INVITED REVIEW

Not so rare Earth? New developments in understanding the origin of the Earth and Moon.

Kevin Righter *

Mailcode KT, NASA Johnson Space Center, 2101 NASA Parkway, Houston, TX 77058 USA

* corresponding author: tel.: +1 281 483 5125 ; fax: +1 281 483 5347.
  email address: kevin.righter-1@nasa.gov (Kevin Righter)

Abstract

A widely accepted model for the origin of the Earth and Moon has been a somewhat specific giant impact scenario involving an impactor to proto-Earth mass ratio of 3:7, occurring 50-60 Ma after T_0, when the Earth was only half accreted, with the majority of Earth's water then accreted after the main stage of growth, perhaps from comets. There have been many changes to this specific scenario, due to advances in isotopic and trace element geochemistry, more detailed, improved, and realistic giant impact and terrestrial planet accretion modeling, and consideration of terrestrial water sources other than high D/H comets. The current scenario is that the Earth accreted faster and differentiated quickly, the Moon-forming impact could have been mid to late in the accretion process, and water may have been present during accretion. These new developments have broadened the range of conditions required to make an Earth-Moon system, and suggests there may be many new fruitful avenues of research. There are also some classic and unresolved problems such as the significance of the identical O isotopic composition of the Earth and Moon, the depletion of volatiles on the lunar mantle relative to Earth's, the relative contribution of the impactor and proto-Earth to the Moon's mass, and the timing of Earth's possible atmospheric loss relative to the giant impact.

1. Introduction

The origin of the Earth and the Moon has been linked since the giant impact hypothesis took hold in the mid 1980s (Hartmann and Davis, 1975; Cameron and Ward, 1976). The previously popular ideas of capture, fission, and co-accretion were all pushed aside in favor of the idea that a giant impact between the Earth and a proto-planet produced a debris ring that later accreted to form the Moon (e.g., Wood, 1986). The maturation of that idea in the ensuing decades led to the concept of a late giant impact between the Earth and a Mars-sized impactor (Benz et al., 1986; Cameron, 1997), which could explain the high angular momentum of the Earth-Moon system. These models solved many long standing problems such as the total angular momentum of the Earth-Moon system, the low Fe content of the Moon, and the older age of the Moon relative to the Earth. At the time of the conference on the Origin of the Earth and Moon (December 1998), a paradigm was advanced that the impact involved a 3:7 impactor : proto-Earth mass ratio, the impact occurred when the Earth was half accreted, with this time determined to be 50-60 Ma after T_0, and the Earth so accreted was dry, with a later addition of water (Origin of the Earth and Moon, R.M. Canup and K. Righter, eds, 2000).

Although the giant impact model successfully explains many aspects of the Earth-Moon system, there have been several new developments that have challenged the paradigm. First, the idea that Earth accreted wet has been explored in greater detail. Second, the idea of magma oceans (as hot as 4000 K and as deep as 50 GPa for Earth) on many differentiated bodies has taken hold and has implications for the thermal state of proto-planets. Third, technical advances, additional measurements, and modelling of W isotopes have led to major revision of the timing of accretion and core formation. Fourth,

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physical models for the dynamics of accretion have better revealed mixing between 0.5 and 4 AU in the inner solar system (Chambers, 2001). And finally, additional models of impact processes have explored "grazing" impacts that lead to bodies that survive impacts and continue on paths within the inner solar system (Asphaug et al., 2005; Bottke et al., 2006). Also, the original late "giant" impact model, with an approximately 1:10 impactor to protoEarth ratio, returned into favor as well (Canup and Asphaug, 2001; Canup, 2004a,b), making plausible several different dynamic situations, rather than one unique one.

2. Hot molten early Earth – other planets too?

By 1999, the idea that the early Earth had experienced extensive melting and a magma ocean had just taken hold. Hard evidence for this idea, although proposed by some based on thermal models for accreting planets and the energetics of large impacts (e.g., Hostetler and Drake, 1978; Kaula, 1979; Stevenson, 1988; Tonks and Melosh, 1990), was limited to work on siderophile elements (e.g., Walter et al., 2000). And although there were suggestions that early high temperatures might explain aspects of mantle geochemistry (Murthy, 1991), and theory also predicted that core formation might depend upon the presence of silicate melt to facilitate mobility (G.J. Taylor, 1992), it took some time for experimentalists to explore this concept. Specifically, the concentrations of Ni, Co, Mo, W, P, and other siderophile (iron-loving) elements in the Earth's mantle could be explained by equilibration between metallic liquid and silicate melt at mid-mantle depths (e.g., Li and Agee, 1996; Ohtani et al., 1997; Righter et al., 1997).

Since then new evidence for magma oceans has been considered from a number of perspectives. Greenwood et al. (2005), based on high resolution oxygen isotopic measurements of angrites, eucrites, lunar and martian meteorites and terrestrial samples, argued that the oxygen isotopic homogeneity within various individual parent bodies (**Fig. 1**) was most likely influenced by the presence of magma oceans early in the history of the solar system. Boyet and Carlson (2005) measured ^{142}Nd (derived from the alpha decay of ^{146}Sm with a half life of 103 Ma) in chondrites, eucrites and terrestrial samples and noted that the anomalies in terrestrial samples relative to chondrites could only have been derived if the Earth differentiated within 30 Ma of the start of the solar system (T_0). Similarly rapid differentiation timescales have been proposed for Mars based on Nd and Hf isotopes (Borg et al., 2003; Foley et al., 2005).

Physical modelling of the metal-silicate segregation process has also suggested hot and rapid timeframes for differentiation. Rubie et al. (2003), pursuing the likelihood of two different physical mechanisms in establishing metal-silicate equilibrium in planetary interiors, concluded that one mechanism will dominate due to its speed and efficiency. An inefficient mechanism is that of a convecting molten mantle equilibrating with a ponded metal layer at its base. This scenario takes too long and would require > 100 Ma to equilibrate a terrestrial mantle like Earth's, much longer than a lifetime of a magma ocean. On the other hand, a very efficient mechanism is one where droplets of metallic liquid fall through a magma ocean equilibrating along the path. Droplets break apart above a certain threshold size (due to shear forces acting on the two liquids), and this helps to promote equilibration since the scale of equilibration is always lower than this threshold value. A molten mantle can equilibrate on the time frame of tens of

thousands of years, very similar to the lifetime of a terrestrial magma ocean (**Fig. 2**; Rubie et al., 2003). Independent modelling of Hoink et al. (2006) has resulted in slightly different timescales for metal-silicate equilibrium (perhaps due to slightly different modelling parameters), but still very rapid (tens to thousands of years). Crystallization times of magma oceans are also short – as short as 1000 years (Rubie et al., 2003, and references within). The rapid timeframe for metal-silicate equilibrium and magma ocean crystallization is consistent with Hf-W Moon formation dates around 4540 Ma, but ages of lunar anorthosites are still 80 Ma younger, at 4460 Ma (Norman et al., 2003). Perhaps additional samples from the Moon (e.g., meteorites; Korotev, 2005) will yield new chronologic constraints that can help to resolve this discrepancy.

Noble gas models for the origin of Earth's atmosphere also accommodate the concept of an early magma ocean, as the ocean provides a way of storing noble gases away in the interior before an event that drives off the primordial atmosphere (e.g., Pepin and Porcelli, 2002; Porcelli and Pepin, 2000). In this way, the secondary atmosphere may reflect an early (partly) solar contribution that is later released upon loss of the primary atmosphere. Finally, recent modelling of the evolution of the martian mantle has included an early magma ocean stage (Elkins-Tanton et al., 2003; Borg and Draper, 2003; Reese and Solomotov, 2006). Although there is not a siderophile element signature in the martian mantle that is suggestive of an extensive (hot and deep) magma ocean on Mars like there is for Earth (compare Righter and Drake, 1999 for Earth to Righter et al., 1998 for Mars), it still indicates a transformation of thought about the thermal states of early terrestrial planets. Siderophile element modelling for all four of the individual bodies for which we have samples - Earth, Moon, Mars and Vesta – leads to the conclusion that

they experienced magma oceans of some depth (e.g., Righter, 2002; Moon; Righter et al., 1998; Mars; Righter and Drake, 1997; Vesta; Righter, 2003; Earth). Our understanding of thermal conditions on planets and proto-planets is not complete, but will ultimately be dependent upon the sum of energies derived from impacts, radioactive heating, core formation, and surface heating from atmospheres (e.g., see Rubie et al., 2007); there are ways of melting a large size range of objects with these mechanisms.

A hot early inner solar system would have other effects as well. For example, the redox state of every planet is different (as are the FeO mantle content and core sizes; **Figs. 3A,B**) and could have been set by thermal gradients (e.g., **Fig. 3C**) or gradients in volatiles such as H_2O or CO_2 (or both). For the Earth, Wanke (1981) argued that it accreted dry and reduced first, and then the accreting material changed to more oxidized, bringing in water and volatiles. Recent variations of this kind of model have been proposed (Galimov, 2005; Wade and Wood, 2005) where the inner solar system started out reduced, and was then oxidized by water. And Gaillard et al. (2004) argue that the inner solar system started oxidized and was reduced sequentially inwards by H_2. Finally, Palme (2000) considers compositional gradients in the inner solar system and notes that while there may be zoning of some elements and isotopes such as the alkalis and Cr, it is not clear if they formed directly from the nebula, or if these trends were established by movement of materials after accretion.

3. New constraints on the timing of core formation and differentiation

Early solar system chronology has undergone revision due in part to the development of new mass spectrometers that allow much higher precision measurements

than before. As a result of this high precision, detailed isotopic studies for specific elements in bulk chondrite samples, have identified interferences from pre-solar grains, and/or heterogeneities in the distribution of p- or r-process isotopes in the solar nebula. Because small grains can contain isotopic anomalies that pre-date our solar system, and the new high resolution measurements can detect this contribution, it has been necessary to reconsider assumptions used for the initial isotopic compositions of our solar system. For example, Ranen and Jacobsen (2006) have identified excesses of ^{137}Ba and ^{138}Ba in carbonaceous chondrites. Andreasen and Sharma (2006) have found a 100 ppm deficit in ^{144}Sm in the Allende and Murchison carbonaceous chondrites. Carlson and Boyet (2006) have proposed that because carbonaceous chondrites contain a significant deficit of ^{142}Nd, our initial ^{142}Nd/^{144}Nd ratios should be based on ordinary or enstatite chondrite which are not affected. And this issue may also affect the long standing problem with the decay constant of ^{176}Lu (Albarede et al., 2006). These are a few examples of the potential impact this issue has on early solar system chronology. Because of substantial developments in this field in the last few years, several systems will be the focus of detailed discussions below (also see Table 1).

3.1 Hf-W chronometry

The initial measurements of ^{182}W in terrestrial, lunar, and meteoritic samples resulted in identical values for terrestrial and chondritic samples, and very radiogenic values for lunar samples (Lee and Halliday, 1996, 1998; Lee et al., 1997), leading to the idea that the Moon formed early and the Earth completed its formation later. Subsequent work by new research groups as well as the original groups, led to revision of both of

these fundamental observations (e.g., Jacobsen, 2005), and caused re-assessment of the timing of formation of the Earth and Moon.

Early measurements were done using isotope dilution and TIMS on mass spectrometers such as the P54 (e.g., Halliday et al., 1998; Lee and Halliday, 1996; Harper et al., 1995). This technique was difficult to develop, and only a few early measurements were reported by Harper et al. (1995) and Harper and Jacobsen (1996), followed by a very extensive data set derived by Lee and Halliday (1996, 1998) and Lee et al. (1997). Development of multi-collector ICP-MS allowed several additional research groups to measure ^{182}W in meteoritic and planetary materials (Yin et al., 2002; Kleine et al., 2002; Ireland et al., 2003; Schoenberg et al., 2002; Quitte and Birck, 2004). These groups all obtained very different values for the initial ^{182}Hf/^{180}Hf for the solar system, affecting ultimately the interpretation of W isotope data bearing on the age of the Moon. Because early measurements had shown that terrestrial samples had the same ^{182}W values as chondritic samples, terrestrial samples became the common standard for W isotopic work. However, the new measurements of chondrites by additional research groups resulted in chondritic values that were significantly different from terrestrial (e.g., Kleine et al., 2004; **Fig. 4**). Details of the analytical conditions leading to the differences of values have not been discussed in the literature extensively, but were the topic of a spirited exchange at the Davos Goldschmidt meeting in 2002 (e.g., Halliday et al., 2002; and discussion in Halliday, 2004a). As a result, terrestrial values are now more radiogenic than chondritic, creating the possibility that the Earth and Moon both record isotopic values indicating accretion within 60 Ma of T_0.

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The other development requiring revision to interpretation of W isotopic work was the realization of the importance of production of ^{182}W by cosmogenic (i.e., by cosmic rays) derived from ^{182}Ta at the lunar surface (Leya et al., 2000). Relatively Ta-rich samples could become enriched in ^{182}W, and this source could contribute a significant percent of the radiogenic ^{182}W measured in lunar samples. After the correction for ^{182}Ta, many lunar samples exhibited only minor ^{182}W anomalies (**Fig. 5**). Thus the revised situation for the Moon was the opposite to that of the Earth, and smaller anomalies removed requirements of a Moon derived while the Earth was half-accreted (e.g., Kleine et al., 2004).

Now that various research groups have apparently come to agreement on initial W isotopic ratios, the importance of cosmogenic derived ^{182}W, and the cause of the analytical discrepancies, more extensive modelling has been attempted. Halliday (2004b) shows that a giant impact (impactor:target mass = 1:9) at 55 Ma can explain the W isotopic values if the impactor and target have Hf/W = 15 and 26% of the Theia's core equilibrated with the BSE. They also showed that the data could also be explained by the same impact, but with Hf/W=5, and only 4% of Theia's core equilibrated with the BSE (**Fig. 6**). The disequilibrium between Theia's core and the BSE could result from merging of the cores of the impactor and Theia, without total re-equilibration. However this may not be realistic for two reasons. First, most recent simulations show that the metal and silicate are mixed together extensively after the impact (Canup, 2004b). And second, this scenario would lead to widespread re-equilibration as metal rained through molten silicate (Rubie et al., 2003; Hoink et al., 2006). These two constraints have led Jacobsen (2005) to propose that the giant Moon-forming impact occurred at 32 Ma, the

tungsten isotopic values were reset by re-equilbration, and the Earth and Moon evolved to their present day εW values of +1.9 and +3.2, respectively (Jacobsen, 2005).

So, a range of conditions can explain the W isotopic values and there exists no unique model, given current uncertainties in some of the variables (see below). In fact some models have even combined physical dynamic models for planet growth with W isotopic values. For example, Nimmo and Agnor (2006) couple the late accretion modelling results of Canup and Agnor (2000) with W isotopic measurements and show that the nature of the impactor and proto-Earth are also important to the successful outcome. Similarly, Yuki and Abe (2005) show from coupled modelling of W isotopes, Ni partitioning and planetary dynamics, that either the Earth underwent equilibration in a shallow magma ocean scenario (~50 GPa), or experienced disequilibrium in a deep magma ocean scenario (as deep as the core-mantle boundary). It is clear that coupling of dynamic models and isotopic data holds great potential for understanding the origin of the Earth and Moon. Much additional work remains along these lines of investigation, as the dynamic models become more and more realistic (see below).

Several aspects of the modelling should be highlighted to demonstrate shortcomings or areas where improvement may be made in future efforts. First, many models have assumed that the partition coefficient for W between metallic liquid and silicate liquid (D(W) met/sil), is constant. Although this makes modelling easier, it is not realistic. In fact, although our understanding of D(W) met/sil is likely incomplete (i.e., effects of pressure, temperature and fO_2 are currently being investigated; Cottrell and Walker, 2004; Danielson et al., 2006), the current models predict that as a planet like the Earth grows through the accretion process, D(W) met/sil can vary from values as low as 5

at as high as 100 (**Fig. 7**). Second, various models assume an initial Hf/W that does not change. Not only will this change with changing D(W) met/sil, and thus during accretion, but Hf/W will also change due to silicate fractionation. Righter and Shearer (2003) show that silicate phases – clinopyroxene and garnet – can fractionate Hf and W such that silicate differentiation that occurs after core formation can also contribute to the production of mantle reservoirs with elevated [182]W (**Fig. 8**). No models have yet incorporated this effect. And finally, the issue of extent of post impact core equilibration remains controversial (e.g., Kleine et al., 2004; Jacobsen, 2005), yet has a substantial control on the final outcomes of models.

3.2 Sm-Nd chronometry

As mentioned for the Hf-W system, new mass spectrometers have provided greater resolution and multi-collector capability. This has improved the ability to detect isotopic anomalies in the [146]Sm – [142]Nd system. Measurements by Nyquist et al. (1995) indicated the possibility of small anomalies of [142]Nd that could be explained by the presence of live [146]Sm in the Moon. Similarly, several groups in the past decade have debated the presence of [142]Nd in Archean terrestrial rocks (Caro et al., 2003; Boyet et al., 2003). New measurements (MC-ICP-MS) by Boyet and Carlson (2005) on chondrites, eucrites and terrestrial samples have identified distinct differences between chondrites and Earth, indicating that the silicate Earth may have differentiated as early as 30 Ma after T_0 (**Fig. 9**). Furthermore, they proposed that there is a hidden enriched reservoir in the deep mantle that also formed early, and is the complement to the Earth's depleted mantle. Using similar improved techniques, Rankenburg et al (2006) have measured [142]Nd in a suite of lunar samples (Apollo and lunar meteorites) demonstrating that the

Moon is chondritic, in contradistinction to the BSE which is depleted (**Fig. 10**). They

also place limits on the amount of the Moon which could have come from the impactor –

up to but not more than 80%. The issue of whether the Moon is made from material from

the impactor or the protoEarth is unresolved (e.g., McFarlane, 1989; Warren, 1992), and

remains an important focus for future work. For example, based on Ta-Nb constraints,

Münker et al. (2003) argue that the Moon could be made of up to ~50% of an impactor,

but no more (**Fig. 11**). This conclusion is at odds with that provided by the Sm-Nd work

(Rankenburg et al., 2006), geophysical modelling and all impact simulations to date (see

later section). This important point will be visited later in this paper as well.

3.3 Other isotopic systems – Nb-Zr, Pd-Ag, Tc-Mo

Ironically, as the initial Hf-W isotopic data came out between 1995 and 1999 and

suggesting relatively late formation dates for the Earth, several groups had measured

other short-lived isotopes in terrestrial and lunar samples and found evidence for an older

Earth-Moon system (Table 1). For example, Hauri et al. (2000) proposed a relatively

rapid core formation, 20 to 30 Ma after T_0, based on measurements of ^{107}Ag derived from

the decay of ^{107}Pd (half-life of 6.5 Ma). More recent measurements with higher precision

led Schönbächler et al. (2006) to propose a similar time interval for core formation, from

10 to 29 Ma. Another short-lived chronometer with potential application to core

formation and differentiation in the Earth is ^{97}Tc-^{97}Mo, with a half-life of 2.6 Ma. Yin

and Jacobsen (1998) measured slightly different ^{97}Tc-^{92}Mo ratios in iron meteorites and

the bulk silicate Earth, arguing that terrestrial accretion must have lasted to between 19

and 24 Ma after T_0. A short lived chronometer ^{92}Nb-^{92}Zr is comprised of two lithophile

elements, has a half life of 36 Ma, and the potential to date silicate differentiation in

rocky planets or planetesimals. Münker et al. (2000) found that lunar, terrestrial, and chondrite samples have identical relative abundances of ^{92}Zr. Furthermore, the absence of ^{92}Zr anomalies in the Earth and Moon may indicate their silicate reservoirs (mantles) formed more than 50 Ma after T_0, but may also be related to the very small fractionation of Nb from Zr (Schönbächler et al., 2005). Clearly each isotopic system yields slightly different constraints and ages. Additional measurements and modelling will hopefully lead to a more consistent picture of the timing of differentiation events in the early solar system and Earth-Moon system.

4. New constraints from impact modelling and planetary dynamics

The very specific models of Cameron (1997, 2000, 2001), requiring an impact occurring when the Earth was half formed and with a 3: 7 impactor to proto-Earth ratio, have given way to a class of models with a range of parameters resulting in a Moon-forming impact. Several new studies with exciting implications for this field will be discussed here. Advances are being made due to increased resolution, improved equations of state, and overall faster computing options allowing testing of more parameters in a shorter period of time.

4.1 Advances in the dynamics of accretion

Dynamic models for planetary accretion involve three main stages for growth once dust aggregates and settles in the protoplanetary disk– a dust to planetesimals stage that takes about 10^5 years, b) a planetesimals to protoplanets (or "oligarchs") stage that takes about 10^6 years, and c) a chaotic growth (large impacts) stage that takes about 10^7 years (Weidenschilling et al., 2000; Kokubo et al., 2000). There has been significant

progress in the last 6 years in understanding this entire process, and the subtleties that can lead to different outcomes of planetary systems.

Tracking the origin of materials through each stage of this process has been the focus of studies by Chambers (2001, 2004, 2005). His interesting findings are that a substantial amount of material (up to 5%) that starts in the innermost inner solar system can become incorporated into a planet that forms in the outermost inner solar system (**Fig. 12**). The opposite is true as well, with similar amounts of material from the outermost part of the inner solar system ending up in the innermost planet. These results are of great interest to those who have proposed that because the Earth and Moon have identical O isotopic values they must have formed in the same part of the nebula (e.g., Drake, 2000). These results would seem to strengthen that argument. However, as is discussed later we really have no idea of what the O isotopic composition of the inner solar system is – if all materials from the Earth inwards have the same O composition, then this argument is irrelevant (e.g., Wasson, 1988). And the dynamic modelling results show that even a relatively small body – 5% of the total mass of a planet – could come from a relatively far distance within the inner solar system and bring with it a significant amount of a volatile species such as water (e.g., Morbidelli et al., 2000). This latter implication will be discussed in the next section on the origin of Earth's water.

Second, although N-body simulations such as those of Chambers (2001), Agnor et al. (1999) and Raymond et al. (2004) can produce systems of terrestrial planets, several factors have been recognized to exert a strong control on the outcomes – gas giant orbits and damping due to dust and gas. When gas giants are in circular orbits, the radial extent of terrestrial planet growth is controlled by the location of Jupiter, since it can clear out

material from as far in as the asteroid belt. If Jupiter is far enough out, it can leave material in the vicinity of the asteroid belt that can later be available in the last stages of planet growth. Elliptical orbits for Jupiter and Saturn result in much larger volumes of accretion and thus more effectively clear out the asteroid belt. This has implications for "wet" planetesimals in particular, since many will form in the asteroid belt region. Damping mechanisms can help to reduce the high eccentricities (e) and inclinations (i) produced in standard N-body simulations. Levison et al. (2005) showed that the dynamical friction introduced by having a population of remnant planetesimals (leftover from impact processes), can lead to planetary systems with e and i more similar to our own. Also, Goldreich et al. (2004) showed that small dust grains can also dampen eccentricities. Incorporating the effects of dynamical friction on the accretion process and exploring differences between circular or eccentric orbits for Jupiter and Saturn, O'Brien et al. (2006) found some fundamental differences in accretion simulation outcomes. For example, dynamical friction lowers the timescale for accretion and also results in terrestrial planet systems that are less dynamically excited compared to previous modeling. Also, the simulations using eccentric Jupiter and Saturn orbits, result in a dearth of water-bearing embryos in the Earth's region, and a late veneer (defined by O'Brien et al. as the fraction of mass delivered after the last large impact) that is much larger than suggested by geochemical evidence. Simulations using circularized orbits for Jupiter and Saturn result in many water-bearing embryos in the Earth's region and a late veneer that is more similar in mass to that suggested in geochemical modeling.

Third, a hybrid class of calculations has been initiated that combines coagulation and N-body code in order to follow joint evolution of planetesimals and oligarchs

(Kenyon and Bromley, 2006; Bromley and Kenyon, 2006). Because leftover planetesimals interact with the largest oligarchs, dynamical friction is high, and the orbits tend to circularize with low e. Also, the lower mass planets end up with more eccentric orbits than the most massive planets, which is similar to our Solar System. This approach will undoubtedly lead to additional insights into the planet formation process (Nagasawa et al., 2007).

Fourth, sweeping resonances can exert a strong control on terrestrial planet formation. In particular, the v5 resonance (when the precession of the line of apsides of an orbit is in phase with the precession of Jupiter's orbit) can play an important role in the transition from oligarchic growth to chaotic growth (Ward et al., 1976; Nagasawa et al., 2000, 2005; Lin et al., 2006). As gas in the disk dissipates, the v5 resonance sweeps inward from Jupiter's orbit, potentially moving material into the terrestrial planet zone. Simulations show that the resonance can shake up orbits of protoplanets, and depending upon the timing of gas dissipation, can lead to orbit crossings and mergers of protoplanets. Once the gas is totally dissipated, the last remnants of the disk can circularize the orbits of the remaining planets (Nagasawa et al., 2007). These calculations can lead to planetary systems with mass distributions, eccentricities and inclinations very similar to our own inner solar System. However, the amount and timing of the damping (gas drag) is critical to the outcomes and remains an active area of research.

And finally, additional constraints on the state of the early Moon come from planetary dynamics. Calculations by Garrick-Bethell et al. (2006) have demonstrated the possibility that the Moon's three principal moment's of inertia could have been caused by

a past high-eccentricity lunar orbit. Specifically, the Moon may have once been in a 3:2 resonance of orbit period to spin period, similar to Mercury's present state. Recognition of past high-eccentricity orbits for the Moon are fundamental to a full understanding of its dynamical history, and would be important to integrate with the results of giant impact modelling, as well as models for the early thermal and tidal evolution of the Moon.

4.2 Advances in impact modelling

First, Canup and Asphaug (2001) used higher resolution SPH (smoothed particle hydrodynamics) calculations (i.e., more particles than previous models – 100,000 vs. 3,000 of previous work) with the Tillotson equation of state for the impacting materials, to show that a Moon-forming impact was possible for the case of a 1:9 impactor to protoEarth mass ratio. The resulting proto-lunar debris disk was 10-20% by mass from the target (protoEarth), hot (2000-3000 K), Fe-depleted (2 to 4% metal), and occurred late in the accretion process. This work loosened the restrictions imposed by the earlier work on the timing and relative mass proportions of the impactor and protoEarth. More recent work by Canup (2004b) utilizes similarly high resolution, but substitutes a revised ANEOS equation of state (M-ANEOS; Melosh, 2000, 2006) that has a more complete treatment of silicate vapor production and molecular vapor species, compared to previous models. These new developments, show that there exists a continuum of impactor: total Earth masses, from early to late in the accretion process, which can satisfy the conditions required to form a terrestrial Moon (Canup, 2004a, b; **Fig. 13**). This conclusion is likely to be revised even more in future studies, but it shows clearly that there are no "special conditions" required to make a Moon, but rather a range of conditions. It should be noted here that despite the range of conditions, the late impact scenarios seem the least

restrictive, because earlier impact scenarios have the conceptual problem of accreting the rest of the Earth (including Fe-rich materials) after the Moon forming event.

Second, new modelling involving low-angle impacts (Asphaug et al., 2006), shows that smaller impactors can survive impacts, do not merge with the target, and can produce some bodies of unusual composition. For example a few simulations have resulted in metal-rich or silicate-rich bodies, all of which are deformed, spun-up, and depressurized (Asphaug et al., 2006; Bottke et al., 2006). These smaller scale impacts have implications for explaining some of the unusual bodies in the asteroid belt, but also stimulating ideas regarding parameters that determine the outcomes of giant impacts such as the impact angle.

Third, coupling the results of late accretion models of Canup and Agnor (2000) with tungsten isotope modelling, Nimmo and Agnor (2006) consider three different physical scenarios of the giant impact: primitive differentiation, mantle equilibration, and core merging. In primitive differentiation, an undifferentiated impactor hits a differentiated target, differentiation occurs during impact, and then the mantles and cores merge with each other without further equilibration. In the case of mantle equilibration, both objects are differentiated, the mantles re-equilibrate and the cores merge. And in the core merging scenario, cores and mantles of the two differentiated objects merge without further re-equilibration. All three of these scenarios produce a different result in terms of W isotopic value of the target, with intermediate, low and high values, respectively, for the three models (Nimmo and Agnor, 2006). These models are simplified with respect to D(W) met/sil and Hf/W as mentioned above, but nonetheless illustrate the range of outcomes in different physical scenarios (**Fig. 14**).

5. Wet Earth – dry Moon: New constraints on the origin of water

A traditional view for the origin of water on Earth is that it was delivered late, or after the main accretion process, by comets (e.g., Delsemme, 1997). Although this origin has appeal because it could also bring in the noble gases and highly siderophile elements that are elevated in the terrestrial mantle, it has recently been challenged by several different lines of evidence. First, measurements of D/H ratios in several Oort type comets (Halley, Hale-Bopp, Hyakutake) are much higher than the D/H measured in Earth's oceans (**Fig. 15**). Second, measurements of oxygen isotopes in ancient (4.4 to 4.0 Ga) zircons from Archean rocks (Valley et al., 2005) yield values higher than those expected from dry mantle processes (**Fig. 16**). Because the presence of water can fractionate O isotopes through a variety of processes involving water, Valley et al. (2005) propose that the ancient zircons were derived from an early Earth that had water at its surface. This study, together with differences between cometary and terrestrial D/H, suggests that water may have been present during the accretion process. Although our understanding of comet D/H may be skewed by the specific kinds of comets for which we have knowledge (other comets may yield different values), the idea of wet accretion has prompted a number of studies on different aspects of the problem.

5.1 How does water survive the accretion stages?

If Earth accreted wet, there has to have been a way for water to survive the three accretion stages mentioned earlier: dust to planetesimal to embryo to protoplanet. Stimpfl et al. (2004) recognizing the potential problem of keeping water in the early stages of accretion, showed that the adsorption of water vapor onto dust grains is a very

efficient process and one which could have left plenty of water available in the next and later stages of accretion. Similarly, Ciesla et al. (2003) showed that the inward migration of phyllosilicates formed by the hydration of chondrules during shock waves in icy regions of the solar nebula, could be a mechanism for hydrating the middle part of the inner solar system (e.g., see discussion of Chambers, 2001, above).

In planetesimals, water may have been stored in a number of ways. First, we know from studies of carbonaceous chondrites that phyllosilicates are plentiful in the matrices. For example, there are saponite and serpentinite intergrowths observed in Tagish Lake (**Fig. 17**; Keller and Flynn, 2001) and other carbonaceous chondrites. These minerals are stable at relatively low temperatures. On the other hand, a new type of hydrated metamorphosed chondrite has been recovered from Antarctica: LAP 04840 is a Rumuruti-type chondrite (oxidized and perhaps related to ordinary chondrite groups) that contains calcic hornblende and biotite (**Fig. 17**). Although some of the hornblende OH site may be occupied by O, F and Cl, there could be a significant amount of water (several wt%) stored in these phases (McCanta et al., 2006) in this kind of chondrite. Finally, magmas can store a substantial amount of water at higher pressures. Planetesimals can achieve sizable pressures of up to a few kilobars for a Vesta-sized body. Using a simple water solubility model for terrestrial magmas, it can be estimated that 3 to 4 wt% water could be dissolved in a chondritic magma ocean on a small planetesimal (**Fig. 18**). These two general conditions of hydrous minerals and melts may both be relevant, as the midplane temperatures can be as cool as 1000 K, and as hot as 2000-3000K due to shock heating (e.g., Boss, 1998; Boss and Goswami, 2006). However, a better understanding of nebular temperatures at the Earth's orbital radius out

to the asteroidal zone, would be particularly interesting with respect to the stabilities of hydrous minerals in general.

The accretion models of Morbidelli et al. (2000) show that some planetesimals can survive late and come from the outer part of the inner solar system (i.e., the location of the current outer asteroid belt). Such planetesimals are likely to be water-bearing. As such, they argue that Earth's water could easily have been delivered by one planetesimal near the end of the main accretion process (i.e., not a late veneer, but late accretion). In all of six different simulations, such an embryo from > 2.5 AU accreted > 30 Ma after T_0, and when Earth was $> 70\%$ accreted.   As discussed above, new simulations of O'Brien et al. (2006) also demonstrate that accretion models in which the orbits of Jupiter and Saturn are treated as circularized result in many water-bearing embryos being available in the region of the Earth.

How is water retained during these processes that involve potentially highly energetic impacts? This is an area of research that has seen some work for chondritic materials, and is just starting for the high temperature conditions at which impacts occur. For example, even at high shock pressures of 90 to 100 GPa, chondrites still retain 50% of their water (see summary diagram from Marty and Yokochi, 2006).   Addressing the question of water loss during giant impacts involving ice-bearing impactors and Earth-sized targets, Canup and Pierazzo (2006) showed that substantial water losses ($> 50\%$) result when the impact angle is low (<30 ˚) and impactor velocities ~ 1.5x higher than the mutual escape velocity of the impactor-target pair.   On the other hand, head-on collisional scenarios tend to have greater water retention.

5.2 From where is the water coming?

Having demonstrated that water is likely to survive the accretion process, we can now consider the many available sources of water during that process. There are several additional possibilities to the late cometary origin, not one unique combination, and it seems that several options are possible given our current state of knowledge. One end member model holds that Earth obtained its water from the early solar nebula (Hayashi et al., 1979), and developed to the current higher D/H ratio by Jeans or hydrodynamic escape (e.g., Ikoma and Genda, 2006; Sasaki, 1990). Another possibility is that its water is actually a mixture of solar nebular and later cometary sources (e.g., Abe et al., 2000, and sources therein). A third view is that Earth's water came from the asteroid belt as represented by the meteoritic samples in our collections (e.g., Engrand et al., 1998; Deloule et al., 1998). Of course, a mixture of all three sources is also possible; for a more thorough discussion of these sources and the issues surrounding them, the topic has been reviewed from several perspectives (Abe et al., 2000; Robert, 2004; Drake, 2005; Marty and Yokuchi, 2006).

If water is plentiful on the Earth, and the origin of the Earth and Moon is linked, then why is the Moon dry? Discussions of the Moon's water generally conclude that it is very dry (e.g., Taylor et al., 1995), or that it may have volatiles at the surface acquired by accretion of volatile-bearing materials over 4.5 Ga sequestered into polar cold traps (see Bussey et al., 2003, for a review of this topic). There are several dynamical constraints that are worth mentioning with respect to the Moon's water budget. First, if the Moon accreted from an impact generated disk, this process would have occurred at low pressures where the solubility of water in hot silicate melt is minimal to none. Furthermore, Pritchard and Stevenson (2000) argue that evaporation in such a proto-lunar

disk would result in volatile loss without isotopic fractionation (as measured for K by Humayun and Clayton, 1995). Finally, a late giant impact origin for the Moon would seem more consistent with our dry Moon since an early Moon forming impact would leave a Moon around the partly-formed Earth, vulnerable to receiving hydrated material while the Earth completed its accretion.

The degree of certainty of any model is tempered by the fact that there are potential water sources in the solar system for which we have no data, such as Jupiter family comets and Kuiper Belt Objects. Future contributions to this issue should look for ways to discriminate between the different scenarios outlined above, as well as integrating these other sources into modeling. In addition, there is a need to link D/H ratios to other volatiles such as N, noble gases, O, and C (e.g., Owen and Bar-nun, 2000).

6. Outstanding problems

Certain aspects of the Earth-Moon system remain unexplained by our current models, either due to lack of data or insufficient models (or both). Four major issues that will be discussed here are the source of the material making the Moon, the volatile element depletions in Moon relative to Earth, the necessity of a late chondritic veneer, and the timing of Earth's possible atmospheric loss relative to the giant impact.

6.1 Did the Moon come from the impactor or the Earth ?

A major conundrum encountered in attempts to unravel the origin of the Earth and Moon has been the conflicting conclusions of the cosmochemistry and planetary dynamics communities (e.g., S.R. Taylor, 1991). On the one hand, cosmochemists see many similarities between the Earth and Moon, such as Mn-V-Cr abundances, O isotopes

and Nb/Ta ratios (Dreibus and Wanke, 2002; Wiechert et al., 2001; Münker et al., 2003, respectively). Recent developments in Nd isotope geochemistry have forced consideration of the idea that the Moon formed out of the depleted terrestrial mantle (Carlson and Boyet, 2006). On the other hand, impact modelling results yield a Moon that is derived nearly entirely from the impactor (Cameron, 1997; Canup and Asphaug, 2001). Some have placed great emphasis on the significance of the O isotopic similarity, arguing that the Earth and Moon came from similar parts of the inner solar system (e.g., Drake, 2000), or that these two bodies re-equilibrated during and after the impact (Pahlevan and Stevenson, 2005). And, many geochemists consider mixed contribution (proto-Earth and impactor) models (e.g., McFarlane, 1989; Warren, 1992). All of these models are subject to our limited understanding of chemical zoning in the inner solar system (e.g., Palme, 2000). For example, we have samples only from the Earth outwards and not a good understanding of even the oxygen isotopic composition of Venus or Mercury (e.g., Wasson, 1988). Thus the significance of the identical oxygen isotopic values for the Earth and Moon is not clear. If the inner solar system is identical, then O isotopes are not a fingerprint of provenance. It is hoped that the Genesis and Stardust missions will help better define the ranges of O (and other) isotopic values of solar system materials (e.g., Wiens et al., 1999; **Fig. 19**).

Furthermore, the current understanding of the existence and temporal stability of narrow vs. wide mixing and accretion zones in the inner solar system is in its infancy. A more sound understanding of this transition will improve origin models. And finally, the fact that dynamic modelling is changing so rapidly makes one wonder if there is a physically plausible scenario whereby the Moon could be derived from material ejected

from the mantle of the Earth, even though all impact simulations to date have predicted a lunar origin from the impactor mantle.

6.2 How did the Moon become volatile element depleted, and is it enriched in refractory elements?

A large group of elements – the volatile elements – is substantially more depleted in the Moon than the Earth's mantle (**Fig. 20;** Ringwood, 1979). A compelling explanation for this difference has been elusive. Many processes can potentially fractionate volatile elements, such as evaporation or condensation. Detailed studies of potassium isotopes in a wide range of planetary and meteoritic materials resulted in an astounding discovery – that K isotopes are not fractionated at all (Humayun and Clayton, 1995). Cadmium, another volatile element with a condensation temperature of 684 K, is also unfractionated in lunar and terrestrial materials (Schediwy et al., 2006), arguing against any major differences between the Earth and Moon. Slight but measurable fractionations in Fe isotopes and differences between the Earth and Moon and other materials led Poitrasson et al. (2004) to propose that these were due to partial vaporization. However, it is not clear that these effects were not caused by some other secondary process such as oxidation (e.g., Williams et al., 2004). Because Li is one of the lightest elements, it is more susceptible to fractionation and has been the focus of many recent studies. Lithium appears to be unfractionated in the Earth and Moon, but a significant amount of variation can be caused by aqueous alteration and magmatic fractionation, making interpretation of this volatile element somewhat difficult (e.g., Magna et al., 2006; Seitz et al., 2005, 2006; Sephton et al., 2006).

Missing so far from this topic is a rigorous model that will predict both major, minor element and isotopic fractionations associated with partial vaporization and recondensation of material during and after a giant impact. Such modelling may also lead to new insights regarding the issue of whether the Moon is enriched in refractory elements (e.g., Taylor et al., 2006; Hagerty et al., 2006). Coupled assessment of refractory and volatile elements for the Earth-Moon system may lead to a better understanding of the conditions during Moon formation after the giant impact.

6.3 Was there a late chondritic veneer?

In an attempt to explain the near chondritic relative abundances of the highly siderophile elements in Earth's upper mantle (HSE; Au, Re and the platinum group elements), Chou (1978) proposed that they were delivered to the Earth after core formation, by late addition of carbonaceous chondrite material. Because such material would also contain volatiles (H_2O, C, noble gases; Marty and Yokuchi, 2006), the late veneer also became a way to bring water after a dry accretion. Some recent dynamic models for terrestrial planet formation even accrete veneer-like masses after giant impacts (O'Brien et al., 2006). However, the geochemical evidence for a late veneer has been challenged and re-assessed over the last decade. First, as discussed above, it seems there are ways of accreting the Earth with water present. Second, with the recognition that many moderately siderophile elements can be explained by high temperature and pressure metal-silicate equilibrium (e.g., Walter et al., 2000), some have questioned how such equilibrium would also affect the HSE. Although isotopic work has placed tight constraints on the near chondritic Re/Os and Pt/Os of the primitive terrestrial mantle (e.g., Brandon et al., 2000; Carlson, 2005), no known volatile-rich chondrite provides a

match to the Os isotopic constraints (Brandon et al., 2005), thus not necessarily requiring a connection between water and HSEs (Drake and Righter, 2002). Furthermore, there are suprachondritic HSE ratios such as Pd/Ir and Ru/Ir in peridotite massifs and xenolith (Becker et al., 2006; Schmidt et al., 2000; Pattou et al., 1996). Advances in understanding D(HSE) metal/silicate have been slow due to the experimental and analytical challenges associated with the HSE (e.g., Righter, 2005). Several recent high PT studies will be discussed here.

In one of the first high PT studies of D(Pd) and D(Pt), Holzheid et al. (2000) showed that both partition coefficients decrease at higher pressures and temperatures, but not enough to explain the mantle concentrations of Pt or Pd by metal/silicate equilibrium. One drawback to their experiments was the unusual K- and Si-rich silicate melt used in the experiments – very different from a peridotite - as well as the use of S-free metal. Focusing only on Pt, Ertel et al. (2006) measured to solubility of Pt in an Fe-free basalt – the diopside-anorthite system eutectic composition – to high pressures and temperatures, and found that D(Pt) does not decrease enough to explain the Pt concentrations of the upper mantle. Again, drawbacks to this study included an FeO-free bulk composition, S-free system, and questionable assumptions about oxygen fugacity in graphite dominated systems. Both of these studies were followed by experiments in systems more closely approaching that of the early Earth.

In a study utilizing a novel technique using secondary ion mass spectrometry to analyze Au in silicate melts, Danielson et al. (2006) demonstrate that D(Au) metal/silicate decreases substantially with increased pressure and temperature. Because they used a peridotite melt and S-bearing metallic liquids, it makes one wonder whether

D(Pd) and D(Pt) could also be lower in systems approaching that of the terrestrial mantle composition. Again focusing only on Pt, Cottrell and Walker (2006) found that D(Pt) decreases substantially at the high PT conditions of an early magma ocean. However, their conclusion rests on an interpretation that the tiny metallic nuggets plaguing many such experiments, were formed upon quench. There is no agreement on this issue, so the general question of HSE solubility at high pressures and temperature remains unresolved. Key to resolution will be the ability to work in nugget-free, uncompromised conditions, and to utilize experimental compositions that approximate those of the early Earth (i.e., peridotite and light element bearing FeNi metallic liquid).

6.4 Is the Earth's atmospheric loss linked to the giant impact ?

Xenon isotope data have shown that the Earth became closed to Xe loss at approximately 100 Ma after T_0 (e.g., Podosek and Ozima, 2000). However, this interpretation depends to some extent upon the inventory and sources of Earth's noble gases (e.g., Genda and Abe, 2005). Traditionally, the solar-like noble gas isotopic patterns for the terrestrial interior led to the idea that Earth acquired its volatiles from a solar nebular source (e.g., Becker et al., 2003; Pepin, 1991). Acquiring solar nebular gas during a protracted terrestrial accretion (100 Ma after T_0) is difficult as nebular gas tends to be dissipated within 1 to 10 Ma after T_0. Two alternatives to a purely solar nebular origin are a) that the Earth's building blocks were meteoritic and some meteorites have a solar component as well (Busemann et al., 2004), or b) that there was solar wind implantation on early planetesimals (Podosek, 2003). A solution to this problem may lie somewhere in the middle, but some of the problems with Earth acquiring nebular gas may be relaxed if the faster timescales for terrestrial accretion are correct (e.g., those

described in sections 3.1 and 3.2). A recent model proposed by Pepin and Porcelli (2006) argues for two episodes of mantle degassing to explain the noble gas data: an early 20-50 Ma event (Moon-forming impact) followed by a later 95-100 Ma event (a later impact), with two distinct and separate reservoirs being formed and isolated from each other over geologic history. Reconciliation of the Xe age of the Earth and it noble gas inventory remains unsolved, and should be the focus of future efforts – particularly those that also address and satisfy geophysical constraints.

7. How rare is the Earth-Moon system?

An understanding of how the Earth and Moon formed will ultimately allow us to address the broader question of "How do you make a habitable planet?". This has been a topic of keen interest to many influential scientists in the 20th century. I was influenced as a teenager by a wonderfully thought provoking book by Isaac Asimov called "Extraterrestrial Civilizations" (1979) in which he laid out the groundwork for understanding how life could evolve on a rocky planet like Earth. He made it seem easy. Later in graduate school I found a similar optimistic attitude in the writings of George Wetherill (e.g., Wetherill, 1996).   At the same time, I encountered views of the possible "special conditions" required for the development of life on a planet (e.g., S.R. Taylor, 1999). However, that viewpoint was popularized by Ward and Brownlee (2000) when they argued that the Earth is indeed so rare that it is the most diverse planet in 10,000 light years. Part of their argument is based on the difficulties involved in making a large Moon around the Earth – one that keeps Earth's obliquity low and thus is climate stable (Williams and Pollard, 2000). In 2000 our understanding was that there exists a narrow

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set of dynamic factors that would allow the formation of our Moon by a giant impact. It is now clear from extended modeling, and the new results discussed in this review, that perhaps the Moon is not so rare and such moons can form over a very broad range of conditions during terrestrial planet accretion. In fact, maybe this is how Pluto's moon Charon formed as well (Canup, 2005). In addition, astronomers are questioning whether small and cool terrestrial-like planets are more common in the universe than previously thought (e.g., Beaulieu et al., 2006; Raymond et al., 2006). Several extrasolar planetary systems with a rocky component (in addition to ice and gas) have been found in recent years (e.g., Lovis et al., 2006; Rivera et al., 2005), and although these are characterized broadly as Earth-like, planets similar to Earth have not yet been found. Nonetheless, the hunt is on, and progress in this field is rapid - perhaps the balance is tipping back to Wetherill and Asimov – time will tell. I hope this review is helpful to those who are trying to keep track of the incredible progress this field is experiencing.

Acknowledgements

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Figure Captions

Figure 1: Comparison of high resolution oxygen isotope data for Moon, Mars, Vesta, and the angrite parent body showing the distinct fractionation lines for each body. Also shown are the terrestrial fractionation line (TFL), mars fractionation line (MFL), angrite fractionation line (AFL) and eucrite fractionation line (EFL) (data are from Greenwood et al., 2004 and Wiechert et al., 2001). Data from these two studies have been replotted at the same scale for ease of comparison. $\delta^{18}O$ SMOW is defined as $[1- (^{18}O/^{16}O$ (sample) - $^{18}O/^{16}O$ (standard mean ocean water or SMOW) $)*1000$. $\Delta^{17}O = \delta^{17}O - 0.5245 * \delta^{18}O$.

Figure 2: Illustration of the dependence of metallic liquid droplet diameter (A) and equilibration distance (B) as a function of silicate melt viscosity. The range of silicate melt viscosities in a terrestrial magma ocean is indicated with double ended arrows. Metallic liquid droplet sizes will not grow above 10 cm in this range due to dynamic shear instabilities generated by one liquid falling through another. Figures are from Rubie et al. (2003).

Figure 3: Variation of mass fraction of metallic core, wt% FeO in mantle, and temperatures derived from siderophile element modelling for each of the inner terrestrial planets as well as the Moon and Vesta. Figures A and B are from Righter et al. (2006) and C is a summary of the studies of Righter et al. (1998, 1999), Righter and Drake (1997) and Righter (2002).

Figure 4: Illustration of the difference in tungsten isotopic composition of carbonaceous and enstatite chondrites from that of the Earth (Kleine et al. 2002). Vertical dashed line is the previous location of the chondritic value (Lee and Halliday, 1996). Epsilon (ε) W is defined as $[^{182}W/^{184}W$ (sample) $/$ $^{182}W/^{184}W$ (standard) - 1] * 10000, so that variations in the tungsten isotopic ratio can be expressed in small, rather than unwieldy numbers.

Figure 5: εW versus $^{181}Ta/^{184}W$ for mineral separates from the high Ti mare basalt 70035. Intersection of the line with the y-axis yields the initial εW of +1.5. This approach was required to correct many of the lunar samples from cosmogenic derived ^{182}W (Lee et al., 2002).

Figure 6: Models of Halliday (2004b) showing how the bulk silicate Earth could have been derived from a giant impact involving 26% and 4% of Theia's (the proto-Earth) core equilibrated with the bulk silicate Earth. In the former case, Hf/W = 15, and the latter case Hf/W = 5, and in both cases D(W) is fixed at a value of 15. Vertical dashed line represents the time (in million years or Myr) at which the impacts occur, relative to the start of the solar system.

Figure 7: Variation of mass accreted to the Earth as a function of time, based on the modelling of Canup and Agnor (2000) (Figure C). Abrupt vertical lines in each curve represent accretion events in which a large mass of material is added to the Earth. The value of D(W) also changes as the planet grows (and pressure and temperature increase), thus also affecting the value of Hf/W in the mantle (A and B, respectively). Clearly a

realistic model must take these changing parameters into account. See Righter (2003) for more details.

Figure 8: Variation of D(Hf) and D(W) with garnet composition. The range of terrestrial mantle garnet compositions are indicated by shaded regions, and illustrate the large potential differences between D(Hf) and D(W). Fractionation of Hf/W in a planetary mantle by garnet and also clinopyroxene is possible, and can be as significant as metal-silicate fractionation. Figure modified from Righter and Shearer (2003).

Figure 9: ε^{142}Nd for chondrites, eucrites and terrestrial samples, as measured by Boyet and Carlson (2005). The difference between terrestrial samples and chondrites indicates that the terrestrial mantle contains an ancient hidden enriched reservoir, possibly in the deep mantle. The simplest model involves differentiation of the Earth as early as 35 Ma after T_0, much earlier than ever proposed previously. Epsilon $(\varepsilon)^{142}$Nd is defined as $[^{142}$Nd/^{144}Nd (sample) / ^{142}Nd/^{144}Nd (standard) - 1] * 10000. Some of the cumulate eucrites (Moore County, Moama, and Binda) have positive values resulting from high Sm/Nd values produced during igneous processes on their parent body.

Figure 10: Measurements of ε^{142}Nd and source ^{147}Sm/^{144}Nd for a suite of Apollo and meteorite lunar samples. The coincidence of the data with a chondritic initial lunar mantle indicates that in contradistinction to the depleted terrestrial mantle, the Moon originates from chondritic material (from Rankenburg et al., 2006).

Figure 11: Plot of Nb/Ta of Moon, present silicate Earth and the range of allowable contributions (in %) of impactor material. The Nb/Ta ratio of the Moon can be explained by either 0% impactor, or up to 50% chondritic impactor with a high Nb/Ta ratio of 20. Also shown are the ranges predicted for geophysical models as horizontal shaded region (Münker et al., 2003).

Figure 12: Results of studies of O'Brien et al. (2006) for four different simulations (EJS 1 to EJS 4) where material from four different parts of the inner solar system (shaded regions) is tracked in the final outcomes of three to four terrestrial planets. Scale at bottom is from 0.25 to 2.25 Astronomical Units (AU). The interesting feature is that material from the outer parts accretes into the innermost planet, and material from the inner parts accretes into the outermost planet. As a result, there is some radial mixing that can cause heterogeneity in planet compositions, outside of that directly caused by narrow feeding zones. The abbreviation EJS refers to eccentric orbits for Jupiter and Saturn (from O'Brien et al., 2006).

Figure 13: Range of impactor masses and total masses in giant impact simulations summarized by Canup (2004a), that can produce a Moon-like satellite around an Earth-like planet. These conditions ranging from an early impactor to late impactor, broaden the range of possible conditions during which a Moon-like satellite can form.

Figure 14: Summary diagram, from Nimmo and Agnor (2006) illustrating the control of timing, impactor state, target state, and equilibration style on the final tungsten isotopic

composition of the target. Core merging scenarios produce the most radiogenic W, impact of an un-differentiated impactor without further equilibration results in moderately radiogenic mantle, whereas mantle equilibration scenarios result in the least radiogenic targets.

Figure 15: Comparison of D/H ratios in clays in ordinary chondrite (LL3) meteorites comets, meteorites, Mars' atmosphere and mantle, carbonaceous chondrites, Earth, interplanetary dust particles (IDP), and protosolar H_2 as represented by Jupiter, Saturn, and the Sun (from Robert, 2001). Mixtures of comets and protosolar sources can provide a match to Earth's oceans, as could entirely meteoritic sources. Alternatively, we do not know the D/H ratios of Jupiter family comets. And, it is possible that Jeans escape of H (mass dependent fractionation at the top of the atmosphere) from the early Earth could have pushed the D/H ratio to higher values but this may require unrealistic amounts of H loss.

Figure 16: 4.0 to 4.3 Ga zircons from the Beartooth Mtns. (Wyoming) and Jack Hills Yilgarn Craton (Australia) have $\delta^{18}O$ higher than the terrestrial mantle (lower right hand corner of figure). Some have interpreted this as evidence for the presence of liquid water on the surface of the Earth very early in history (figure from Valley et al. (2005).

Figure 17: Water storage may be possible at low temperatures, such as in serpentine and saponite in the matrix of the Tagish Lake carbonaceous chondrite (Flynn and Keller, 2001), or at higher temperatures such as in the calcic hornblende- and biotite-bearing R chondrite LAP 04840 (Righter and Neff, 2007).

Figure 18: Calculated water contents in a molten chondritic mantle using the water solubility model of Moore et al. (1998). An asteroid the size of Vesta (500 km diameter) could dissolve as much as 3 wt% water, but smaller asteroids would not dissolve appreciable amounts due to the low pressures.

Figure 19: The oxygen isotopic composition of meteoritic and planetary materials (b) has a relatively small variation compared to proposed solar reservoirs (a). Figure is from Huss (2006). a) Oval field in upper right side was proposed by Ireland et al. (2006) as the location of a solar oxygen isotopic reservoir. Cross within a polygon in lower left side represents the location of the solar oxygen isotopic composition proposed by Hashizume and Chaussidon (2005). TFL is the 'terrestrial fractionation line', and ^{16}OFL is the 16 oxygen fractionation line.
b) Enlarged from dashed box in (a). This region shows the location of the Moon, Mars and Earth fractionation lines, compared to the ^{16}OFL, the terrestrial fractionation line (TFL), and the carbonaceous chondrite anhydrous meteorite (CCAM) line. Other shorter lines parallel to the TFL are possible O isotopic variations on small asteroid size parent bodies. A more complete understanding of the origin of these small variations awaits study of the Genesis and Stardust materials that will hopefully aid in constraining the solar reservoirs, and the degree to which water may effect oxygen isotopic composition.

Figure 20: Depletion of volatile elements in the lunar mantle relative to the terrestrial mantle (Ringwood, 1979). An explanation for this large group of elements has eluded geochemists since initial study of Apollo samples.
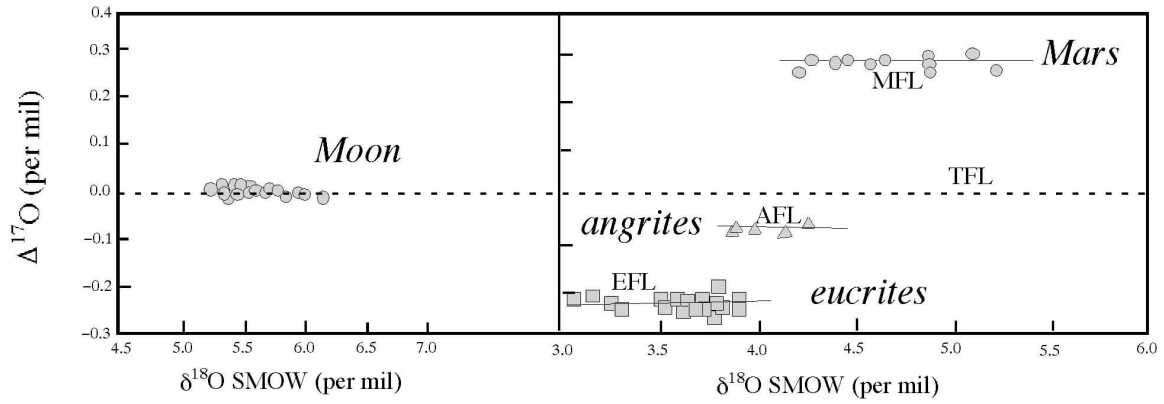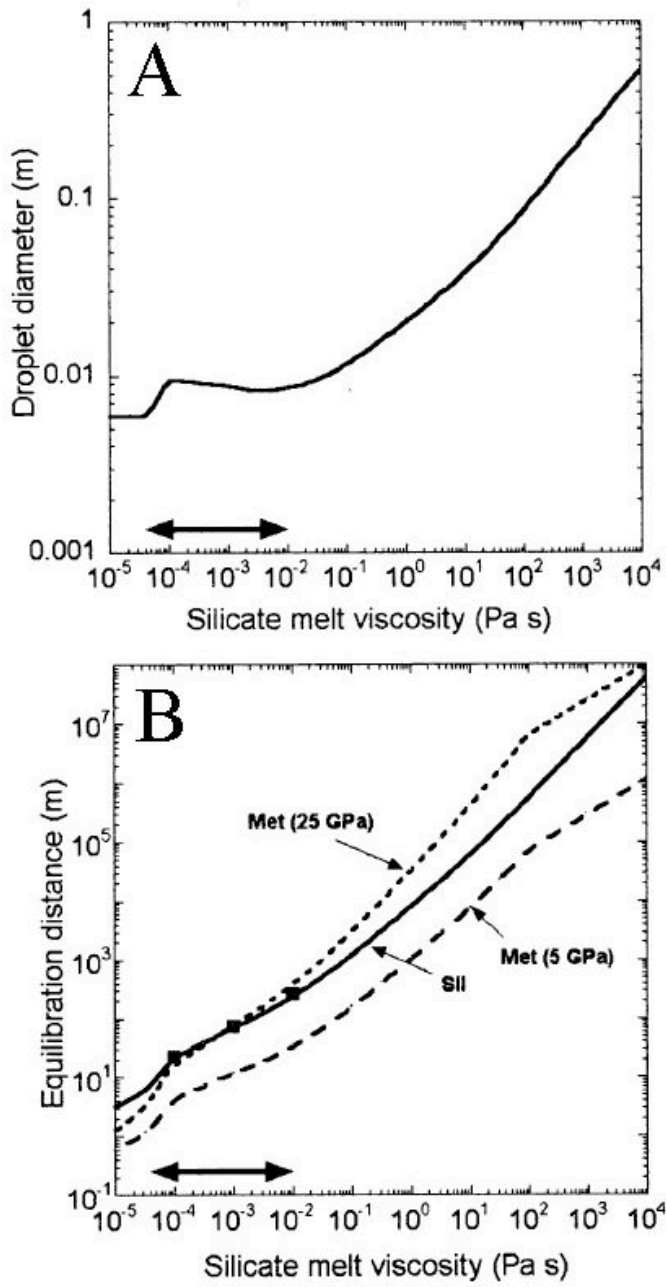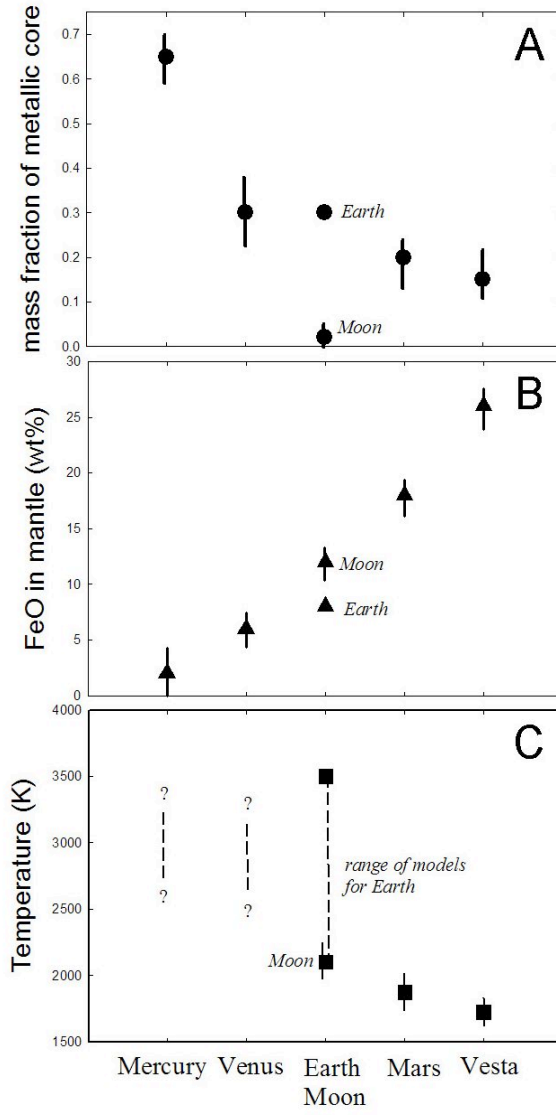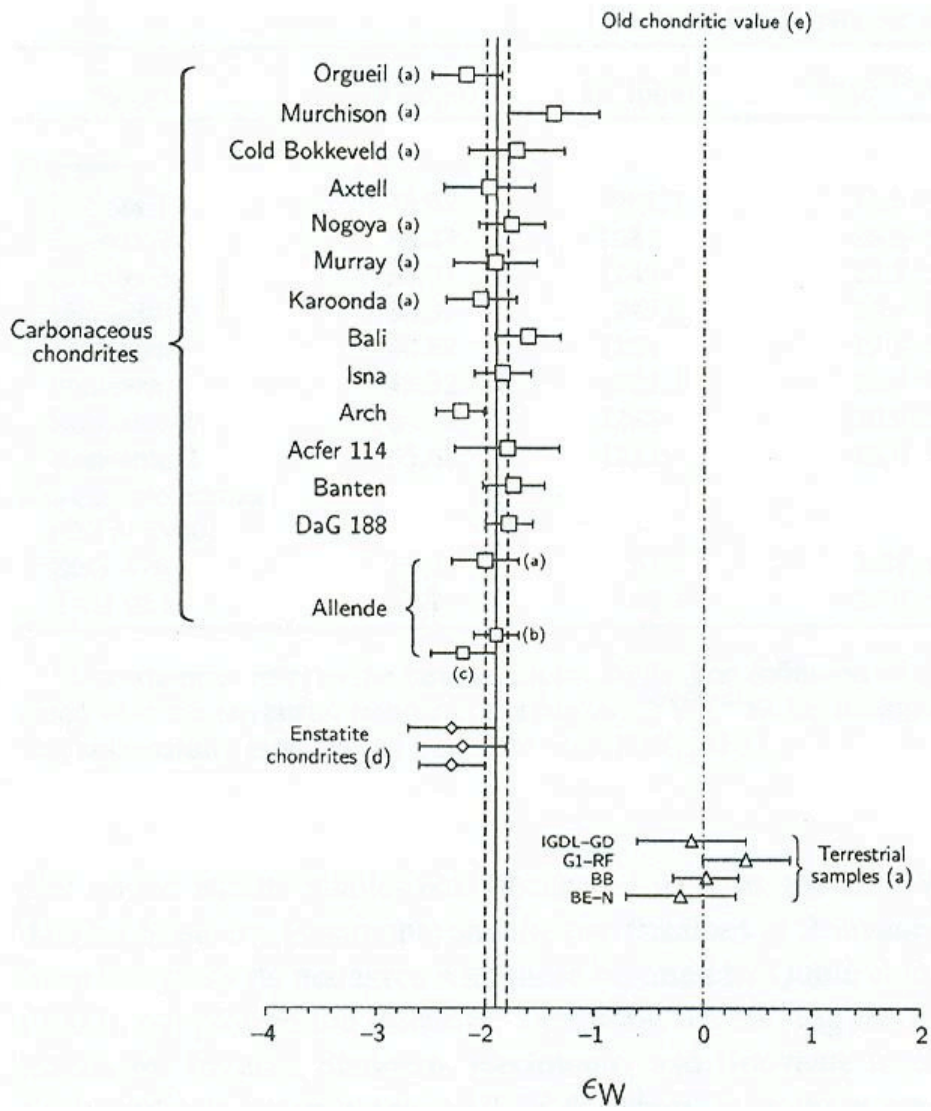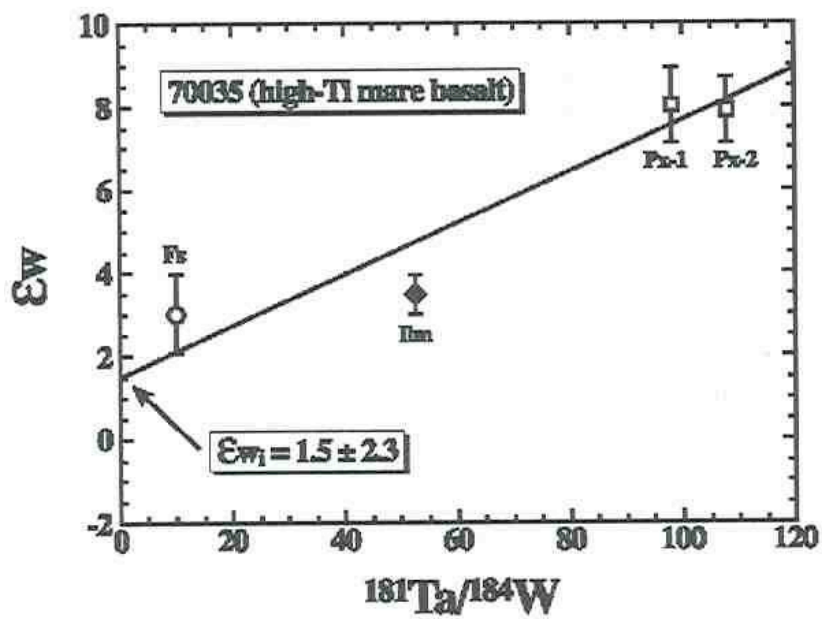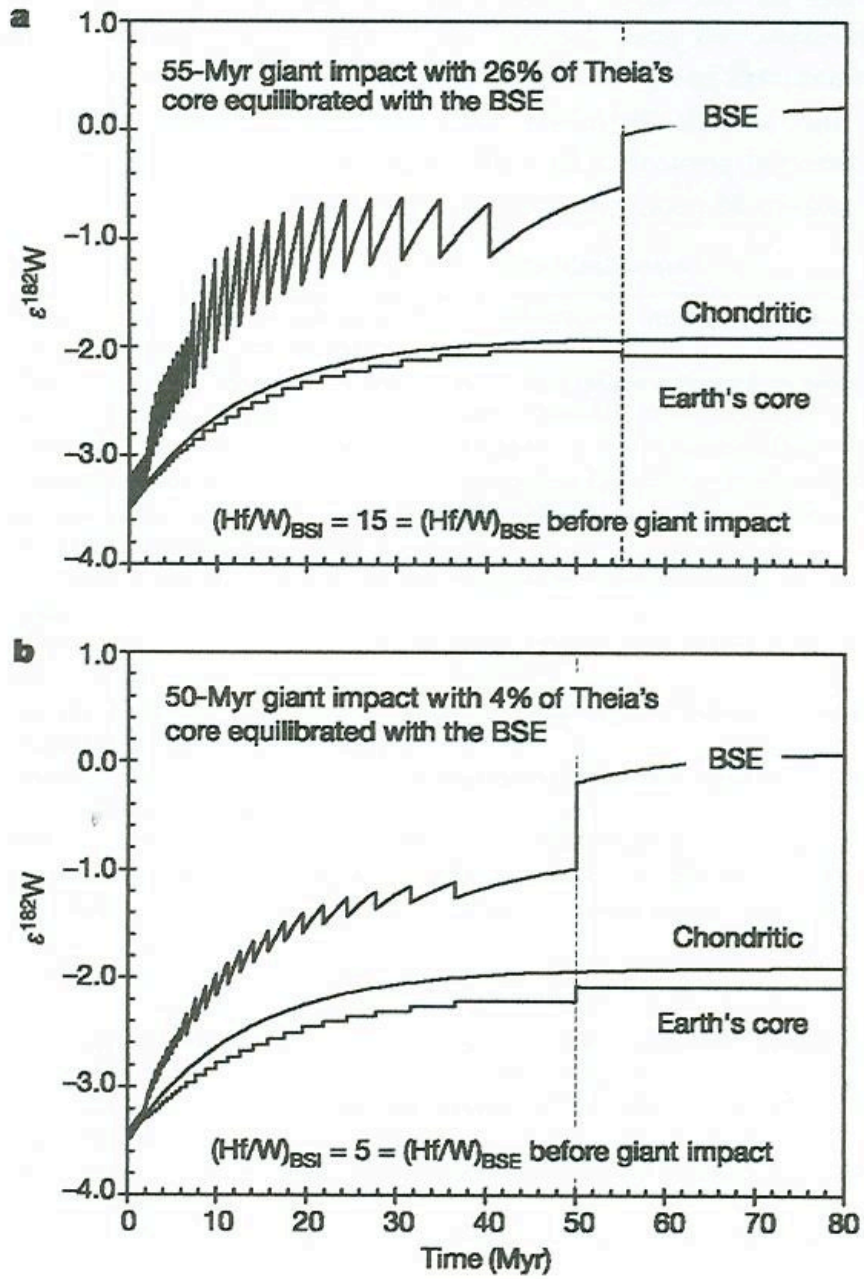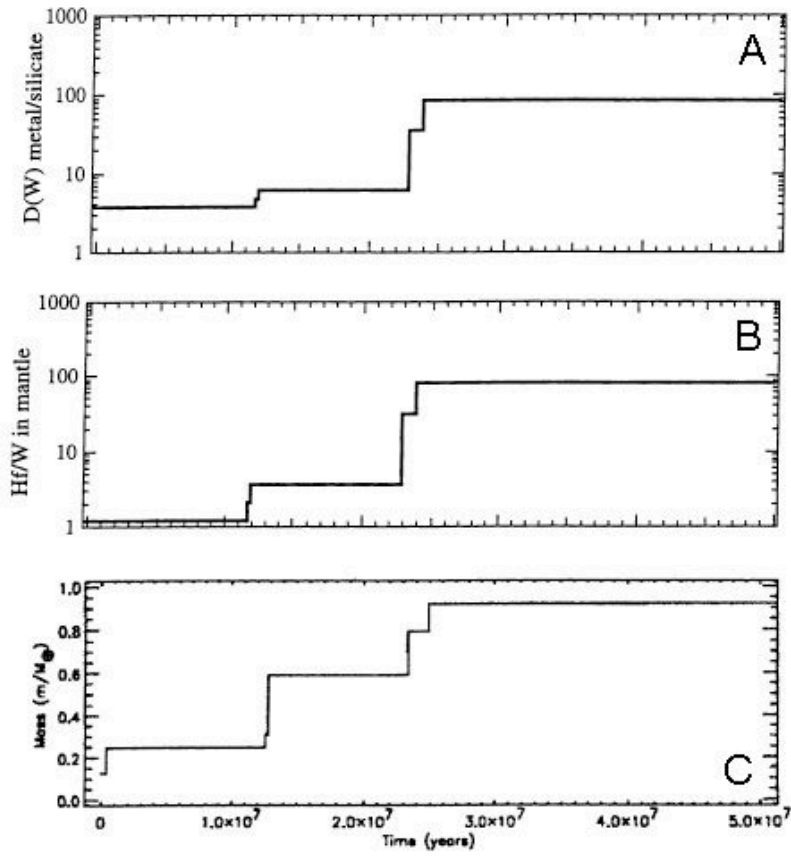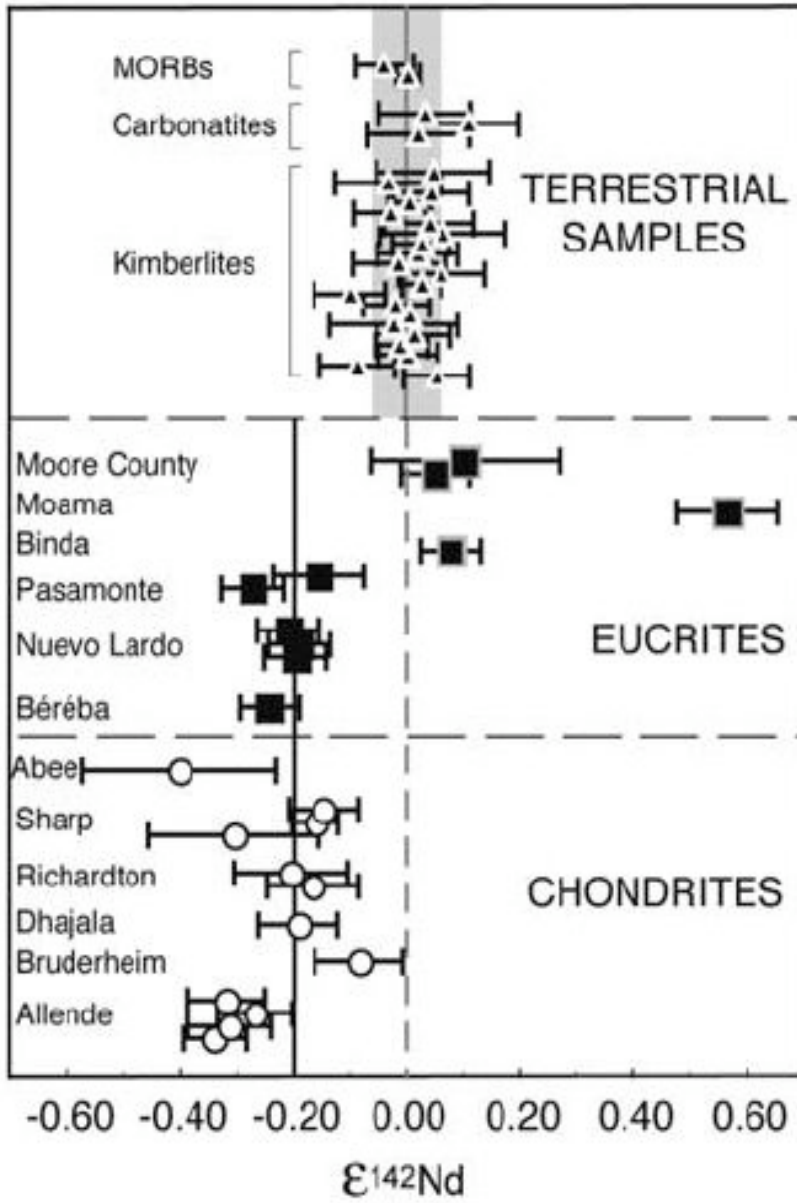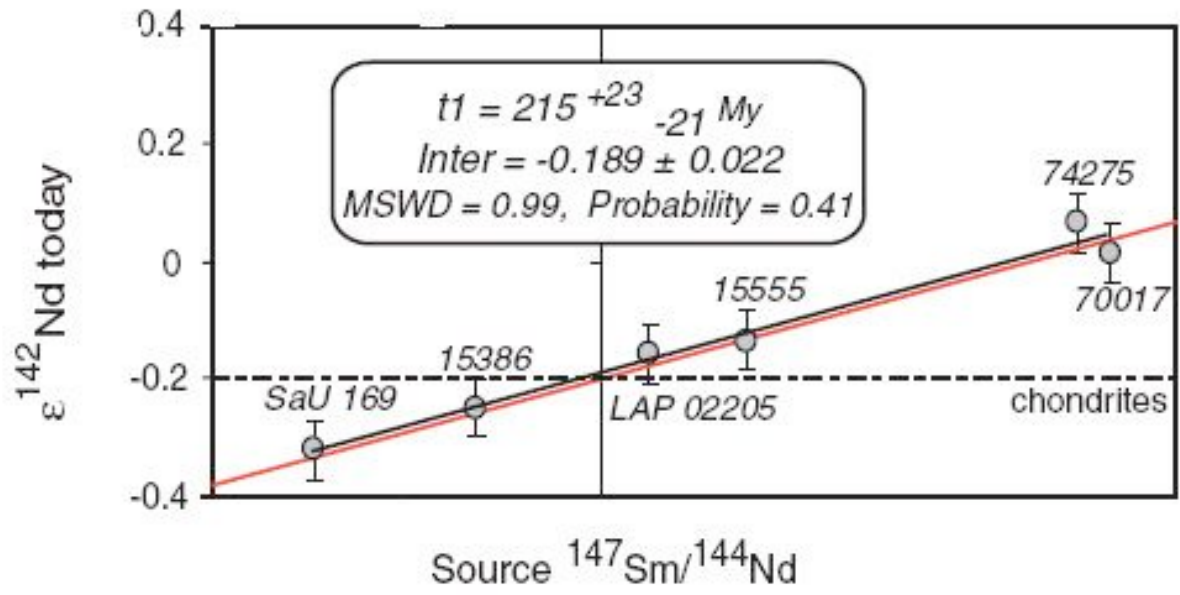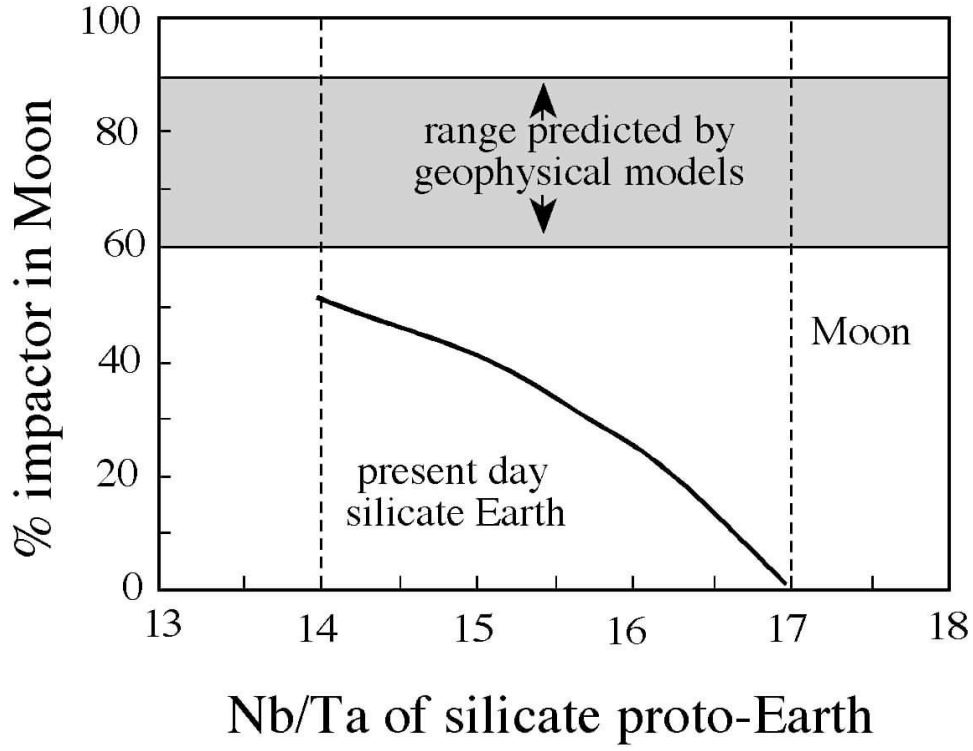
Figure 1

Figure 2

Figure 3

Figure 4

Figure 5

70035 (high-Ti mare basalt)

$\varepsilon_{W_i} = 1.5 \pm 2.3$

55

Figure 6

a. 55-Myr giant impact with 26% of Theia's core equilibrated with the BSE

$(Hf/W)_{BSI} = 15 = (Hf/W)_{BSE}$ before giant impact

b. 50-Myr giant impact with 4% of Theia's core equilibrated with the BSE

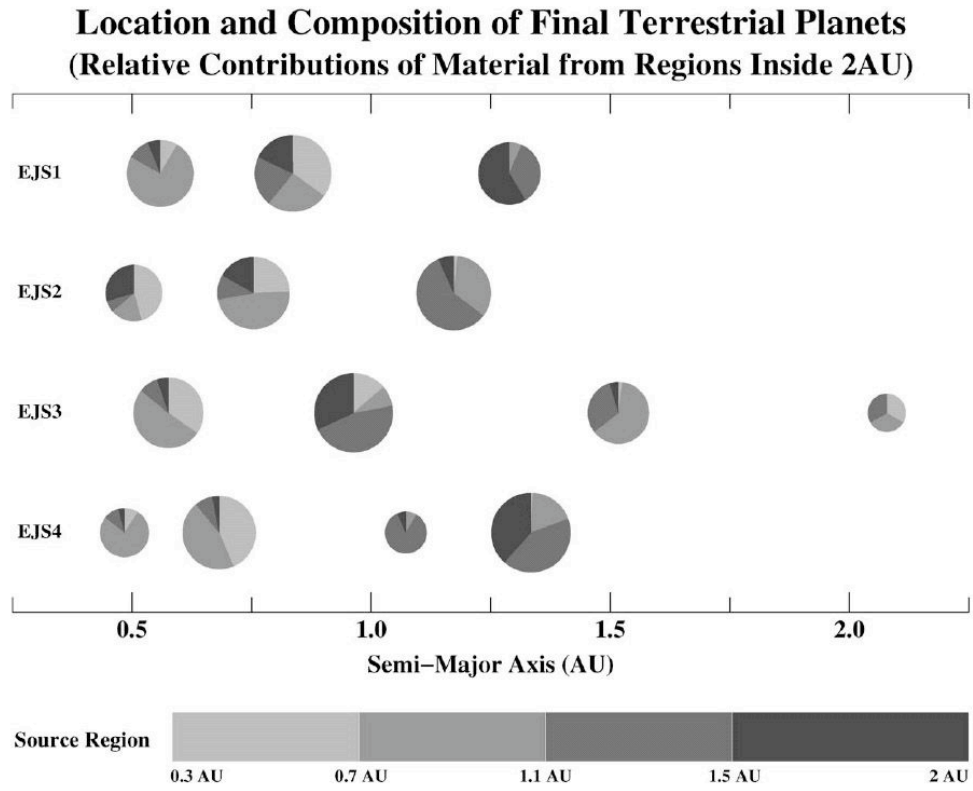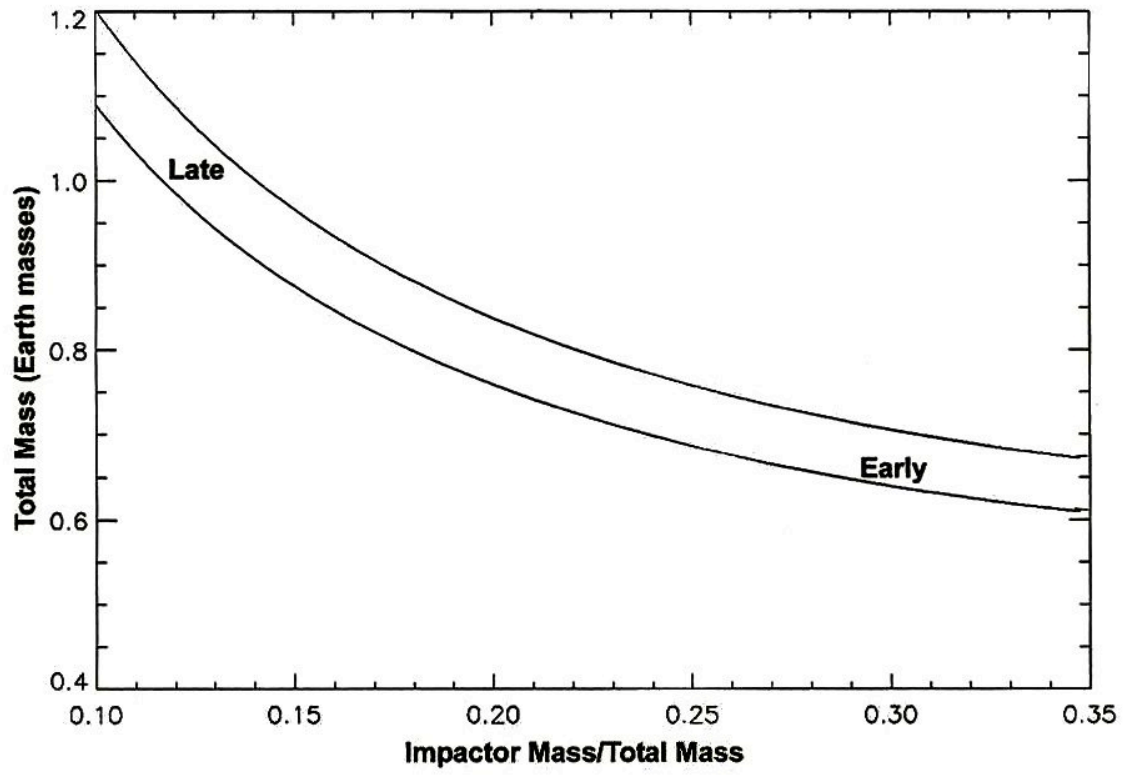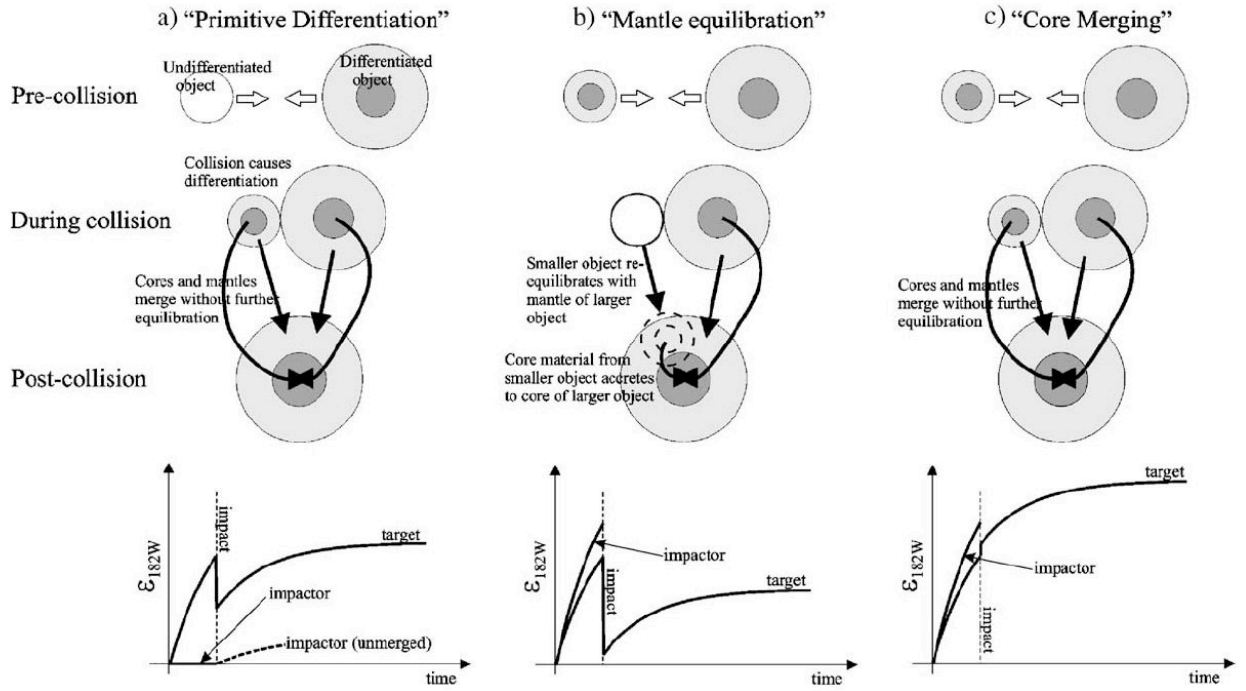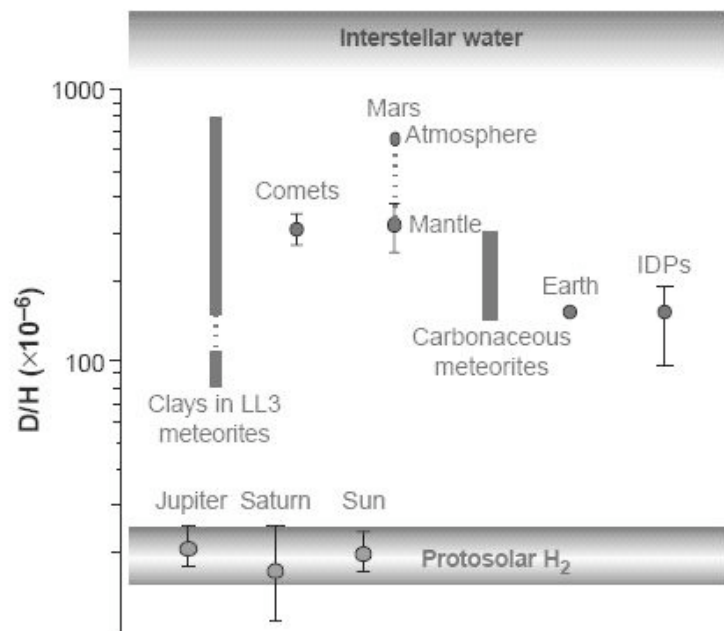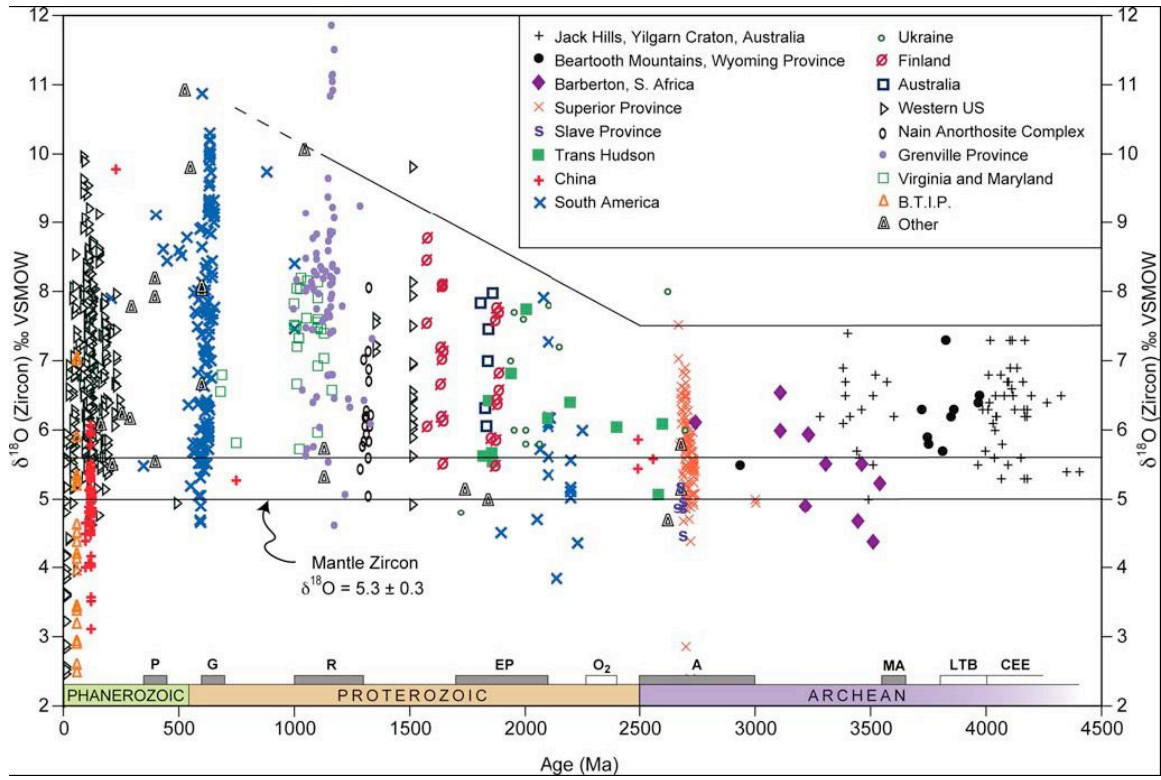$(Hf/W)_{BSI} = 5 = (Hf/W)_{BSE}$ before giant impact
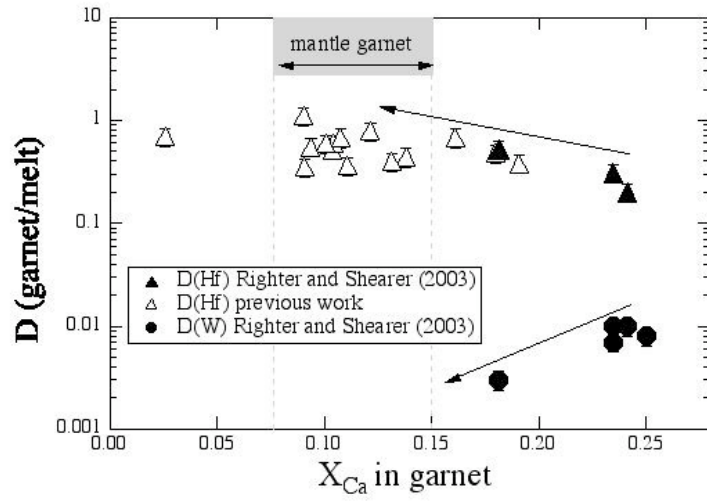
Figure 7

Figure 8

Figure 9

Figure 10

Figure 11

Figure 12

Location and Composition of Final Terrestrial Planets
(Relative Contributions of Material from Regions Inside 2AU)

Figure 13

Figure 14

a) "Primitive Differentiation"   b) "Mantle equilibration"   c) "Core Merging"

Pre-collision

Undifferentiated object   Differentiated object

During collision

Collision causes differentiation

Cores and mantles merge without further equilibration

Smaller object re-equilibrates with mantle of larger object

Core material from smaller object accretes to core of larger object

Cores and mantles merge without further equilibration

Post-collision

ε_{182W} impact   target   impactor   impactor (unmerged)   time

ε_{182W} impactor   impact   target   time

ε_{182W} target   impactor   impact   time

64

Figure 15

Figure 16

Figure 17

Figure 18

Solubility of water with pressure in a molten embryo mantle
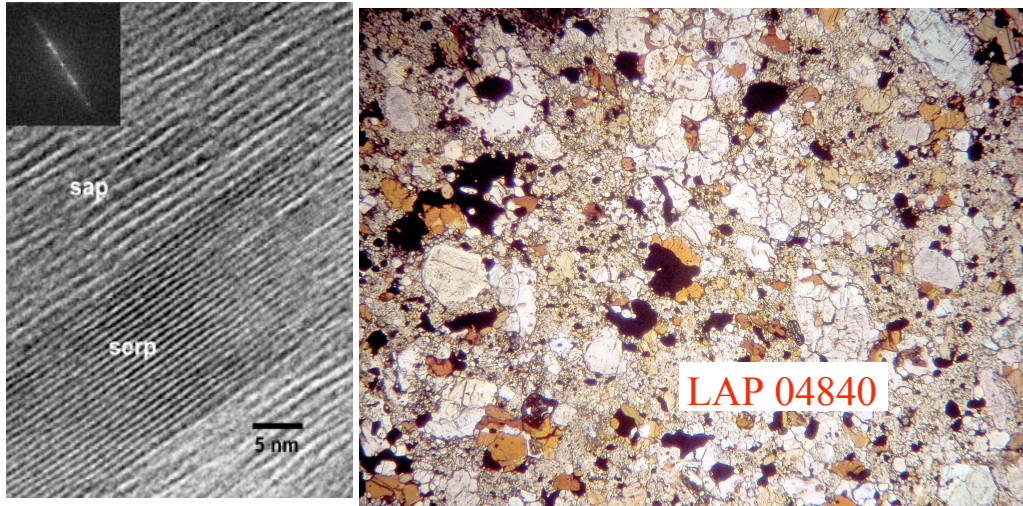
Vesta (500 km)

medium (100 km)

small (20 km)

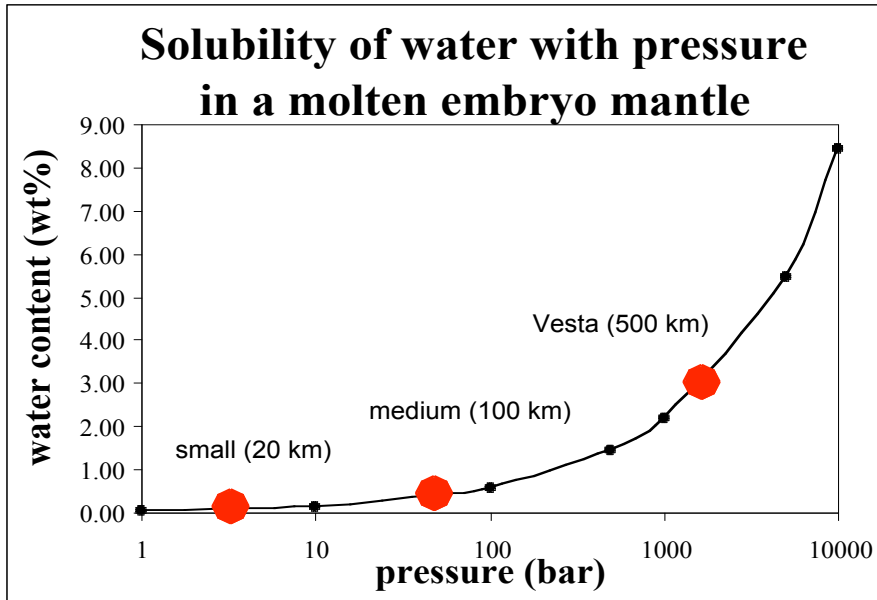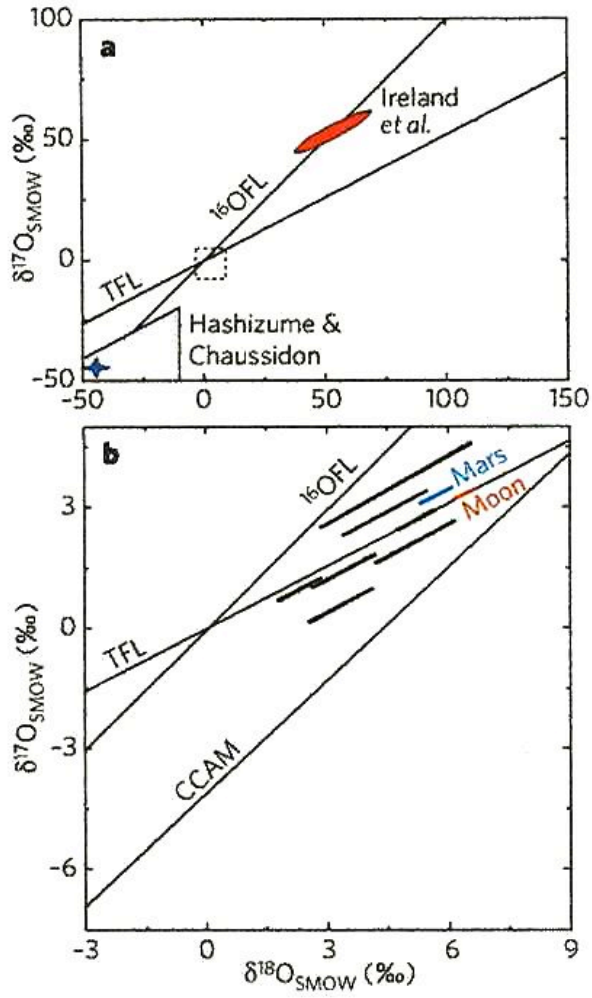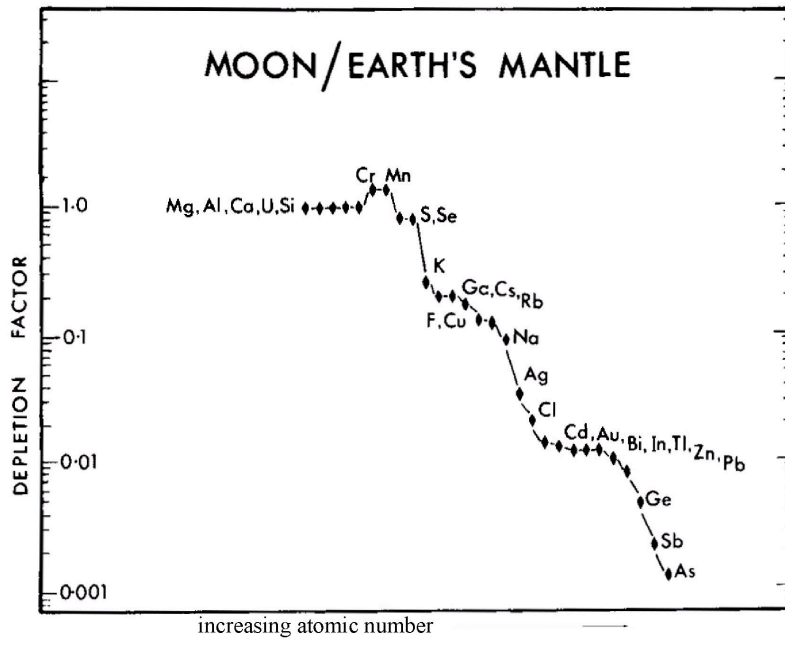water content (wt%)

pressure (bar)

Figure 19

Figure 20

Table 1: Summary of results of short-lived chonometers for the age of the Earth

System	Half life (Ma)	Age range (Ma after To)	Reference
^{182}Hf-^{182}W	9	25-50	1
^{146}Sm-^{142}Nd	10.3	30	2
^{107}Pd-^{107}Ag	6.5	10-29	3
^{97}Tc-^{97}Mo	2.6	19-24	4
^{92}Nb-^{92}Zr	36	>50 (silicates)	5

1) Kleine et al. (2004); 2) Boyet and Carlson (2005); 3) Schönbächler et al. (2006); 4) Yin and Jacobsen (1998); 5) Münker et al. (2000).