Early differentiation of the Earth and the Moon

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We examine the implications of new $^{182}$W and $^{142}$Nd data for Mars and the Moon for the early evolution of the Earth. The similarity of $^{182}$W in the terrestrial and lunar mantles and their apparently differing Hf/W ratios indicate that the Moon-forming giant impact most probably took place more than 60 Ma after the formation of calcium-aluminium-rich inclusions (4.568 Gyr). This is not inconsistent with the apparent U–Pb age of the Earth. The new $^{142}$Nd data for Martian meteorites show that Mars probably has a superchondritic Sm/Nd that could coincide with that of the Earth and the Moon. If this is interpreted by an early mantle differentiation event, this requires a buried enriched reservoir for the three objects. This is highly unlikely. For the Earth, we show, based on new mass-balance calculations for Nd isotopes, that the presence of a hidden reservoir is difficult to reconcile with the combined $^{142}$Nd–$^{143}$Nd systematics of the Earth’s mantle. We argue that a likely possibility is that the missing component was lost during or prior to accretion. Furthermore, the $^{142}$Nd data for the Moon that were used to argue for the solidification of the magma ocean at ca 200 Myr are reinterpreted. Cumulate overturn, magma mixing and melting following lunar magma ocean crystallization at 50–100 Myr could have yielded the 200 Myr model age.

Keywords: Earth; Moon; giant impact; differentiation; $^{142}$Nd and $^{182}$W

1. Introduction

Understanding the formation of planets has been proven to be a crucial task for constraining their later evolution. First, this question is inherently linked to the bulk composition of the planet since designing a scenario for planet accretion and differentiation is often necessary to derive their composition (Allègre et al. 1995; McDonough & Sun 1995). Particularly crucial in this respect are the behaviours of siderophile elements. Namely, the incorporation of these elements in the core will be a function of the accretion scenario and oxidation state of the Earth. This will in turn determine the chemical composition of the mantle. Second, as early differentiation events rapidly follow accretion, a description of these processes is essential for determining the initial conditions of the Earth’s system. The initial

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energy available upon accretion may allow a wholesale differentiation that may be difficult to erase at a later stage. There are now indications that, following magma ocean crystallization, a dense and enriched reservoir could have formed in the lower mantle (Boyet & Carlson 2005; Labrosse et al. 2007), and fluid dynamical constraints on mantle convection have shown that, if an initial density stratification existed in the Earth, then it is likely that, it could have persisted over time scales of several billion years (Davaille et al. 2002). Third, the early evolution of planets will also determine their rate of cooling, and this has implications on stirring rates. A large planet is likely to cool slower, but this can be strongly compounded by the existence of an insulating lid. Another important parameter is the presence of water. If water is a component in the accreting material (as is the case on Earth), the lithosphere becomes more ductile and plate tectonics is more likely to take place. This will ultimately result in faster cooling rates.

In this paper, we examine recent data we have obtained for the Moon and Mars that pertain to the early evolution of the Earth and the Moon, and we discuss their implications for the evolution of the terrestrial planets. This approach further emphasizes that the study of nearby planetary objects can provide a wealth of information about the history of our own planet.

2. New constraints on the age of the Moon and termination of Earth’s accretion

(a) New $^{182}$W data and chronological implications

There are several reasons to argue that the giant impact represents the last significant stage of terrestrial accretion (Canup & Asphaug 2001; A. Morbidelli 2008, personal communication). First, significant accretion following the Moon’s formation could only have involved numerous small impacts because, otherwise, the angular momentum of the Earth–Moon system would have been substantially altered and the Moon might have been lost. Second, if significant amounts of mass had been accreted after the giant impact, there would be a potential problem in accounting for the siderophile elements in both objects. These considerations limit the amount of mass that could have accreted to the Earth after the giant impact to less than 0.05 Earth’s masses (Canup & Asphaug 2001). Third, if significant accretion had taken place after the Moon-forming giant impact, then the oxygen isotopes in the Earth and the Moon could have become different. High-precision oxygen isotope measurements have shown that, within approximately 0.02 ppt (parts per thousand), this is not the case (Wiechert et al. 2001). If we assume that the later incoming material had an approximately 1 per mil deviation in $\delta^{18}$O from the Earth (equivalent to the difference between Mars and the Earth), this limits the mass fraction of material accreted by a giant impact to less than 2 per cent. A similar calculation could be made for $\delta^{17}$O. Thus, it would seem that the age of the Moon would provide an age of termination for the Earth’s accretion (A. Morbidelli 2008, personal communication).

Several previous studies have attempted to derive a meaningful age for the formation of the Moon based on $^{182}$Hf–$^{182}$W chronometry (Lee et al. 1997; Shearer & Newsom 2000; Lee et al. 2002; Righter & Shearer 2003; Kleine et al. 2005; Touboul et al. 2007). One major difficulty has been to tackle the issue of cosmogenic production of $^{182}$W via the reaction $^{181}$Ta(n, $\gamma$)$^{182}$Ta followed by $\beta^-$-decay to $^{182}$W.
This reaction is obviously enhanced in silicate rocks for which Ta/W ratios are higher than in metals. After an initial report of large variations in the $^{182}$W/$^{184}$W of lunar whole rocks (Lee et al. 1997), it was realized that the enhanced $^{182}$W/$^{184}$W ratios in many samples reflect cosmogenic $^{182}$W production (Leya et al. 2000). Attempts to correct for these effects have yielded imprecise results owing to the large corrections required. The most direct method to determine the $^{182}$W/$^{184}$W unaffected by cosmic-ray effects is the W-isotope analyses of Fe metal separated from lunar rocks because these metals do not contain any Ta and hence cosmogenic $^{182}$W (Kleine et al. 2005; Touboul et al. 2007). As shown in figure 1, the new study by Touboul et al. (2007), which used larger sample sizes and focused on the analysis of high-purity metal separates, seems to represent the most self-consistent dataset. Remarkably, the W isotopes are constant among all the investigated lunar samples and similar to terrestrial samples. In the following we examine in more detail the implications of these results.

(i) New $^{182}$W data and lunar mantle evolution

The new $^{182}$W-isotope data for the Moon have three major important features. First, the W-isotope composition of the Moon is distinct from that of chondrites. Second, the $^{182}$W/$^{184}$W ratio of the lunar mantle is identical to that of the bulk silicate Earth (BSE) and, third, despite a significant range in Hf/W in their sources, all lunar samples have identical $^{182}$W, defined as

$$\left(\frac{^{182}W}{^{184}W}\right)_{\text{sample}} - 1\right) \times 10^4.$$

In the following, we discuss in more detail the implications of these features. This discussion represents an extension of the first report by Touboul et al. (2007).

The formation of the Moon by a giant impact implies that its original temperature must have been very high. As a consequence, the initial state of the Moon was so hot that it must have been molten and must have evolved initially as a magma ocean. The magma ocean concept proposed for the Moon is in part based on the existence of complementary Eu anomalies in the lunar crust (anorthosite) and lunar basalts (mare basalts) as well as the existence of the so-called KREEP component assumed to represent the latest stage of magma ocean crystallization. The data of Touboul et al. (2007) provide an important outlook on this question because they can be used to estimate a lower limit for the age of the lunar magma ocean. At face value, the data indicate that the closure of the Hf–W system took place more than 60 Ma after the beginning of the Solar System. In the context of a lunar magma ocean, one could assume that the closure of the system took place when the convection in the magma ocean slowed down (60% crystal fraction; Solomatov (2000), but see also the discussion below). As $^{182}$Hf is short-lived ($t_{1/2} = 8.9$ Ma), the absence of $^{182}$W differences in the lunar rocks implies that the various lunar reservoirs were remixed and isotopically homogenized until after $^{182}$Hf became effectively extinct. This corresponds roughly to a time of at least 60±10 Myr (Touboul et al. 2007), given the range in Hf/W ratios in the sources of lunar rocks (Shearer & Newsom 2000; Righter & Shearer 2003; Kleine et al. 2005). This time scale now seems to be consistent with the $^{142}$Nd data available for the Moon, indicating Sm–Nd closure at ca 200 Ma (Boyet & Carlson 2007), if one
takes the data at face value. One surprising aspect of the Sm–Nd data for the Moon is that it is much younger than the corresponding Sm–Nd age for Mars and the Earth (Caro et al. 2003; Boyet & Carlson 2005; Bennett et al. 2007; Debaille et al. 2007; Caro et al. 2008). This would imply that the Moon, in

Figure 1. Compilation of the W-isotope data for the Moon. Data sources: Lee et al. (2002), Kleine et al. (2005) and Touboul et al. (2007). The data clearly show that the $\varepsilon^{182}$W for the Moon is indistinguishable, within error, from the terrestrial composition. Circles, KREEP-rich samples; diamonds, high-Ti mare basalts; squares, low-Ti mare basalts. WR, whole rock.
spite of its smaller size, had a longer cooling time than Mars and the Earth. This would seem rather paradoxical and has been rationalized by Albarède & Blichert-Toft (2007), who argued that the lunar anorthosite crust formed a lid that insulated the Moon and considerably slowed down its cooling (Martin et al. 2006). Such a stable and early lid was not preserved on Earth and thus the cooling time could have been shorter. It is quite clear that conductive cooling through a boundary layer is much slower than for an unstable boundary layer, which is typically the case when plate tectonics causes the generation of new seafloor. There is, however, a difficulty in the scenario proposed by Martin et al. (2006), as the plagioclase forming the anorthosite crust only appears once the lunar magma ocean has cooled by 500°C and the crystal fraction has reached at least 70 per cent (e.g. Longhi 2003). By that time, the temperature in the lunar mantle would be approximately 1250°C and convection should be more sluggish. Above a 60 per cent melt fraction, convection of the crystal–melt assemblage becomes slower but the mobility of melts at a planetary scale could allow some chemical re-equilibration. Using the scaling of Solomatov (2007) for the case of the Moon, a typical differentiation time scale related to melt percolation would be ca 0.5 Ma, assuming a melt viscosity of $10^{-2}$ Pa s and a 1 per cent melt fraction. Thus, the chemical and isotope equilibration would effectively stop shortly after the 60 per cent crystal fraction is reached. This simple calculation does not consider the effect of the compositionally driven melt convection that could take place in the magma ocean cumulate pile. Thus, it is not clear how Nd isotopes would still effectively equilibrate at a planetary scale to yield a ca 200 Myr isochron (Boyet & Carlson 2007). There seems to be a contradiction between the predicted rapid cooling times and Sm–Nd data that calls upon a re-examination of this question. One possibility of explaining the young $^{146}\text{Sm}^{142}\text{Nd}$ apparent ‘age’ of the Moon is that thermal solid-state convection and later differentiation would have disturbed the original $^{146}\text{Sm}^{142}\text{Nd}$ isochron derived from magma ocean crystallization. In the following, we explore this possibility with a simple box model and its implications for the actual age of lunar differentiation. In this respect, the lunar data could be compared with the terrestrial data. For several long-lived isotope systems, we know that, in open and continuously differentiating systems, the slope of isochrons does not directly date a differentiation event (Allègre et al. 1995; Kellogg et al. 2002; Rudge 2006). Rather, the slope of a mantle isochron is a function of both differentiation and later mixing. Although the history of the Moon does not involve plate tectonics and thorough mixing, magmatic activity in the Moon lasted until ca 3 Ga ago. Thus, it is worth exploring models of continuing differentiation that would yield a 200–250 Myr isochron in the $^{146}\text{Sm}^{142}\text{Nd}$ system.

To approach this question, we have used a simplified statistical model including melting of the lunar mantle, radioactive decay and mixing. The model runs over a number of steps and, at each step, the melt is assumed to mix with a randomly chosen component already present in the lunar mantle (or crust). This model attempts to simulate the various processes affecting the lunar mantle: magma ocean crystallization; cumulate overturn; melting; and contamination. The essential processes involved are thus fractionation due to melting (or crystallization) and mixing. As shown in figure 2, this statistical model can
produce an array in the $^{142}\text{Nd}/^{144}\text{Nd}$ versus $^{147}\text{Sm}/^{144}\text{Nd}$ diagram that mimics the observations in the Sm–Nd data. The slope of this array does not represent the age of fractionation since, in this model, Sm–Nd fractionation starts at 70 Myr and proceeds until 3.5 Gyr ago. Rather, it is a function of the successive steps of fractionation and mixing. Arguably, there is more dispersion in the model array in figure 2, which indicates that the model does not perfectly match observations. On the other hand, the $^{142}\text{Nd}$ dataset for lunar rocks is still limited and more data should enable a better assessment. The basic conclusion of this simplistic modelling is that the apparent 200 Myr age for the Moon does not necessarily represent the age of crystallization of the lunar magma ocean. Rather, it reflects both crystallization of the lunar magma ocean and later processes that have both induced Sm–Nd fractionation and mixing.

(ii) Exploring scenarios for Moon–Earth equilibration following the giant impact

As mentioned above, an important aspect of the $^{182}\text{W}$ isotope in lunar rocks is that the $\varepsilon^{182}\text{W}$ values in the lunar and terrestrial mantles are identical while their Hf/W ratios are not. Using U/W, Th/W and Th/U of the lunar and terrestrial mantles, Touboul et al. (2007) estimated the following values: $(\text{Hf/W})_{\text{Moon}} = 26.4 \pm 1.5$ and $(\text{Hf/W})_{\text{Earth}} = 18 \pm 5$. As explained in Touboul et al. (2007), there are several ways of rationalizing this important observation. However, in order to derive time constraints on the giant impact, one needs to define a scenario for the interaction of the impactor with the Earth. One reasonable expectation is that, prior to the impact, the impactor was differentiated into a mantle and a core. Thus, upon impact, it is possible that the impactor’s core merged almost entirely with the Earth’s core. In this

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case, the equilibration would have involved an equilibration between the Earth’s and the impactor’s mantle. Alternatively, the impactor’s core might be entirely or partially mixed with the Earth’s mantle during the impact, in which case metal–silicate interaction will be the dominant process (Halliday 2004; Kleine et al. 2004; Nimmo & Agnor 2006). Another important aspect is the age of core differentiation for the impactor and the proto-Earth. Depending on the age of differentiation (Nimmo & Agnor 2006) and the degree of metal–silicate equilibration, the $\epsilon^{182}$W could be either highly radiogenic or simply close to chondritic values. Furthermore, the oxidation state of the impactor and the proto-Earth (Halliday 2004) will strongly influence the degree of Hf–W fractionation and hence the $\epsilon^{182}$W of the mantle. A model including all these aspects would have too many unconstrained parameters (Halliday 2004; Kleine et al. 2004; Nimmo & Agnor 2006). To make this problem more tractable while assessing the implications of the new data of Touboul et al. (2007), we have considered a model for the giant impact using only two parameters. First, we consider the W-isotope difference between the proto-lunar mantle and the proto-Earth, and we use this as the first parameter in the model. The second parameter of this model is the degree of re-equilibration ($f$) between the proto-Earth’s mantle and the proto-lunar mantle which is defined as

$$f = 1 - \frac{\Delta \epsilon^{182} W_{\text{after impact}}}{\Delta \epsilon^{182} W_{\text{before impact}}}.$$  

With this definition, $f=0$ means no equilibration, while $f=1$ means full equilibration. Our approach was to consider two-stage models for the evolution of the impactor because we will focus only on the latest event in Earth’s accretion (i.e. the giant impact and Moon formation). As argued by A. Morbidelli (2008, personal communication), it is very likely that the Moon represented the latest giant impact, while there could have been a number of preceding ‘giant impacts’. It would seem more logical to model the Earth’s accretion with a multi-stage model, but in our case, since we are only dealing with the last event, considering only a mean composition for the Earth and the impactor is sufficient. Our modelling enables us to consider only two parameters to represent a large range of possible models, including variable degrees and processes of equilibration between the proto-Earth and the proto-Moon, and the previous history of the impactor and the proto-Earth (including variable core sizes, Hf/W ratios or the age of proto-core formation). It considerably extends the number of possibilities considered in Touboul et al. (2007).

The results of the modelling are shown in figures 3 and 4. What this modelling shows is that there is a priori a large range of possibilities that would satisfy the observations. As shown in figure 3, an important feature is that the theoretical lower age for the earliest formation of the Moon is ca 37 Ma, as obtained from a two-stage model (Touboul et al. 2007), which assumes a chondritic initial $^{182}$W/$^{184}$W of the Moon. However, two lines of evidence suggest that this lower limit is highly improbable. First, it is unlikely that the Moon had a chondritic initial W-isotope composition because the Moon formed mostly from high Hf/W mantle material. Second, the difference in both Hf/W and initial $\epsilon^{182}$W between the lunar and terrestrial mantles must have been such that the $\epsilon^{182}$W evolved fortuitously to identical present-day values.

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Another important result of this modelling is that it shows that some degree of equilibration between the proto-lunar mantle and the proto-Earth is very likely. Depending on the Hf/W ratio in the impactor’s mantle, the accretion and differentiation history of the impactor, and the degree to which the impactor core material was remixed with the proto-lunar material, the $^{182}\text{W}$ of the proto-lunar mantle could in principle be either lower or higher than the $^{182}\text{W}$ of the proto-Earth. However, since the present-day Hf/W of the lunar mantle is higher than that of the Earth, the $^{182}\text{W}$ of the lunar mantle had to be lower than the $^{182}\text{W}$ of the Earth, if the giant impact occurred before 62 Myr (Figure 4). This is only possible if the impactor’s mantle was highly oxidized (and hence had a relatively low Hf/W) or if there has been more silicate–metal...
equilibration between the proto-lunar mantle and the impactor core than with the Earth’s mantle. Given that the numerical simulations of the giant impact suggest that most of the impactor’s core directly merges with the Earth’s core, this seems implausible. This scenario would also seem unlikely because the impactor’s mantle should have had a highly radiogenic $^{182}\text{W}$, given that the impactor most probably differentiated early. Consequently, it is probable that the $^{182}\text{W}$ similarity of the terrestrial and lunar mantles reflects equilibration in the aftermath of the giant impact, unless the Moon predominantly consists of terrestrial material.

(iii) **Isotope equilibration of the Earth and Moon after the giant impact**

The identical $^{182}\text{W}$-isotope composition of the lunar and terrestrial mantles suggests either that the Moon mainly derives from the material of the Earth’s mantle or that there has been isotope equilibration between the lunar magma disc and the Earth. The first of these options appears inconsistent with the results from numerical simulations of the giant impact, all of which indicate that more than approximately 80 per cent of the Moon is derived from the impactor. For future simulations, it will be essential to determine whether there are scenarios in which the Moon could be formed entirely from terrestrial material. In the case of an equilibration between the lunar magma disc and the Earth, one important question is whether the equilibration process that has been proposed for O isotopes by Pahlevan & Stevenson (2007) could also be valid for W. The condition for having isotope equilibration between the Earth and the lunar magma disc is to reach a temperature in the disc that will be high enough to allow volatilization of W. A key factor in this respect will be the oxygen fugacity because, as pointed out by Fegley & Palme (1985), WO$_3$ is far more volatile than the reduced species WO$_2$. If the temperature in the disc reaches 2000 K, then it is quite likely that W will be volatilized, and this could ultimately lead to a loss of W as has been argued for Mo by O’Neill (1991). Although O’Neill (1991) argues that the case for W loss by volatility is difficult to make based on a thermodynamic calculation, the temperature of the lunar magma disc might have been substantially higher than the temperature of 1400 K assumed in the calculation of O’Neill (1991). For a higher temperature, a significant fraction of W might have been lost as WO$_3$. Volatilization of W would also facilitate isotope equilibration of W isotopes between the Earth and the proto-Moon. An additional implication could be that the higher Hf/W of the Moon (26.4 ± 1.5 as opposed to 18 ± 1.6) is due to the volatile loss of W rather than core formation.

In brief, while one cannot fully ascertain that there was W-isotope equilibration between the Moon and the Earth at the time of the giant impact, the mechanism described by Pahlevan & Stevenson (2007) could be plausible. More detailed modelling will be needed to ascertain this process. It will also be essential to evaluate in future dynamical models if the Moon could have been formed largely from terrestrial mantle material.

(b) **Chronology of terrestrial accretion and astronomical implications**

An important implication of the new W-isotope data for the Moon is that the age of the Moon and the last episode of terrestrial accretion are younger than previously thought (ca 30 Myr). The new data of Touboul et al. (2007) indicate that the last stage of Earth’s accretion occurred later than 50 Ma. As the age of
the oldest lunar rocks is approximately 112 ± 40 Myr (Carlson & Lugmair 1988; Norman et al. 2003), one can estimate the age of the giant impact and the last stage of terrestrial accretion to be between 50 and 150 Myr. This estimate is compatible with earlier estimates for the age of the Earth based on U–Pb as well as on I–Xe systematics (Allègre et al. 1995; Ozima & Podosek 1999). It has been argued by Harper & Jacobsen (1996) that the elevated U/Pb ratio of the Earth’s mantle is due to the loss of volatile Pb rather than removal of Pb into the Earth’s core. This was based on the comparison of the Pb abundance of the Earth’s mantle with the depletion of lithophile elements with similar volatility. We have re-examined this question by considering trends in chondrites between Rb and Pb, which have a similar condensation temperature in a gas of solar composition (Allègre et al. 2001). Based on Pb and Rb abundances in chondrites, the estimated \( \frac{^{238}\text{U}}{^{204}\text{Pb}} \) ratio ranges between 1 and 1.4 in the bulk Earth (BE; figure 5). If no Pb is segregated into the Earth’s core, then the \( \frac{^{238}\text{U}}{^{204}\text{Pb}} \) ratio

Figure 4. (Caption opposite.)
Figure 5. Derivation of a bulk Earth $^{238}\text{U}/^{204}\text{Pb}$ ratio based on the Pb–Rb systematics for meteorite data. The meteorite data are from Wasson & Kallemeyn (1988). The data for the primitive mantle are from Hofmann (1988). The Pb/Rb versus Rb data for the meteorites define a negative trend reflecting the greater volatility of Rb relative to Pb ($y = -0.6965x + 2.2777$, $R^2 = 0.8331$). The BE composition is marked. If one assumes that Pb and Rb are not incorporated in the core, the BSE composition should lie on the trend (once Rb is corrected for the mass of the core). The observed value of the BSE plots well below this trend. $(\text{Pb/Rb})_{\text{predicted}}/(\text{Pb/Rb})_{\text{observed}}$ defines the degree of depletion of Pb due to the incorporation of Pb in the core. One can then infer the degree of depletion of Pb due to core formation.

Figure 4. (Opposite.) (a) A schematic showing the various pools of W involved in the make-up of the lunar and terrestrial mantles. The vertical double arrows show the $\Delta\varepsilon^{182}\text{W}$ before impact (long arrow) and after equilibration (short arrow). (b–e) Time evolution diagrams of $\varepsilon^{182}\text{W}$ for the BSM and BSE, illustrating re-equilibration scenarios as labelled in figure 3. The solid and long dashed curves show the evolution of the lunar and terrestrial mantles, respectively, calculated using their present-day $\varepsilon^{182}\text{W}$ combined with their Hf/W ratio. The solid vertical lines indicate the re-equilibration of W isotopes at different times of giant impact. Equation (2.1) together with input parameters is used to set a $\Delta\varepsilon^{182}\text{W}$ prior to equilibration. A complete model of the impactor and the proto-Earth and their equilibration would involve too many unknown parameters. The dotted curves give a hypothetical simplistic example of pre-giant-impact $\varepsilon^{182}\text{W}$ evolution for the impactor and proto-Earth reservoirs. The chondritic evolution (grey curves) is also shown for reference. (b) An unrealistic scenario with a giant impact at 32 Myr involving essentially materials from the impactor core ($\Delta\varepsilon^{182}\text{W}_{\text{initial}} = 2.4\varepsilon$ and $\Delta\varepsilon^{182}\text{W}_{\text{final}} = 0.9\varepsilon$). (c) The earliest possible scenario (37 Myr) that assumes an undifferentiated impactor (or a re-equilibration between its core and its mantle) and no re-equilibration between the Earth and lunar materials in the aftermath of the giant impact ($\Delta\varepsilon^{182}\text{W}_{\text{initial}} = 0.5\varepsilon$ and $\Delta\varepsilon^{182}\text{W}_{\text{final}} = 0.5\varepsilon$). (d) A scenario with a giant impact at 45 Myr that requires a terrestrial mantle initially slightly more radiogenic than the lunar mantle. This implies that the impactor mantle had a lower $\varepsilon^{182}\text{W}$ than the mantle of the proto-Earth prior to partial equilibration ($\Delta\varepsilon^{182}\text{W}_{\text{initial}} = 1\varepsilon$ and $\Delta\varepsilon^{182}\text{W}_{\text{final}} = 0.2\varepsilon$). (e) A scenario with a giant impact at 62 Myr that requires a complete W-isotope re-equilibration between the terrestrial mantle and proto-lunar materials to yield identical initial $\varepsilon^{182}\text{W}$ for the BSM and BSE ($\Delta\varepsilon^{182}\text{W}_{\text{initial}} = 2\varepsilon$ and $\Delta\varepsilon^{182}\text{W}_{\text{final}} = 0\varepsilon$). (f) A scenario with a giant impact at 88 Myr. The terrestrial mantle has to be slightly less radiogenic than the lunar mantle, and this scenario therefore requires a partial equilibration between the terrestrial mantle and a more radiogenic impactor mantle ($\Delta\varepsilon^{182}\text{W}_{\text{initial}} = -2\varepsilon$ and $\Delta\varepsilon^{182}\text{W}_{\text{final}} = -0.06\varepsilon$). CHUR, chondritic uniform reservoir.
of the BSE should be equal to the $^{238}\text{U}/^{204}\text{Pb}$ ratio of the BE. Yet, the BSE has a $^{238}\text{U}/^{204}\text{Pb} \sim 8–9$, which indicates that a significant part of the U–Pb fractionation is related not only to volatile depletion but also to core segregation. As argued by Jacobsen & Harper (1996), the continuous model age of core formation is almost identical to the two-stage model age for U–Pb, such that, if Pb was indeed partitioned into the Earth’s core, the U–Pb model age of ca 70–100 Ma should closely reflect the time of terrestrial core formation. This is consistent with the new constraints on the age of the giant impact based on W isotopes. Based on the observation that the Hf–W and U–Pb model ages for the formation of the Earth’s core are different, Wood & Halliday (2005) argued that Pb was removed into the Earth’s core by late sulphide segregation. The new W-isotope data for lunar rocks, however, reveal that there must have been ongoing core formation until after 50 Ma, such that there is no contradiction between the U–Pb and Hf–W systems.

The W results also have important astronomical implications for the dynamics of the Solar System. As shown by the models of O’Brien et al. (2006), an early termination of terrestrial accretion (less than 30 Myr) requires an eccentric orbit for Jupiter and Saturn (the EJS model). However, this is not the preferred model of O’Brien et al. (2006) because this initial condition leads to a rather complete clean-up of water-rich planetesimals in the region that is supposed to deliver water to the Earth at a late stage. Their preferred model is to have a circular orbit for Jupiter and Saturn (the CJS model). Remarkably, this model leads to a later termination for the Earth’s accretion (more than 70 Myr) and allows the delivery of water to the Earth. It is also worth pointing out that an initially circular orbit is easier to justify from a dynamical viewpoint (A. Morbidelli 2008, personal communication).

3. Constraints on the early differentiation of the Earth

The early differentiation of the Earth’s mantle is currently a highly debated topic and a recent overview is given in Bourdon & Caro (2007). There is now clear evidence for an early terrestrial mantle differentiation event (Caro et al. 2003) and the remaining questions are related to the exact timing of this differentiation as well as the composition of the BE. The recent analyses of $^{142}\text{Nd}$ in meteorites (Boyet & Carlson 2005) have revealed that (i) the composition of the Earth’s mantle is distinct from that of meteorites and (ii) the two main classes of chondrites (ordinary and carbonaceous chondrites) also have a distinct $^{142}\text{Nd}$-isotope composition despite having quite similar Sm/Nd ratios. This latter observation can apparently be rationalized by a difference in the abundance of r- and s-process nuclides (Andreasen & Sharma 2006; Carlson et al. 2007). If we assume that the BSE has a chondritic composition in Nd isotopes, then the positive $\epsilon^{142}\text{Nd}$ measured in terrestrial rocks relative to ordinary chondrites can be explained by early segregation of an enriched reservoir (with a low Sm/Nd), which would have led to a higher Sm/Nd in the remaining portion of the mantle. In order to produce a positive $^{142}\text{Nd}$ anomaly, this differentiation would have had to take place early enough in the Earth’s history. In the following, we examine the consistency of the W-isotope constraints on the timing of the formation of the Moon inferred by Touboul et al. (2007) and the hypothesis of an early mantle differentiation of Boyet & Carlson (2005).
(a) The age of early Earth differentiation

As argued by Boyet & Carlson (2005), if the Earth has a chondritic Nd-isotope composition, then the difference in $^{142}\text{Nd}$ between the Earth and chondrites yields an age for mantle differentiation of less than 30 Myr. The actual age of differentiation given by this system is likely to be significantly less than 30 Myr. Boyet & Carlson (2005) have used the $^{143}\text{Nd}$ in mid-ocean ridge basalts (MORB) to estimate the maximum degree of Sm–Nd fractionation in the early Earth. As shown in their fig. 3, if the early depleted mantle (EDM) had formed later than 30 Myr, the mean $^{143}\text{Nd}$ of the observable silicate Earth (=EDM) should be greater than approximately 10.5 units. We know that this is not the case and this estimate is therefore an upper limit, as clearly stated by Boyet & Carlson (2005). An additional reason why this is an upper limit is that the effects of later continental crust (CC) extraction are ignored (as is the ocean island basalts (OIB) reservoir, which also has a lower Sm/Nd on average than the MORB reservoir). Here, we attempt to refine this estimate by taking into account the fact that the CC has to be included for estimating the composition of the EDM. A conceptual model for this calculation is illustrated in figure 6a. If one takes into account the Nd in the CC, one can calculate a new age for mantle depletion based on the following equations (CHUR, chondritic uniform reservoir):

\[
\frac{\text{today}}{\text{EDM}} \frac{^{143}\text{Nd}}{^{144}\text{Nd}} = \left( \left( \frac{^{143}\text{Nd}}{^{144}\text{Nd}} \right)_{\text{CC}}^	ext{today} + \left( \frac{^{147}\text{Sm}}{^{144}\text{Nd}} \right)_{\text{EDM}}^	ext{today} \right) e^{\lambda_{147} T_{\text{CC}}} - 1,
\]

\[
\frac{\text{today}}{\text{CC}} \frac{^{143}\text{Nd}}{^{144}\text{Nd}} = \left( \left( \frac{^{143}\text{Nd}}{^{144}\text{Nd}} \right)_{\text{EDM}}^	ext{today} + \left( \frac{^{147}\text{Sm}}{^{144}\text{Nd}} \right)_{\text{CC}}^	ext{today} \right) e^{\lambda_{147} T_{\text{CC}}} - 1,
\]

\[
\frac{\text{today}}{\text{DM}} \frac{^{143}\text{Nd}}{^{144}\text{Nd}} = \left( \left( \frac{^{143}\text{Nd}}{^{144}\text{Nd}} \right)_{\text{EDM}}^	ext{today} + \left( \frac{^{147}\text{Sm}}{^{144}\text{Nd}} \right)_{\text{DM}}^	ext{today} \right) e^{\lambda_{147} T_{\text{CC}}} - 1,
\]

\[
\frac{^{143}\text{Nd}}{^{144}\text{Nd}}_{\text{EDM}} = \left( \frac{^{143}\text{Nd}}{^{144}\text{Nd}} \right)_{\text{EDM}}^	ext{today} + \left( \frac{^{147}\text{Sm}}{^{144}\text{Nd}} \right)_{\text{EDM}}^	ext{today} \left( e^{\lambda_{147} T_{\text{EDM}}} - e^{\lambda_{147} T_{\text{CC}}} \right),
\]

\[
\frac{^{142}\text{Nd}}{^{144}\text{Nd}}_{\text{EDM}} = \left( \frac{^{142}\text{Nd}}{^{144}\text{Nd}} \right)_{\text{EDM}}^0 + \left( \frac{^{144}\text{Sm}}{^{144}\text{Nd}} \right)_{\text{EDM}}^0 \left( e^{-\lambda_{146} (T_0 - T_{\text{EDM}})} \right),
\]

\[
\frac{^{142}\text{Nd}}{^{144}\text{Nd}}_{\text{CHUR}} = \left( \frac{^{142}\text{Nd}}{^{144}\text{Nd}} \right)_{\text{CHUR}}^0 + \left( \frac{^{144}\text{Sm}}{^{144}\text{Nd}} \right)_{\text{CHUR}}^0 \left( 1 - e^{-\lambda_{146} (T_0 - T_{\text{EDM}})} \right),
\]

\[
\frac{^{147}\text{Sm}}{^{144}\text{Nd}}_{\text{EDM}} = \alpha_{\text{DM}} \left( \frac{^{147}\text{Sm}}{^{144}\text{Nd}} \right)_{\text{DM}} + \left( 1 - \alpha_{\text{DM}} \right) \left( \frac{^{147}\text{Sm}}{^{144}\text{Nd}} \right)_{\text{CC}},
\]

\[
\frac{^{143}\text{Nd}}{^{144}\text{Nd}}_{\text{EDM}} = \alpha_{\text{DM}} \left( \frac{^{143}\text{Nd}}{^{144}\text{Nd}} \right)_{\text{DM}} + \left( 1 - \alpha_{\text{DM}} \right) \left( \frac{^{143}\text{Nd}}{^{144}\text{Nd}} \right)_{\text{CC}}.
\]
The values of the input parameters are given in the legend of figure 6. We assume a chondritic evolution prior to the formation of the EDM reservoir. These eight equations are nonlinear and can be solved iteratively for the eight unknowns assuming starting input parameters $T_{EDM}$ and $\alpha_{DM}$, the fraction of Nd in the depleted mantle. This latter parameter is used to calculate the concentration of Nd in the depleted mantle. The main results of this calculation are that, given that the concentration of Nd in the depleted mantle should be approximately 1 ppm at most (Hofmann 1988; Salters & Stracke 2004; Workman & Hart 2005), the age of early mantle depletion has to be less than 10 Ma after the formation of calcium–aluminium-rich inclusions (CAI; figure 6b). Essentially, this means that the formation of a hidden reservoir complementary to the EDM has to occur before the Earth is completely accreted, which is unrealistic. This conclusion is very similar to what is discussed below in §3b based on the observations of $^{142}$Nd.

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in Martian meteorites. A major conclusion is that it is more plausible to have the loss of enriched material prior to complete accretion rather than due to magma ocean differentiation as argued by Boyet & Carlson (2005). Having an early mantle depletion at 20–30 Myr would require a prohibitively high Nd concentration in the depleted mantle.

Even if we ignore this mass-balance calculation and take the estimate of Boyet & Carlson (2005) for the differentiation of the early Earth, it is quite clear that this early event (less than 30 Ma) should have taken place prior to the Moon-forming giant impact (more than 50 Myr, see §2b). If that earlier event itself resulted from another giant impact (as is most likely in order to produce large-scale mantle differentiation), then the Earth should have remained quite hot (i.e. close to its solidus temperature) just prior to the Moon-forming impact (Tonks & Melosh 1993). In this case, the calculations of Tonks & Melosh (1993) show quite clearly that the entire Earth must have been molten during the second giant impact. It is likely that the giant impact would have violently disturbed an early segregation. First, much of the impactor’s core most probably merged with the terrestrial core either as a single event or by Rayleigh–Taylor instabilities, and this must have seriously disrupted any dense layer at the base of the mantle. Second, the heat provided by the segregation of the impactor’s core is likely to strengthen convection- and buoyancy-driven instabilities. All this would appear inconsistent with the idea of the preservation of an earlier enriched reservoir stored at the bottom of the lower mantle, unless the viscosity and density contrasts were large enough to avoid remixing of the enriched dense layer. Obviously, this point will need to be explored more thoroughly with dynamical models. Yet the new timing for the formation of the Moon does not lend strong support for the very early storage of an enriched reservoir. In addition, our mass-balance calculation on the Sm–Nd budget casts some serious doubts on the plausibility of a hidden reservoir. In the following we examine this point further and propose an alternative model.

(b) The Nd-isotope composition of the terrestrial planets

In this section, we examine whether one can assume that the bulk composition of planets in Sm–Nd is chondritic on the basis of the new Nd-isotope data for Martian meteorites (Caro et al. 2008). The new data of Caro et al. (2008) shown in an $^{142}\text{Nd}$ versus $^{147}\text{Sm}/^{144}\text{Nd}$ diagram in figure 7 clearly show that the Martian bulk composition is probably not chondritic. It must be noted that enriched shergottites have an $^{142}\text{Nd}$ equal to that of ordinary chondrites but they have lower $^{147}\text{Sm}/^{144}\text{Nd}$ ratios and their $^{147}\text{Nd}$ and $^{176}\text{Hf}$ are both negative (−7 and −13, respectively), which is a typical crustal signature. On this basis, it would seem that $^{142}\text{Nd} \sim −20$ ppm cannot be representative of a bulk composition for Mars.

If the intersection of the mantle compositions of Mars, the Moon and the Earth represents the mean composition of terrestrial planets, one would infer that the composition of these planets is superchondritic in Sm/Nd, unless Mars has also experienced an early episode of mantle differentiation leading to a hidden reservoir with a composition identical to that of the Earth. Caro et al. (2008) have argued that this is an unlikely possibility. To make this case more convincing, we explore more systematically scenarios whereby the common point for terrestrial planets shown in figure 7 can be explained by an early
differentiation event as argued for the Earth by Boyet & Carlson (2005). Unlike Debaille et al. (2007), we assume that the bulk composition of Mars lies on the planetary isochron, implying that there is no hidden reservoir on Mars.

Figure 7. (Caption opposite.)
One question is whether the intersection with the Earth is likely to represent the bulk composition of Mars. If the bulk composition of Mars lies above the intersection with the terrestrial composition (figure 7a), then the chondrite–Mars Sm–Nd fractionation event would have to post-date the accretion of Mars dated at 0–10 Myr, as shown by Kleine et al. (2002) and Nimmo & Kleine (2007). This putative event would also have to pre-date the large-scale mantle differentiation event dated at 40 Myr (Caro et al. 2008), which seems unreasonable. If the bulk composition of Mars is below the terrestrial composition, then the fractionation in Sm–Nd must be older than the Solar System, which is impossible. Thus, in both cases, it seems likely that the intersection with the terrestrial composition represents the bulk Martian composition (figure 7a).

Thus, one may conclude that the observations of Caro et al. (2008) imply that the Sm–Nd fractionation observed in Martian meteorites must have taken place either before or immediately after accretion. However, if one calls upon the segregation of a hidden reservoir, the Sm–Nd fractionation must have been strikingly identical to that observed in the Earth. Although there has not been a physically plausible scenario proposed by Boyet & Carlson (2005) to explain their hidden reservoir, it is not clear why the hidden reservoir on Earth would lead to identical Sm/Nd than is observed. This would also be inconsistent with the data of Rankenburg et al. (2006).

A different line of argument has been used by Debaille et al. (2007), who have argued that the linear array formed by shergottites in an $\epsilon^{142}$Nd versus $\epsilon^{143}$Nd diagram cannot be explained by a superchondritic mantle identical to that of the Earth. For their argument, they have assumed that the $\epsilon^{143}$Nd value for bulk Mars would be the same as that of the depleted terrestrial mantle (i.e. $\epsilon^{143}$Nd $\sim$ 10.7, which is the value used by Boyet & Carlson (2005)). However, the choice of this particular composition for Earth (or Mars) is not justified since we know that both magma ocean crystallization and later crustal extraction have modified the $^{143}$Nd-isotope composition while $^{142}$Nd is only affected by the earlier event. As a consequence, the argument of Debaille et al. (2007) is not conclusive and an alternative solution is discussed below.

The $^{146}$Sm–$^{142}$Nd systematics for lunar rocks are also difficult to reconcile with a hidden reservoir model that would be consistent with both Nd and W observations. A reasonable scenario would be that the Moon formed at 50 Myr
(the earliest age possible based on the W isotopes) with a bulk chondritic composition and that the Nd isotopes would equilibrate during the proto-Moon stage as has been argued for O, Si and W isotopes (Georg et al. 2007; Pahlevan & Stevenson 2007; Touboul et al. 2007). As shown in figure 7b, the intersection of the lunar mantle isochron with the terrestrial Nd-isotope composition would fall at a higher Sm/Nd than is observed. Thus, this scenario cannot explain the observations. The only realistic explanation is then to have the Moon formed entirely from the Earth’s mantle, but this scenario would be in conflict with inferences about the iron abundance in the Moon (O’Neill 1991). It would also contradict dynamical models that predict that the Moon is formed mostly from the impactor (Canup & Asphaug 2001), although more recent models have shown that the mass fraction of the proto-Earth in the Moon could be as large as 0.5 (Canup 2004).

One may note that there is a substantial uncertainty in the intersection of Mars, the Moon and the Earth in the Sm–Nd diagram. In principle, the $^{147}\text{Sm}/^{144}\text{Nd}$ value required to produce a 20 ppm offset in $^{142}\text{Nd}$ between chondrites and planets is 0.21, which is slightly higher than the value given by the intersection. This discrepancy could come from several sources of uncertainties. First, one may note that the $^{142}\text{Nd}$ of carbonaceous chondrites is obtained based on a correction using the $^{148}\text{Nd}$ isotope (Carlson et al. 2007), which introduces an additional source of uncertainty. If one considers only the difference between planets and ordinary chondrites, then the difference in $^{142}\text{Nd}$ becomes 17 ppm. Another source of uncertainty is the initial $^{146}\text{Sm}/^{144}\text{Sm}$ ratio ($=0.008\pm1$) used to determine the slope of the 4.568 Ga isochron in the $^{146}\text{Sm}$–$^{142}\text{Nd}$ diagram. This slope is slightly steeper if $^{146}\text{Sm}/^{144}\text{Sm}$ is increased to 0.009, and $^{147}\text{Sm}/^{144}\text{Nd}$ needs to be increased to 0.207 to match a difference of approximately 17 ppm between ordinary chondrites and the Earth.

An important implication is that the Sm–Nd fractionation observed in Mars, the Earth and the Moon would pre-date the accretion and the high Sm/Nd would characterize the chemical composition of the inner Solar System. As there is no direct meteorite sample from this inner Solar System region, it is difficult to test that inference. Because Sm and Nd are both lithophile and refractory, the expectation is that the Sm/Nd ratio of the bulk planets should be identical to that of chondrites. There are reports of rare-earth element (REE) fractionation under specific nebular conditions. One could consider the following two hypotheses: (i) given that chondrules have systematically higher Sm/Nd than bulk chondrites (Krestina et al. 1999; Amelin & Rotenberg 2004), a preferential accumulation of chondrules in the inner Solar System region should, in principle, lead to a higher Sm/Nd in terrestrial planets, and (ii) as the Earth, Mars and, possibly, the Moon-forming impactor formed from precursory planetesimals, the impact history of these differentiated objects could have led to the removal of a crust with a low Sm/Nd.

The observation that each chondrite group has a characteristic size distribution of chondrules has led to the hypothesis that chondrules have been sorted by a dynamical process (Kuebler et al. 1999). Furthermore, modelling of turbulence in the solar nebula has also shown that the abundances of chondrules themselves could be enhanced by the process of turbulent concentration (Cuzzi et al. 2001). More importantly, the prediction is that chondrules should be
concentrated in the terrestrial planet regions (Cuzzi et al. 1998), while porous aggregates would be concentrated in the outer planet regions. In essence, this is exactly what our inference would be, based on the $^{146}\text{Sm}–^{142}\text{Nd}$ systematics, suggesting that the offset in the Sm/Nd ratio in terrestrial planets could be due to an accumulation of chondrules with a higher Sm/Nd. In practical terms, the calculated concentration of chondrules can increase by a factor of 10 with a high probability (Cuzzi et al. 2001). Using the histogram shown in figure 8, this is enough to shift the $^{147}\text{Sm}/^{144}\text{Nd}$ from 0.1967 to 0.206.

One important question that arises is what process would have caused the Sm/Nd to be greater than chondritic in the first place. This represents a shift of only 5 per cent and several processes can a priori be proposed. From a cosmochemical viewpoint, the relative volatility of rare earths has been explored in variable conditions, including oxidizing and reducing environments (Boynton 1975; Lodders & Fegley 1993; Pack et al. 2004). In general, there can be an enhancement of Sm or Nd volatility in oxidizing and reducing conditions, but the expectation is that other REE will then be far more fractionated (i.e. Eu, Ce or Yb), and this extreme fractionation is not observed in terrestrial mantle peridotites (Jagoutz et al. 1979). Thus, fractionation due to volatility, although it has been observed in CAIs and special chondrules (Pack et al. 2004), does not seem to be the process to explain the difference between bulk chondrites and chondrules.

There are only a few high-precision REE datasets on chondrules (Krestina et al. 1999; Amelin & Rotenberg 2004) and these studies do not reveal the origin of the fractionation. High Sm/Nd ratios are observed in ordinary chondrites that also have a low Sm/Nd phosphate phase. If one assumes that the REE were

![Figure 8. A histogram showing the $^{147}\text{Sm}/^{144}\text{Nd}$ for bulk chondrites (narrow white bars) and chondrules (black bars) (adapted from Caro et al. 2008). The high Sm/Nd found in chondrules could be a result of REE redistribution during the formation of low Sm/Nd phosphate, but the Sm/Nd does not correlate with degree of metamorphism. Further investigations of the Sm–Nd systematics in chondrites are required to prove that the concentration of chondrules can effectively explain the high Sm/Nd in the inner planets. The mean chondrule $^{147}\text{Sm}/^{144}\text{Nd}$ is 0.21, while the bulk chondrite $^{147}\text{Sm}/^{144}\text{Nd}$ is 0.1967.](http://rsta.royalsocietypublishing.org/)

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redistributed during the ‘late’ formation of the phosphate on the parent body, then the high Sm/Nd ratio in chondrules would not be a primary feature and thus would not be relevant to explaining the high Sm/Nd of planets. However, it is worth noting that chondrules in carbonaceous chondrites that do not have a phosphate phase also have a high Sm/Nd ratio. Furthermore, as shown in figure 8, the chondrules of one of the least metamorphosed ordinary chondrites (Krymka, LL 3.1) also have a high Sm/Nd ratio, such that this observation is not linked to thermal processing and phosphorus redistribution.

However, it is worth pointing out that recent studies of type 1 chondrules (Libourel & Krot 2007) that include relicts of planetesimal mantles could give clues on the observed Sm–Nd fractionation. Although there are very few of these chondrules, perhaps because many chondrules have been entirely molten, these observations provide important insights about the origin of chondrules. If the chondrules were part of differentiated parent bodies, then it is likely that they experienced melt extraction leading to depletion in Nd relative to Sm. One could speculate that these observations could explain the higher Sm/Nd in chondrules, but further documentation will need to be provided in the future.

Clearly, there are potential difficulties in explaining the high Sm/Nd in Earth, Mars and the Moon by accumulation of chondrules, and further investigations are needed to test this hypothesis.

The second hypothesis involving impact erosion of a differentiated crust was also proposed by Caro et al. (2008) and is discussed at greater length by O’Neill & Palme (2008). Numerous models (e.g. Benz & Asphaug 1999; Agnor & Asphaug 2004; Asphaug et al. 2006) have reported that collisions are commonly non-accretionary and lead to erosion of the outer layers of a growing planet. The inference that one might make here is that this process should have taken place at a very early stage (within the first few million years after the beginning of the Solar System) and have led to the removal of an early formed crust on the planetesimals that ultimately formed the Earth, Mars and the Moon. One additional constraint brought by the $^{142}$Nd data is that the age and composition of the removed crust would have been similar to the precursory material of the Earth, Mars and the Moon. This requirement would seem easier to fulfil at an early stage (when bodies have not grown significantly yet) rather than at a late stage once the planets are formed (see above). Our present state of knowledge does not allow us to come to any conclusion about the actual process that would have led to a chemical depletion in the bulk composition of the Earth, Mars and the Moon.

4. Summary

The new $^{182}$W data for the Moon provide new constraints on the timing of the Moon-forming giant impact (later than 50 Myr). Based on these data, one can also estimate that the solidification of the magma ocean took place later than 60 Myr. This solves the apparent conflict with $^{142}$Nd age constraints for the differentiation of the lunar magma ocean. However, because the $^{146}$Sm–$^{142}$Nd system is sensitive to fractionation until 4.1–4.2 Gyr, the relatively young age given by this system (200–250 Myr) for magma ocean crystallization probably reflects later events, including melting, mixing, convection and contamination. A simple model including these processes shows that this hypothesis is perfectly plausible.
A wide exploration of models for the giant impact and the subsequent W-isotope evolution suggests that there must have been some degree of W-isotope equilibration between the Earth and the Moon following the impact. Most probably, part of the core of the impactor partially re-equilibrated with the Earth’s and the Moon’s mantle.

We have shown, based on Sm–Nd mass balance, that the hidden reservoir scenario proposed by Boyet & Carlson (2005) based on the difference in $^{142}$Nd abundance between the Earth and chondrites needs to be questioned. This Sm/Nd fractionation associated with the formation of a hidden reservoir has to take place prior to the accretion of the Earth to match the Nd-isotope observations. Based on the new data for Martian meteorites, we further hypothesize that the Sm/Nd composition of terrestrial planets could be higher than that of chondrites. This process could result from an enrichment in a chondrule component by sorting or, alternatively, by impact erosion of the outer layers of planetesimals forming the Earth and Mars. Deciding between these alternatives will require deeper investigations of both processes.

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