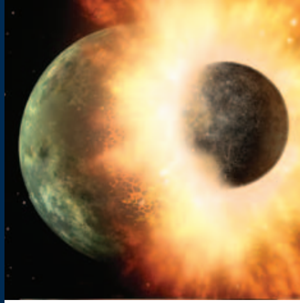
Chronometry of Meteorites and the Formation of the Earth and Moon



Artist's impression of a planetary collision.
IMAGE COURTESY OF NASA

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The planets of the Solar System grew by collisions, starting with the aggregation of tiny dust particles within the solar nebula and culminating in giant collisions between large planetary bodies. These giant impacts occasionally caused the formation of satellites such as the Earth's Moon. Our understanding of planet formation is based on information from various sources, including meteorites – leftovers from the earliest stages of planet formation – and samples from the Earth and Moon. By combining results from isotopic dating of these materials with dynamic modelling of the solar nebula and planet formation, researchers can reconstruct the accretion and early evolution of planetary bodies during the first ~100 million years of Solar System history.

KEYWORDS: meteorites, early Earth, Moon, core formation, radioisotopes, chronology

INTRODUCTION

The Solar System formed ~4.6 billion years ago by gravitational collapse of a localized, dense region of a large interstellar molecular cloud. Contraction of this cloud led to the accretion of a central star surrounded by a rotating disk of gas and dust, termed the solar nebula. The planets accreted from this nebula by the aggregation of dust into larger and larger bodies, a process that occurred in several distinct stages. Settling of dust to the disk's midplane was followed by growth into planetesimals ~1 km in size. Collisions among these planetesimals resulted in the rapid formation, in less than 1 million years (My), of about a hundred Moon- to Mars-sized embryos at 1 astronomical unit (AU). The final stage of planet formation occurred on a more protracted timescale of several tens of millions of years and was characterized by large collisions among the embryos, one of which is thought to be responsible for the formation of the Moon (e.g. Chambers 2004).

Collisions during accretion and the decay of short-lived radionuclides caused melting of planetary interiors, leading to a separation of liquid-metal blobs that sank to the centre of the planet, forming a core (e.g. Stevenson 2008). As a consequence, all major bodies of the Solar System and many smaller objects, such as the parent bodies of iron meteorites, are differentiated. The chondritic meteorites were derived from bodies that never melted, however, and as such provide invaluable hands-on samples for investigating the evolution of the solar nebula prior to the onset of planet formation.

Current thinking on the accretion and early evolution of inner Solar System planets is largely based on isotopic analyses of extraterrestrial matter. These provide key information on the timescale of planetary accretion, on the parameters that controlled planetary evolution, and on the consequences of collisional growth on the composition and differentiation of the planets.

DATING THE EARLY SOLAR SYSTEM

Two Kinds of Chronometers

Both long-lived and short-lived radioisotopes are used for dating the processes that occurred in the early Solar System (TABLE 1). Long-lived radioisotopes (e.g. ^{238,235}U–^{206,207}Pb, ^{147}Sm–^{143}Nd) provide absolute ages, but for processes that occurred ~4.5 billion years ago only U–Pb ages are sufficiently precise. Short-lived radioisotopes (e.g. ^{26}Al–^{26}Mg, ^{182}Hf–^{182}W) are unstable nuclides that existed at the beginning of the Solar System but have since decayed away. Their former presence can thus only be detected via the isotopic composition of their daughter products. Short-lived nuclides are powerful dating tools (TABLE 1) because the abundance of their daughter isotopes changed rapidly over short time intervals, such that the precise measurement of the daughter's isotopic composition can provide relative ages with high temporal resolution. The key assumption is that the short-lived radionuclides were homogeneously distributed throughout the solar nebula. Whether or not this was the case has been a matter of much debate. The broad consistency of the chronology obtained from different short-lived nuclides (Nyquist et al. 2009) and the observation that the initial Mg isotope composition of chondrules is consistent with the ^{26}Al–^{26}Mg evolution of the solar nebula (Villeneuve et al. 2009) provide evidence for a homogeneous distribution of extinct nuclides. However, there is some isotopic variability among inner Solar System planets, reflecting variable proportions of various nucleosynthetic components (e.g. Trinquier et al. 2009). How this affects the short-lived nuclides is currently unclear.

TABLE 1 IMPORTANT CHRONOMETERS OF THE EARLY SOLAR SYSTEM

Parent	Daughter	Half-life (My)	Application
^{26}Al	^{26}Mg	0.73	Relative ages of CAIs and chondrules
^{53}Mn	^{53}Cr	3.7	Relative ages of meteorites and their components
^{182}Hf	^{182}W	8.9	Core formation in planetesimals and terrestrial planets
^{146}Sm	^{142}Nd	103	Mantle differentiation in terrestrial planets
^{235}U	^{207}Pb	703	Absolute age of CAIs; core formation in the Earth
^{238}U	^{206}Pb	4468	

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The Hf–W System

The decay of ^{182}Hf to ^{182}W is particularly useful for studying planetary accretion and core formation (e.g. Kleine et al. 2009). During core formation, tungsten (W) preferentially partitions into the metal while hafnium (Hf) remains in the silicate mantle. Over time the mantle accumulates radiogenic ^{182}W through the decay of ^{182}Hf, while the W isotope composition of the core remains constant. Provided that core formation took place while ^{182}Hf was extant, the mantle will evolve a ^{182}W excess compared to the undifferentiated planet, whereas the core maintains the W isotope composition acquired at the time of core formation and thus develops a ^{182}W deficit. When calculating core formation ages, it is assumed that the Hf–W isotopic evolution of the bulk, undifferentiated planet was similar to that of undifferentiated chondritic meteorites. The basis for this assumption is that Hf and W are refractory elements that were not strongly fractionated by processes within the solar nebula. As such, they should occur in constant relative abundances throughout the Solar System. We will see later, however, that for larger bodies such as the Earth this assumption may not be strictly valid.

FROM DUST TO PLANETESIMALS

Making Chondrites from Nebular Gas and Dust

Chondrites contain Ca–Al-rich inclusions (CAIs), whose composition is similar to that expected for the first refractory nebular condensates and whose age is the oldest known among materials formed within the Solar System. Consequently, CAIs are commonly used to define the age of the Solar System, and the timing of all other events is given in reference to the time of CAI formation (Fig. 1). While CAIs have been precisely dated using the U–Pb system (Bouvier and Wadhwa 2010), their exact age remains somewhat uncertain due to U isotope variations among CAIs (Brennecka et al. 2010). Nevertheless, a combined high-precision U and Pb isotope study of an Allende CAI provides an absolute age of 4567.2 ± 0.5 million years before present (Ma) (Amelin et al. 2010), an age that is consistent with those obtained from short-lived chronometers (Kleine et al. 2009; Krot et al. 2009; Nyquist et al. 2009) and provides the current best estimate for the age of the Solar System.

CAIs probably formed close to the young Sun and were subsequently transported outwards to several astronomical units (AU) where the chondrites accreted (Ciesla 2010). This outward transport was most effective for those CAIs that formed within the first ~10^5 years, such that these CAIs were preferentially preserved. This is consistent with ^{26}Al–^{26}Mg chronometry evidence that CAIs in chondritic meteorites formed over a period of only 20,000 years (Jacobsen et al. 2008).

Chondrules are the major components of most chondrites and are thought to have formed by rapid heating, melting and cooling of dust aggregates within the solar nebula. ^{26}Al–^{26}Mg chronometry indicates that most chondrules in ordinary chondrites formed ~2 million years (My) after CAIs. Chondrules in carbonaceous chondrites appear to have formed over a longer time span, some as late as 4–5 My after CAIs (e.g. Krot et al. 2009). Further, there appears to be peaks in the distribution of Al–Mg ages for individual chondrules from a given chondrite group, suggesting that chondrules from a single parent body formed in distinct events over a period of several million years (Villeneuve et al. 2009). However, chondrules should have been rapidly remixed in a turbulent solar nebula, such that the distinct physical and chemical properties of a given chondrule population could only be preserved by rapid accretion following their formation (Alexander 2005). Reconciling dynamic models of the solar nebula with the chronology of chondrule formation will be an important step towards fully understanding the accretion of chondrite parent bodies.

Remnants of the First Protoplanets

Although chondrites contain some of the oldest material in the Solar System, they are not derived from the first planetary objects. The chronology of chondrules suggests that most chondrite parent bodies accreted more than ~2 My after CAI formation (Krot et al. 2009). In contrast, Hf–W chronology of magmatic iron meteorites – samples from the metallic cores of protoplanets – indicates that accretion and differentiation of their parent bodies must have occurred within ~1 My of CAI formation (Fig. 1). Iron meteorites were thus derived from protoplanets that were already differentiated before chondrules formed and chondrite parent bodies accreted (Kleine et al. 2009).

While iron meteorite parent bodies were traditionally viewed as samples from the metal cores of small bodies (diameter ~20–200 km), it is now thought that they were derived from much larger bodies that were disrupted by hit-and-run collisions (Yang et al. 2007). Iron meteorites may thus be remnants of some of the earliest planetary embryos, which perhaps had formed within the terrestrial planet region and were later scattered in the asteroid belt (Bottke et al. 2006).

^{26}Al and the Evolution of Planetesimals

The meteorite chronometry summarized above indicates that planetesimal accretion started at about the same time as CAIs formed and continued over at least 4–5 My. The following few million years were dominated by a variety of parent-body processes, including the extrusion of lava flows on some asteroidal bodies, peak metamorphism in the ordinary chondrites, and aqueous alteration on some carbonaceous chondrite parent bodies. Although these processes took place in very different thermal regimes, they all occurred ~4–10 My after CAI formation (Fig. 1).

These coeval but different thermal histories of meteorite parent bodies probably reflect their different initial ^{26}Al abundances. There is an inverse correlation between peak temperatures reached inside planetesimals and their accretion age (Fig. 2). Since the variation in accretion times is

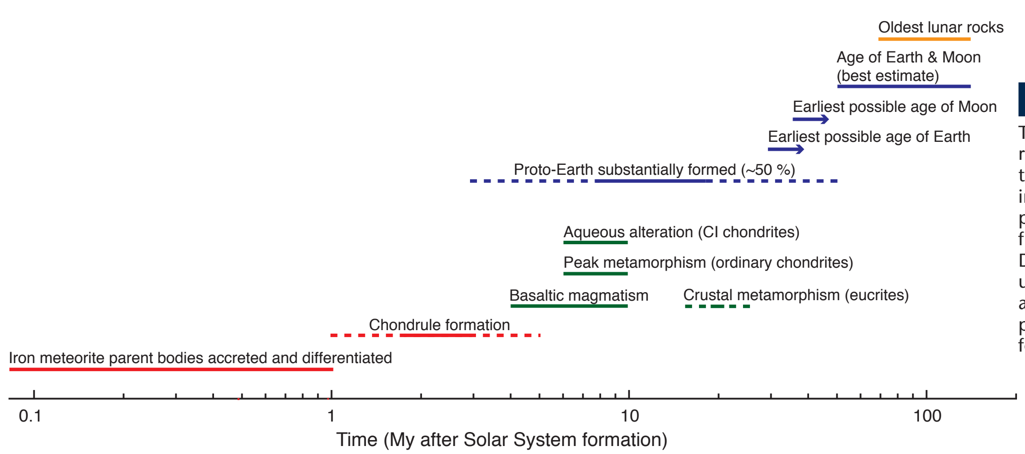

Figure 1 Timescales (logarithmic) for the formation and evolution of inner Solar System planets as determined from isotopic dating. Dashed lines indicate uncertain timescales; arrows indicate earliest possible time of formation.

of the order of several ^{26}Al half-lives, the meteorite parent bodies contained different initial ^{26}Al abundances. In the iron meteorite parent bodies, abundant ^{26}Al facilitated melting and core formation, while in the chondrite parent bodies too little ^{26}Al remained to cause melting. The different thermal histories of the ordinary and carbonaceous chondrites may also reflect variable initial ^{26}Al, but the presence of water in the latter may also have played a role.

ACCRETION OF THE EARTH

Link between Accretion and Core Formation

In the current view of planet formation, Earth accreted most of its mass by collisions with Moon- to Mars-sized embryos. These collisions were highly energetic and raised the temperature of Earth's interior by thousands of degrees, causing widespread melting and the formation of magma oceans. In these oceans, metal easily separated from molten silicate and segregated towards the centre. Thus, accretion and core formation were intimately linked and, as metal segregation is thought to happen much faster than accretion, the rate of Earth's accretion can be determined by dating the formation of Earth's core. Such age constraints can be obtained from the Hf–W and U–Pb systems.

Dating the Earth's Core

Samples from Earth's crust and mantle have uniform W isotope compositions and exhibit small excesses in ^{182}W compared to undifferentiated meteorites. The most straightforward interpretation of this ^{182}W excess is that a fraction of Earth's core formed during the lifetime of ^{182}Hf (Jacobsen 2005; Kleine et al. 2009).

The Pb isotope composition of the bulk silicate Earth provides evidence for a major U/Pb fractionation associated with Pb loss from Earth's mantle 50–150 My after CAI formation (Wood and Halliday 2010). It has been a matter of much debate as to whether Pb was lost to the core or to space. More extreme is the proposal by Albarède (2009) that the U–Pb systematics of the bulk silicate Earth reflect the late addition of volatile-rich material. However, recent experimental data show that Pb is siderophile under the conditions of Earth's core formation, such that the elevated U/Pb ratio of Earth's mantle at least in part resulted from core formation (Wood and Halliday 2010).

The isotopic data have traditionally been interpreted within the framework of an exponential model for Earth's accretion (Halliday 2006). While collisions during Earth's accretion were stochastic, Wetherill (1986) predicted that these collisions occurred at an exponentially decreasing rate. For this reason the exponential model is often used when calculating accretion timescales (Fig. 3). In such models, age information is given as the mean-life of accretion, τ, corresponding to the time taken to accrete 63% of Earth's mass (Jacobsen 2005). The mean-life is the inverse of a rate constant for exponential change.

The end of accretion is not well defined in such a model as the final ~10% of Earth's mass was added by a single collision, which was also responsible for the formation of the Moon. The end of accretion has been determined using a two-stage model, which assumes that the core segregated from a fully formed Earth in a single instant. However, Earth's core formed continuously during the collisional growth of the Earth and not by a single event at a well-defined point in time. The two-stage model nevertheless provides useful age information as it corresponds to 95% growth of Earth's mass in the exponential model (Rudge et al. 2010); as such the two-stage model age provides a reasonable approximation of the time when Earth's accretion terminated in this model.

Two Ways to Make Sense of Tungsten and Lead Isotopes

Using the exponential growth model and assuming that core formation was an equilibrium process, the Hf–W data give a mean-life of accretion of ~11 My, while the end of accretion as given by the two-stage model is ~30 My. Using the same model, the U–Pb data give timescales that are a factor of 2 to 5 times longer. This disparity in calculated accretion timescales has been a matter of much debate, and Figure 3 illustrates two ways to match the Hf–W and U–Pb timescales.

Consistent core formation timescales are obtained when more general accretion models than the exponential model are used (Fig. 3). For instance, a rapid accretion of 63% of Earth's mass in ~1.5 My followed by a more protracted late accretion matches both the Hf–W and U–Pb observations in an equilibrium model (Rudge et al. 2010). It is a common misconception that in such equilibrium models core formation must have terminated ~30 My after CAI formation (Jacobsen 2005; Albarède 2009). About 95% of the Earth has accreted by 30 My in the exponential model for Hf–W, but more general accretion models reach the same 95% accretion point at much later times. For instance, about 95% of the Earth is accreted at 52 My after CAI formation in the equilibrium model shown in Figure 3c.

Another way to match the Hf–W and U–Pb timescales is to invoke disequilibrium during core formation (Fig. 3b). Both systems provide consistent timescales in the exponential model if only 40% of the cores of accreted objects first re-equilibrated with Earth's mantle before entering Earth's core (Rudge et al. 2010).

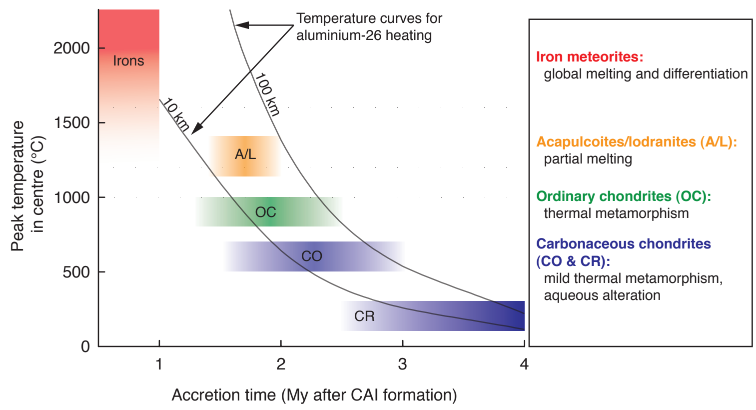

Figure 2 Accretion age versus peak temperatures reached inside meteorite parent bodies. Solid lines show temperatures reached in the centre of planetesimals 10 and 100 km in radius due to heating by ^{26}Al decay, assuming a homogeneous distribution of ^{26}Al throughout the Solar System, with an initial abundance equal to that in CAIs.

Equilibrium or Disequilibrium Core Formation?

Many of the planetary bodies that accreted to the Earth were pre-differentiated into mantle and core. Assessing the degree of equilibration during Earth's core formation thus requires an understanding of what had happened to the cores of accreted bodies during their descent through Earth's mantle. Efficient equilibration would have occurred if the impactor core completely emulsified as small metal droplets in the terrestrial magma ocean (Fig. 4a). This is likely to have happened in the case of small impactor-to-target ratios (Rubie et al. 2007). What happened during large collisions is poorly understood, however. Owing to the enormous length scales involved, the degree of equilibration is difficult to quantify (Rubie et al. 2007). However, Dahl and Stevenson (2010) showed that during large collisions the impactor metal cores would not have completely emulsified but would have sunk down into Earth's core as unequilibrated blobs (Fig. 4d). In this case, some memory of the accretion and differentiation history of Earth's precursor planetary bodies is retained in the isotopic and chemical composition of Earth's mantle.

It has been thought likely that the distribution of siderophile elements in Earth's mantle reflects equilibrium partitioning between metal and silicate in a deep terrestrial magma ocean. This view was based on the observation that the depletions of siderophile elements in Earth's mantle provide such an excellent match to those expected from partitioning experiments (Fig. 4b). However, disequilibrium core formation models provide an equally good match to the observed siderophile element abundances in Earth's mantle (Fig. 4e) (Rudge et al. 2010). Thus, the distribution of siderophile elements in Earth's mantle provides no additional information on the extent of core–mantle equilibrium.

Tying Models of the Earth's Core to an Accretion Timescale

The discussion up to this point illustrates that there are many different ways to interpret the isotopic record of Earth's accretion and core formation. In general, it is not possible to unequivocally choose a particular accretion model and timescale as the most probable scenario (Kleine et al. 2009; Rudge et al. 2010). Nevertheless, the isotopic data allow some constraints to be placed on the fraction of Earth's mass that must have accreted by a certain time.

The Hf–W system mainly constrains Earth's early accretion, while the U–Pb system is mainly sensitive to the late stages of Earth's accretion (Figs. 4c and 4f). For instance, in equilibrium core formation models, the Hf–W data require that at least 85% of the Earth was accreted by ~35 My after CAI formation but provide no information on when the last 14% was added. In contrast, the U–Pb isotopic data contain little information regarding Earth's early accretion but show that addition of the final 10% of Earth's mass must have begun by 120 My after CAI formation.

These bounds on Earth's accretion timescale change depending on the degree of disequilibrium during core formation. Figure 4f shows an example of a disequilibrium model with the degree of equilibrium being close to the minimum value constrained by the Hf–W data. In this example, the Hf–W data require that ~50% of Earth's mass had accreted by ~60 My after CAI formation, while the other half could have been added any time later. However, combined with the U–Pb constraints, the last 10% of Earth's mass must have been added by ~140 My after CAI formation.

Dating the Moon and Earth's Terminal Accretion

The Moon is thought to have formed at the very end of Earth's accretion when a Mars-sized body collided with the proto-Earth. Determining the age of the Moon thus provides an independent constraint on how quickly the Earth formed. Dating the giant Moon-forming impact directly is difficult, however. The earliest possible time of Moon formation is given by a Hf–W two-stage model age of ~36 My after Solar System formation (Touboul et al. 2007). The latest possible time of Moon formation is given by the age of the ferroan anorthosites; they are the oldest

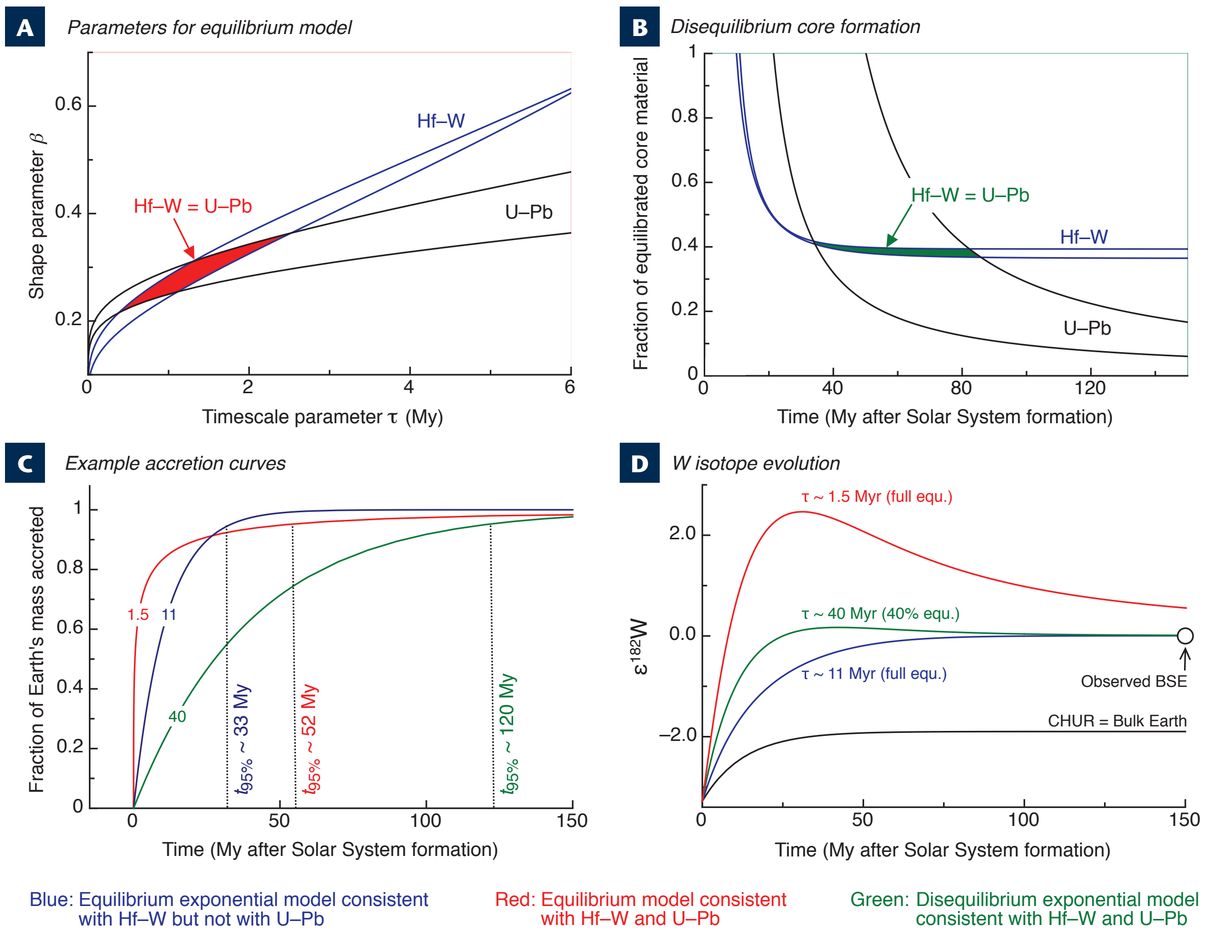

Figure 3 Two ways to match the Hf–W and U–Pb timescales for Earth's accretion. (**A**) Parameters τ and β for the Weibull accretion model of the form $M(t) = 1 - \exp\{-(t/\tau)^{\beta}\}$, where M(t) is the fraction of Earth's mass accreted at time t, τ is the mean-life of accretion, and β is the shape parameter of the Weibull function. For β = 1 the exponential model is recovered. There is a region of overlap (red) where Hf–W and U–Pb provide consistent timescales. (**B**) Effect of disequilibrium during core formation on the calculated accretion timescales. The region of overlap (green) is for ~40% equilibration, where Hf–W and U–Pb give identical results in an exponential model. (**C**) Examples of accretion curves for the Earth. The numbers on the curves refer to values for τ. The end of accretion, as given by the time of 95% accretion, is different in these three models. (**D**) W isotope evolution of Earth's mantle for the three accretion models shown in (C). BSE = Bulk Silicate Earth, CHUR = Chondritic Uniform Reservoir, full equ. = full equilibration, 40% equ. = 40% equilibration. $\varepsilon^{182}W$ is the deviation from the W isotope composition in parts per 10,000; the present-day $\varepsilon^{182}W$ of the BSE is 0 by definition.

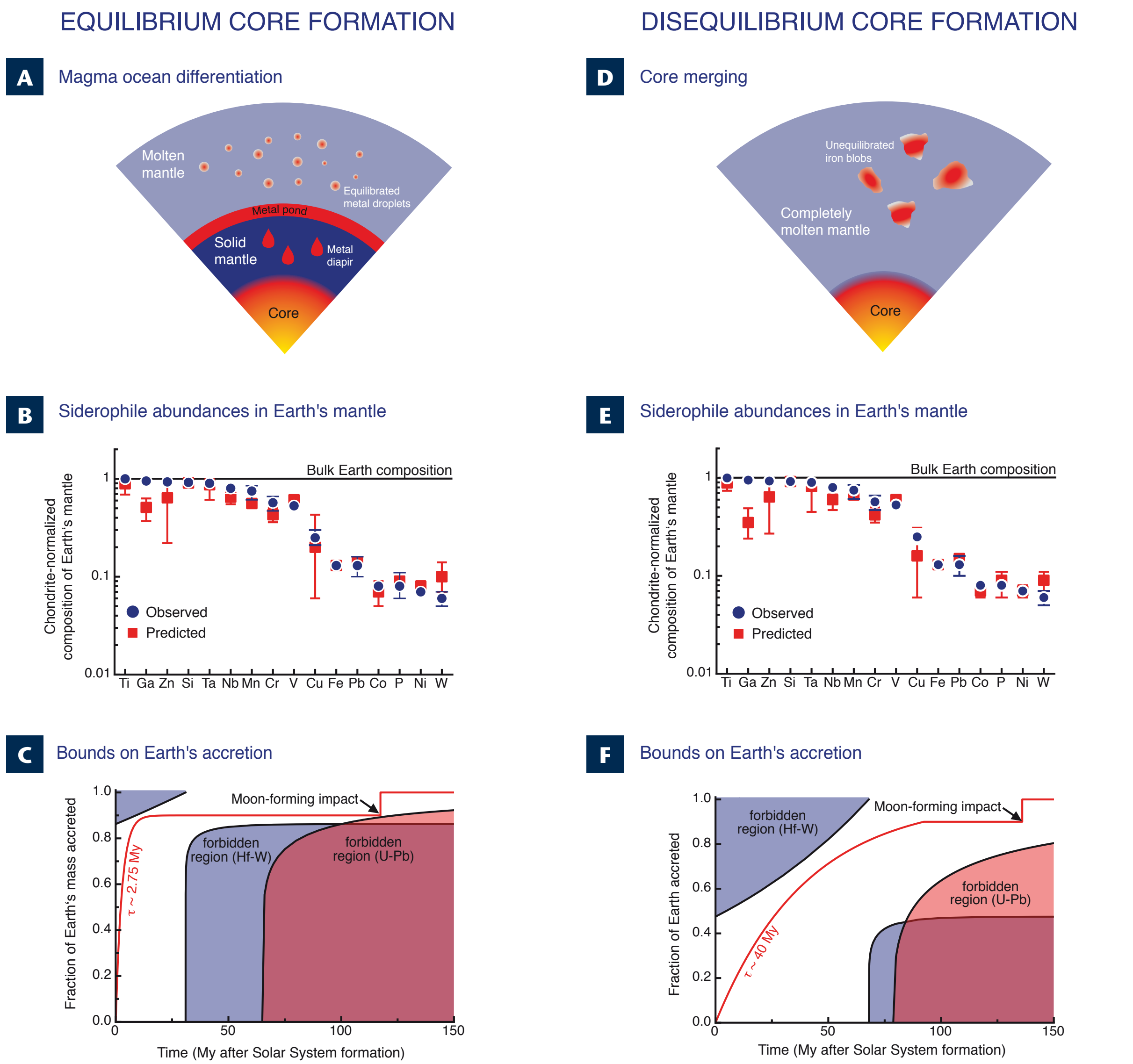

FIGURE 4 (**A**) Equilibrium core formation: sinking metal droplets aggregate to a metal pond at the base of a magma ocean; from there, the metal descends as diapirs through a mostly solid lower mantle. (**B**) Present-day depletions of some siderophile elements in Earth's mantle compared to those predicted based on metal–silicate equilibration in a deep magma ocean. (**C**) Bounds on Earth's accretion in equilibrium core-formation models. Shaded areas are forbidden. One possible accretion curve is shown. (**D**) Disequilibrium during core formation: some fraction of the metal core does not emulsify as small metal droplets but descends through the mantle as large iron blobs, leaving little opportunity to equilibrate with the magma ocean. (**E**) Same as (B) but assuming that only 40% of the newly accreted metal core equilibrated with Earth's mantle. (**F**) Same as (C) but for only 40% equilibration.

lunar rocks and formed ~4.46 billion years ago (Norman et al. 2003).

Another way to deduce the earliest possible time of Moon formation is to use the indistinguishable W isotope compositions of the silicate Earth and Moon, which suggest that the Moon formed more than ~50 My after Solar System formation (Touboul et al. 2007). This age is consistent with most Hf–W model ages for the Earth, as well as with the U–Pb age of the termination of Earth's core formation (FIG. 1).

A Non-Chondritic Earth?

One of the basic assumptions when using isotopic data to model Earth's differentiation is that refractory elements occur in precisely chondritic relative abundances (see above). However, Earth's ^{142}Nd – ^{182}W record is more easily understood if Earth has non-chondritic relative abundances of refractory elements. The small ^{142}Nd excess of Earth's mantle relative to chondrites suggests differentiation of the silicate Earth well before 30 My (Boyet and Carlson 2005), that is, before Earth's core had fully segregated and before formation of the Moon. It seems implausible, however, that a signature of an early differentiation within Earth's mantle remained isolated during ongoing core formation and also survived the giant Moon-forming impact. This enigma is easily resolved if Earth is non-chondritic, because then the ^{142}Nd excess of Earth's mantle does not require mantle differentiation prior to formation of the Moon (Kleine et al. 2009). The Moon and Mars may also have non-chondritic compositions for refractory elements (Caro et al. 2008).

O'Neill and Palme (2008) proposed that Earth may have acquired a non-chondritic composition by collisional erosion of early-formed crust on precursor planetary bodies. Since crust is enriched in incompatible elements and has a lower-than-chondritic Sm/Nd ratio, preferential removal of crustal material leads to bulk planetary bodies with higher-than-chondritic Sm/Nd. In this case the ^{142}Nd excess of Earth's mantle relative to chondrites reflects the Nd isotope composition of the bulk Earth and does not require an early differentiation of its mantle. A non-chondritic composition of the Earth affects the Hf–W chronology of core formation and results in more protracted timescales compared to those calculated for a chondritic Earth. For instance, the equilibrium two-stage model age is ~40 My instead of ~30 My after CAI formation (Kleine et al. 2009).

SUMMARY AND OUTLOOK

A lot happened in the first few million years of the Solar System. About 4567 My ago, CAIs formed close to the young Sun, defining the start of the Solar System. It took less than ~1 My to accrete and differentiate the first protoplanets, evidence of which are the iron meteorites. These may have been derived from bodies that were disrupted by hit-and-run collisions early in Solar System history. Chondrite parent bodies had a less violent history, perhaps because they formed a few million years later and further out in the Solar System. Although the chondrite parent bodies assembled late and in an orbit beyond Mars, some chondritic meteorites contain CAIs that formed close to the active young Sun. This leaves little doubt that there has been substantial outward transport, over several AU, of material. While we understand how CAIs were incorporated into small bodies in the outer asteroid belt and beyond, the processes by which chondrules formed and then accreted to chondrite parent bodies are still unclear. Further progress in this area will require reconciling dynamical models of the solar nebula with results from meteorite chronometry.

After the rapid formation of planetary embryos, accretion slowed down, and the Earth and Moon were fully formed no earlier than ~50 My after Solar System formation. Collisions during terrestrial planet formation may have led to preferential erosion of early-formed crust from the embryos, causing non-chondritic compositions, even for elements that long were thought to occur in constant relative proportions throughout the Solar System. Collisions during Earth's accretion also led to melting and core formation, but exactly how the metal segregated towards Earth's core remains unclear. It has been thought likely that metal–silicate equilibration took place between metal droplets and the surrounding molten silicate in a magma ocean, but there is now evidence that during giant impacts a fraction of the impactor's metal core directly merged with Earth's core, leaving no opportunity to equilibrate within the terrestrial magma ocean. A memory of the differentiation of Earth's precursor planetary bodies is thus retained in the chemical and isotopic composition of Earth's mantle.

Our understanding of how the Earth formed is thus changing from the relatively simple picture of a chondritic Earth that internally differentiated by a one-stage metal–silicate equilibration process to a more complex picture of a dynamically evolving planet involving collisional erosion and metal–silicate disequilibrium. The implications for compositional models of the Earth are profound and open up new avenues of research that will likely lead to a new understanding of how the Earth formed and first evolved.

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