H and ^{15}N enrichments (see Alexander 2005). Recently, Busemann et al. (2006) found micrometer-sized hotspots in the matrices of several CR chondrites where the $^{15}N/^{14}N$ and D/H ratios exceeded even those found in IDPs. They inferred that the isotopically anomalous organic grains in both matrices and IDPs formed in the cool edges of the protosolar disk, or more plausibly in the protosolar molecular cloud.

Spectroscopic studies of ecliptic and nearly isotropic comets and laboratory studies of cometary IDPs show that comets, like chondrite matrices, contain submicrometer crystals of FeO-poor olivines and pyroxenes and amorphous silicate grains of comparable size (Bradley 2003, Harker et al. 2005, Nuth et al. 2005, Wooden et al. 2006). The same FeO-poor minerals are common in disks around protostars and may have similar origins (Natta et al. 2006). Cometary silicate grains in IDPs mostly have Solar System oxygen isotopic compositions and formed in the solar nebula (Messenger et al. 2003). Abundant enstatite whiskers and platelets in chondritic IDPs are probably vapor phase condensates, and the platelets show intergrowths of ortho and clinoenstatite, indicating cooling at ~1000 K/hour from 1300 K (Bradley et al. 1983). IDPs contain several kinds of amorphous silicates, but the most abundant are grains of glass with embedded metal and sulfide (called GEMS) that resemble amorphous interstellar grains (Bradley et al. 1999, Rietmeijer 2002).

The similarity of the silicate, refractory grain, and organic materials in cometary IDPs and primitive chondrite matrices suggests that the chondrite matrices formed from silicate dust that resembled the dust where comets formed and that the silicate and organic components at the two sites had closely related origins. Three features suggest that matrix silicates and chondrules formed in related high-temperature events: both cooled in hours from >1300 K, matrix and chondrules are chemically complementary in carbonaceous chondrites (Bland et al. 2005), and the minerals in the matrices and chondrules are identical in Kakangari and the EH3 chondrites. Note that the chondrules in carbonaceous and ordinary chondrites are very diverse and could not have been formed simply by melting their bulk matrix material. Nevertheless, type I chondrules in these chondrites have igneous rims of forsterite and enstatite, which are abundant in their matrix rims. This suggests that chondrite matrix may be a mixture of silicate dust grains from all the regions where the associated chondrules formed.

Achondritic and Chondritic Clasts

Chondrites are not simply collections of particles that formed in the solar nebula or in presolar environments, they also contain rock fragments from chondritic bodies and more rarely from igneously differentiated bodies. The chondritic fragments are different from those that accumulate on asteroidal surfaces over several billion years, which are found embedded in regolith breccias as the latter are shocked and have diverse sizes. Many type 3 chondrites that are not regolith breccias contain so-called dark clasts, which are 100–1000 μm fragments of aqueously altered chondritic material. Their sizes are correlated with those of the accompanying chondrules (i.e., clast sizes increase in the sequence CH<CO<CR<CV, just as their chondrule sizes increase). Some clasts are coated with matrix material, and in CR, CH, and CV chondrites, the clasts are clearly more altered than the host material (Scott & Krot 2003). This suggests that some dark clasts accreted with chondrules from the nebula.

Chondrites also contain rare rock fragments that come from bodies that accreted before the chondrites (e.g., Kennedy et al. 1992, Hutchison et al. 2005). These may have formed closer to the protosun inside 2 AU, whereas the altered chondritic fragments probably formed in the outer part of the asteroid belt. One chondrite, Kaidun, is composed almost entirely of millimeter-sized chondritic fragments that resemble carbonaceous, enstatite, and ordinary chondrites, with a few additional fragments of differentiated bodies (Zolensky & Ivanov 2003). Kaidun appears to have formed in the solar nebula at a location where few chondrules were present and fragmental materials accumulated until they accreted by the same process that concentrated chondrules and other chondritic ingredients, e.g., turbulent accretion (Cuzzi et al. 2001).

CHRONOLOGY

Because the median lifetime of protostellar disks is ~3 Myr (Haisch et al. 2001), we should anticipate that a time resolution of at least 1 Myr is necessary to provide useful chronological information about solar nebula processes. However, long-lived nuclides such as ^{87}Rb and ^{147}Sm that are used to date the Solar System have such long half-lives (1–100 Gyr) that their time resolution is generally limited to >20 Myr. Two types of chronometer have been developed to solve this problem. The ^{207}Pb-^{206}Pb method relies on the decay of ^{235}U and ^{238}U, which have half-lives of 0.7 and 4.5 Gyr, respectively, and in favorable cases it can resolve age differences as small as

0.1–0.5 Myr solely from Pb isotopic analyses (Baker et al. 2005, Amelin 2006). The second type of chronometer is based on nuclides with half-lives of <100 Myr that are now considered to be extinct, which potentially can provide relative ages of early Solar System objects. Although many short-lived nuclides have been studied in chondrites since the discovery of ^{129}Xe from ^{129}I (Reynolds 1960), uncertainties about their origin and distribution in the Solar System and the effects of secondary processing in asteroids have, until recently, precluded their use as reliable chronometers for determining the formation ages of chondritic components.

Short-Lived Radionuclides in the Early Solar System

Short-lived nuclides can provide important constraints on early Solar System chronology, nuclide formation in stars, particle irradiation in the solar nebula, and heat sources for melting asteroids. Thus, possible chronologic interpretations of short-lived nuclide data are commonly affected by constraints from other fields. Thirteen short-lived nuclides have been detected in chondritic components and other extraterrestrial samples (**Table 2**). Their origins, which for certain nuclides are controversial, are reviewed by Goswami et al. (2005) and Wadhwa et al. (2006b). In almost all cases,

Table 2 Properties of short-lived nuclides that once existed in chondrites, their components, and other meteorites

Nuclide	Half-life (Myr)	Daughter	Estimated initial solar abundance		Analyzed objects
^{7}Be	1.5×10^{-7}	^{7}Li	^{7}Be/^{9}Be	6×10^{-3} [a]	CAIs[a]
^{41}Ca	0.10	^{41}K	^{41}Ca/^{40}Ca	1.4×10^{-8}	CAIs
^{36}Cl	0.3	^{36}S	^{36}Cl/^{35}Cl	4×10^{-6} [b]	CAIs, chondrules, alteration minerals
^{26}Al	0.72	^{26}Mg	^{26}Al/^{27}Al	5.8×10^{-5}	CAIs, chondrules, achondrites
^{10}Be	1.5	^{10}B	^{10}Be/^{9}Be	$\sim$6–9 $\times 10^{-4}$ [b]	CAIs
^{60}Fe	1.49	^{60}Ni	^{60}Fe/^{56}Fe	5–10 $\times 10^{-7}$	Chondrules, achondrites
^{53}Mn	3.7	^{53}Cr	^{53}Mn/^{55}Mn	$\sim 9 \times 10^{-6}$	CAIs, chondrules, achondrites, alteration minerals
^{107}Pd	6.5	^{107}Ag	^{107}Pd/^{108}Pd	$\sim 5 \times 10^{-5}$	Iron meteorites, pallasites
^{182}Hf	8.9	^{182}W	^{182}Hf/^{180}Hf	1.1×10^{-4}	CAIs, chondrites, igneous meteorites, planets
^{129}I	15.7	^{129}Xe	^{129}I/^{127}I	10^{-4}	Chondrules, chondrites, igneous meteorites, alteration minerals
^{92}Nb	36	^{92}Zr	^{92}Nb/^{93}Nb	$10^{-5} - 10^{-3}$	Chondrites, igneous meteorites
^{244}Pu	82	Fission products	^{244}Pu/^{238}U	7×10^{-3}	CAIs, chondrites, igneous meteorites
^{146}Sm	103	^{142}Nd	^{146}Sm/^{144}Sm	8×10^{-3}	Chondrites, achondrites

Sources: Kita et al. (2005) and McKeegan and Davis (2003); additional data from Hsu et al. (2006), Bizzarro et al. (2005), Tachibana et al. (2006), Kleine et al. (2005).

[a]Inferred abundance of Chaussidon et al. (2006b) needs confirmation.

[b]Inferred initial abundance for analyzed objects.

the presence of the short-lived nuclide is well established by a correlation between the excess of the daughter nuclide and the concentration of a stable nuclide of the parent element. The initial ratio of $^{26}Al/^{27}Al$, for example, can be determined from the slope of a graph of analyses of $^{26}Mg/^{24}Mg$ against $^{27}Al/^{24}Mg$. If two samples formed from a reservoir in which the $^{26}Al/^{27}Al$ ratio had been homogenized, their relative ages can then be calculated from the inferred initial $^{26}Al/^{27}Al$ ratios and the 0.72 Myr half-life:

$$\Delta t_{1-2}(\mathrm{Myr}) = 0.72/\ln 2 \times \ln[(^{26}\mathrm{Al}/^{27}\mathrm{Al})_2/(^{26}\mathrm{Al}/^{27}\mathrm{Al})_1].$$

The nuclides ^{53}Mn, ^{182}Hf, and ^{146}Sm have inferred initial solar concentrations that are compatible with those estimated for interstellar material owing to continued addition of nucleosynthetic contributions from numerous evolved stars (Meyer 2005). Thus, these nuclides should have been homogeneously distributed in the interstellar medium before the solar nebula formed. However, the inferred initial concentrations of ^{41}Ca, ^{36}Cl, ^{26}Al, and ^{60}Fe exceed the values expected from continuous nucleosynthesis and can only be explained by the addition of freshly synthesized stellar matter to the protosolar cloud or solar nebula disk, or by formation through energetic particle interactions in the disk or protosolar cloud.

For ^{60}Fe, there are no suitable projectiles and targets for the irradiation mechanism to operate, and for the concentrations observed, the only plausible origin is in a nearby supernova (or possibly an asymptotic giant branch star) just before the formation of the Solar System (Mostefaoui et al. 2005, Tachibana et al. 2006). In the case of ^{10}Be and ^{7}Be (whose presence requires confirmation), formation in stars is not possible as the light elements Li, Be, and B are destroyed during nuclear burning in stellar interiors. Assuming that ^{10}Be and ^{7}Be were both present in CAIs, they probably formed by irradiation at the inner edge of the solar nebula disk where magnetic fields reconnected and accelerated protons and helium nuclei to energies of 10 MeV or more (Gounelle et al. 2006a). Such an origin would imply that CAIs formed at the inner edge of the disk <0.1 AU from the protosun, possibly in the so-called X region (Shu et al. 1996). However, ^{10}Be could also have been acquired by trapping of galactic cosmic rays in the protosolar cloud (Desch et al. 2004).

If the short-lived nuclides like ^{41}Ca, ^{26}Al, and ^{53}Mn formed predominantly in an evolved star outside the solar nebula, as Meyer (2005) and others infer, and were well mixed in the solar nebula before the chondritic components formed, they can provide useful constraints on the relative formation ages of these components. However, if these short-lived nuclides formed largely by irradiation in objects in the solar nebula, as Gounelle et al. (2006a) claim, relative ages inferred from the initial abundances of short-lived nuclides will be uncorrelated with absolute Pb-Pb ages (see Gounelle & Russell 2005). These conflicting claims can be tested by modeling to reproduce the inferred initial solar abundances listed in **Table 2** and also by testing whether the relative formation ages calculated from the inferred initial abundances and Pb-Pb dating are consistent. Fortunately, the wide variety of radionuclides with diverse half-lives, volatilities, and metal-silicate preferences provides numerous tests for the concordance of the chronometers. Below, we review constraints on the formation ages of chondritic components from short-lived nuclides and the Pb-Pb isotopic system.

Early Solar System Chronology

To test whether the ^{207}Pb-^{206}Pb method and the short-lived nuclide systems are capable of providing useful chronological information about chondritic components and early Solar System processes, we require chondritic components that formed at high temperatures and escaped significant metamorphism and alteration or igneously formed meteorites that cooled relatively rapidly (in $<10^6$ years). In addition, the samples must have escaped any subsequent impact heating effects that caused isotopic exchange between minerals and perturbed or reset the radiometric clocks. Because of the prevalence of impacts over 4 Gyr and the extensive metamorphism and alteration of chondrite parent bodies that lasted typically for 10–100 Myr, few meteorite samples satisfy these criteria: chondrules and CAIs in a few type 2 and 3 chondrites, a few fine-grained, basaltic meteorites (angrites and eucrites), rapidly cooled achondrites called ureilites, and strongly metamorphosed meteorites called acapulcoites. **Figure 5**

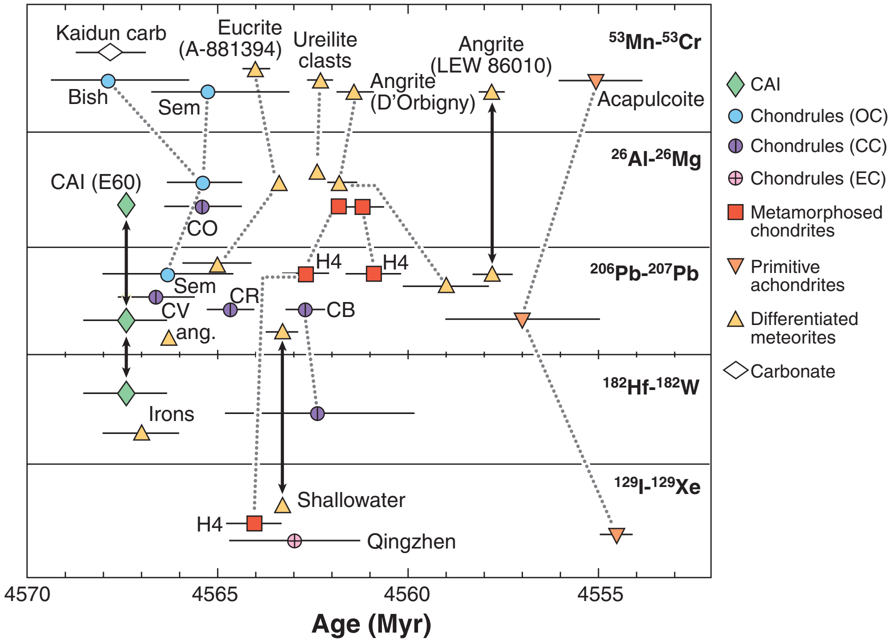

Figure 5

Early Solar System chronology inferred from radiometric ages of CAIs, chondrules in ordinary and carbonaceous chondrites (OC, CC), metamorphosed chondrites (H4), primitive achondrites (acapulcoite), differentiated meteorites (eucrites, ureilites, and angrites), and alteration minerals. The Mn-Cr, Al-Mg, and Pb-Pb data and diagrams are taken from Kita et al. (2005) who list data sources. Additional data: Hf-W (Kleine et al. 2005, Markowski et al. 2006); I-Xe data and Shallowater (Whitby et al. 2002, Gilmour et al. 2006); Kaidun carbonate (Hutcheon et al. 1999). Vertical arrows connect the samples used to anchor the short-lived chronometers (Mn-Cr, Al-Mg, Hf-W, and I-Xe) to the absolute ages derived from Pb-Pb dating. Dashed lines connect data from the same meteorite. 2σ error bars for short-lived chronometers do not reflect additional errors in the ages of the anchors. Key: Bish, Bishunpur; Sem, Semarkona.

shows that the inferred formation ages for these select samples based on ^{207}Pb-^{206}Pb, ^{53}Mn-^{53}Cr, ^{26}Al-^{26}Mg, ^{182}Hf-^{182}W, and ^{129}I-^{129}Xe isotopic systematics are generally consistent. Valuable comparative studies of different chronometers using diverse objects have been provided by Lugmair & Shukolyukov (2001), Shukolyukov & Lugmair (2002), Zinner & Göpel (2002), McKeegan & Davis (2003), and Kita et al. (2005). Relative ages from the short-lived chronometers are anchored to the absolute ages from the Pb-Pb method using CAIs for the Al-Mg and Hf-W systems. For the Mn-Cr and I-Xe systems, achondrites are used as anchors, as CAIs, which are deficient in the volatile elements Mn and I, are readily contaminated so that their precise initial $^{129}I/^{127}I$ and $^{53}Mn/^{55}Mn$ ratios have not been determined directly. Recent reviews of early Solar System chronology have focused on chondritic components (Kita et al. 2005), alteration minerals in chondrites (Krot et al. 2006c), differentiated meteorites (Wadhwa et al. 2006a), planets (Halliday & Kleine 2006), nebular timescales (Russell et al. 2006), and implications for the formation of asteroids and Jupiter (Scott 2006).

If the chronometers are reliable, differentiated meteorites generally formed a few million years after CAIs, which are the oldest objects, and chondrules, but there are exceptions. For example, the two angrites, NWA 1296 and SAH 99555, have $^{206}Pb/^{207}Pb$ ages of 4566.2 $\pm$ 0.1 Myr (Baker et al. 2005), showing that they formed 1.0 $\pm$ 0.6 Myr after CAIs (Amelin et al. 2002) and 1.5 $\pm$ 0.6 Myr before chondrules in CR chondrites (Amelin et al. 2004). In addition, the $^{182}W/^{184}W$ ratios of iron meteorites show that, except for the silicate-rich group IAB irons, molten metallic Fe,Ni and silicate were separated in their parent asteroids less than 1–1.5 Myr after CAIs formed. This can be inferred because Hf isotopes are concentrated in silicate, whereas the W isotopes are concentrated in metallic Fe,Ni, and the $^{182}W/^{184}W$ ratios in irons and are almost identical to the inferred initial value for CAIs (Kleine et al. 2005). When corrections are made for cosmic ray exposure effects in meteoroids, metal-silicate separation occurred 0.3 $\pm$ 1.5 Myr after CAIs formed (Kleine et al. 2005, Markowski et al. 2006). (Initial claims that some irons were significantly older than CAIs were made before the correction for cosmic-ray effects was adequately evaluated.)

The Pb-Pb and short-lived isotope chronometers can also be tested for consistency by dating secondary minerals formed by alteration in asteroids (Krot et al. 2006c). Isotopic data for the I-Xe and Mn-Cr systems both show that carbonates, magnetite, fayalite, and altered chondritic clasts formed up to 10 Myr after CAIs. The youngest secondary minerals are the Kaidun carbonates, which have Mn-Cr ages that are indistinguishable from those of CAIs (**Figure 5**). Thus, these chronometers are consistent and suggest that alteration in asteroids started within 1 Myr of CAI formation.

One concern about the radiometric chronology for the early Solar System is that Baker et al. (2005) concluded that CAIs were 2.5 $\pm$ 0.7 Myr older than Amelin et al. (2002) inferred. Their conclusion was based on $^{206}Pb/^{207}Pb$ ages for two angrites of 4566.2 $\pm$ 0.1 Myr and model initial $^{26}Al/^{27}Al$ ratios inferred from $^{26}Mg/^{24}Mg$ ratios, which implied the angrite parent body melted 3.3–3.8 Myr after CAIs formed. How this discrepancy will be resolved is not clear. Nevertheless, in general, the short-lived nuclides, ^{26}Al, ^{53}Mn, and ^{182}Hf, and the Pb-Pb ages appear to provide a reasonably consistent chronology. This suggests that ^{26}Al and ^{53}Mn were sufficiently well mixed in the early Solar System and that they do provide useful formation ages. Additional evidence for homogeneity of ^{26}Al comes from precise measurements of the bulk Mg isotopic composition of diverse chondrites (Bizzarro et al. 2004). Casting aside major concerns about isotopic homogeneity, we review below the formation ages of CAIs and chondrules.

Formation Ages of Chondrules and CAIs

Pb/Pb and Al-Mg isotopic data show that the first objects to form and survive in the Solar System were CAIs (Amelin et al. 2002, Kita et al. 2005). The isochron defined by a suite of Allende CAIs on a plot of $^{26}Mg/^{24}Mg$ vs. $^{27}Al/^{24}Mg$ shows that CAIs probably formed in $<10^5$ years before any other class of planetary materials with an initial $^{26}Al/^{27}Al$ value of 5.8×10^{-5} (Bizzarro et al. 2004, 2005; Thrane et al. 2006). Internal isochrons for most individual CAIs in pristine chondrites give slightly lower initial $^{26}Al/^{27}Al$ ratios of 4–5 $\times$ 10^{-5} (MacPherson et al. 1995). This difference may arise because the isochrons for whole CAIs date Al-Mg fractionation during condensation, whereas internal isochrons for minerals in individual CAIs date subsequent melting and solidification 0.2–0.4 Myr later (Russell et al. 2005). Possible explanations for the higher $^{26}Al/^{27}Al$ ratio of 7×10^{-5} obtained by Young et al. (2005) are discussed by Kita et al. (2005) and Halliday & Kleine (2006). Rare CAIs and refractory minerals with nonmass fractionation isotopic effects commonly accompanied by large mass fractionation effects (FUN-type inclusions) have initial $^{26}Al/^{27}Al$ ratios that are >10 times less than those in normal CAIs, presumably because they formed before ^{26}Al was introduced (Sahijpal & Goswami 1998). Other CAIs with lower than canonical inferred $^{26}Al/^{27}Al$ ratios but otherwise normal isotopic compositions are generally thought to have been reprocessed in the nebula or parent asteroid (see MacPherson 2003).

As noted above, the most precise absolute ages of CAIs are those of Amelin et al. (2002) who determined a Pb-Pb age for two Efremovka inclusions of 4567.2 $\pm$ 0.6 Myr. [Note that the quoted errors do not reflect uncertainties in the decay constants of ^{235}U and ^{238}U, which generate additional errors of 9 Myr. However, these cancel out when considering age intervals 4.57 to 4.4 Ga (Amelin 2006). Decay constant errors are probably not significant for the other chronometers shown in **Figure 5** except possibly for ^{53}Mn, which has a half-life of 3.7 $\pm$ 0.4 Myr.]

Evidence for live ^{26}Al in ferromagnesian chondrules in LL and CO chondrite groups indicates that they formed 1–3 Myr after CAIs, as their inferred initial $^{26}Al/^{27}Al$ ratios are 2.5–10 times lower than the CAI value (Kita et al. 2005). The 1–3 Myr period was confirmed by Pb-Pb dating by Amelin et al. (2002, 2004), who derived Pb-Pb ages of 4566.7 $\pm$ 1.0 and 4564.7 $\pm$ 0.6 Myr for chondrules in CV and CR chondrites, respectively. This shows that 2.5 $\pm$ 0.8 Myr elapsed between the formation of CV CAIs and CR chondrules. Additional confirmation comes from Al-Mg studies of relict CAIs in chondrules that indicate a period of $\geq$2 Myr between CAI and chondrule formation without any assumption of nebula homogeneity for $^{26}Al/^{27}Al$ (Krot et al. 2006b). Al-Mg and Pb-Pb age data for chondrules have been summarized by Wood (2005) and are shown in **Figure 6**. Because of large uncertainties in the Al-Mg ages

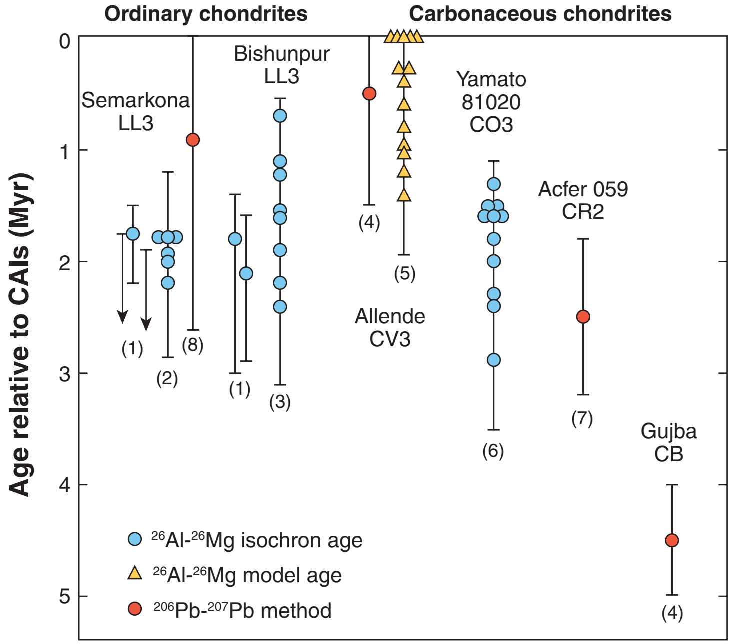

Figure 6

Formation ages of chondrules in ordinary and carbonaceous chondrites relative to CAIs as inferred from the Al-Mg and U-Pb isotopic systems given the Pb-Pb age of CAIs from Amelin et al. (2002) and an initial $^{26}Al/^{27}Al$ ratio in CAIs of 5×10^{-5} [after Wood (2005)]. Both chronometers suggest that most chondrules formed 1–3 Myr after CAIs, although CB chondrules formed 4.5 Myr after CAIs. Sources of data: (1) McKeegan et al. (2000), (2) Kita et al. (2000), (3) Mostefaoui et al. (2002), (4) Amelin et al. (2004), (5) Bizzarro et al. (2004), (6) Kurahashi et al. (2004), (7) Amelin et al. (2002), (8) unpublished data of Amelin and coworkers cited by Kita et al. 2005.

(typically ~0.5 Myr), the total duration of chondrule formation in a specific chondrite group could be much shorter than the ~1 Myr spread in ages suggests.

Mn-Cr dates for chondrules are shown in **Figure 5** (Kita et al. 2005). Although some are nominally older than CAIs, the larger error bars, which do not include errors in the angrite anchor age used to tie the Mn-Cr and Pb/Pb systems, suggest that the two isotopic systems can give reliable formation ages. Most I-Xe data for chondrules, including those in the Semarkona chondrite, which are not shown, show a much wider spread of ages than those from Pb/Pb and Al-Mg systematics owing to the mobility of I and Xe during parent body processing. However, Gilmour et al. (2006), who derive a new age for isotopic closure in the Shallowater achondrite standard of 4563.3 ± 0.4 Myr, suggest that the I-Xe ages for chondrules in EH3 chondrites by Whitby et al. (2002), shown in **Figure 5**, may represent true formation ages.

The CV chondrites appear to contain the oldest chondrules, consistent with the abundance and relatively large sizes of many CAIs in these chondrites (Cuzzi et al. 2003). Although some CV chondrules appear to be as old as CAIs (**Figure 6**)

(Bizzarro et al. 2004), their ages are model ages for whole chondrules rather than isochron ages for individual chondrules and so may reflect Al-Mg fractionation in chondrule precursor material prior to chondrule crystallization (Russell et al. 2005). The chondrite groups with the youngest chondrules are those in the CB chondrites, which as noted above, contain unusual chondrules and metal grains and may have formed in a planetary-scale impact (Rubin et al. 2003, Krot et al. 2005b).

Timing of Metamorphism and Differentiation in Asteroids

The idea that parent bodies of igneous meteorites accreted and differentiated before chondrules formed was a surprise to many, but this is entirely consistent with thermal models for asteroids heated by decay of ^{26}Al and ^{60}Fe. Given the inferred early solar abundances of these nuclides (**Table 2**), any asteroid with a chondritic bulk composition larger than 20–30 km in diameter would have been melted by radiogenic heat provided that the asteroid formed <1–1.5 Myr after CAIs (Hevey & Sanders 2006, Bizzarro et al. 2005). Chondritic parent bodies, which are thought to have been larger than this size limit, therefore accreted >1.0–1.5 Myr after CAIs formed or they would have melted (assuming that thermal buffering by ice was not significant). If one assumes instead that chondrites accreted before the parent bodies of the differentiated meteorites because they are more primitive, the inferred initial ^{26}Al/^{27}Al ratios of chondrules could be taken as evidence that ^{26}Al was not the heat source that melted asteroids and that impacts, for example, were responsible (Kunihiro et al. 2004). However, the low ^{182}W/^{184}W ratios of irons indicating metal-silicate fractionation in <1–1.5 Myr are entirely consistent with the 0.7 Myr half-life of ^{26}Al, and they exclude a major role for impact heating (Scherstén et al. 2006).

If the parent bodies of igneous meteorites were melted largely by heat from ^{26}Al and ^{60}Fe because they accreted quickly, as argued above, the formation ages of chondrules should be correlated with the thermal history of their parent bodies. Provided that the chondritic parent bodies were large enough and accreted quickly after their chondrules formed, we should expect to find that the maximum metamorphic temperature reached in a group of chondrites would be inversely correlated with the chondrule formation age for that group. (The metamorphosed samples would have resided deep inside the parent body, whereas the least metamorphosed samples, suitable for chondrule dating, were located in the cooler surface.) Although we have limited statistics, the data support such a correlation and are remarkably consistent with the thermal modeling of Bizzarro et al. (2005) for asteroidal heating by ^{26}Al and ^{60}Fe (**Figure 7**).

THERMAL PROCESSING IN THE SOLAR NEBULA

Without meteorites we would not have suspected that nebula dust was heated predominantly by brief localized events, rather than nebula-wide heating, and that heating lasted for at least 4 Myr. It seems possible that transient heating events are also common in other protoplanetary disks, but as yet there is no astronomical evidence for the kind of heating events necessary to account for chondrules and other chondritic components (Connolly et al. 2006).

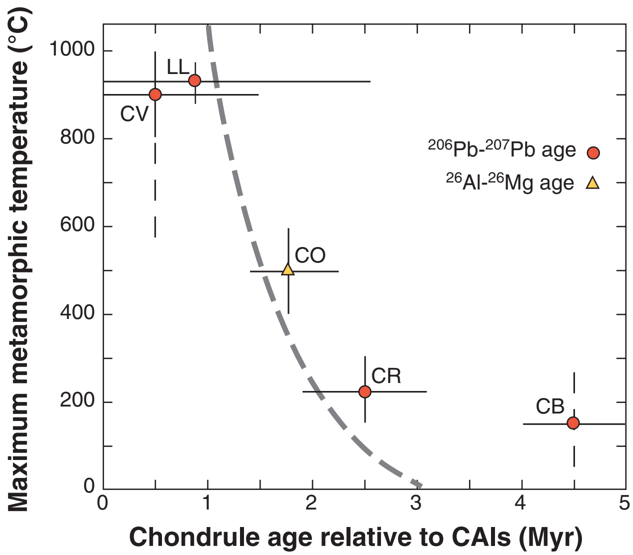

Figure 7

Maximum metamorphic temperatures of chondrite groups decline with increasing chondrule formation age consistent with radioactive heating of chondrite parent bodies. The dashed line shows calculated peak temperatures from Bizzarro et al. (2005) for asteroids >40 km in radius assuming accretion immediately followed chondrule formation and initial ^{26}Al and ^{60}Fe concentrations from Bizzarro et al. (2004) and Mostefaoui et al. (2005), respectively. See **Figure 4** legend for chondrule ages. Peak temperatures: LL (Huss et al. 2006), CO (Jones & Rubie 1991), CR (Keil 2000), CV [Allende 650°C (Huss et al. 2006), metamorphosed CVs (Schoenbeck et al. 2006, Krot and Hutcheon 1997)], CB (this work).

Transient heating mechanisms for both CAIs and chondrules have been reviewed by Ciesla (2005) and Connolly et al. (2006). Chondrule formation models have been evaluated by Rubin (2000), Connolly & Desch (2004), and CAI formation models have been evaluated by Wood (2004). Below, we summarize the four major types of heating processes that have been invoked: formation by impacts into hot or cold planetesimals, heating in the hot inner solar nebula, the X-wind model, and finally, shock heating. **Figure 8** summarizes what seem to be the most plausible mechanisms for heating dust in the disk and making CAIs, chondrules, and the crystalline silicate grains that are so abundant in chondrite matrices and comets.

Chondrules from planetesimals. Since the realization that chondrules probably formed after the first planetesimals had accreted and melted, several workers have argued that planetesimals played a role in forming chondrules (Kleine et al. 2005, Hevey & Sanders 2006). Several possible roles have been proposed for planetesimals; most imply that chondrules were volumetrically insignificant in the Solar System (Sears 1998). Sears (2004) proposes that massive collisions on large, volatile-rich, chondrule-free carbonaceous asteroids resulted in plumes of melt droplets, vapor, dust, and fragments that were gradually deposited on to the asteroid surfaces. Weidenschilling et al.

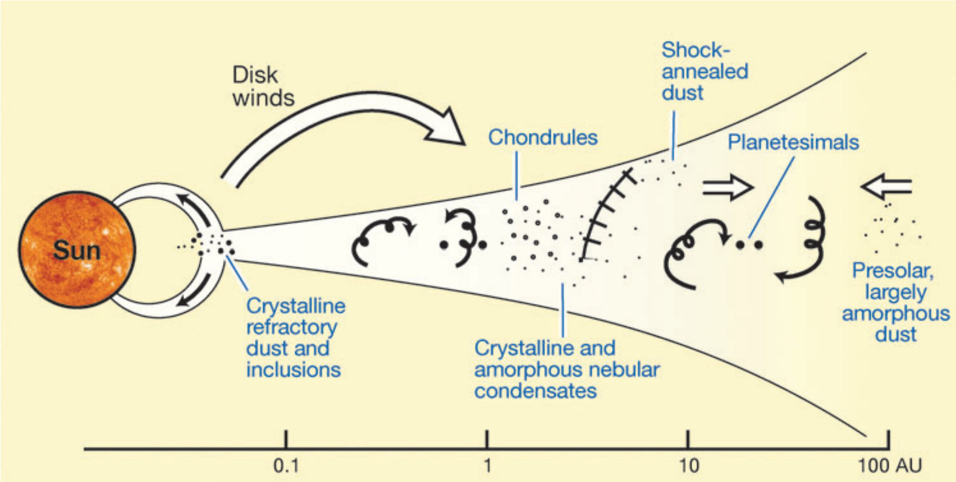

Figure 8

Schematic diagram showing how thermal processes in the nebular disk may have processed amorphous interstellar silicate dust before it accreted into chondritic and cometary planetesimals (after Scott & Krot 2005b). Dust that accreted at 2–3 AU into chondrite matrices and at >10 AU into comets contains traces of refractory dust, which were dispersed from the inner edge of the disk by disk winds and turbulence, and presolar silicates and oxides, which escaped thermal processing. Chondrite matrices and comets contain abundant forsterite and enstatite crystals that may have condensed as a result of heating by the same heat source that made chondrules. Debris from collisions between early-formed planetesimals was also added to the nebula.

(1998) suggest that planetesimal impact debris was melted in the nebula by shocks produced by Ceres-sized bodies and that the melt droplets accreted on to the surfaces of preexisting asteroids or into newly formed planetesimals. Lugmair & Shukolyukov (2001) and Sanders & Taylor (2005) appeal to collisions between molten bodies to disperse melt droplets into the nebula, which then accrete onto existing bodies or into newly formed ones. Finally, Krot et al. (2005b), building on earlier impact models by Rubin et al. (2003), suggest that the unique chondrules in CB chondrites formed from an impact plume of vapor and melt droplets produced by a collision between planetary embryos in the asteroid belt.

In general, most of these models predict that chondrites come from the surface layers of bodies that lack chondrules, or from bodies that accreted in the vicinity of asteroids that melted. However, the first suggestion is difficult to reconcile with the lack of breccias containing chondrule-free and chondrule-bearing materials (e.g., Bischoff et al. 2006), especially as many meteorites are breccias made of materials from different depths in their parent bodies. The second proposal appears contrary to the work of Wetherill & Inaba (2000), who found that early-formed planetesimals would quickly grow into planetary embryos that would dynamically excite any nearby late-formed planetesimals and prevent them from accreting as separate bodies. In addition, igneously differentiated asteroids, besides Vesta, appear to be rare (Burbine et al. 2002).

Planetesimal impact models for making chondrules need further study, but at present they do not appear to have general application. The most plausible applications for impacts are in forming chromite-rich chondrules, which are found only in breccias composed largely of metamorphosed chondrite material (see Scott & Krot 2003), and CB chondrules (Krot et al. 2005b).

Hot solar nebula. The traditional heating source for explaining thermal processing of chondrites and their components has been the hot innermost nebula. This has been invoked to explain various features of chondrites, e.g., depletions of moderately volatile elements in chondrites, to produce a small fraction of zoned metallic Fe,Ni grains in CH chondrites (Meibom et al. 2000), and to form CAIs (Cassen 2001),

It is plausible that fine-grained CAIs may have formed early in the innermost region of the nebula disk, as condensation of REE, for example, would have taken years (Palme & Boynton 1993). Nevertheless, refractory aggregates had to have been periodically removed quickly to ensure that chondrites contain diverse objects that formed at very different nebula temperatures. The X-wind model described below offers one solution to this problem.

X-wind model. The X-wind model of Shu et al. (1996, 2001) and the related bipolar jet model of Liffman & Brown (1996) are outgrowths of models initially devised to understand bipolar outflows from protostars. These authors propose that CAIs and chondrules were both formed at the inner edge of the disk and delivered back to the disk. The inner edge of the disk is truncated at the corotation radius where the Keplerian rotation rate matches that of the protostar. At that point, magnetic fields direct the inward flowing nebula gas, which has been heated and ionized, directly onto the protostar (**Figure 9**). According to the X-wind model, the magnetic field around the inner edge of the disk directs a major portion of the inward-flowing gas

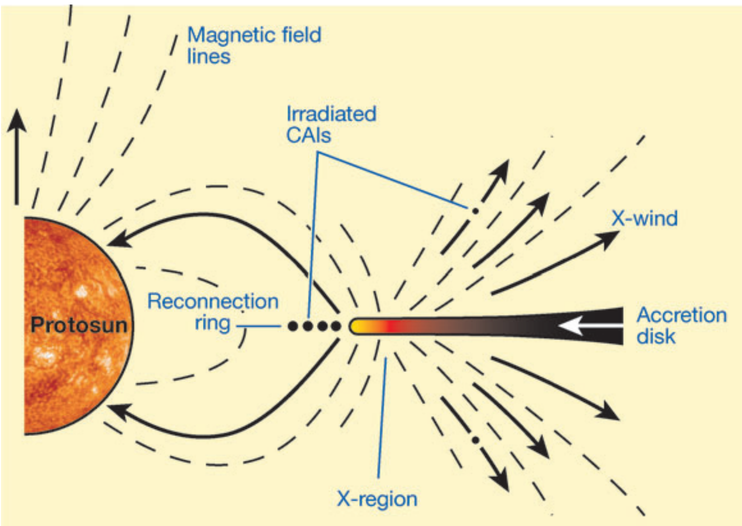

Figure 9
Schematic diagram showing how CAIs (or proto-CAIs) may have formed at the inner edge of the disk (~0.06 AU) where they were irradiated and then transported out to the asteroid region by the X-wind. Model and sketch from Shu et al. (2001).

away from the protostar so that it falls back onto the disk, far from the protostar. Shu et al. (1996, 2001) infer that proto-CAIs were manufactured between the inner edge of the disk and the protosun in the reconnection ring where magnetic reconnection events and impulsive flares analogous to those observed in the modern Sun heated and irradiated refractory particles. Because the magnetic field intensity varied, the position of the X-point fluctuated, causing the X-wind to periodically launch CAIs and chondrules, which were formed by melting disk solids, back onto the disk. In the X-wind model, short-lived nuclides such as ^{26}Al, ^{53}Mn, and ^{10}Be were formed as a result of energetic particle bombardment of proto-CAIs in the reconnection ring (e.g., Chaussidon & Gounelle 2006).

Wood (2004) has evaluated models for CAI formation and concludes that the inner edge of the disk is the most plausible location where solids could have been evaporated and recondensed to form CAIs. The requirement for a short early period of CAI formation favors an origin during the earliest stage in the formation of the protostar, when accretion to the protostar was most rapid and temperatures in the inner disk were highest. Despite problems with some aspects of the X-wind model, Wood (2004) concludes that a model like the X-wind model offers the best explanation for making CAIs and putting them back in the disk. Because detailed calculations of the expected thermal histories of chondrules have not yet been made for the X-wind model, it is unclear whether chondrule formation at the X-point is feasible. One problem is that micrometer-sized dust and chondrules would have been launched on different trajectories, therefore this model cannot readily explain close links in chemistry and mineralogy between chondrules and matrix.

Aside from the short-lived nuclides, firm evidence for nebula irradiation of chondritic ingredients as opposed to regolith exposure is sparse. Possible examples include silica-rich grains embedded in organic matter with ^{18}O/^{16}O and ^{17}O/^{16}O ratios of $\sim 10^{-1}$ (Aleon et al. 2005), and a rare dark inclusion in an ungrouped carbonaceous chondrite, Ningqiang, which has radiation-damaged silicates (Zolensky et al. 2003).

Nebula shocks. Although a very large number of mechanisms have been proposed for making chondrules, only one has been shown to satisfy most of the constraints: shock heating of nebula gas and entrained silicates (Rubin 2000, Connolly & Desch 2004, Ciesla 2005, Connolly et al. 2006). Several groups have shown that millimeter-sized silicate grains in the nebula could have been heated during passage of a shock wave and cooled at the rates inferred from laboratory experiments (Iida et al. 2001, Desch & Connolly 2002, Ciesla & Hood 2002, Desch et al. 2005). The nebular shock model predicts that most chondrules have porphyritic textures, as observed. Micrometer-sized dust could have been evaporated and recondensed in regions where gas densities were high, so that matrix silicates could have been formed by the same heat source that made chondrules. In addition, shocks may have been responsible for making igneous-textured CAIs from fine-grained refractory material when the nebula gas was ^{16}O-rich.

There is no observational evidence for shocks in protoplanetary disks, and a single source of shocks that could have operated over several million years has not been demonstrated. However, it is possible that shocks were generated in a variety of different ways. Among possible sources of multiple shocks are gravitational instabilities in the disk (Boss & Durisen 2005) asteroids in eccentric orbits (Ciesla et al. 2004, Hood et al. 2005), proto-Jupiter, and planetary embryos.

DISCUSSION

The chronology inferred above for the formation of chondrite parent bodies and for differentiated asteroids requires a solar nebula in which CAIs were preserved for several Myr and planetesimals were accreting for the whole of this period.

How Were CAIs Preserved in the Nebula for Millions of Years?

A major reason for questioning a CAI-chondrule age difference of several million years has been the belief that CAIs would have spiraled into the protosun because of gas drag in a small fraction of this time (e.g., Wood 1996). In the absence of turbulence, meter-sized bodies approach the Sun at ~1 AU/century; millimeter-sized chondrules and centimeter-sized CAIs take 10^5 and 10^4 years, respectively, to migrate 1 AU (Weidenschilling 1977). One explanation is that CAIs accreted into planetesimals that were small enough so that they did not overheat and could be disaggregated subsequently (Weidenschilling et al. 1998). However, this is incompatible with the fragile shapes of some CAIs (Wood 2005), and the evidence from iron meteorites that the first planetesimals were not made of CAIs. Two other proposals invoke mechanisms to transport CAIs outward. Disk winds may have launched CAIs from the inner edge to the periphery (Liffman & Brown 1996; Shu et al. 1996, 2001). Second, outward radial diffusion in a weakly turbulent nebula may have counteracted inward drift owing to gas drag (Cuzzi et al. 2003, 2005). Although the degree of turbulence in disks is uncertain, the median lifetime of 3 Myr for protostellar disks suggests that other disks besides the solar nebula were sufficiently turbulent to have preserved small particles for several million years (Haisch et al. 2001, Natta et al. 2006).

Why Did Planetesimal Accretion Last for Several Million Years?

It is possible that the first planetesimals that formed in the inner Solar System were located close to the Sun <2 AU, and that differentiated asteroids and meteorites are collisional fragments of bodies that were subsequently tossed into the asteroid belt (Bottke et al. 2006). In this case, some process delayed the onset of planetesimal accretion in the asteroid belt for more than a million years. Some accretion models assume that planetesimal formation was delayed because chondrules and other chondritic components simply took longer to assemble into planetesimals further from the Sun (e.g., Grimm & McSween 1993, Ghosh et al. 2006). But the chondrule ages suggest that the chondrules that accreted in the asteroid belt were not even formed until 1 Myr after CAI formation. Conceivably, planetesimal accretion was delayed in the asteroid belt inside the snow line until there were sufficient chondrules (Scott 2006).

Understanding planetesimal accretion in the inner Solar System requires more detailed studies of asteroid geology from spacecraft and sample return missions, as well as theoretical models that can accommodate the meteorite chronology. We do not know whether kilometer-sized planetesimals were formed by collisional growth or gravitational instabilities—their origin is "one of the great unsolved problems of planet formation" (Johansen et al. 2006).

SUMMARY POINTS

1. The Allende chondrite is not a pristine chondrite, as was once believed. It was severely altered by fluid-assisted metamorphism in its parent asteroid. Fewer than a dozen relatively pristine chondrites escaped the ravages of asteroidal processing and preserve the best record of nebula processes.
2. High-precision radiometric Pb-Pb and Al-Mg dating and high-resolution ion probe analysis of minerals and components in pristine chondrites combined with formation ages for selected differentiated meteorites based on both short- and long-lived isotopes together provide a consistent outline chronology for the first few million years of Solar System history. ^{26}Al was sufficiently well homogenized in the solar nebula that inferred ^{26}Al concentrations provide useful relative formation ages.
3. Ca-Al-rich inclusions were the first objects to form in the solar nebula: chondrules formed 1–4 Myr later.
4. ^{26}Al was the major heat source that melted asteroids. Chondrites come from bodies that accreted 1–4 Myr after CAIs formed when ^{26}Al was no longer abundant enough to melt asteroids. Iron meteorites come from bodies that accreted <1 Myr after CAIs formed.
5. Matrix material in pristine chondrites is largely composed of submicrometer-sized forsterite and enstatite crystals and amorphous silicates, which are all present in comets. Various links between matrix and chondrule silicates suggest that the crystalline silicate dust was generated from amorphous silicate by the same heat sources that melted chondrules.
6. CAIs and refractory dust probably formed close to the protosun at high ambient temperatures and were widely distributed by nebula turbulence. Most chondrules were probably formed by shock heating in the nebula at much lower ambient temperatures near the midplane where solid particles had already been concentrated.
7. Asteroids and meteorites preserve an exquisite record of planetesimal formation and accretion, but deciphering this record requires geological studies and in situ analyses of asteroids by spacecraft as well as sample return missions and further dynamical studies of the evolution of the asteroid belt.

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