

Geochemical Constraints on the Origin of the Earth and Moon

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As originally conceived, the giant impact theory for the origin of the Moon made several predictions: (1) the Moon's complement of metal resides in Earth's core; (2) similarities in some siderophile-element abundances between the Earth and Moon exist because the Moon inherited them from Earth; (3) low abundances of volatiles in the Moon are the result of thermal processing of terrestrial materials; and (4) Earth passed through a magma ocean stage as a result of the giant impact. The geochemical and geophysical evidence now suggests that (1) the Moon has a small core, so prior removal of siderophiles into Earth's core is not required; (2) the similarity in W abundances between the Earth and Moon is coincidental; (3) alkalis in the Moon were not simply inherited from Earth; and (4) there is evidence from terrestrial mantle lherzolites that Earth may never have had a magma ocean. These observations are sufficient to cast doubt on the giant impact hypothesis.

1. INTRODUCTION

A possible relationship between Earth and the Moon, based on the chemical compositions of both planets, was extensively discussed at the Kona conference on the Origin of the Moon in 1984. The chemical results, as presented in several articles in *Origin of the Moon* (1986, Lunar and Planetary Institute, Houston) may be broadly summarized as follows:

1. The lower density of the Moon requires a lower bulk Fe content than that of Earth. However, the higher FeO content of lunar basalts implies that the silicate portion of the Moon has a higher FeO content than that of Earth.

2. Estimates for the refractory element content of the Moon range from 2–4× CI, compared to 2.75× CI for Earth's mantle. Relative abundances of refractory elements in both bodies appear to be chondritic. In addition, the abundances of Cr and V are similar in the mantles of The Earth and Moon, but different from CI chondrites.

3. The abundances of siderophile elements in the lunar mantle decrease with increasing siderophilicity, i.e., in the sequence Fe, W, Co, Ga, P, Ni, Mo, Re. This is strong circumstantial evidence for the presence of a small lunar core.

4. Moderately volatile elements are depleted in the Moon relative to Earth. Sodium, K, and Zn are all depleted in the Moon. A notable exception is Mn, which is higher in the Moon than in the Earth's mantle.

5. Highly volatile elements (e.g., Cd, Tl) are depleted in the Moon by about a factor of 50, relative to the Earth, with the exception of Cs. The reason for the comparatively high Cs abundance in the Moon is unknown.

6. The most abundant element in the Earth and Moon

is O. And, within analytical error, the Earth and Moon are identical in terms of their O isotopic compositions.

Taken as a group, these chemical characteristics were believed most compatible with a giant impact origin for the Moon, whereby a Mars-sized planetesimal struck Earth and ejected protolunar material into Earth orbit. Although some specific points were difficult to reconcile with this collisional ejection hypothesis, this model was favored by most researchers. In part this may have been because the model offered a high degree of flexibility in explaining the composition of the Moon, as reflected in the wide range of estimates for the contribution of the projectile to the bulk Moon. For example, *Ringwood* (1990) believed that terrestrial material dominated the Moon and estimated, on the basis of siderophile elements, that there had been less than ~1% chondritic material added to the Moon. *MacFarlane* (1989), in comparing the compositions of the Earth and Moon to various chondrite groups, concluded that Earth probably contributed no more than 50% of the protolunar material. *Newsom and Taylor* (1989) believed that the Moon's composition was dominated by the impactor. And *O'Neill* (1991) calculated that Earth contributed 70% of the present lunar mass. And, of course, this multiplicity of views concerning the relative contributions of the target and impactor complicates any discussion of lunar origin.

Since the Kona conference in 1984, several new analytical techniques have been developed or refined, and application of these new methods are providing new insights into the question of the lunar origin:

1. The precise determination of the Cr isotopic composition of lunar samples confirmed the isotopic similarity of lunar and terrestrial material, as shown by *Lugmair and*

Shukolyukov (1998), who showed that the $^{53}\text{Cr}/^{52}\text{Cr}$ is variable for inner solar system materials, but identical in the Earth and Moon.

2. The analysis of the W isotopic composition of meteorites, lunar, and terrestrial rocks must be considered as the most important geochemical parameter applied to problems of lunar origin that has been made since the Kona conference. The implications of these new $^{182}\text{W}/^{184}\text{W}$ data (*Lee et al.*, 1997) will be discussed in a separate section.

3. The isotopic composition of K has been precisely measured, and it was found that terrestrial and lunar samples are indistinguishable in their $^{41}\text{K}/^{39}\text{K}$ ratios (*Humayun and Clayton*, 1995).

4. In the last 10 years it has become increasingly clear that metal/silicate partition coefficients may strongly depend on pressure and temperature. It has been suggested that the high concentrations of Ni and Co in the mantle of Earth are the result of silicate equilibration with metal at the base of a terrestrial magma ocean, created by the impact of a Mars-sized object during the formation of the Moon (*Li and Agee*, 1996; *Righter et al.*, 1997).

5. The Os isotopic composition of xenoliths of the terrestrial mantle have been measured. Osmium isotopes in mantle samples approach those of chondrites as the bulk compositions of the xenoliths themselves approach estimates of the bulk silicate Earth (*Meisel et al.*, 1996). Therefore, there is a correlation between siderophile elements (Re/Os) and lithophile elements such as Al and REE. The most fertile xenoliths have Os isotopic compositions that are closest to chondritic.

Here we will examine both old and new geochemical observations in order to evaluate the giant impact hypothesis. First we will review the most relevant older data, and then we will discuss several different geochemical topics in more depth.

2. BULK CHEMISTRY AND PHYSICAL PROPERTIES

It is generally agreed that the Moon and Earth differ in substantive ways. The Moon is more reduced than Earth, but the Moon has a lower Mg# [molar $\text{Mg}/(\text{Mg} + \text{Fe}) \times 100$] and higher FeO than Earth's mantle. The Moon's uncompressed density is lower, and its moment of inertia is significantly higher. We shall discuss these noncontroversial aspects of the Earth and Moon before turning to those properties that are less agreed upon. Often we will compare chemical abundances in the Earth and Moon to those of chondrites, and in these discussions we will use the data compiled by *Newsom* (1995). Unless stated otherwise, we will not normalize elemental abundances to elements such as Mg or Si. Thus, normalizations to CI chondrites have not been corrected for volatile loss.

We will often refer to the bulk silicate Earth (BSE), distinguishing between the bulk Earth (including the core) and its silicate portion. Strictly speaking, this composition is well

known only for Earth's upper mantle. However, we know of no substantive evidence that the lower mantle has a different composition than the upper mantle. And current models of mantle convection favor good communication between the upper and lower mantle (e.g., *Davies*, 1988). Hereafter, we assume that the bulk compositions of the upper and lower mantle are indistinguishable.

2.1. Oxygen Isotopes

The O isotopic compositions of different meteorite groups are usually different. Thus, O isotopes have become the "industry standard" of the meteoritics community for determining the provenance of rocky materials within the solar system (e.g., *Clayton*, 1993), even if the geographic location of that provenance is unknown. The Earth and Moon have essentially identical O isotopic signatures, indicating that they were derived from isotopically similar materials. We note, however, that enstatite chondrites, the most reduced solar system materials, and CI chondrites, the most oxidized meteorites, fall on or near the same fractionation line as the Earth and Moon (e.g., *Clayton*, 1993). Thus, it is unclear whether the O isotopic signature of the Earth and Moon is generally indicative of all rocky materials within 1 AU of the Sun or whether it is a local property of the Earth-Moon system. For this reason alone, it would be important to know the O isotopic compositions of Venus and Mercury. Unfortunately, the prospects for such measurements in the foreseeable future are poor.

2.2. Redox State

2.2.1. Earth. The overall redox state of the Earth is not known precisely, but this particular aspect of the Earth is not a very useful parameter. This is because different portions of the Earth record different f_{O_2} , and thus appear to be in disequilibrium. Earth's surface and atmosphere are highly oxidizing, as indicated by the presence of water and oceans. Earth's mantle is more reduced, having an f_{O_2} near that of the QFM (quartz-fayalite-magnetite) oxygen buffer (e.g., *Ringwood*, 1979). And *Christie et al.* (1986) have argued that, in detail, Earth's upper mantle is somewhat more reducing than QFM and that QFM-1 (one \log_{10} unit below QFM) is a better approximation. In detail, a fairly wide variation in inferred f_{O_2} can be found within the mantle-derived sample suite, but we prefer to use estimates based on basalts, such as mid-ocean ridge basalts (MORB), that sample large mantle volumes.

The Earth's core, of course, must be considerably below QFM and even below IW (iron-wüstite). In general Fe metal is only stable at conditions below IW, since the activity of FeO in bodies of chondritic composition is below unity. However, this observation is tempered by the presence of Ni, which can persist in the metallic form to redox states near or exceeding QFM. Thus, at redox states of IW-1 or below, Fe-rich metal is possible, but at redox states above IW-1 metal becomes increasingly Ni-rich.

It is of interest, however, to calculate what f_{O_2} would be necessary to account for the current depletion in FeO from chondritic, assuming that the mantle and core were once in equilibrium. Assuming ideality and metal/silicate equilibration at low pressure, this calculation implies that the core formed from the mantle at around IW-2.5 (Jones and Drake, 1986; Holzheid and Palme, 1996). This is quite reducing and, if the core and current mantle were never in equilibrium, could be a conservative upper limit to the true f_{O_2} that pertained during core formation. This latter possibility has been explicitly incorporated into the bulk compositional models of Wänke (1981), who postulated that, during core formation, there was very little FeO in the silicate portion of Earth. In the Wänke model, nearly all FeO is accreted and added to the mantle after the main epoch of core formation is over. Evidence for this scenario comes from the high abundances of siderophile elements such as W and Mo in Earth's mantle (Wänke, 1981). We will discuss this model and alternative models of core formation in more detail below.

To summarize, the Earth's different physical domains have different redox states and are apparently in disequilibrium. These domains appear to have maintained approximately the same redox conditions over most of Earth's history. The oldest sediments (about 3.9 Ga) yield evidence of liquid water on Earth's surface, and the oldest basalts imply a redox state near QFM [3.8 Ga (Delano, 1993); see also Canil (1999)]. These Archean basalts were already low in FeO compared to lunar mare basalts, implying that core formation was essentially complete by 4 Ga. Some have argued (in order to explain "future" Pb isotopic compositions) that core formation has continued over Earth's entire history (e.g., Oversby and Ringwood, 1971). However, there is no actual evidence that core formation has occurred since ~4 Ga (Newsom et al., 1986). Conversely, it has been suggested that some plume basalts may contain trace amounts of core materials entrained at the core-mantle boundary (Walker et al., 1995). Still, the dominant situation appears to be disequilibrium between reservoirs that have communicated poorly since the time of Earth's formation.

2.2.2. Moon. Comparatively, the redox state of the Moon appears quite simple. Lunar basalts are very nearly saturated in Fe metal. Iron-nickel metal is a common late-stage crystallization product of lunar basalts, and lunar basaltic liquids do not gain or lose significant FeO when they are contained in high-purity Fe metal capsules (Walker et al., 1977). In addition, lunar basalts and eucrites have similar FeO concentrations. Taken together, these observations imply that the lunar f_{O_2} is similar to the eucrite parent body, which has been experimentally determined to be IW-1 (Stolper, 1977).

Another point of view can also be taken. Since the lunar mantle contains nearly 500 ppm Ni and correspondingly high Co contents, Seifert et al. (1988) concluded that the lunar core is Ni-rich, indicating a slightly higher f_{O_2} than defined by pure FeO-Fe equilibrium (~IW + 0.5).

These two sets of observations and calculations probably

bracket the true redox state of the Moon. And we, the authors, do not agree as to which is most likely correct. However, the differences in these estimates are not large, making it likely that the redox state of the Moon is between IW and IW-1.

2.3. Mg#, CaO/Al₂O₃, and Total Iron

2.3.1. Earth. The Mg# of the Earth is well constrained from mantle xenoliths to be about 89 (O'Neill and Palme, 1998; BVSP, 1981; Taylor, 1982). And the FeO content used above to calculate the f_{O_2} of core-mantle equilibration (~IW-2.5) was ~8 wt%. If the Earth's core contains ~85 wt% Fe [assuming ~10 wt% of a light element (e.g., Jeanloz, 1990) and ~5 wt% Ni], Earth's total Fe is calculated to be ~31 wt%. This translates into a chondrite-normalized concentration of ~1.7× CI, compared to 1.55× CI for Mg, assuming a Mg-free core. This calculation assumes that the bulk Earth should have an approximately chondritic Fe/Ni ratio.

It is assumed that the Ca/Al ratio of Earth is chondritic (e.g., Palme and Nickel, 1985). However, Ca/Al in primitive spinel lherzolites from the mantle are often superchondritic (Palme and Nickel, 1985). This is probably because these rocks have lost small amounts of low-degree (<5%) partial melts, which typically have low Ca/Al ratios (Baker et al., 1995). The best estimates for CaO and Al₂O₃ in the BSE are 3.8 and 4.7 wt% respectively (O'Neill and Palme, 1998).

2.3.2. Moon. The Mg# of the Moon is not well known. Because we do not have lunar materials that escaped the differentiation of the lunar magma ocean, the composition of the bulk Moon must be reconstructed, and this is difficult. Still, most models of lunar composition converge on an Mg# of 80–85 (e.g., Taylor, 1982; Ringwood et al., 1987; Jones and Delano, 1989). This translates into a bulk lunar FeO content of about 13 wt%, significantly higher than that of Earth's mantle. Jones and Hood (1990) and Mueller et al. (1988) found that this amount of FeO was necessary to simultaneously reconcile the lunar geophysical properties of density and moment of inertia. In models that postulated that the Moon had the same chemical composition as the Earth's upper mantle, the Moon was not dense enough. Another means of increasing the Moon's density (i.e., increasing the size of a hypothetical iron metal core; see below) causes the moment of inertia to be too small. Consequently, it has become generally accepted that the Moon has 1.5–1.6× more FeO than Earth's mantle. However, the FeO/MgO of the lunar mantle is unlikely to be uniform. The existence of the lunar Mg suite, which contains lithologies with Mg# ≥ 90, argues that there must be a lunar reservoir having an Mg# similar to that of Earth (e.g., Warren, 1986). Therefore, it seems to us that deriving total FeO (and Mg#) from bulk geophysical properties is the least model-dependent approach.

Similarly, the total Fe content of the Moon is also uncertain. However, a maximum Fe content may be calculated by taking 13 wt% FeO in the silicate Moon and assuming

a pure Fe core that constitutes 5 wt% of the Moon. This translates into a total Fe content of 14.6 wt% or $\sim 0.8 \times \text{CI}$, at least a factor of 2 lower than that for the Earth. If the lunar core is Ni rich ($\sim 45\%$ Ni; *Seifert et al.*, 1988), then cores larger than 1 wt% result in superchondritic Ni/Fe ratios in the Moon. Regardless, it seems inescapable that the bulk Fe (bulk FeO) contents of the Earth and Moon are different.

The Ca/Al ratio of the Moon is also presumed to be chondritic. *Jones and Delano* (1989) calculated a wide variety of models for the Moon's bulk composition and found that bulk CaO and Al_2O_3 concentrations were fairly insensitive to the exact model. Consequently, we adopt the *Jones and Delano* (1989) values of lunar CaO and Al_2O_3 as 3.0 and 3.7 wt% respectively.

2.4. Geophysical Properties

2.4.1. Earth. The moment of inertia of the Earth about its polar axis, $C/(\text{MR}^2)$, is 0.3305, which should be compared to the value of 0.4 expected for a homogeneous sphere (*Verhoogen et al.*, 1970). This smaller value is because of the mass concentrated in the Earth's core. Similarly, the calculated average density of the Earth (5.517 g/cm^3), which is too high for rock alone, indicates that Earth must contain a dense metallic component. These inferences are confirmed by seismographic observations that also indicate that there is a large metallic core at the center of Earth (e.g., *Macelwane*, 1951).

2.4.2. Moon. In contrast to the Earth, the Moon's moment of inertia of 0.3905 is very close to what would be expected for a homogeneous sphere (*Hood*, 1986; *Hood and Jones*, 1987). Further, the Moon's average density of 3.344 g/cm^3 is consistent with that expected for ultramafic rocks whose mineralogy is dominated by olivine with an Mg# of ~ 90 .

Therefore, based on these observations, it is reasonable to expect that the Moon either has no core or has only a very small core. Initially, it was believed that there was no core present in the Moon; and this observation was the driver behind models that postulated that the Moon had been substantially derived from the silicate portion of Earth (e.g., *Ringwood*, 1979). Effectively, in this scenario, the Moon's complement of metal now resides in the Earth's core.

However, further investigation indicates that the Moon probably does have a small core $\sim 200\text{--}500 \text{ km}$ in radius (*Hood and Jones*, 1987; *Mueller et al.*, 1988; *Jones and Hood*, 1990). This calculation is sensitive to many assumptions, particularly the bulk silicate composition, the assumed differentiation history of the Moon, and the current thermal state of the lunar interior. However, permuting these various parameters results in a series of models, most of which require a small metallic core. And as long as this core is $300\text{--}450 \text{ km}$, its size would also be consistent with electromagnetic sounding data that place size limits on the lunar core (*Hood et al.*, 1999).

In summary, we reiterate that a small lunar core is consistent with measurements of lunar density and moment of inertia and is actually helpful in modeling these parameters

in a self-consistent way (*Hood and Jones*, 1987). There are also good geochemical reasons for invoking a small lunar core that we will discuss in more detail below.

2.5. Summary

The Earth and Moon have similar O-isotopic compositions and therefore are probably closely related in terms of the materials that comprise them. However, there are also significant chemical differences, particularly in Mg# and redox state. If the Moon formed by a giant impact upon Earth, there must have been significant amounts of impactor material incorporated into the Moon to account for these chemical differences. In an extreme version of this model, the composition of the Moon could be dominated by the composition of the impactor (e.g., *McFarlane*, 1989). In the latter case, any chemical similarities between the Earth and Moon would be largely coincidental.

3. VANADIUM, CHROMIUM, AND MANGANESE

There are indeed apparent chemical similarities between the Earth and Moon. The elements V, Cr, and Mn were considered crucial by *Wänke and Dreibus* (1986) in this regard. All three elements are depleted in the Earth's mantle relative to chondrites. And *Wänke and Dreibus* (1986) assumed that V, Cr, and Mn all partitioned into Earth's core early (prior to the formation of the Moon) under very reducing conditions. Subsequently, small amounts of oxidized material were added to Earth's mantle, so that it acquired its current redox state. Thus, *Wänke and Dreibus* (1986) considered the depletions of Cr, Mn, and V to be a uniquely terrestrial signature. Analyses of meteorites considered to be of martian origin (the so called "SNC" meteorites) and samples from the eucrite parent body (hereafter, EPB; probably asteroid 4 Vesta) indicated an absence of this kind of fractionation.

Ringwood et al. (1991) bolstered this view by providing experimental evidence that V, Cr, and Mn would not partition into the core of a Mars-sized planet. Therefore, these authors postulated that the depletions of V, Cr, and Mn in the Moon could not be due to depletions inherited from the impactor (in a giant impact scenario) but must reflect their depletion in Earth during core formation at very high pressures ($\sim 1000 \text{ kbar}$) and very reducing conditions. The abundances of V, Cr, and Mn in the Moon may therefore provide clues to its origin.

But there is little evidence that Mn partitions into Fe-Ni metal, even at reducing conditions (*Rammensee et al.*, 1983). *O'Neill and Palme* (1998) noted that the ratio of the two similarly volatile elements, Mn and Na, is chondritic in the Earth's mantle. Additional loss of Mn into the core is therefore cosmochemically unlikely. The low concentrations of Mn and Na in the Earth are the result of a general depletion in moderately volatile elements. Such depletions are not unusual. Carbonaceous chondrites are increasingly

depleted in Mn and Na going from type 1 to type 3, reflecting nebular conditions during the formation of solid matter. Manganese depletion in the Earth is almost certainly not the result of an indigenous terrestrial process.

The situation is different for V and Cr. The depletion of these two elements in the Earth cannot be ascribed to volatility-related depletions. Chromium is similar in volatility to Mg; and V is nominally considered even less volatile. But, although V is generally considered a refractory element, it is less enriched in Ca,Al-rich inclusions from the Allende meteorite than other more refractory elements such as Al and REE (Wänke *et al.*, 1974a). This leads to a bulk depletion of V in CV chondrites, such that the Al/V ratio in CV chondrites is 182 (Wasson and Kallemeyn, 1988), whereas the CI ratio is 159 (Palme and Beer, 1993). On the other hand, the BSE has an Al/V ratio of 291 (O'Neill and Palme, 1998), a value clearly larger than observed in bulk meteorites. This would argue against volatility as the cause of the terrestrial V depletion. But Shaffer *et al.* (1991) showed that the volatilization rate of V was sensitive to f_{O_2} and therefore could be a function of the local dust/gas ratio during condensation. Alternatively, some V may reside in Earth's core, as suggested by the experiments of Rammensee *et al.* (1983) and Ringwood *et al.* (1991). Consequently, interpretation of the Earth's V depletion is not straightforward.

The most model-independent element of the three appears to be Cr. Chromium is the most fractionated, as seen in the Mg/Cr ratio, which is 87 in the mantle of Earth and 36 in CI chondrites. Chromium in the Earth's mantle may indeed be possibly ascribed to early fractionation into the Earth's core, as suggested by several authors. Consequently, we will concentrate the remainder of our discussion on Cr.

First, it is significant that the Cr isotopic compositions of the Earth and Moon are the same (Lugmair and Shukolyukov, 1998). As in the case of O, the Cr isotopic signature of the Earth and Moon indicates that both bodies formed from similar materials. And in the context of a giant impact, it appears that large contributions (more than approximately two-thirds) from an isotopically dissimilar impactor are not allowed (Lugmair and Shukolyukov, 1998). Of course, this estimate depends in detail on the exact magnitude of the impactor's Cr isotopic anomaly.

Chromium is a compatible element and that makes estimating its abundance in the bulk Moon problematic. A lunar Cr abundance of 2200 ppm (Ringwood *et al.*, 1987) was primarily derived from the Cr/Fe correlation among mare basalts (Seifert and Ringwood, 1987; O'Neill, 1991). However, mare basalts may be derived from rather differentiated source regions, as demonstrated by their ubiquitous Eu anomalies (i.e., denoting plagioclase fractionation). Thus, the Moon may have a large reservoir of cumulate opx and olivine that has not been sampled. In calculating the lunar Cr content, the amount of Cr in this reservoir is very important but not well known.

We have recalculated the Cr content of the Moon in two different ways that complement earlier calculations. The first makes use of the good Cr vs. V correlation in lunar samples

(Seifert and Ringwood, 1987; Haskin and Warren, 1991) and the assumption that V is present at $2.2\times$ CI in the Moon (124 ppm), along with other refractory elements. This calculation yields 2500 ppm Cr in the bulk silicate Moon and a lunar Cr/V ratio of 20. If V is truly depleted relative to other refractory lithophile elements, then bulk lunar Cr will be correspondingly lower. For comparison the BSE Cr/V ratio is 30 and the Cr content of the BSE is 2540 ppm (e.g., O'Neill and Palme, 1998).

A second method is to assume that the Apollo 15 green glass (an olivine-normative composition with a fairly flat REE pattern and a small, negative Eu anomaly) was only in equilibrium with olivine at the time of its eruption and use an appropriate olivine/liquid partition coefficient for Cr to arrive at the Cr content of the lunar mantle. In this calculation we assume that Cr in the crust is negligible and that green glass A (Delano, 1986a) was produced by 20% partial melting. For a melt of Apollo 15 green A composition (Delano, 1986a) and a Cr speciation dominated by Cr^{2+} , the $^{ol/liq}D_{Cr}$ is ~ 0.5 (Hanson and Jones, 1998). This translates into a Cr content of 2000 ppm for the bulk silicate Moon.

These calculations are clearly oversimplified. But several different approaches, with very different assumptions, converge on Cr abundances of 2000–2500 ppm; and the 2500 ppm value may be an upper limit if V is actually depleted compared to other refractory elements. Here, we will adopt the Ringwood *et al.* (1987) value of 2200 ppm that is intermediate to our two calculations.

(As an aside, we return to the possibility of a "hidden" reservoir that is enriched in Cr and Mg and that is possibly the source of the lunar highland Mg suite. However, our adopted value for bulk lunar Cr of 2200 ppm is not very different from the Cr content of olivine-rich Mg-suite rocks. For example, the dunite 72415 has a Cr content of 2300–2500 ppm. Consequently, it is not obvious to us that ignoring the Mg-suite reservoir in our calculation is a serious problem.)

For a bulk lunar MgO content of 37% and our adopted Cr concentration, the Mg/Cr ratio of the Moon is ~ 100 and similar to that of Earth's mantle (~ 86). Within the error of the calculation, the Mg/Cr ratios of the mantles of the Earth and Moon are the same. Thus, although Mg/Cr shows little variation in chondrites, Cr appears to be significantly and similarly depleted in both the Earth and Moon. This may be the best evidence for a terrestrial origin of the Moon. However, as we shall see below, at least one other element, W, has similar depletions in the Earth and Moon, but that this is a coincidence.

4. ALKALI METALS AND THEIR ISOTOPES

4.1. Alkali Metals

The alkalis in the Moon are depleted relative to Earth, although the exact amounts of these depletions are subject to controversy.

4.1.1. Earth. *Dreibus et al.* (1976) presented Li-Zr systematics for terrestrial basalts, indicating that the Li/Zr ratio of the Earth is ~ 0.1 . If refractory lithophiles such as Zr exist at $2.75\times$ CI in the Earth (*Jagoutz et al.*, 1979), the Li concentration is 1 ppm. Both *Taylor and McLennan* (1985) and *Ringwood* (1991) give a Na concentration of ~ 2500 ppm for the bulk silicate Earth (BSE). The K content of the Earth is constrained from both ^{40}Ar systematics and from correlations with nonvolatile, incompatible elements from igneous rocks. *McDonough et al.* (1992) calculate a K content for the BSE of 240 ± 30 ppm. This number is based on a terrestrial K/U ratio of 1.3×10^4 and a BSE U concentration of 21 ppb. In addition, from correlations of K and Rb, a BSE Rb concentration of 0.64 ppm can be derived. *Jones and Drake* (1993), while accepting the *McDonough et al.* (1992) calculation of K and Rb, were not convinced that the BSE Rb/Cs of 28 estimated by these authors was correct. This was because *McDonough et al.* (1992) chose to invoke an unsampled mantle reservoir possessing chemical properties that they found convenient. *Jones and Drake* (1993) used mass balance arguments and the Rb/Cs ratios of continental crust and oceanic mantle to calculate a BSE Rb/Cs ratio of 40, which translates into a BSE Cs concentration of 16 ppb. These calculated abundances for Li, Na, K, Rb, and Cs correspond to depletions relative to CI chondrites of 0.72, 0.50, 0.43, 0.27, and 0.086 respectively. Thus, the alkalis are somewhat depleted in the Earth, and these depletions appear to increase with increasing volatility (*Kreutzberger et al.*, 1986).

4.1.2. Moon. The *Dreibus et al.* (1976) Li-Zr correlations for lunar rocks translate into a lunar Li concentration of 0.35 ppm, assuming refractory lithophile elements in the Moon exist at $\sim 2.2\times$ CI (*Ringwood et al.*, 1987; *Jones and Delano*, 1989). Similarly, a Rb concentration of 0.086 ppm and a K concentration of 36 ppm for the Moon are derived from the lunar Ba/Rb and K/Ba ratios of 60 and 7 respectively (*BVSP*, 1981; *Wänke*, 1981). The Rb/Cs ratio of 21 for mare basalts and KREEP yields a lunar Cs concentration of 4.1 ppb. And taking Taylor's Na/K ratio of 7.2 for the bulk Moon (*Taylor*, 1980) translates into a bulk Na concentration 260 ppm. This calculation for Na is difficult since Na may have acted somewhat compatibly during early lunar petrogenesis. For example, anorthositic highland samples may contain nearly as much Na as KREEP. And using the *Taylor* (1980) Na value in this calculation results in a Na/K ratio that is slightly subchondritic. Therefore, we prefer to use the Na estimate of *Ringwood et al.* (1987) of 450 ppm and will do so in subsequent discussions. These calculated abundances for Li, Na, K, Rb, and Cs correspond to depletions relative to CI chondrites of 0.24, 0.090, 0.066, 0.038, and 0.021 respectively.

4.1.3. Discussion. Earth. Figure 1 shows terrestrial alkali depletions plotted vs. the square root of their atomic masses. This linear correlation is considerably better than that based on 50% condensation temperatures (e.g., *Wasson*, 1985) or the boiling temperatures for the pure elements and is suggestive of fractionation by a diffusive transport process or Jeans escape (i.e., fractionations that depend on

$M^{-0.5}$). Conceptually, this idea has attractive aspects. If depletions in "volatile" elements were actually due to a transport process, as opposed to simple volatility, a long-standing contradiction might be explained: If a moderately volatile element is modestly depleted, a highly volatile element ought not to have condensed at all. In fact, what is observed in chondrites is that the highly volatile elements are more depleted than moderately volatile elements, but not to the extent thermodynamics would predict.

However, if the systematics of Fig. 1 are due to a nebular process, then alkali abundances of chondrites should have the same functional form. In fact, this is not observed. Although there are aspects of the alkali patterns for Earth and for carbonaceous chondrites that are similar, imagination would be required to equate them. The strongest similarity, in terms of the slope of the depletion trend, is between the Earth and one of the experiments of *Kreutzberger et al.* (1986), where a suite of alkalis were volatilized from a $\text{Di}_{75}\text{-An}_{25}$ silicate liquid.

The *Kreutzberger et al.* (1986) experiment cannot be indicative of a transport process *per se*. The charges themselves are homogeneous with respect to the alkali elements (M. J. Drake, personal communication). The $M^{0.5}$ parameterization therefore reflects some process at the liquid-gas interface (or, possibly, it is merely a convenient parameterization). Consequently, it is tempting to interpret the terrestrial alkali pattern as due to partial volatilization of dust or

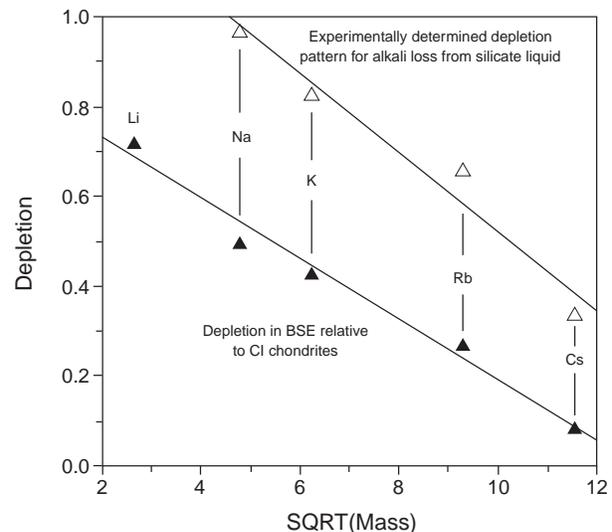


Fig. 1. Alkali depletions in the Earth vs. the square root of their atomic weight (solid symbols). It is generally believed that alkali depletions in the Earth are due to volatility. However, condensation temperatures are not available for all alkalis. Because of this, we have used the square root of atomic weight as a parameterization of alkali depletions. Shown for comparison are depletions produced in a 1-bar volatility experiment (open symbols; *Kreutzberger et al.*, 1986). Both patterns have similar slopes and could possibly reflect the operation of similar processes; but the exact interpretation of the terrestrial depletion pattern is still uncertain (see text).

early nebular condensates. But this hypothesis, too, has its difficulties. The data of *Humayun and Clayton (1995)* constrain the depletion process to be one that does not fractionate K isotopes (see below); for most volatile-loss scenarios, this is difficult (although not impossible).

Alternatively, there could be a fractionation that is indigenous to Earth that would produce the observed relationship between abundance and mass. Metallization of K, Rb, and Cs has been suggested as a possible mechanism for incorporating these elements into the core (e.g., *Bukowinski, 1976*). However, in this case, the inclusion of Na and Li by the parameterization would have to be accidental.

We currently have no unambiguous explanation for the systematics of Fig. 1. We tentatively prefer some partial volatilization scenario. Conceivably, there are nebular processes that could totally volatilize all the K from some dust grains but leave other grains largely unaffected. Alternatively, if the devolatilized dust and gas were to maintain chemical equilibrium, isotopic fractionation might be avoided (*Humayun and Clayton, 1995*). But in this case, we are again faced with the thermodynamic problem of how a suite of elements with rather different condensation temperatures can be depleted to similar degrees.

Moon. Figure 2 shows the lunar alkali pattern normalized both to the BSE (top) and to H chondrites (bottom). Lunar alkalis are depleted and fractionated relative to those of Earth. However, because Cs is enriched compared to what would be expected on the basis of volatility alone (*Kreutzberger et al., 1986; Wulf et al., 1995*), other factors must

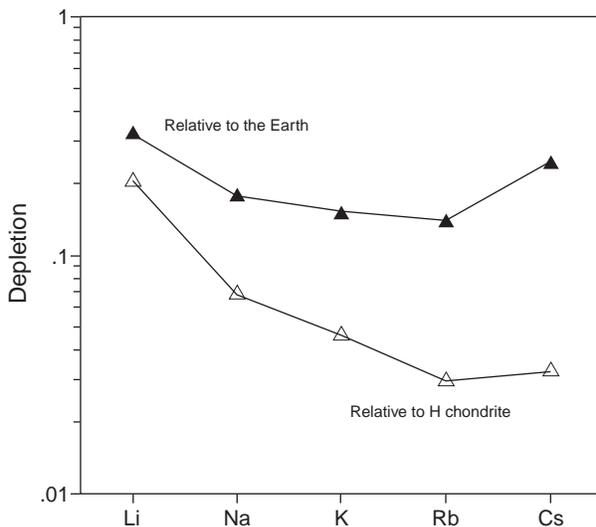


Fig. 2. Lunar alkali abundances normalized to those of the Earth and to an H chondrite. The Rb/Cs ratios of the Earth and Moon make derivation of the Moon from the Earth problematic. If the Moon were derived from the Earth, but depleted in alkalis by a high-temperature event, the most volatile element, Cs, should be the most depleted and this is not observed. But as noted by *O'Neill (1991)*, lunar Rb and Cs abundances may be best explained by minor addition of chondritic material to the proto-Moon. The lunar Rb/Cs ratio is very similar to that of an H chondrite.

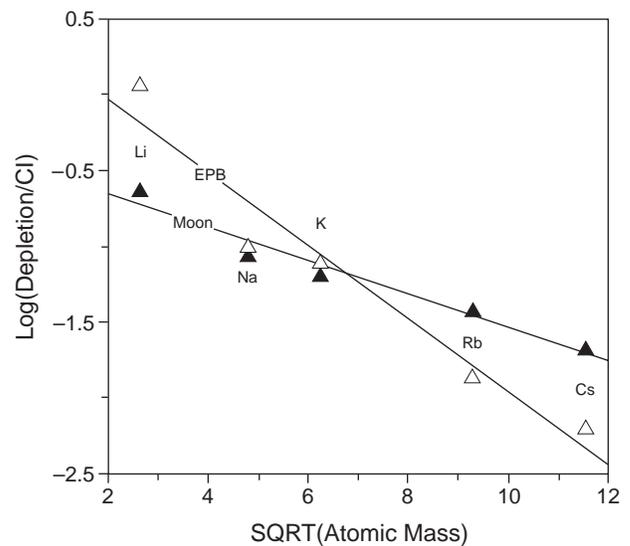


Fig. 3. Log depletion vs. square root of atomic weight for alkalis in the Moon (solid symbols) and the eucrite parent body (open symbols). Note the difference in functional form from Fig. 1. While alkalis in the Moon and EPB are not identical, they are broadly similar. Because we have no reason to believe that the EPB suffered a giant impact, we question whether the Moon did either.

have played a role in establishing lunar alkali abundances. *O'Neill (1991)* suggested that addition of 4% H-chondrite material, in addition to volatile loss from terrestrial parental materials, could explain the observed lunar alkali abundance pattern. Figure 2 shows that, indeed, the lunar Rb and Cs abundances are in H-chondritic relative proportions (i.e., flat pattern). And, if this H-chondrite model is correct, approximately two-thirds of the lunar K budget came from that source as well. But, again, lunar alkalis cannot be directly related to terrestrial alkalis in any simple way.

Figure 3 compares the alkali depletions for the Moon and the EPB. The EPB relative depletions were calculated by taking the alkali/Ba analyses of *Tera et al. (1970)* and assuming that all these elements are incompatible during silicate partial melting. Absolute depletions were calculated assuming that Ba in the EPB is $2 \times CI$. Note the change of scale from Fig. 1, which is linear, whereas the ordinate of Fig. 3 is a log scale. The square root of atomic weight parameterization is not as good for these bodies as for Earth, but the main point of Fig. 3 is that the alkali depletions in the Moon and EPB are more similar than they are different. Currently we have no reason to suspect that an event as energetic as the hypothetical giant impact ever affected the EPB. Consequently, the similarity of elemental depletion patterns between the Moon and the EPB does not seem to require a giant impact origin for the Moon.

Finally, we note that some of the experimental alkali depletions produced by *Kreutzberger et al. (1986)* appear similar to the Moon and the EPB. For short experiments, like that shown in Fig. 1, depletion is best plotted vs. $M^{0.5}$. For long experiments with greater alkali loss, the best pa-

parameterization is $\log(\text{depletion})$ vs. $M^{0.5}$. The shorter experiments appear “Earth-like,” whereas the longer experiments appear “EPB-like.” Thus, it is possible that a single loss mechanism, operating to differing degrees, can explain the observed variation in alkali depletion patterns for the terrestrial planets.

4.2. Constraints from Potassium Isotopes

The foregoing discussion has mostly neglected the important observation that the K isotopic composition of solar system materials is the same everywhere (*Humayun and Clayton, 1995*). The implication of this observation is that the K depletions in the rocky bodies of the inner solar system must have acquired these depletions in a manner that does not fractionate isotopes. This rules out fractional evaporation, diffusion, and Jeans escape as viable depletion mechanisms. It also appears to rule out the hypothetical transport process that was alluded to above as a means of explaining the systematics of Fig. 1.

The slope of the line in Fig. 1 corresponds to a $\sim 4\%$ per mass unit fractionation, if all alkalis acted as isotopes of a single hypothetical element ($\%$, “per mil,” is a measurement in parts per thousand). This implies, based on Fig. 1, that the terrestrial $^{41}\text{K}/^{39}\text{K}$ ratio should be 8% lighter than CI, which is far outside the analytical errors of the *Humayun and Clayton (1995)* analyses ($\sim 1\%$), and fractionations of this magnitude are not observed.

The lack of a difference between the K isotopic compositions of the Earth and Moon also implies that, if the Moon was derived from terrestrial materials, then the observed depletion in lunar K cannot be due to a process that fractionates isotopes. Possibly, though, the addition of unfractionated H-chondrite material (*O’Neill, 1991*) after depletion/fractionation masks isotopic changes. To a certain extent, we can evaluate this possibility. Even if two-thirds of the Moon’s K was added after the hypothetical depletion/fractionation process (Fig. 2), it should be possible to detect the fractionations associated with a giant impact if the K isotopic fractionation from this process was greater than 6% . For comparison, in simple Rayleigh fractionation, it is expected that a $6.5\times$ depletion (the observed depletion) would cause a $\sim 60\%$ effect (*Humayun and Clayton, 1995*). And if one-third of the Moon’s K were made of this material, the effect would be observable. If two-thirds of lunar K were added as a chondritic veneer, the original depletion would have actually been larger, with concomitant increases in isotopic fractionation. Again, the K isotopic compositions of the Earth and Moon are identical within analytical uncertainty.

4.3. Summary

In many respects, lunar alkalis are similar to those of other small, volatile-depleted bodies of the solar system, such as the EPB. There is no evidence from this suite of elements that the Moon was primarily derived from terres-

trial materials. In particular, the K isotopes of lunar rocks are unfractionated relative to Earth and CI chondrites, whereas Rayleigh fractionation following a giant impact should have resulted in isotopically heavy K in the Moon relative to Earth. We conclude that there is no evidence of a giant impact in a suite of elements that is expected to be highly sensitive to volatile element fractionations. There is no reason why conclusions regarding alkali elements, in particular the highly volatile element Cs, should not be extended to other elements with similar volatility (Br, Pb, Tl, etc.). These elements are depleted in lunar basalts relative to terrestrial basalts by nearly a constant factor (0.026; *Wolf and Anders, 1980*).

5. THE LUNAR HAFNIUM/TUNGSTEN RATIO AND THE SIGNIFICANCE OF $^{182}\text{W}/^{184}\text{W}$ RATIOS IN LUNAR SAMPLES

The work of *Lee and Halliday (1995, 1996)* has demonstrated the usefulness of the ^{182}Hf - ^{182}W system for dating the separation of metal from silicates in planetary and protoplanetary environments. The decay of ^{182}Hf ($t_{1/2} = 9$ m.y.) to ^{182}W produces a ^{182}W excess in systems with high Hf/W ratios. Because Hf is lithophile and W is siderophile, metal that separated from silicates early in solar system history has lower than chondritic $^{182}\text{W}/^{184}\text{W}$ ratios. And the silicates that are complementary to this metal have high Hf/W ratios and have correspondingly higher $^{182}\text{W}/^{184}\text{W}$ ratios. These excesses or deficiencies in ^{182}W are commonly expressed as ϵ units, which are deviations from the chondritic value expressed as parts in 10^4 . For example, iron meteorites and chondritic metals, with Hf/W ratios near zero, typically have $\epsilon(^{182}\text{W})$ values of -3 to -4 , whereas eucrites, with Hf/W ratios $\sim 20\times$ CI, have $\epsilon(^{182}\text{W})$ values of $+30$ to $+40$.

5.1. Terrestrial and Lunar Tungsten Isotopes

The $^{182}\text{W}/^{184}\text{W}$ ratio in terrestrial samples is indistinguishable from bulk chondrites [$\epsilon(^{182}\text{W}) = 0$ (*Lee and Halliday, 1995*)]. Because W and Hf are both refractory elements, we do not expect them to fractionate significantly during nebular condensation. However, W and Hf are observed to be fractionated by a factor of ~ 20 in the BSE, presumably by terrestrial core formation. Consequently, if the bulk Earth has chondritic Hf/W and the BSE has chondritic $^{182}\text{W}/^{184}\text{W}$ but superchondritic Hf/W, then separation of W from Hf by core formation must have occurred late, more than 50 m.y. after the formation of chondrites (*Lee and Halliday, 1995*), when all ^{182}Hf had decayed.

In contrast to Earth, there is a clear $^{182}\text{W}/^{184}\text{W}$ excess in most of the lunar samples analyzed by *Lee et al. (1997)*. As discussed in detail by *Lee et al. (1997)*, the most reasonable explanation for these ^{182}W excesses is that the Moon formed and differentiated before all ^{182}Hf had decayed. However, *Wieler and coworkers (personal communication, 1999)* have recently presented calculations indicating that,

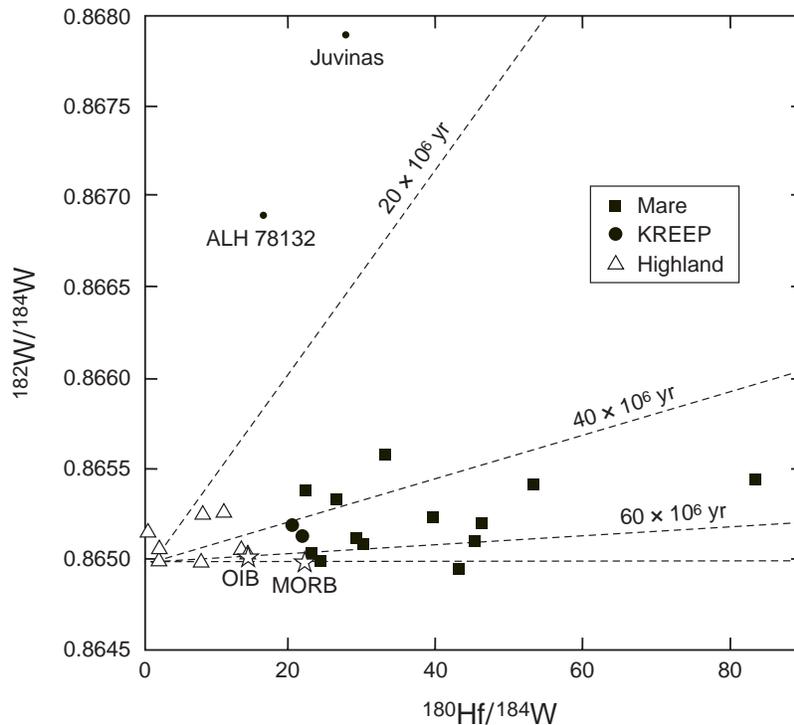


Fig. 4. Tungsten-182/tungsten-184 vs. $^{180}\text{Hf}/^{184}\text{W}$ for lunar samples. Also shown are data for two eucrites (ALH 78132 and Juvinas) and two terrestrial basalts (OIB and MORB; stars). Dashed lines are isochrons with the time of formation dating from that of chondrites. Most mantle-derived lunar samples show a positive (radiogenic) ^{182}W anomaly. It is difficult for the Moon to have inherited its W from the Earth, whose W is not distinguishable from chondritic.

because the cosmic-ray-exposure ages of lunar rocks can be large, ^{182}W -isotopic analyses of lunar samples must be corrected for neutron capture on ^{181}Ta . But even after appropriate corrections are applied to the data of *Lee et al.* (1997), ^{182}W excesses remain, although the sizes of the anomalies are reduced. Consequently, we will discuss the *Lee et al.* (1997) data as they were published. In Fig. 4 the lunar samples analyzed by *Lee et al.* are plotted on a $^{182}\text{W}/^{184}\text{W}$ vs. $^{180}\text{Hf}/^{184}\text{W}$ diagram. Lines with constant $^{182}\text{Hf}/^{180}\text{Hf}$ ratios at 20, 40, and 60 m.y. after chondrite formation are indicated, assuming a $^{182}\text{Hf}/^{180}\text{Hf}$ ratio of 2.61×10^{-4} at the beginning of the solar system. Data for two eucrites, Juvinas and ALH 78132, and two terrestrial basalts, MORB and ocean island basalts (OIB), are shown for comparison (*Lee and Halliday, 1997*).

In models where the Moon formed from Earth's mantle after terrestrial core formation (e.g., *Ringwood, 1979*), one should not expect excess ^{182}W in lunar samples. Lunar samples should lie along a horizontal line with terrestrial basalts (Fig. 4). If, however, the formation of the Moon resulted in a significantly higher lunar Hf/W ratio, it is possible that a small $^{182}\text{W}/^{184}\text{W}$ anomaly that cannot be resolved in terrestrial rocks could be transformed into an analytically resolvable anomaly in lunar rocks. This is essentially the interpretation of *Lee et al.* (1997).

Let us explore this scenario in more detail. In Fig. 4 the terrestrial OIB sample is slightly above the average terrestrial ratio of 0.86500, but with overlapping error bars. The mare basalt plotting on the same line (60×10^6 yr) has a clearly resolvable anomaly. Thus, one way to produce the observed lunar anomalies from young (i.e., ~ 60 m.y. after

chondrites), terrestrial mantle material would be to drastically increase the lunar Hf/W ratio during the formation of the Moon. Later, after all ^{182}Hf has decayed, lunar samples would all plot along the 60-m.y. line. Subsequently, the Hf/W ratios must then be shifted back to the presently observed values without changing the $^{182}\text{W}/^{184}\text{W}$ ratio. But because Hf and W both act incompatibly during silicate differentiation, it is hard to make large changes in the Hf/W ratio. Consequently, even though the scenario we have just given is mathematically possible, it appears physically implausible. In effect, this scenario requires a "hidden reservoir" for either W or Hf in the Moon.

We can discuss this possibility more quantitatively. The basic prerequisites for understanding the W-isotopic evolution of the Moon and its significance for a terrestrial origin are (1) knowledge of the average W/Hf ratio of the Earth and Moon and (2) the extent of early Hf/W fractionation in the Moon.

5.2. Lunar and Terrestrial Hafnium/Tungsten Reservoirs

Wänke et al. (1974b) found a correlation of La with W in a large number of lunar samples of different origin, and the lunar La/W ratio was about $20 \times \text{CI}$. *Rammensee and Wänke* (1977) experimentally determined the metal/silicate liquid partition coefficient of W and concluded that, in order to achieve the observed W depletion, the Moon would require a much more massive metal core than was compatible with lunar density. From this it was inferred that the observed depletion of W in the Moon was achieved during

core formation in the Earth, supporting a terrestrial origin for the Moon (*Rammensee and Wänke, 1977*).

The constancy of La/W ratios in lunar samples suggests a similar behavior of the two elements during global lunar differentiation. Crystallization of the lunar magma ocean resulted in the formation of the anorthositic highland crust, the residual mafic lunar interior, and a small residual melt layer. This residual melt possessed high concentrations of incompatible elements that did not partition into the mafic cumulate layer or the anorthositic crust and was enriched in K, REE, and P, leading to the acronym KREEP for this component (*Hubbard et al., 1971*). After the formation of the Moon, impacts tapped the KREEP reservoir and distributed KREEP-rich material on the surface of the Moon, in particular at the Apollo 14 and (in part) 15 sites (*Warren, 1985*, and references therein). The relative abundances of

the incompatible elements in all KREEP-containing samples is surprisingly uniform (*Palme and Wänke, 1975; Warren and Wasson, 1979*), strongly indicating that the abundance sequence of these incompatible elements reflects a Moon-wide differentiation event.

In Fig. 5 the KREEP-rich soil sample 14163 represents the KREEP component. Elements in the upper part of Fig. 5 are arranged in order of decreasing CI-normalized abundance. The sequence thus reflects the degree of incompatibility during crystallization of the lunar magma ocean.

Mare basalts formed by remelting of the mafic cumulate layer. This layer must have had an incompatible-element pattern complementary to KREEP. Elements such as U or Ba are so incompatible that they can hardly be accommodated in the mafic cumulate layer; elements more compatible with mafic silicates, such as Yb and Sc, have higher

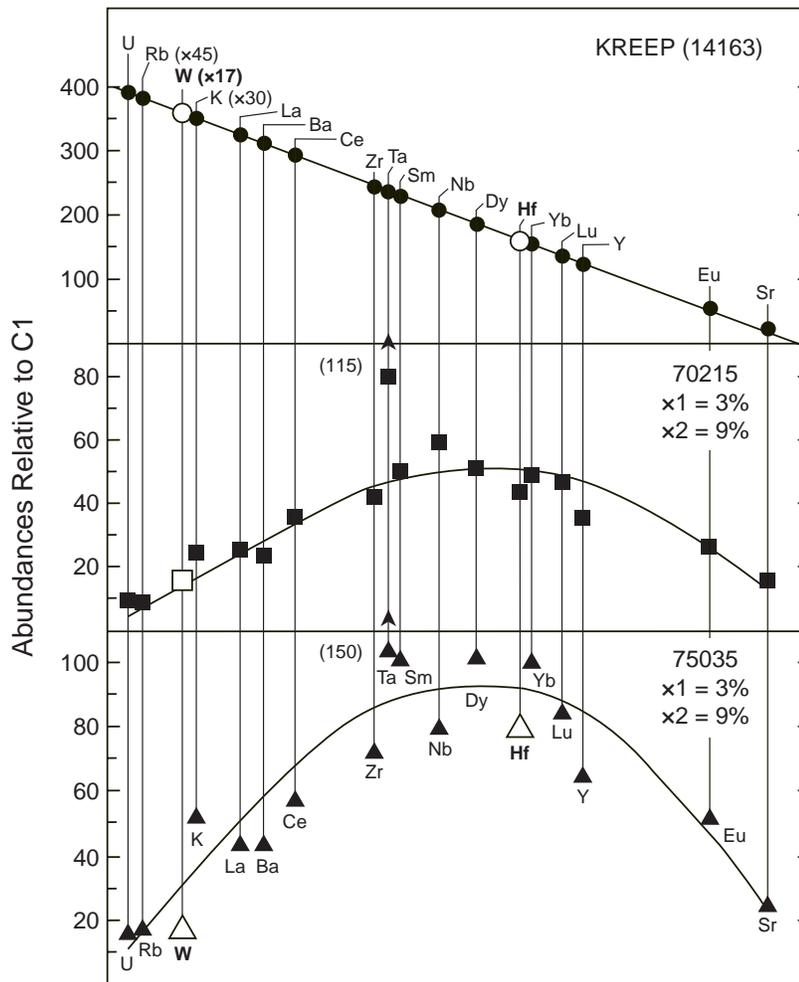


Fig. 5. Trace-element abundances (relative to CI chondrites) for lunar samples. Two complementary reservoirs for Hf and W exist on the Moon, KREEP and the cumulates from the lunar magma ocean. For KREEP the elements are ordered in increasing incompatibility and are believed to be the result of 97% fractional crystallization of the lunar magma ocean. The mare basalts 70215 and 75035 can be modeled by 9% partial melting of the cumulates. Tungsten is more incompatible than Hf during lunar petrogenesis and if lunar ^{182}W anomalies are generated from terrestrial material by producing regions of extreme Hf/W ratios, KREEP should have no ^{182}W anomaly. KREEP does have a positive ^{182}W anomaly, making derivation of the Moon from terrestrial material unlikely.

concentrations in mare basalts and lower concentrations in KREEP. In a second event, partial melting of the cumulate layer produced the characteristic patterns of mare basalts, as shown for two Apollo 17 basalts in Fig. 5 [see *Wänke et al.* (1974b) and *Palme and Wänke* (1975) for details].

The complementary relationship between KREEP and mare basalts allows us to “quantify” the incompatibility of W. The abundance of W in Apollo 14 soil 14163 and in the two mare basalts can be fitted into this scheme by placing W between La and U in Fig. 5. Thus, in the lunar environment W is more incompatible than La and less incompatible than U. The positions of the moderately volatile elements Rb and K in Fig. 5 are determined in the same way. From Fig. 5 it is also clear that the Hf/W ratio of the bulk Moon is bracketed by KREEP (from below) and mare basalts (from above).

Palme and Rammensee (1981) noticed a good correlation of W and U in lunar rocks with systematically higher ratios of U/W in KREEP-dominated rocks and lower U/W ratios in mare basalts, as expected from Fig. 5. An average lunar U/W ratio of 1.93 was calculated. This leads to a W depletion factor, $(W/U)_{\text{Moon}}/(W/U)_{\text{CT}}$, of about 22 \times . (The depletion factor of 17 \times indicated in Fig. 5 is based on less-accurate La/W ratios and older CI normalizing values.) From this U/W ratio and the CI abundances of refractory elements, an average lunar Hf/W of 25.2 is calculated, corresponding to a $^{180}\text{Hf}/^{184}\text{W}$ ratio of 29.7.

A similar picture is obtained for Earth. In Fig. 6 we have plotted the average incompatible element contents of continental crust normalized to the BSE. Terrestrially, W is apparently the most incompatible element, with the largest crust/mantle ratio of 69, using the data of *Newsom et al.* (1996) with 1.1 ppm for BCC (bulk continental crust) and

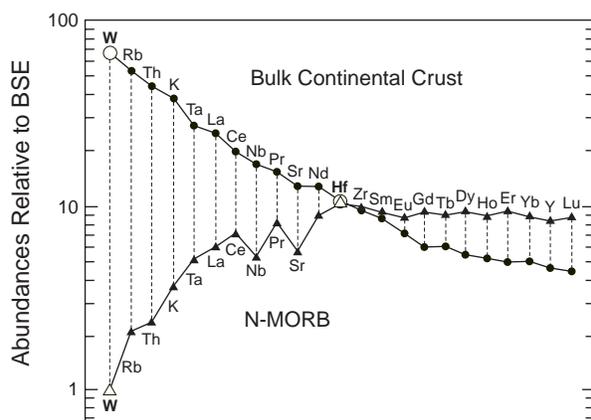


Fig. 6. Trace-element abundances in the Earth’s continental crust and oceanic upper mantle, normalized to the bulk silicate earth (BSE). Though produced by a different process than on the Moon (*Hofmann*, 1988), the Earth also has complementary reservoirs, with the continental crust resembling KREEP and the MORB mantle resembling the lunar cumulate mantle.

16 ppb for the BSE, significantly above Rb with a corresponding factor of 53 (BSE, *McDonough and Sun*, 1995; BCC, *Taylor and McLennan*, 1985). The high extraction efficiency of W into the crust is compensated by low W in the depleted mantle as sampled by MORB. The similarity of Figs. 5 and 6 reflects the strongly incompatible behavior of W in the Earth and Moon. The depletion of W in the Moon (22 \times) is within error the same as that of the terrestrial mantle (21 \times). But although this similarity in Hf/W has been used to support a common origin for the Earth and Moon, the presence of a W-isotopic anomaly on the Moon and its absence on Earth cannot be reconciled with a completely terrestrial origin for the lunar W.

In summary, there is no evidence for a “hidden reservoir” for either W or Hf in the Moon. The basaltic samples analyzed by *Lee et al.* (1997) fit comfortably into the scheme outlined above. They are either representative of the depleted mantle or the residual KREEPy liquid from the crystallization of the lunar magma ocean. We argue, therefore, that the geochemistry of W and Hf in the Moon is well understood.

The logic seems inescapable. If the Moon formed from the terrestrial mantle late (~50 m.y. after chondrites), the Hf/W ratio of the Moon must increase so that any remaining ^{182}W contributes significantly to the lunar W budget. There are only reasonable two ways to do this: (1) extraction of W to the lunar core or (2) extreme silicate differentiation. The first scenario is ruled out since the Hf/W ratios of the Earth and Moon are the same; and the second scenario makes a simple prediction that is not fulfilled. One of the complementary lunar reservoirs, either KREEP or the mafic cumulates, should have no W-isotopic anomaly. In this case, the prediction is that the low Hf/W reservoir (KREEP) should have no anomaly. In fact, KREEP-rich basalts do have a W isotopic anomaly of at least 2ϵ (*Lee et al.*, 1997). We see no way to make the Moon solely from the terrestrial mantle.

If one wishes to reconcile the W-isotopic evidence of the Earth and Moon with the giant impact hypothesis for the formation of the Moon, more complicated models are required. There are two general kinds of such models:

5.2.1. Early formation of the Moon. The Moon could have formed by a giant impact earlier — for example, 35–40 m.y. after chondrite formation, as indicated in Fig. 4. This, however, requires that the core of Earth formed before the giant impact. But core formation in the Earth earlier than ~50–60 m.y. after chondrites would produce a $^{182}\text{W}/^{184}\text{W}$ anomaly in the mantle that is presently not observed. Later incoming material and additional core formation, after the formation of the Moon, must then have diluted the ^{182}W excess on Earth (but not on the Moon!) to an extent that would not allow detection. If, for example, the Moon had formed from Earth 40 m.y. after the formation of iron meteorites, the Earth and Moon could have had a $^{182}\text{W}/^{184}\text{W}$ ratio of 0.86520 (about 2ϵ units above the chondritic value, at a Hf/W ratio of 20) to explain the presently observed lunar anomalies. To reduce the $^{182}\text{W}/^{184}\text{W}$ ratio to

the present value in the Earth would require the later removal of at least 50% of the W present at that time into Earth's core and the addition of the same amount of W with chondritic $^{182}\text{W}/^{184}\text{W}$.

This model may have physical difficulties. The later addition of accretional components to the Earth required by the model occurred (according to present models of Earth accretion) by impacts of $\sim 10^{26}$ g bodies, so-called embryos (Wetherill, 1994). If these embryos had differentiated into core and mantle with correspondingly elevated $^{182}\text{W}/^{184}\text{W}$ ratios in their silicates (Taylor and Norman, 1990), then the iron cores would have to completely disaggregate and re-equilibrate with the silicates during impact in order to avoid any ^{182}W excess in Earth. But in many of the simulations of the impact process (Benz and Cameron, 1990; Benz et al., 1989; Cameron and Canup, 1998) there is little interaction between the impactor core and the target mantle. Thus, for this general class of models to work, the late-accreting impactors would probably have to be of a different size than is currently envisioned.

However, regardless of the exact physical means by which W is added to Earth after the Moon's separation, highly improbable coincidences are required. If W and Hf in the Moon today are representative of the terrestrial mantle prior to the Moon's formation, then terrestrial radiogenic W must have been sequestered (to the core?) and nonradiogenic W must have been added to Earth's mantle. These additions and subtractions must be done in such a way that the Hf/W ratio remains constant. Therefore, the W added must exactly counterbalance the W that was removed. Whether this delicate balance can be achieved is problematic. It seems more likely to us that the similarity of Hf/W between the Earth and Moon is coincidental.

5.2.2. Contribution of the impactor. Of course, there may have been contributions from the impactor to the Moon. In fact, in some models the impactor dominates the bulk lunar composition (e.g., Newsom and Taylor, 1989), although it is not clear to us that this type of model is testable. An undifferentiated projectile would not provide isotopically unusual W. But if, as discussed above, the projectile were differentiated into a core and mantle, the mantle should have a significantly enhanced $^{182}\text{W}/^{184}\text{W}$ ratio. The core of the impactor with the complementary low $^{182}\text{W}/^{184}\text{W}$ is required to separate from the impactor mantle in order to preserve the ^{182}W anomaly of that mantle and import it into the Moon. Presumably the ultimate fate of that core is the core of Earth (e.g., Benz et al., 1989). Earth's mantle should also receive radiogenic W from the impactor mantle, although presumably less than the Moon. Therefore, this model's attractiveness may possibly be enhanced if the impactor core is allowed to equilibrate or partially equilibrate with the terrestrial mantle. If the impact occurred 50 m.y. after chondrites (or later), then ^{182}Hf would be essentially all decayed, and we cannot expect lunar differentiation to have had any influence on lunar W isotopic composition. We would expect a uniform, elevated $^{182}\text{W}/^{184}\text{W}$

ratio in the Moon. But this scenario seems unlikely given the observed variability in lunar $^{182}\text{W}/^{184}\text{W}$ measurements.

Note the dichotomy within a single scenario: The impactor core is not allowed to equilibrate with protolunar material but may be required to equilibrate with Earth's mantle. To prevent the former it would be convenient if the impactor core remained intact. To facilitate the latter it would be convenient if the core were dispersed. Thus, the fate of the core of a differentiated impactor is a very important question that needs to be addressed by detailed physical modeling.

Of course, it may be that, despite the similarities between the Earth and Moon that we have discussed, the origins of the Earth and Moon are not strongly coupled. In models of a separate origin of the Moon (indistinguishable from models where the entire Moon comes from the impactor), $^{182}\text{W}/^{184}\text{W}$ ratio of the Moon would have to be below that of the mare basalts and above that of KREEP, something around 0.86537, which would, with a $^{180}\text{Hf}/^{184}\text{W}$ ratio of 30, correspond to core formation in the Moon (or in the impactor) at ~ 40 m.y. after the formation of chondrites.

We note as an aside that it is possible that the W-isotopic anomaly of the Moon is significantly larger than estimates based on the currently analyzed samples. Jones (1998) presented evidence that mixing has occurred between a radiogenic lunar mantle and a nonradiogenic crust. His preferred age for the Moon (~ 34 m.y. after chondrites) was derived from the sample with the largest ^{182}W anomaly (15555). But if 15555 itself contains a nonradiogenic W "contaminant," the Moon could be older still. More lunar rocks need to be analyzed to discover the most radiogenic W isotopic signature. However, it is not out of the question that the Moon is significantly older than we currently believe.

5.3. Summary

In summary, the W-isotopic compositions of the Earth and Moon do not support a terrestrial origin of the Moon, although it is possible to construct models that are compatible with W isotopes and the giant impact hypothesis. However, these models are complex and alternately appear to require either additions of undifferentiated or differentiated projectiles that either equilibrate with some reservoirs but not others. Models where the impact occurs early (~ 35 m.y. after chondrites) only work if later-accreting material removes the signature of radiogenic W from Earth's mantle. Models where the impact occurs late (>50 m.y. after chondrites) are best accommodated by an impactor that differentiated early (e.g., Taylor and Norman, 1990). These late impact models work best if the impactor and target (Earth) equilibrate their W isotopes. However, we do not favor late impacts and inherited anomalies because of the observed variation in $^{182}\text{W}/^{184}\text{W}$ within the lunar sample suite. We agree with Lee et al. (1997) that the overall variation in lunar

$^{182}\text{W}/^{184}\text{W}$ is best explained by decay of ^{182}Hf within the Moon.

6. IRON, COBALT, AND NICKEL AND THE NATURE OF THE LUNAR CORE

The major element whose abundance is best known in both the Earth and the Moon is FeO. Based on analyses of lunar rocks and on geophysical data, it seems well established that the FeO content of the silicate fraction of the Moon is about 50% higher than the FeO content of Earth's mantle (see above discussion). Cosmochemically iron is associated with Ni and Co. The three elements Fe, Co, and Ni will therefore be discussed together. Estimates for the bulk silicate Earth and the bulk silicate Moon are summarized in Table 1.

According to Table 1 the silicate part of the Moon has about 50% more FeO than Earth's mantle, the Co content of the lunar mantle is similar to that of Earth's mantle, and the Ni content of the lunar mantle is a factor of 4 below that of Earth's mantle. As the metal/silicate partition coefficients increase in the sequence from Fe through Co to Ni, it is clear that the abundances of these elements (relative to chondrites) in the lunar mantle decrease with increasing siderophilicity.

If the Moon is predominantly made of material with the composition of the present terrestrial upper mantle, then an addition of ~5 wt% FeO from the impactor is required to satisfy the FeO content of the lunar mantle — even more if a lunar core is considered (Table 1). For example, with a devolatilized CI projectile (27.3% Fe; *O'Neill, 1991*), this leads to about 18% of impactor component in the Moon's silicate composition.

$$\text{Fe}_{\text{Earth-mantle}} \times 0.82 + \text{Fe}_{\text{impactor}} \times 0.18 = \text{Fe}_{\text{silicate.Moon}} \quad (1)$$

If the impactor was undifferentiated, as the above calculation implies, it follows that at least 2900 ppm Ni and 136 ppm Co must have been added to the Moon in the same event. These Ni and Co additions are not strongly depen-

dent on the type of impactor, as long as the projectile is of broadly chondritic composition. The impactor contribution of Ni and Co significantly exceeds the presently observed concentration in the Moon (Table 1).

This extra Ni and Co could have been extracted into a lunar core. To accommodate this Ni and Co requires an ~1 wt% lunar core with some 40% Ni (*O'Neill, 1991*). Higher impactor contributions or lower FeO in the proto-Earth mantle imply larger cores. Using this CI-addition model, we can calculate the composition of the lunar core and calculate core/mantle "partition coefficients" of 6.2, 111, and 851 for Fe, Co, and Ni respectively. Experimentally determined metal/silicate liquid partition coefficients (calculated relative to a partition coefficient of 6.2 for Fe) give 215 for Co and 2533 for Ni respectively (*Holzheid et al., 1997*). These values are considerably larger than we just calculated. Thus, for a given Fe distribution between lunar core and mantle, there is too much Ni and Co in the lunar mantle. A larger core with less Ni and more Fe would make the fit worse, as this implies higher core/mantle ratios for Fe and correspondingly higher effective metal/silicate partition coefficients for Ni and Co. Thus, as we saw in our discussion of W isotopes (above), making the Moon by adding undifferentiated chondritic material to terrestrial mantle material is difficult.

To avoid the problem of excess Ni and Co it has been suggested that the impactor was differentiated into core and mantle and that the core of the impactor mixed with that of the Earth without reequilibrating with silicates (*Taylor and Norman, 1990*). The mantle of the impactor would then have to have significantly higher FeO contents than the Earth's mantle in order to produce the elevated FeO of the Moon. A larger fraction of differentiated impactor material is required to produce the Moon's FeO content compared to the addition of chondritic matter, because (by definition) some Fe has been used to make a core. But, as discussed above, an impactor that is differentiated into core and mantle may have a substantial $^{182}\text{W}/^{180}\text{W}$ excess in its mantle (see section on W isotopes). If the impactor core does not reequilibrate with silicate, a significant W anomaly may be imposed on Earth's mantle.

Another possible solution to the excess Ni and Co problem would be to have lunar core formation occur, not from a totally molten Moon, but from a partially molten Moon. *Jones and Delano (1989)* used this concept to explain the high Mg# source for the lunar Mg-suite lithologies. Again, taking the data from Table 1 at face value, we can perform the same mass-balance calculation but with three reservoirs (metal, olivine, silicate liquid) rather than two. If we take the metal/silicate liquid partition coefficients given above and arbitrarily assume 20% silicate partial melting during core formation, we can calculate what values of D_{Co} (olivine/silicate liquid) and D_{Ni} (olivine/silicate liquid) would be necessary to explain the "excess" Co and Ni. For this calculation, we require that 20% silicate liquid and 80% olivine must account for the lunar silicate NiO and CoO and

TABLE 1. Iron, Ni, and Co abundances in the Earth and Moon.

	Fe %	Co ppm	Ni ppm	Fe/Ni (oxides)	Ni/Co (oxides)
Earth mantle	6.35	102	1860	34	18
Lunar mantle	10.1	90	470	205	5.2
Bulk Moon*	10.1	159	4430		
Earth/Moon	0.65	1.17	4.23		
CI	18.23	506	10770	16.9	21.3

Sources: Earth: *O'Neill and Palme (1998)*; Moon: Fe, *Jones and Delano (1989)*; Ni and Co, *Delano (1986b)*; CI: *Palme and Beer (1993)*.

* 82% Earth mantle + 18% CI-volatile free impactor (27.3% Fe).

that the addition of 1 wt% metal to this silicate mixture constitutes bulk lunar Ni and Co. The values of $D(\text{olivine/silicate liquid})$ that we calculate are $\sim 3\text{--}4$ for both elements. We do not expect that $D_{\text{Ni}}(\text{olivine/silicate liquid})$ should equal $D_{\text{Co}}(\text{olivine/silicate liquid})$, but considering the oversimplified nature of the calculation, we consider core formation from a partially molten Moon a viable solution to the Ni-Co problem.

In summary, the higher FeO content of the lunar mantle requires addition of Fe from an impactor if the Moon is primarily made of Earth mantle material. Because of constraints from W isotopes, the additional FeO cannot easily come from an undifferentiated projectile. However, addition of $\sim 18\%$ CI-like material to Earth's mantle would satisfy FeO mass balance. If the FeO content of the impactor mantle resembled that of Mars or the EPB ($\sim 19\%$ Fe), we would need about 30% impactor material to make the Moon from Earth's mantle. The present-day abundances of Ni and Co are better explained if core formation occurred in a partially, rather than a totally molten Moon.

6.1. Other Siderophile Elements

As mentioned above, the abundances of Fe, Co, and Ni in the silicate Moon decrease with increasing siderophilicity. This trend extends to the more siderophile elements, such as Mo, Ge, and Re as shown by *Newsom* (1984). There is a clear trend: the greater its siderophilicity, the greater an element's depletion in the lunar mantle. Thus, the pattern of siderophile elements in the Moon appears to require the presence of a lunar core. Assuming that the Moon dominantly formed from terrestrial materials, the extent of the contamination of the Moon with an impactor is unclear and depends strongly on the poorly known size of the actual lunar core. A larger impactor contribution to the proto-Moon would require a larger core, if the impactor were chondritic. However, as we have argued above, this model is acceptable for the Moon but may have undesired consequences for terrestrial W isotopes.

Calculations by *Jones and Hood* (1990) show that it is difficult to model the lunar siderophile-element pattern by separation of metal from a lunar bulk composition having the siderophile-element abundances of the terrestrial upper mantle. Trivial amounts of metal will deplete terrestrial siderophiles to lunar abundances and no commonality between different siderophile-element models was found (i.e., each siderophile element required a different model). In contrast, great commonality (i.e., intersection of model solutions) was found if a modified chondritic source was assumed (modified in the sense that phosphorus was somewhat volatile depleted). *Jones and Hood* (1990) found intersections between different siderophile-element solutions when the degree of silicate partial melting was $\sim 20\%$ and the size of the metallic core was $3\text{--}5\%$ wt%. This solution is in general agreement with that found for Ni and Co above. And although the *Jones and Hood* (1990) core size was larger than the 1 wt% assumed above, *Jones and Hood*

(1990) assumed more reducing conditions than *O'Neill* (1991), with commensurately less Ni in the metal phase, requiring a larger core.

6.2. Summary

In summary, the lunar siderophile-trace-element pattern appears to require core formation in the Moon. The best fits to the siderophile-element data are models that assume core formation from an essentially chondritic, partially molten Moon. Starting compositions for the Moon that resemble the present-day upper mantle do not fit the siderophile data well. The core sizes ($3\text{--}5\%$ wt%) required to deplete approximately chondritic levels of siderophiles in the silicate portion of such a Moon (to the levels that are currently inferred) are compatible with core sizes estimated from the most recent geophysical data ($1\text{--}3\%$ wt%). These conclusions from other siderophiles are in general agreement with those from W isotopes, provided core formation on the Moon occurred ~ 35 m.y. after chondrites. The only problem for this scenario is the lack of a linkage with the Earth. We reiterate that W-isotopic systematics imply that terrestrial core formation appears to have occurred considerably later than on the Moon. If the giant impact formed the Moon and triggered core formation in the Earth, then this should not be the case.

7. EVIDENCE FROM TERRESTRIAL MANTLE XENOLITHS

The hypothesized giant impact would have been the most important thermal event Earth ever experienced. The kinetic energy gained during such an impact is sufficient to melt most of the Earth and to vaporize the rest (*Melosh*, 1990). But impacts do not distribute their energy uniformly and the physical state of the Earth after the giant impact is uncertain. If for no other reason, the exact geometry of the impact will likely play an important role. Grazing impacts will impart less energy to the Earth than a dead-center collision. Even so, the giant impact is widely believed to have produced a terrestrial magma ocean, and it is possible that this large degree of partial melting facilitated the separation of Earth's core (*Li and Agee*, 1996; *Richter et al.*, 1997). If core formation did indeed occur immediately following the giant impact, the Earth would almost certainly have been left in a totally molten state, since the core-formation event would have also released large amounts of potential energy (e.g., *Jones and Drake*, 1986).

In addition, most models of accretion in the early solar system favor the growth of a relatively small number of embryonic planetesimals (e.g., *Wetherill*, 1994). As these planetesimals accrete into planets, they do so violently, because of their size. All these considerations, taken as a whole, argue that Earth began life very hot and perhaps mostly molten.

In an important paper, *Jagoutz et al.* (1979) noted that several fertile mantle xenoliths (spinel lherzolites) from around the world had major-element abundances that ap-

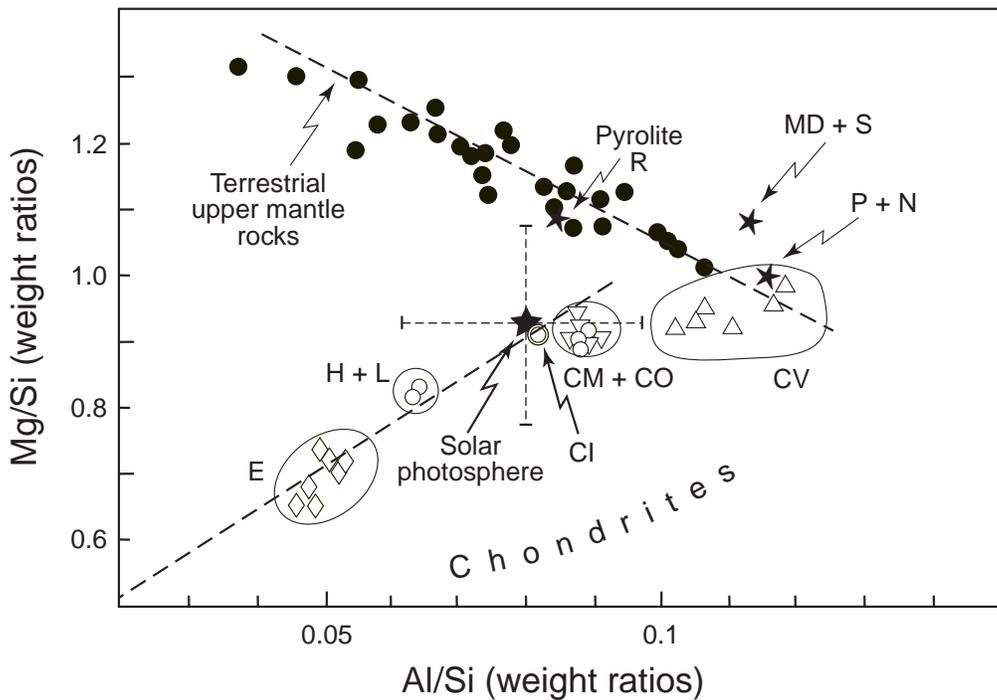


Fig. 7. Magnesium/silicon vs. Al/Si for mantle peridotites and chondrites. As peridotites (solid circles) become more fertile (decreasing Mg/Si and increasing Al/Si), they approach a trend defined by chondrites (open symbols) that was produced by nebular processes. In turn, our best estimates for the composition of the BSE [Pyrolite R = Ringwood (1979); MD + S = McDonough and Sun (1995); P + N = Palme and Nickel (1985)] fall near the intersection of these two trends, as do some of the mantle peridotites. This suggests that some portions of the terrestrial mantle have survived without much igneous processing. In turn, this sheds doubt on whether the Earth ever passed through a magma ocean stage.

proached a trend defined by chondritic meteorites (Fig. 7). They reasoned that these xenoliths were nearly unprocessed samples of the BSE and used them to reconstruct the BSE composition. Jones and Drake (1986), in their review of terrestrial core formation, were impressed by the Jagoutz *et al.* (1979) observation. Consequently, Jones and Drake (1986) deliberately kept the degree of silicate partial melting in their core formation models small, fearing that large degrees of silicate melting would lead to differentiation, eradicating the primitive sample suite identified by Jagoutz *et al.* (1979).

Ringwood (1990) stated his opposition to the giant impact hypothesis by noting that, if the Earth had ever been totally molten, it would tend to crystallize in such a way that it would be stratified and stable against convective mixing. Tonks and Melosh (1990) countered that any terrestrial magma ocean would have convected turbulently and crystallizing solids would remain suspended and not settle, as in a normal magma chamber. Thus, a magma ocean need not imply differentiation.

Jones (1996) reiterated Ringwood's concern in a different way. He noted that, even if the magma ocean turbulently convected and crystals remained suspended, crystallization would eventually proceed to a point where the mantle was

mostly solid and could not continue to convect in a turbulent manner. At this point or at some later time, the residual liquid may not remain in the mantle but may erupt as basaltic magma, depleting the mantle of its basaltic components. Thus, an important question is whether the "primitive" samples identified by Jagoutz *et al.* (1979) could have survived a terrestrial magma ocean. Or can mantle mixing effectively undo the differentiation we expect the magma ocean to have done?

One means of answering this question is to assess the degree to which fertile mantle lherzolites approach chondritic compositions, and one of the most interesting studies utilized Os isotopes. Rhenium-187 decays to ^{187}Os with a $t_{1/2}$ of 50 G.y. Meisel *et al.* (1996) measured the Os-isotopic compositions of a suite of mantle xenoliths possessing varying degrees of fertility. Figure 8 shows the correlation between $^{187}\text{Os}/^{188}\text{Os}$ and two different indexes of fertility, Al content and Lu content. As Al and Lu approach the values we expect for the BSE, the Os-isotopic composition approaches chondritic. Thus, for the most fertile lherzolites, the time-integrated Re/Os ratio of these mantle samples has remained within $\pm 3\%$ of chondritic over the lifetime of the solar system.

It seems unlikely to us that differentiation in a magma

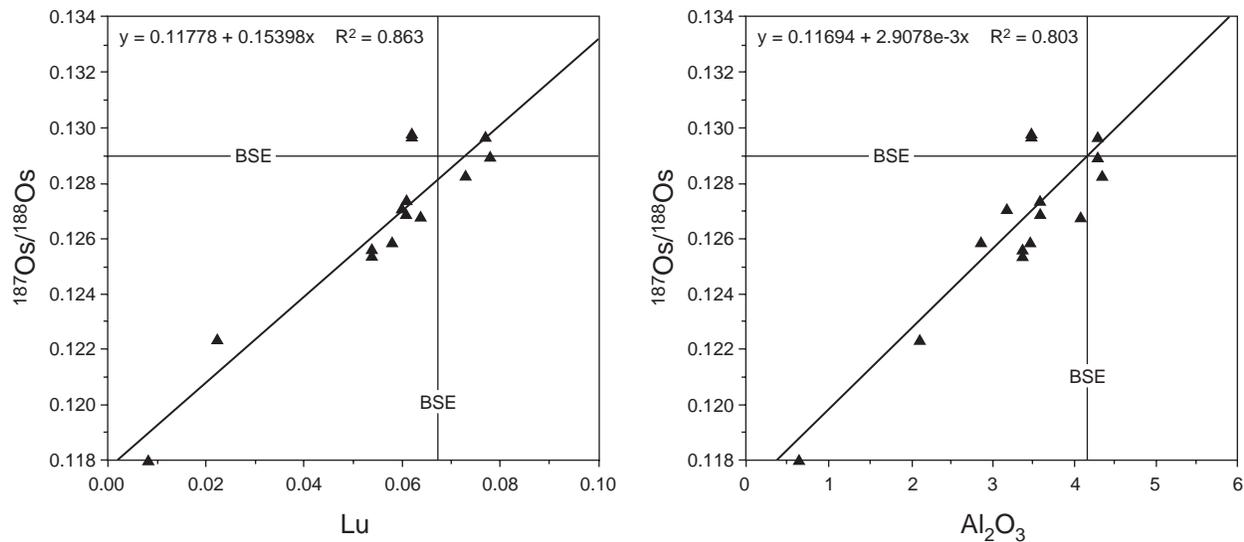


Fig. 8. Osmium-187/osmium-188 vs. Lu and Al_2O_3 in mantle peridotites. As mantle peridotites become more fertile and their Lu and Al abundances approach that of the BSE, their $^{187}\text{Os}/^{188}\text{Os}$ ratio approaches chondritic. For the most fertile peridotites, this means that their time-integrated Re/Os ratio has remained within $\sim\pm 3\%$ of chondritic. The $\text{Re}-\text{Os}$ system has apparently suffered minimal igneous processing over the age of the Earth. Again, this sheds doubt on whether the Earth ever experienced a magma ocean.

ocean and subsequent homogenization (perhaps by mantle convective processes over the age of the Earth) could have conspired to produce such an Os isotopic outcome. During basalt production Os behaves as a compatible element and remains in the residue, whereas Re behaves moderately incompatibly and is concentrated in the basaltic liquid. (Thus, the Os isotopic composition of the continental crust is quite radiogenic because the crustal Re/Os ratio is high.) Considering the ease with which Re and Os can fractionate during igneous processes, it seems more reasonable to us that the Re and Os in fertile lherzolites have never been much fractionated. We therefore assume that they were added as a late veneer (along with other highly siderophile elements) after core formation had ceased and were then mixed into Earth's mantle in the solid state.

In addition, another aspect of the *Meisel et al.* (1996) study appears to have gone underappreciated. The good correlations between the xenoliths' Os isotopic compositions and their Lu and Al concentrations appear to imply that the veneer was mixed into a mantle that was approximately chondritic and homogeneous (except for its previously removed metal component). Otherwise, if the late veneer had been mixed into a heterogeneous, nonchondritic mantle, there is no reason for the correlations seen in Fig. 8 unless later remixing were complete.

However, there is evidence that mixing of the veneer itself into the mantle was not as complete as was originally thought by *Jones and Drake* (1986). *Spettel et al.* (1991) have analyzed several suites of spinel lherzolites; and Ni and Ir analyses of these samples are shown in Fig. 9. Nickel/Ir ratios vary from sample to sample (which may reflect our ability to acquire a representative sample) and, more impor-

tantly, from locality to locality. Sample localities are designated by either solid or open symbols depending on the average Ni/Ir ratio of that locality. With few exceptions, individual samples from a given locality either have an Ir content of 2–3.5 ppb and a Ni/Ir ratio ≥ 600 or have an Ir content of 3.5–6 ppb and a Ni/Ir ratio ≤ 600 . Almost all variation in Ni/Ir in Fig. 9 is due to variability in Ir and is much larger than analytical uncertainty ($\sim 10\%$ relative). Because Ir acts compatibly during normal igneous processes, we take this variation to reflect the completeness of mixing of the late veneer (*Spettel et al.*, 1991). Although more work on this issue is needed, we take the *Meisel et al.* (1996) and *Spettel et al.* (1991) data to mean that the early terrestrial mantle was largely undifferentiated at the time the veneer was added. Mixing was apparently incomplete for Ir and, consequently, significant mantle heterogeneities that were present at the time the veneer was added are unlikely to have been totally erased by subsequent mixing.

But if the early mantle did differentiate, then it had re-mixed and was homogeneous by the time the Ir, Re, and Os were added. So when did addition of this late veneer occur? This is a difficult question to answer. Because Re and Os in fertile lherzolites have not obviously fractionated from their chondritic ratio, there is no event to date. However, we believe it unlikely that the veneer was added after the late heavy bombardment recorded on the Moon (~ 3.8 – 3.9 Ga). Remixing of Earth's mantle would have to have been complete by that time.

However, we see no evidence of large, early differentiation. We believe that, except for the removal of metal to the core, large regions of the mantle of the early Earth were left

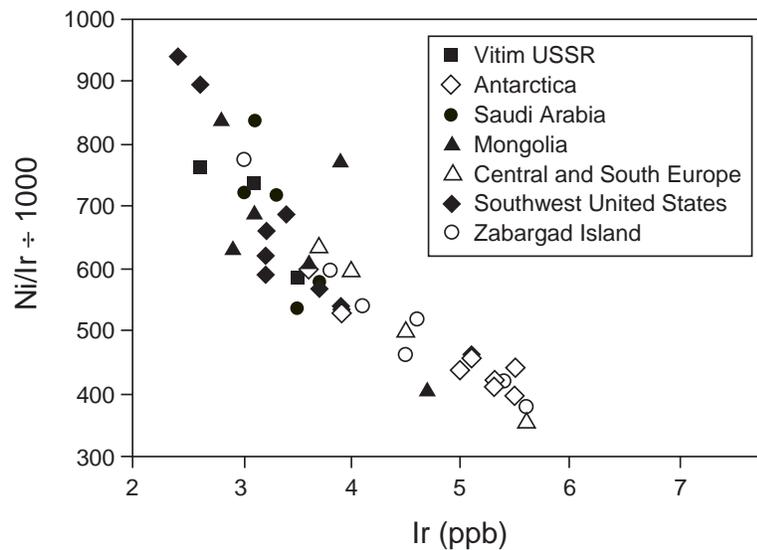


Fig. 9. Nickel/iridium vs. Ir for spinel lherzolites from several different localities (Spettel *et al.*, 1991). These localities are given in the legend. A given locality has been assigned either a solid symbol or an open symbol based on the average Ni/Ir ratio of that locality. Variations in Ni/Ir are almost entirely associated with variation in Ir. Individual samples do not span the entire range of the data and, consequently, the solid and open symbols are highly segregated. We interpret these data to reflect original heterogeneity from the mixing of the late veneer into the Earth's mantle (see text).

relatively unprocessed and that some remnants of this pristine mantle “survive” today. We use “survive” in quotations because even the most pristine mantle lherzolites have had some removal of highly incompatible elements, presumably by extraction of silicate liquid (e.g., Hirschmann *et al.*, 1998). These extractions have sometimes been minor enough that major-element abundances and the Re-Os system have not been much affected. This is presumably because Re and Os either reside compatibly in silicate (or oxide), as Os is presumed to, or reside in minor sulfides that are not very soluble in low-degree silicate partial melts.

Thus, if the Earth differentiated and was then remixed (to the hand-specimen scale), this remixing is required to be so complete that we cannot detect it. We see no clear evidence that either differentiation or remixing occurred. Clearly, if this general conclusion is correct, it is difficult for Earth to have passed through a magma ocean stage. And if the Earth did not have a magma ocean, we believe it is reasonable to question the giant impact hypothesis as well. As alluded to above, if some mantle samples have survived unprocessed and undifferentiated, we may have to not only rethink the giant impact hypothesis but the general problem of accretion in the inner solar system.

8. CONCLUSIONS

We have reviewed several important aspects of the chemical composition of the Earth and Moon. In summary: (1) It appears that the bulk compositions of the Earth and Moon are different. (2) The depletions of volatile alkali elements in the Moon cannot simply be the residues that remain following the partial volatilization of terrestrial materials. (3) Taken at face value, ^{182}Hf - ^{182}W systematics would indicate that the Moon is older than Earth. Or, stating it more precisely, core formation on the Moon must have predated core formation on Earth. (4) The presence of a

significant lunar core (1–3 wt%) has obviated the need for the Moon's metal component to reside in the Earth, even though the Moon must be depleted in Fe relative to chondrites. (5) It is not evident from our sampling of the terrestrial mantle that the Earth ever passed through a magma ocean stage.

Although none of these observations actually disproves the giant impact hypothesis, we find it disquieting that the obvious consequences expected of a giant impact are not observed to be fulfilled. In its original conception, the giant impact predicted that (1) the Moon's depletion in volatile elements occurred as a result of the impact (Wänke and Dreibus, 1986); (2) the Moon, being derivative from the Earth, should be younger; and (3) the thermal consequences of the impact would result in a terrestrial magma ocean (Melosh, 1990; Tonks and Melosh, 1990). Other simple predictions, such as that the Moon should have no core and that the Mg# of the Moon should match that of the Earth, were known or suspected to be unfulfilled, even at the time the giant impact hypothesis was proposed. However, it was believed at the time that contributions from the impactor might resolve these discrepancies. Contributions from the impactor are notoriously difficult to disprove. In their extreme limit, some impactor-contribution models may not even be testable. However, as we have shown above, an analysis that includes both consequences for Earth, as well as for the Moon, may preclude some models or render them unlikely.

Our current best estimate is that the Moon formed from material having nearly chondritic abundances of siderophile elements at ~35 m.y. after chondrites. We also favor a partial magma ocean for the Moon, rather than complete melting. The main problem with our overall scenario is that it does not accommodate most versions of the giant impact hypothesis very well. It may, however, be compatible with the circumterrestrial disk model of Weidenschilling *et al.*

(1986). In the Weidenschilling model, silicate is preferentially trapped in a circumterrestrial disk compared to iron metal. Metallic cores or pieces of metallic cores, being stronger and less easily disrupted than silicate, preferentially pass through the disk. Thus, when the disk finally accretes into a moon, that moon is depleted in metal.

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