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Earth-Moon System, Planetary Science, and Lessons Learned

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1. INTRODUCTION

The origin and evolution of the Moon is now well understood. Many bizarre theories were proposed to account for our satellite before these were swept away by the data from the Apollo missions. We now understand with the benefit of hindsight that we are looking at a very ancient object whose surface has changed little for the past 3.5 b.y.

Nevertheless the Moon remains a unique body in the solar system. The Moon formed as the result of a glancing collision with the Earth of a Mars-sized impactor late in the accretional history of the Earth when both bodies had differentiated into metallic cores and silicate mantles. The impactor is derived from the neighborhood of the Earth, as shown by similarities in O and Cr isotopes. Our satellite is mostly derived from the mantle of the impactor rather than from the mantle of the Earth. The Moon melted, perhaps entirely and a thick (50 km?) feldspathic crust, dated at 4460 ± 40 Ma· floated on the ocean of magma. The closest analogue may be the crust of Mercury. The interior crystallized into cumulate zones of differing mineralogy from which the mare basalts were later derived by partial melting.

The impact history of the lunar surface has provided insights into planetary accretion, the planetesimal hypothesis and the bombardment history in the inner solar system. Debate continues whether the concentration of impact-produced basins constitutes a "cataclysm" or the tail end of accretion. A notable example and a target for future missions is the 2500 km diameter South Pole-Aitken basin that may provide insights into the nature of the deep lunar crust.

Remaining significant problems include whether the Moon has a small metallic core, while the deep interior structure is not well understood. New seismic and heat flow data are an urgent priority. The depletion of the Moon in volatile elements is well established but may reflect earlier processes in the solar nebula rather than being due to the giant impact. Likewise,

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the depletion of siderophile elements and the Hf-W isotopic systematics may reflect processes in precursor planetesimals. Meanwhile, the bulk lunar composition for major elements (e.g., Mg, Al, Ca, Si) is not well constrained. Neither the Earth nor Moon resemble the composition of the primordial rocky component of the solar nebula (CI) nor that of the common meteorites. This is indicative of much fractionation in the inner solar nebula occurring during formation and accretion of planetesimals.

In the following sections we discuss the advances in lunar science, both from the Apollo results and particularly from the more recent Clementine ant Lunar Orbiter Missions. We discuss the place of the Moon in the solar system, and conclude with our assessment of how to go about exploring solid planets and satellites on the basis of lessons learned from experience with the Moon.

2. ADVANCES IN LUNAR SCIENCE OVER THE PAST 10 YEARS

2.1. Planetary geochemistry: the Moon in the context of the solar system

2.1.1. Processes in the early nebula. In the current model for the early evolution of the solar nebula, (e.g., Stevenson and Lunine 1988; see review by Taylor 2001a) gas, water and elements volatile <1100K were driven out along a "snow line" at about 5 AU and so were depleted in the inner nebula. This process was most likely associated with violent solar activity (e.g., T Tauri stage) in the earliest stages of nebular evolution. The related buildup of material at the "snow line" allows for the rapid (10^5 year) runaway growth of a 10–20 Earth mass core and enables the gravitational capture of gas before the nebula is dispersed, so allowing early growth of Jupiter. The cores of Saturn, Uranus and Neptune probably formed in the same region of enhanced density and were dispersed by the prior growth of Jupiter (Thommes et al. 1999). Thus the giant planets formed at a significantly earlier stage than the inner planets that accreted later from the surviving dry refractory planetesimals after the gaseous components of the nebula had been dissipated.

The dry and refractory nature of the planetesimals is reflected by the primitive anhydrous mineralogy of chondrites (Brearley and Jones 1998) and by the differing oxygen isotopic ratios of ice and the primary meteoritic minerals (Young and Russell 1998). Ice was present far out in the asteroid belt, and its subsequent melting produced the aqueous alteration present in the CI chondrites. The water now present in the Earth and Mars (and Venus?) accreted later to the planets either by the later drift back of icy planetesimals or from comets that had formed originally in the region of Jupiter (Cyr et al. 1998, 1999; Morbidelli et al. 2000).

Timescales for the formation of planetesimals were considered by Carlson and Lugmair (2000) who concluded that "melting and internal differentiation took place on a timescale of a few million years" after T_o (4566 Ma used here; 4570 Ma is an alternative). The implication of this is that the planetesimals accreting to form the terrestrial planets were already differentiated into metallic cores and silicate mantles, with far-reaching implications for the interpretation, for example, of the Hf-W isotopic system.

These planetesimals were subsequently assembled over the succeeding 10–100 m.y. into the four inner planets, with the Moon forming during a giant collision of a body about the size of Mars late in the accretional history of the Earth (Canup and Asphaug 2001; Canup 2004).

Relative to the elemental abundances in the primitive solar nebula, that are represented by the CI carbonaceous chondrites (hereafter CI), the Moon is more strongly depleted in volatile elements than the Earth and is completely dry (late addition from cometary impacts might account for the reported presence of ice at the poles: Nozette et al. 2001, but see Campbell et al. 2003). However this depletion is similar to that observed in the parent body of the HED meteorites (4 Vesta), and in the angrites. Much of the volatile element depletion and the bone-dry state of the Moon is often attributed to the high temperature conditions during the giant impact. However the low value of the lunar initial $^{87}Sr/^{86}Sr$ ratio, LUNI argues for very early loss of volatile Rb relative to refractory Sr, consistent with the early depletion of volatile elements in the solar nebula, rather than during the impact. Moreover, the lack of fractionation in the abundance patterns of the refractory element patterns to CI are not consistent with condensation from vapor at temperatures of several thousand Kelvin that certainly existed in some regions affected by the giant impact (Cameron 2000, his Figs. 7, 8) nor indeed of much exposure to temperatures above 1100K. The K isotopes show no fractionation either between the Moon and other inner solar system bodies, including meteorites, relative to CI, indicating that the Moon did not accrete by condensation from a hot impact-induced vapor (Humayun and Clayton 1995).

Thus the question arises whether the bone-dry and volatile depleted nature of the Moon is related to the giant impact, or is a relic from the primordial dry inner solar nebula, inherited from the impactor? "Harold Urey thought of the Moon as a primitive object, so it is ironic that the Moon may after all preserve some memory of the earliest nebula during the long journey of the precursor material from primitive dust grains to the curious composition of the Moon" (Taylor 2001b).

2.1.2. Lunar interior. The lunar radius is 1738 km, lunar density is 3.3437 ± 0.0016 g/cm^3 and the moment of inertia (I/MR^2) is 0.3931 ± 0.0002 (Konopliv et al. 1998). The seismic velocity model of Nakamura (1983) modeled upper, middle and lower mantle zones with constant velocities with boundaries at 270 and 500 km. Below about 1000 km, P and S-waves become attenuated. P-waves are transmitted though this region of high attenuation, but S-waves are missing, possibly suggesting the presence of a few percent of partial melt.

Khan et al. (2000) have provided a reinterpretation of the Apollo seismic data from an inversion of the lunar free oscillation periods. However, their lower mantle seismic velocities are higher than either those of Nakamura et al. (1983), Goins et al. (1981) or Kuskov et al. (2001). Until new seismic data become available, the detailed structure of the mantle will remain uncertain and this remains one of the most important measurements to be made in the future.

2.1.3. Lunar core. Although there is a common belief that the Moon possesses a small iron core, there is in fact no decisive evidence, and the moment of inertia, magnetic sounding, seismic data, thermal evolution, and lunar rotation could equally be interpreted as indicating a Moon with a silicate core. Indeed the free oscillation study of Khan and Mosegaard (2001) are consistent with a silicate core. Core-free models are also consistent with the moment of inertia value and the allowable density range in the lower mantle for a core-free Moon is 3.49–3.52 g/cm^3 (Kuskov and Kronrod 2000; Kuskov et al. 2001).

Data from the Lunar Prospector magnetometer have refined the estimate for the size of a metallic core to one with a radius of 340 km (Hood et al. 1999). The revised moment of inertia value places narrow limits: 220–370 km diameter for an Fe core and 350–590 km diameter for an FeS core (Konopliv et al. 1998). A metallic core is consistent with the estimate of the metallic iron abundance in the Moon of 2.5 ± 2% (Dyal et al. 1977), lunar magnetism (Hood et al. 1999) and the depletion of trace siderophile elements in order of their metal-silicate partition coefficients in the lunar mantle (Righter and Drake 1996; Righter 2002). This is evidence of metal segregation somewhere, but perhaps this occurred in precursor bodies prior to the formation of the Moon and the lunar mantle siderophile element abundances, with their evidence of metal-silicate fractionation, might thus be inherited.

The two current models for the interior of the Moon (Fig. 7.1) have implications for the evolution of the Moon. It seems difficult to form a metallic core without involving whole Moon melting, consisting, in the deepest mantle, mostly of olivine and orthopyroxene cumulates from the magma ocean. These are likely to contain few heat sources, making it difficult to account for the possible presence of partial melt.

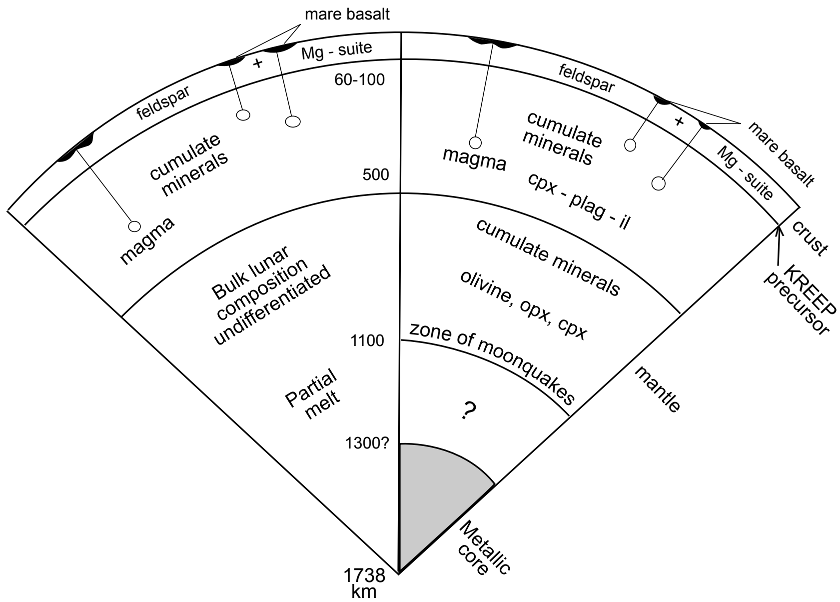

Figure 7.1. Two alternatives for the internal structure of the Moon. On the left, only half the Moon melted and differentiated and the deep interior has primitive lunar composition. Some partial melting has occurred due to the presence of K, U, and Th. This model is consistent with the lunar free oscillation periods (Amir Kahn, Univ. of Copenhagen personal communication). On the right, the Moon was totally melted and differentiated, forming a small metallic core (adapted from Taylor 2001).

In the alternative model, the absence of a core could imply half-Moon melting, with a primitive deep interior consisting of silicate. This might contain sufficient K, U and Th to generate some partial melt (1–2% is required), consistent with the geophysical interpretation. Thus our knowledge of the deep lunar interior and the question of a core must await further data. It is clear, however, that answers to these questions are crucial to understanding the origin, composition and early history of the Moon.

2.1.4. Lunar crustal structure. Geophysical mapping carried out by the Clementine and Lunar Prospector missions have resulted in greatly improved estimates of crustal thicknesses and topography (Hood and Zuber 2000) and in a better understanding of the strength of the crust. Figure 7.2 shows the mantle uplift beneath the Orientale basin. Similar mantle uplift is seen beneath lunar farside basins There is a significant negative correlation of age and mantle relief (Neumann et al. 1996, their Fig. 7b) suggesting that the mantle was somewhat weaker in earlier (pre-4.0 b.y.) times. However the relaxation is not complete and both surface topography and mantle uplift are preserved even in most of the ancient basins. It is clear that the crust has been strong and able to sustain such structures without much relaxation for over 4 b.y. This places constraints of the amount of crustal igneous activity after the solidification of the lunar crust. Figure 7.3 shows the uplift beneath the South Pole-Aitken Basin as well as that partially superimposed later uplift resulting from the excavation of the Apollo basin. The SPA structure, where most of the upper crust is missing has been preserved for over 4.1 b.y.

On the basis of the FeO and Th abundances measured by the Clementine and Lunar Prospector missions, Haskin (1998), Jolliff et al. (2000), Korotev (2000), and Wieczorek and Phillips (2000) divided the crust into three major terranes: (1) the Feldspathic Highlands Terrane (FHT) (2) the Procellarum KREEP Terrane (PKT) and (3) the South Pole-Aitken

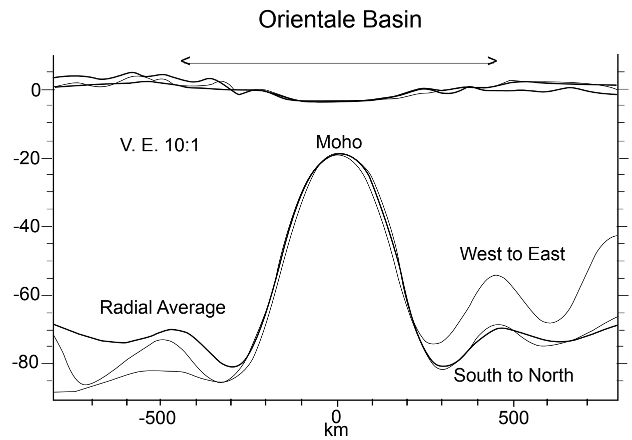

Figure 7.2. Average radial profiles of surface topography and Moho depth, assuming a 1.7 km thick mare load for Mare Orientale (adapted from Hood and Zuber 2000, Fig. 2).

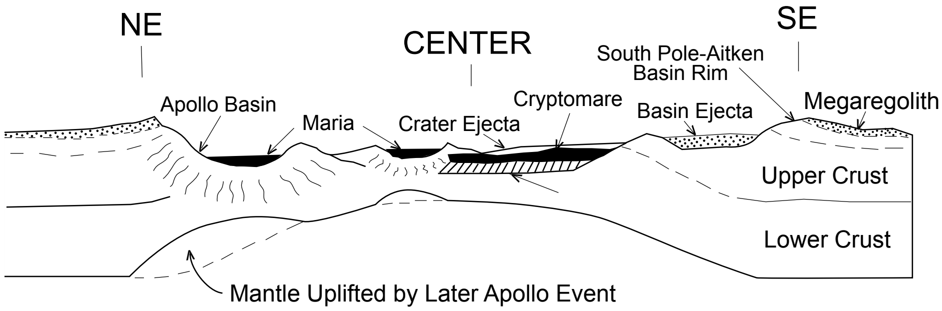

Figure 7.3. Cross-section of the South Pole-Aitken Basin (adapted from Pieters et al. 1997, Fig. 2).

(SPA) Terrane. These represent (1) the formation of the feldspathic lunar crust by accumulation from the magma ocean, (2) intrusion into the crust (or mixing into the crust and underlying mantle) of the residual KREEP liquid from the last stages of crystallization of the magma ocean and (3) the subsequent excavation of the ca. 2500 km diameter South Pole-Aitken basin, an event that stripped off most of the upper crust over that region and whose ejecta contributed significantly to the thickness of the farside anorthositic crust, north of the basin (FHT,A; Plate 1, Jolliff et al. 2000) (Figs. 7.4, 7.5).

2.1.5. Crustal thickness. Revisions of the Apollo seismic data indicate a thickness of 45 km (Khan et al. 2000) at the Apollo 12 and 14 sites rather than the earlier estimates of 60 km. The farside crust is thicker and Wieczorek (pers comm) estimates an average crustal thickness of 52 km (8.7% of lunar volume). The thickness of the lunar crust and the abundance of anorthosite in the crust are critical parameters constraining the bulk composition of the Moon. The plagioclase-rich upper crust extends to a depth of at least 30 km (from the central peak data, Wieczorek and Zuber 2001) so that decreasing the crustal thickness increases the relative proportions of the upper Al_2O_3 rich crust.

2.1.6. Siderophile element depletion. The depletion of Fe in the Moon is accompanied by a depletion of all siderophile elements (Righter and Drake 1996) that is correlated with

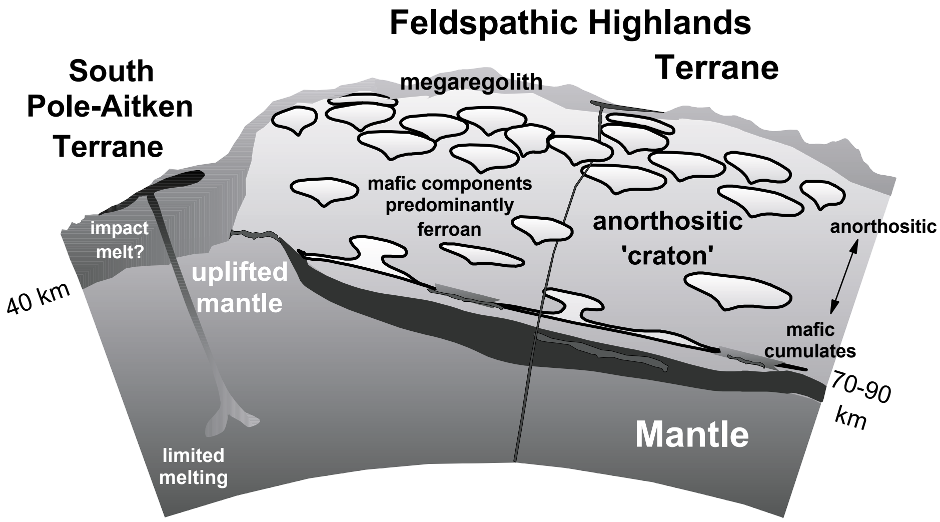

Figure 7.4. Schematic cross section of the Feldspathic Highlands Terrane (FHT) and the South Pole-Aitken Terrane (SPA) (after Jolliff et al. 2000, Fig. 13).

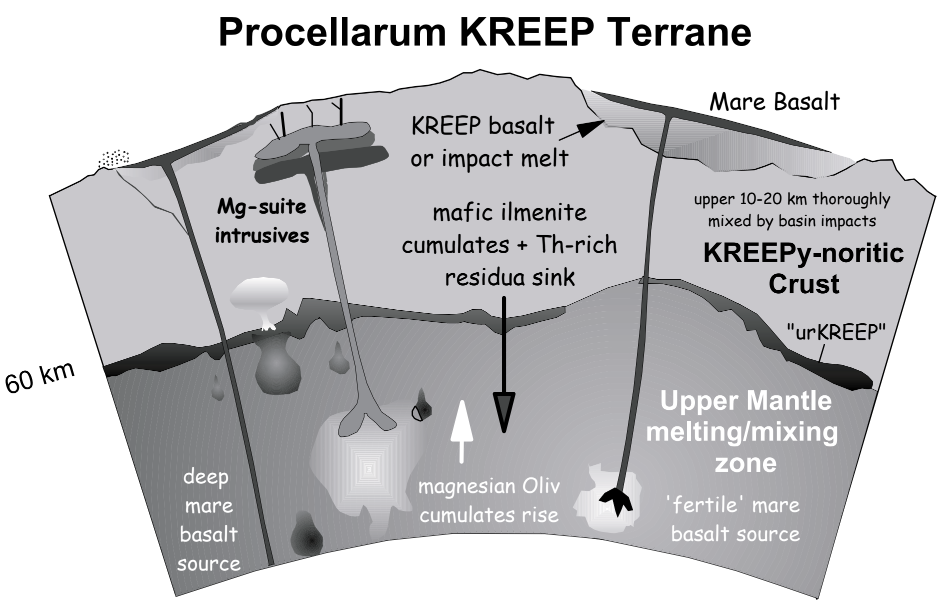

Figure 7.5. Schematic cross section of the Procellarum KREEP Terrane (PKT) (after Jolliff et al. 2000, Fig. 12).

increasing metal-silicate partition coefficients (Co-W-Ni-Mo-Au-Pd-Re-Ir) (Fig. 7.6), but not with increasing volatility (Re-W-Ir-Mo-Pd-Au-Co-Ni) (Newsom 1986).

In the current model for lunar origin, the impactor core was accreted to the Earth following the collision. Various scenarios are possible: (a) the depletion of the siderophile elements occurred during core formation in the impactor or other precursor bodies and so predates the evolution of the Moon. Dates derived from W-Hf isotopic systematics may be difficult to assess in this scenario. (b) Melting of the Moon during accretion resulted in the formation of a small core which incorporated the remaining lunar budget of siderophile elements in order of their metal-silicate partition coefficients (c) Only half the Moon was melted and metal segregation under strongly reducing conditions accomplished the depletion of the remaining siderophile elements in the upper mantle, no core being formed.

2.1.7. The highly volatile elements. The Moon is dry, with less than one ppb water, except for some possible amounts trapped in permanently shadowed craters at the lunar south pole (Nozette et al. 1994; Simpson and Tyler 1999; Feldman et al. 1998, 2001, but see Campbell et al. 2003). Rare gases are indeed rare on the Moon. Those present in the lunar samples either originate from less volatile radiogenic parents or are derived from the solar wind (Ozima and Podosek 1983, Swindle et al. 1986). The absence of any indigenous noble gases, like that of indigenous H_2O, in the Moon is a first-order observation.

The highly volatile elements Bi, Tl, Cd, Br, Se, Te and In, condense below about 800 K (at 10^{-3} atm). They are depleted in the Moon by large factors relative to CI abundances and by factors of about 40 relative to their abundances in the Earth (Fig. 7.7). Many meteorites also record the depletion of the volatile elements. This depletion roughly correlates with condensation temperatures (e.g., Bi, Tl; see Wulf et al. 1995).

2.1.8. Volatile elements. There was a general depletion of volatile elements throughout the inner nebula before the formation of the inner planets (Taylor 2001a). (Fig. 7.8). Venus and the Earth have similar K/U ratios, but these are a factor of six lower than that for C1. The SNC meteorites from Mars have higher K/U ratios, indicative of a volatile element budget about 50% higher than that of the Earth, but Mars is still strongly depleted in comparison with the primordial solar-nebular values. However, the potassium isotopes show no variation in inner solar system materials thus showing no evidence of exposure to Raleigh fractionation processes involving evaporation or condensation (Fig. 7.9).

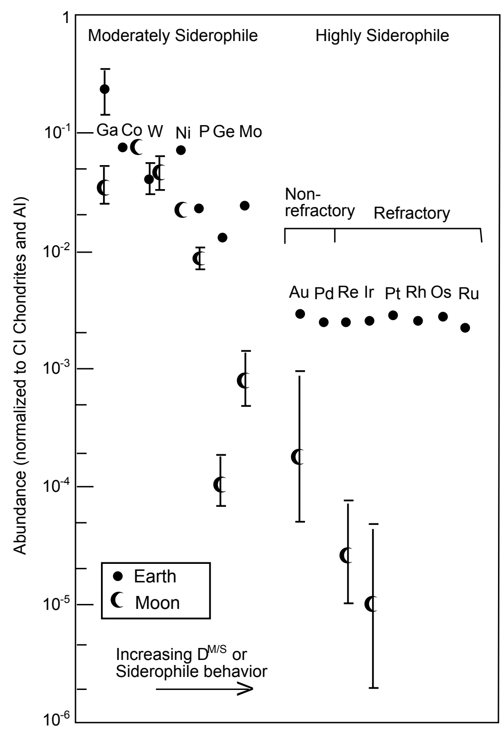

Figure 7.6. The depletion of the siderophile elements in the Earth and the Moon, relative to CI abundances (normalized to refractory Al), plotted in order of increasing metal/silicate partition coefficients. The greater depletion of Ga and Mo in the Moon is due the fact that they are also volatile elements. Courtesy H. E. Newsom, University of New Mexico (adapted from Taylor 2001).

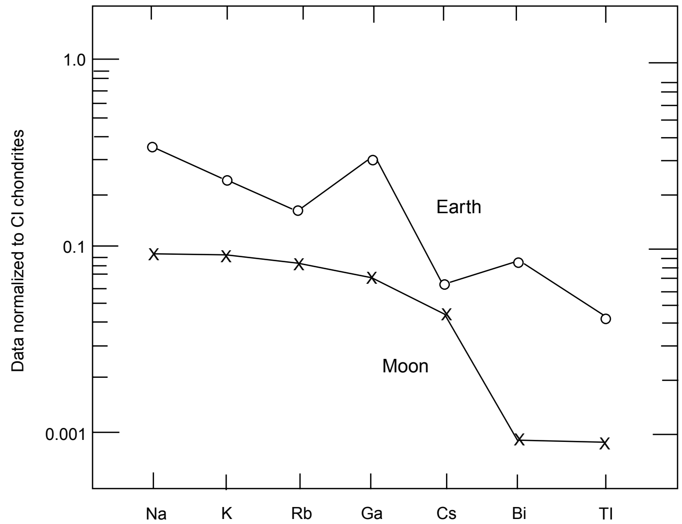

Figure 7.7. The depletion of volatile elements in the Earth and Moon, relative to the abundances in carbonaceous chondrites (CI).

Another moderately volatile element is Mn, which has the useful property that ^{53}Cr is derived in part from short-lived ^{53}Mn. The Earth and Moon have similar $^{53}Cr/^{52}Cr$ ratios. These contrast with higher values in meteorites (Lugmair and MacIsaac 1995) suggesting that ^{53}Mn was depleted relative to refractory Cr in the inner solar system and that there was a heliocentric gradient in Mn abundances.

Basic questions hang over the issue of volatile-element depletion in the Moon. These include where and when this depletion occurred and whether there was more than one stage of element loss. As noted earlier, it is tempting to ascribe the depletion in the Moon, relative to the Earth, as resulting from the giant Moon-forming impact; however, Venus, Earth, and Mars are all depleted, although not to the same extent as the Moon. Basaltic meteorites from Vesta, a very early object, show similar depletions of volatile and siderophile elements to lunar basalts. This, coupled with the low initial strontium isotopic values in the Moon, and the absence of evidence for high temperature (>1100 K) processing of proto-lunar material may indicate that most loss of volatiles occurred in the precursor events and that the Moon mostly inherited its dry volatile-depleted nature.

2.1.9. Vanadium, chromium and manganese. These elements have often been assigned genetic significance because of the claimed similarity between the terrestrial and lunar abundances Indeed, Cr isotopes ($^{53}Cr/^{52}Cr$) are similar between the Earth and Moon (Lugmair and Shukolyukov 1998), but much of this debate becomes irrelevant if most of the Moon is derived from the silicate mantle of the impactor, rather than from that of the Earth. An extensive discussion is provided by Jones and Palme (2000) that illustrates the many uncertainties inherent in trying to estimate V, Cr and Mn abundances in the Earth and Moon.

Ruzicka et al. (1998, 2000) note that "terrestrial volcanic rocks have one to two orders of magnitude lower Cr abundances than do mare basalts or eucrites." Much of this difference is probably due to the differences in redox state between the Earth and Moon with most Cr being trivalent on Earth compared with the divalent state of Cr in the Moon (Papike and Bence 1978). Ruzicka et al. (1998, 2000) have pointed out that the supposed similarity between the Earth and the Moon for V, Cr and Mn is based on model-dependent assumptions about bulk composition and conclude that "contrary to general belief, Cr data do not provide good evidence for a chemical link between Earth and Moon." (Ruzicka et al. 1997, p. 857) (Fig. 7.10). This figure shows the clear differences between the Moon and HED parent body on the one hand, and Mars and the Earth on the other (Taylor and Esat 1996). Papike (1998) has noted the increase in Mn/Fe in olivines and pyroxenes with distance from the Sun, presumably due to the volatile nature of Mn (see comments on Cr isotopes).

This apparent similarity between the composition of the Moon and that of the HED parent body (4 Vesta) raises the question whether such bodies that are depleted in volatile and siderophile elements were more common in the early solar system than previously supposed.

2.1.10. Refractory lithophile elements. There has been a longstanding controversy over the question whether the Moon is significantly enriched in refractory elements relative to the Earth. Fig. 7.11 shows one assessment (Taylor 1982) (see also section 2.1.12). The super-refractory elements Zr, Hf, Y, and Sc do not appear to be fractionated in the Moon relative to the other refractory elements, in contrast to their behavior in some meteorite minerals (e.g., hibonite) and in CAI refractory inclusions where they may be separated from the lesser refractory elements such as the REE.

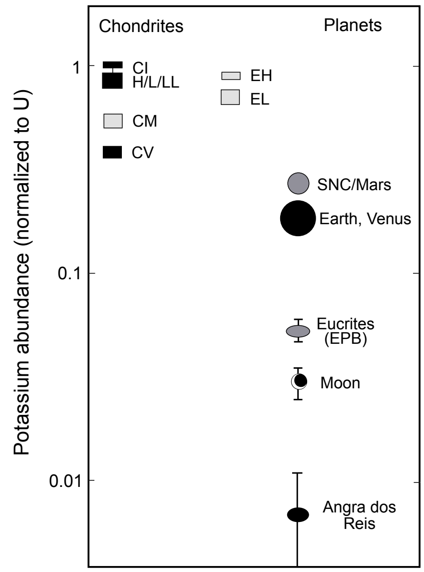

Figure 7.8. The depletion of potassium (normalized to CI and U, a refractory element) in various chondritic meteorites and solar system bodies. Courtesy Munir Humayun and Robert Clayton, University of Chicago (adapted from Taylor 2001).

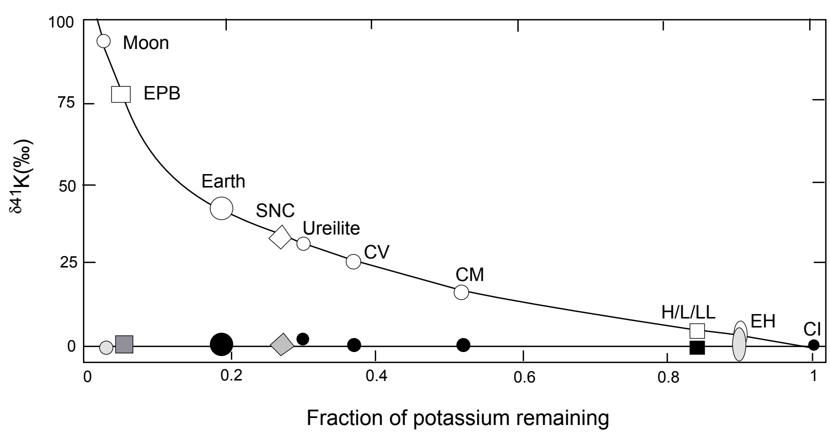

Figure 7.9. Potassium isotopic compositions for planets and meteorites. The bottom line shows the values measured. The top curve shows the calculated values that would be expected if the potassium depletion were due to loss by evaporation, beginning with a CI composition and assuming Raleigh distillation. Courtesy Munir Humayun and Robert Clayton, University of Chicago (adapted from Taylor 2001).

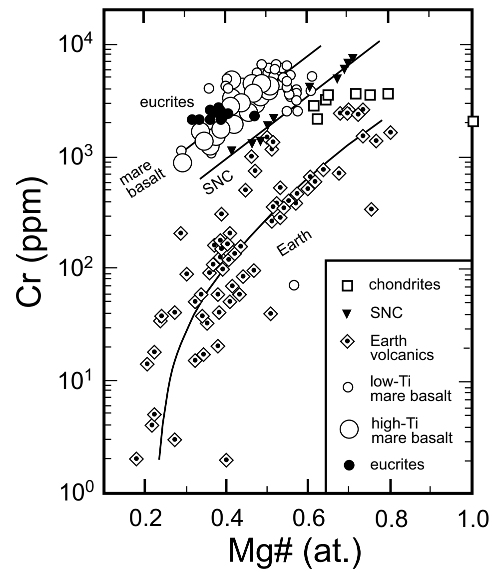

Figure 7.10. The relationship between the abundance of Cr and the Mg# (Mg/Mg + Fe) for mare basalts, eucrites and terrestrial basalts, compared to chondrites. This shows the similarity between eucrites from 4 Vesta and the Moon and their significant difference in Cr content from that of the Earth. Courtesy Alex Ruzicka, University of Tennessee, Knoxville (adapted from Taylor 2001).

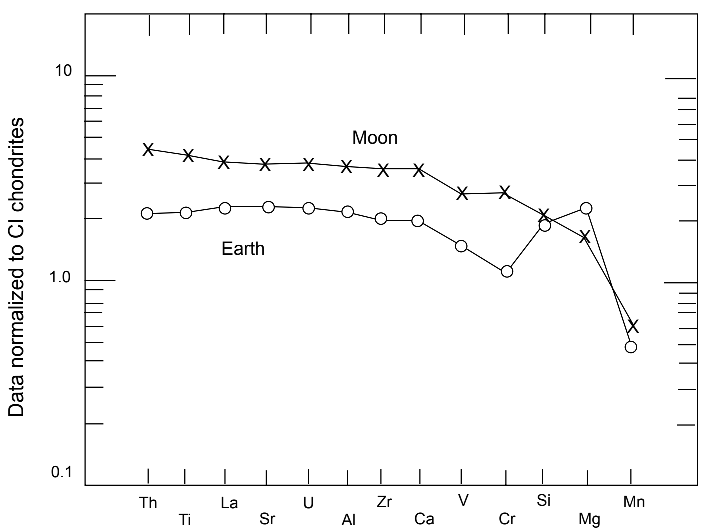

Figure 7.11. The abundances of the refractory lithophile elements in the Earth and Moon, plotted relative to the abundances in CI chondrites.

The most extreme conditions are recorded by the refractory inclusions (CAI) in meteorites. These show depletions and enrichments within the REE group based on relative volatility (Fig. 7.12). In the extreme case of the Group II inclusions, both the most volatile (Eu, Yb, Ce) and the most refractory (Sc, Y, Lu, Er) are depleted, presumably as a consequence of repeated cycles of evaporation and condensation.

In contrast, the bulk planetary REE patterns of the Earth and Moon show no depletions or enrichments (except those related to crystal-liquid fractionation) and are depleted only in

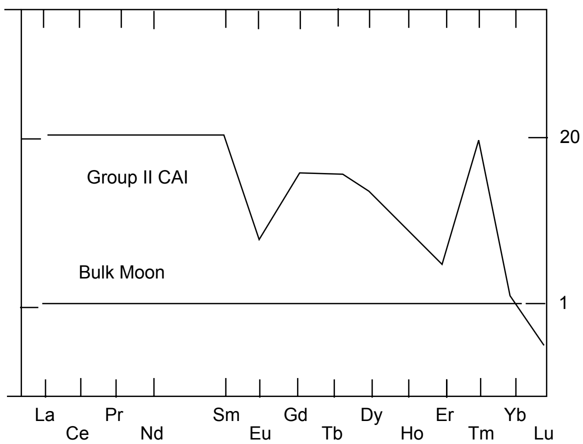

Figure 7.12. The abundances of the REE in the Moon and in Group II CAI plotted relative to CI.

elements more volatile than the REE. Thus the REE patterns in the Moon show no sign of loss of the more volatile members of the group. These span a considerable range in volatility from moderately refractory (e.g., Eu, 1290 K) to very refractory elements (e.g., Er, 1676 K). [The temperatures given are for 50% condensation at 10^{-3} atm from Fegley (1986) and Larimer (1988)]. The lunar REE patterns are mostly sub-parallel to the CI abundances, with minor excursions attributable to crystal-liquid fractionation in igneous melts within the Moon (Fig. 7.13). The most striking deviations are the enrichments and depletions of Eu, due to its separation from the other trivalent REE. This effect occurs in the highly reducing lunar environment, where the smaller Eu^{2+} ion mimics the geochemical behavior of Sr^{2+}. We can thus attribute the behavior of relatively volatile Eu in the Moon to crystal-liquid equilibria because Yb, of about the same volatility, shows no difference from its neighbors on chondrite-normalized plots.

The most volatile of the lithophile refractory trace elements are Sr (1275K) and Ba (1227K). (The temperatures listed are for 50% condensation at 10^{-3} atm.) However, these elements are not depleted relative to the other refractory elements. It thus appears that all elements with condensation temperatures above 1100–1200 K (at 10^{-3} atm) are present in their cosmic abundance proportions. Thus the material now in the Moon was apparently not subjected to temperatures in excess of 1100–1200 K (i.e., was not vaporized) (Taylor 1983, 2001b).

2.1.11. Uranium and thorium abundances. Uranium and thorium are refractory elements and so their abundances correlate with those of Al, Ca and Ti. The terrestrial U abundance of about 20 ppb is tightly constrained by an interlocking set of isotopic and chemical abundance ratios (K/U, K/Rb, Rb/Sr, Sm/Nd).

The bulk lunar alumina value (6%) of Taylor (1982) implies a lunar uranium value of 33 ppb. The value of 6.9% alumina of Kuskov (1997) yields a bulk lunar uranium abundance of 38 ppb. Another way to get at the refractory element abundances is via the lunar K/U ratio that averages about 2500. Drake (1996) gave a minimum estimate of 27 ppb uranium based on the assumption that all the lunar potassium was concentrated in the highland crust. This value must be a minimum for two reasons. First, there is enough potassium and uranium retained in the deep interior to provide the typical values of 500 ppm potassium and 200 ppb uranium in mare basalts. Secondly, the bulk of the potassium and uranium in the Moon ends up being concentrated in the final residual melt from the crystallization of the magma ocean, that is trapped beneath the anorthositic crust.

From the Lunar Prospector and sample data, Jolliff et al. (2000, 2001) and Haskin et al. (2001) estimate that the bulk Th value for the Moon is 0.142 ppm, compared to a value of 0.125

ppm (Taylor 1982) that for Th/U = 3.8 gives a U value of 37 ppb. Although bulk lunar uranium values as low as 19 ppb (Rasmussen and Warren 1985) have been proposed, these fail to provide enough uranium (or other refractory elements) for geochemically reasonable lunar models (see detailed discussion by Taylor 1986a).

However, U, Th, K and the other incompatible elements in KREEP appear to be concentrated in the Procellarum KREEP Terrane. The implication of this strong segregation is that little information about the bulk Moon abundances for any of these elements can be derived either from returned samples or from remote sensing measurements across the lunar surface. Integration of global data and assessment of variations with depth, as can be done with crater ejecta, are required. The same comment applies to local heat flow measurements, unless a major lunar grid was established, or some other technique for establishing the global heat flux is developed. Estimates for the abundances of the trace refractory lithophile elements may be best constrained by their correlation with major refractory elements such as Al, Ca and perhaps Ti.

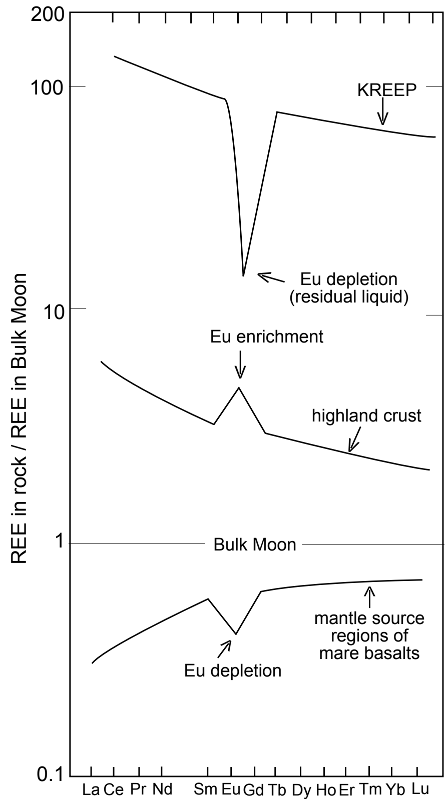

Figure 7.13. The abundances of the rare earth elements in the source regions of the mare basalts, the highland crust and KREEP, relative to the bulk Moon concentrations. These patterns result from the preferential entry of divalent europium (similar in radius to strontium) into plagioclase feldspar. This mineral floats to form the highland crust, and so depletes the interior in Eu. Mare basalts subsequently erupted from this region deep within the Moon bear the signature of this earlier depletion in Eu. KREEP is the final residue of the crystallization of the magma ocean. It is strongly depleted in Eu due to prior crystallization of plagioclase, and is enriched in the other rare earth elements, and other trace elements (e.g., K, U, Th, Ba, Rb, Cs, Zr, P) that are excluded from olivine, pyroxene and ilmenite during the crystallization of the major mineral phases of the magma ocean (adapted from Taylor 2001).

2.1.12. Estimates of bulk lunar compositions of the refractory elements. As noted above, these elements are not fractionated within the group, except in the extreme cases recorded in the CAIs, an accurate determination of the bulk lunar composition of one (e.g., Al) will yield values for the others. The essential point about the bulk lunar composition is that it (a) must contain concentrations of Ca and Al that are sufficiently high to account for the thick feldspathic crust (about 8% of lunar volume) and (b) must allow plagioclase to have precipitated early enough to extract Sr and Eu from the mare basalt source region. Data from the Clementine Mission established the alumina-rich composition of the highland crust, requiring that relative to the Earth, there is an enrichment of refractory elements (e.g., Ca, Al, Ti, REE, U, Th) in the Moon (e.g., Lucey et al. 1995).

Another way to get at the problem was carried out by Hood and Jones (1987) and by Kuskov and coworkers (e.g., 1997, 1998). These workers calculated the phase relations in the lunar mantle, using the constraints of the lunar moment of inertia, density and seismic velocities, using the interpretation of the seismic velocity data by Nakamura (1983). Kuskov and Konrad (1998) then used this thermodynamic modeling to determine bulk composition, density and temperature distribution in the lunar mantle. They concluded that the bulk alumina content of the silicate portion of the Moon lies between 5.5 and 7.0% and that "the Moon is depleted in MgO and is enriched in refractory elements compared to the terrestrial mantle" and "bears no genetic relationship to the terrestrial mantle as well as to any of the known chondrites" (Kuskov and Konrad 1998, p 302).

The Moon has 50% more FeO than the Earth's mantle and in current models, is derived from the mantle of the impactor. But the terrestrial planets themselves vary in composition, the Earth having higher Al/Si and Mg/Si ratios than CI (Drake and Righter 2002). Thus the accretion of the inner planets did not sample a uniform population of planetesimals and these did not match the compositions of the common classes of meteorites. Improved lunar seismic data and heat-flow measurements are needed to resolve all these questions.

However, the question turns critically on the thickness and alumina content of the highland crust and on the depth of melting of the magma ocean and the bulk lunar composition for Al, Ca, Mg and Si remains uncertain. The issue will remain open until these questions are resolved. If the Moon turns out to be significantly enriched in refractory elements, this produces a difficult cosmochemical problem. No adequate mechanisms have yet been proposed to produce such an enrichment.

2.1.13. Oxygen isotopes. The Moon and Earth are indistinguishable on an oxygen three-isotope plot (Clayton and Mayeda 1975; Wiechert et al. 2001), but this feature is shared by the enstatite chondrites (Clayton et al. 1984) (Fig. 7.14). Both on this account, and their extremely reduced nature, enstatite chondrites are favorite candidates for forming the Earth (Javoy 1995). However, their content of volatile elements, their high K/U ratios, and low Mg/Si and Al/Si ratios, make it quite certain that neither the Earth nor the Moon is made of enstatite chondrites. Thus it is probably coincidental that these three objects share the same oxygen isotopic composition. Similarity does not imply identity, a well-known philosophical trap. The similarity between the Earth and Moon in oxygen isotopes has sometimes been used to argue for derivation of the Moon from the Earth. However the probable explanation is that this oxygen isotopic composition is characteristic of material in the primitive solar nebula during the accretion of the Earth. Thus the oxygen isotopic similarities between Earth and Moon, although essential

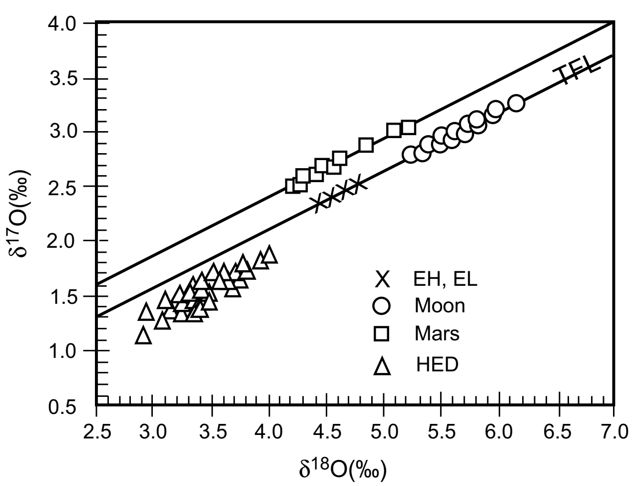

Figure 7.14. The oxygen isotope plot for the Moon, Mars HED and EH and EL meteorites. The data for the Moon plot on the terrestrial fractionation line (adapted from Wiechert et al. 2001, Fig. 2).

in providing motivation for searching for the impact-trigger hypothesis (Hartmann 1986), are by no means definitive proof of a terrestrial origin of the Moon.

2.1.14. Magma oceans. The thick anorthositic crust, the reciprocal Eu anomalies in the crust and the source regions of the mare basalts and the strong near-surface concentration of many incompatible elements, are consistent with large-scale melting of the Moon. The abundances of the incompatible elements in the highland crust require concentration from a large volume of the Moon, are consistent with enrichment due to crystal-liquid fractionation and are not related to volatility (Taylor 1973). This observation ruled out earlier suggestions that the refractory nature of the Moon was due to heterogeneous accretion (Papanastassiou and Wasserburg 1971; Gast 1972). In magma ocean models, KREEP forms the residual melt layer sandwiched between the top of the cumulate pile and the bottom of the anorthositic crust (Taylor and Jakes 1974). However KREEP, with its very high abundances of incompatible elements, including the REE, Zr, Hf, Ba, U, Th, and other refractory lithophile elements, appears to be segregated in the Procellarum KREEP Terrane on the lunar near-side (Jolliff et al. 2000).

The decisive test of the magma ocean hypothesis is the abundance of anorthosite in the lunar crust. The Clementine data show that very large areas of the crust "are composed exclusively of anorthosite as predicted by the magma ocean hypothesis" (Lucey et al. 1995). A lengthy paper on this subject by O'Hara (2000) disputes the magma ocean hypothesis. Crucial to his argument is his statement that "there is no positive europium anomaly in the average lunar highland crust" (O'Hara 2000 p.1545, see also p 1551) and this leads him to deny the existence of a magma ocean from which the crust formed by plagioclase flotation. This claim that there is no overall enrichment of Eu in the average highland crust is based on the data from the Apollo 16 site as interpreted by Korotev and Haskin (1988), but the Apollo 16 site is more mafic than much of the feldspathic highlands.

Subsequent extensive coverage of the farside highlands by the Clementine and Lunar Prospector missions has revealed that they are dominated by Ca-rich anorthosite, as discussed by Lucey et al. (1995). This highly feldspathic nature of the highlands is reinforced by studies of the lunar farside (e.g., Tompkins and Pieters 1999). Although recent data have led to new estimates of surface Al_2O_3 contents as high as 30% (Lucey et al. 1995; Wieczorek and Zuber 2001), an estimate of the integrated crustal Al_2O_3 content by Jolliff and Gillis (2002) of 25% is very similar to the earlier estimate of 24.6% (Taylor 1975).

All basalts derived from the lunar interior display a depletion in Eu, traditionally ascribed to an overall depletion of their source regions by the early crystallization of olivine and pyroxene, which exclude Eu^{2+} relative to REE^{3+}, and by the later crystallization of plagioclase, which concentrates Eu^{2+} relative to REE^{3+}, now in the highland crust (McKay 1982; McKay et al. 1990). Even the samples that come from the deepest mantle sources (probably ~400 km) have been involved in the magma ocean melting. The lunar orange and green glasses are among the most primitive lunar samples available. They possess low $^{238}U/^{204}Pb$ ratios and this is often cited as evidence that they represent samples of the primitive lunar interior. However, the glasses all possess the small but diagnostic signature of depletion in Eu that indicates that they come from a fractionated source (Papike et al. 1998). A sole exception is some USSR Luna 24 VLT basalt samples where only milligram-size amounts were available, raising serious questions about how representative were the samples that were analyzed (Neal and Taylor 1992).

The lunar highland crust that has been sampled can be interpreted as a mixture of a two component system, anorthosite and KREEP (or rocks whose incompatible element signatures are dominated by KREEP), as is apparent from the REE patterns (Fig. 7.15). If serial magmatism was the process responsible for forming the lunar crust (Walker, 1983), or if multiple igneous events occurred during the formation of the highland crust, one might expect to see the wide diversity of REE patterns that are observed in terrestrial igneous rocks, not the monotonous regularity that is observed in the REE patterns of rocks from the lunar highlands

There is thus general agreement (e.g., Warren 1985 and earlier chapters of this book) that the magma ocean concept remains the most viable hypothesis to explain the geochemical evolution of the Moon.

2.1.15. Depth of melting. Two possible alternatives for the initial depth of the magma ocean are a depth of 500 km (about half lunar volume) or total melting of the Moon. In the first case, the deeper parts of the Moon are composed of undifferentiated material of bulk lunar composition that might contain enough K, U and Th to induce a small (1–2%) partial melting apparently seen in the seismic models; in the second case, they will consist of the earliest phases to crystallize from the magma ocean (Mg-rich olivine and orthopyroxene cumulates) and a small metallic core. These regions are likely to be barren of the heat-producing elements. The Apollo seismic velocity data do not clearly distinguish between these two models and new seismic data are required to resolve this issue.

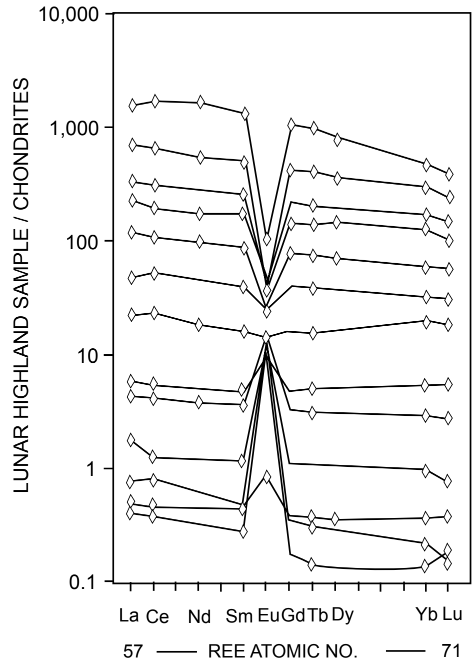

Figure 7.15. The abundances of the REE in a wide variety of highland samples, normalized to CI. There is a smooth transition from high values in KREEP with large depletions in Eu, to low concentrations in ferroan anorthosites which show strong enrichment in Eu due to its entry into plagioclase. The parallel nature of the patterns argues for a uniform petrogenesis. (adapted from Vaniman et al. 1991, Fig. 8.8).

There is very little sign of any features on the lunar surface that can be ascribed to either expansion or compression. This has been used to constrain the initial depth of melting of the Moon to about 200 km (Solomon and Chaiken 1976). However, their limits of ±1 km change to lunar radius apply only after the end of the massive bombardment at about 3.8 Ga that is responsible for the major morphologic features of the lunar surface (DE Wilhelms, pers. comm. 1998). Another assessment of the problem reveals that "the Moon could have been initially >50% molten (with the remainder relatively close to the solidus) and yet experienced little volume change over the last 3.8 b.y." (Kirk and Stevenson 1989). In attempting to specify the initial thermal state of the Moon (molten or partially molten), Pritchard and Stevenson (2000) concluded that the widely cited restriction of Solomon and Chaiken (1976) on the depth of the magma ocean is "insufficiently precise to infer a useful constraint." The other major constraint on the depth of melting comes from the requirement in the magma-ocean hypothesis to generate the thick feldspathic crust. The revised crustal thickness models do not provide fresh constraints on the depth of melting except that deep melting of at least half the Moon is still required.

2.1.16. Mare basalts. Although the familiar dark patches of mare basalt cover 17% of the lunar surface, their volume is trivial (Head 1976). The thin sheets of the visible maria constitute only 1% of the volume of the lunar crust or <0.1% of the Moon. Fragments of intrusive gabbros are rare in highland samples. Thus even allowing for subcrustal intrusions and cryptic maria, the lunar basalts amount at most to 1% of the Moon. Over 25 distinct varieties are known implying derivation from local regions. Accordingly the thermal requirements for the melting of the mare basalts are modest and melting must be very localized. Although there are many disputes in detail, there is broad agreement that the mare basalts were derived from zones of cumulate minerals that crystallized from the magma ocean.

Snyder et al. (2000) discuss the chronology of and isotopic constraints on the evolution of the Moon. Many complexities arise in detail but the limited sampling must engender caution. They note that "only a return to the Earth's Moon and return of samples from previously unsampled regions will allow us to answer many of the outstanding questions in lunar petrology and geochemistry." The lack of field control and the complex nature of the highland breccias make it difficult to distinguish among local (or trivial) and regional processes.

2.2. Discussion of ideas about the origin of the Earth-Moon system

2.2.1. Pre-Apollo ideas. It is difficult now, in the light of our present knowledge, to recall the state of ignorance existing before the Apollo missions cast their revealing light. Before spacecraft exploration, facts about the Moon were restricted to information about the lunar orbit, angular momentum and density. Speculations about composition and origin were unconstrained. Statements that running water carved the lunar rilles, that many of the large craters were volcanic calderas, that ice was present in a permafrost layer, that the maria were sediments, were full of dust, or were a few million years old, that the lunar highlands were composed of granite or covered with volcanic ash-flows, that a 'lunar grid' reflected tectonic stress patterns, that tektites came not only from the Moon, but from the crater Tycho, and that the Moon was essentially a primitive undifferentiated object, were all published in the serious scientific literature.

The Apollo data immediately swept away the earlier speculations. The maria were shown to be composed of basaltic lava flows, derived by partial melting from the mantle, with ages between 3–4 Ga. The postulated calderas were all due to meteoritic impact, the lunar highlands were mainly composed of anorthosite, not granite, the highland plains were ejecta sheets from large basin impacts, not siliceous volcanics, the 'lunar grid' was an artifact of the overlapping basin ejecta patterns, and tektites were terrestrial. Thus in view of our fundamental ignorance about the nature of the Moon, it is not surprising that none of the pre-Apollo theories of lunar origin survived.

The origin of the Moon had been long debated (for a review see Brush 1988, 1996). Pre-Apollo models for the origin of the Moon can be grouped into four separate categories that included (1) capture from an independent orbit, (2) fission from a rapidly rotating Earth, (3) formation as a double planet and (4) disintegration of incoming planetesimals. None of these notions survived the tests imposed of the returned lunar samples.

Hypotheses in which the Earth captured an already formed Moon have severe inherent dynamical problems and provide no obvious explanation for the exotic lunar geochemistry.

Fission hypotheses that derive the material for the Moon from the Earth's mantle encounter two basic difficulties (a) the angular momentum of the Earth-Moon system, although large, is insufficient by a factor of about 4 to allow for rotational fission (Durisen and Gingold 1986) and (b) there is no acceptable mechanism for removing this excess angular momentum following lunar formation. Despite the prediction that the chemistry of the Moon should bear some recognizable signature of the terrestrial mantle, the Moon contains, for example, 50% more iron and has distinctly different trace siderophile element signatures. It also contains lower amounts of volatile elements and probable higher concentrations of refractory elements. The hypothesis thus failed the test requiring an identifiable chemical signature of the terrestrial mantle in the Moon, despite heroic attempts by geochemists to fit the lunar compositional data to that of the silicate mantle of the Earth.

The Earth and Moon have distinctly different siderophile element patterns (Newsom 1986). Claims that the siderophile elements in the Moon show a unique terrestrial signature have been shown to be erroneous (Ruzicka et al. 1998, 2001). The similarity in V, Cr, and Mn abundances in the Earth and Moon, sometimes claimed as definitive proof of lunar origin from the terrestrial mantle (Ringwood 1986), has been shown to result from incorrect models of the bulk lunar composition (Ruzicka et al. 1998, 2001). The eucrite class of basaltic meteorites, formed close to T_o, shows similar depletions in volatile and siderophile elements to lunar basalts (Ruzicka et al. 1998, 2001).

Double-planet models that form the Moon and Earth in association possess the twin difficulties of failing to account for the angular momentum of the Earth-Moon system and of readily accounting for the density difference. They do, however, account for the similarity in oxygen isotopes between Earth and Moon. A variation of these models involved the break-up of differentiated planetesimals as they came within the Roche limit. This process was postulated to result in a circum-terrestrial ring of broken-up silicate debris, while the more coherent metallic cores of the planetesimals either accreted to the Earth or escaped. Such scenarios do not solve the angular momentum problem. Moreover, the proposed break-up of the incoming planetesimals may be difficult (Boss et al. 1990). The previous two hypotheses envisage processes that might be thought to be rather general features of planetary accretion and lead to the presence of moonlike satellites around the other terrestrial planets (e.g., Venus), where they are conspicuously absent.

2.2.2. Post-Apollo views. Any theory of the origin of the Moon must satisfy chemical constraints imposed by the unique lunar composition and by the dynamical constraints of angular momentum. Such a theory must be physically plausible as well (i.e., satisfy the rules of orbital mechanics, with boundary conditions preferably derived from conditions prevailing during the accretion of planets). Few theories proposed for the origin of the Moon can simultaneously satisfy all these conditions, and the acceptability of any theory is entirely dependent on such agreement between theory and observation.

Two first-order observations have been made on lunar composition that have significantly influenced the debate on lunar origin. Siderophile element abundances and Fe/Si depletion indicate that the Moon is composed of matter derived from the mantle of a planet. The depletion of volatile elements has generally been regarded as an important indication of the degree of thermal processing of lunar matter during formation of the Moon (Ringwood 1970, 1992; Hartmann 1986; Stevenson 1987), although Taylor (1979, 1987, 2001a,b) has argued that the volatiles were to a large extent depleted prior to lunar formation. Understanding the nature of volatile element loss from the Moon is of critical importance in determining the origin of the Moon (see discussion in Jones and Palme 2000).

Tidal calculations have often been used to assess the history of the lunar orbit, but attempts to determine whether the Moon was once very much closer to the Earth (for example, near the Roche limit at about 18,000 km), which would place significant constraints on lunar origins, produce non-unique solutions (Williams 1989, 2000; Boss and Peale 1986). During the Late Proterozoic at about 620 m.y., the mean lunar distance was 58.4 ± 1.0 Earth radii, so the Moon was only marginally closer to the Earth (The present mean Earth-Moon distance is 60.3 Earth radii). The current rate of retreat of the lunar orbit predicts that the Moon should have been close to the Roche Limit about 1.5 b.y ago. There is no evidence of such a catastrophic close approach of the Moon 2000 m.y. ago (Williams 1989, 2000; Boss and Peale 1986). Tidal deposits are known in Archean sediments about 3200 m.y. old, indicating that lunar tides, apparently similar in magnitude to those of today, were present at that ancient epoch (Eriksson 1977; Eriksson and Simpson 2000). Current theories of lunar origin form the Moon close to the Earth. If so, then the evolution of the Moon's orbit probably has varied with time, with the Moon receding from the Earth more quickly when it was closer.

2.2.3. The large impact model for the origin of the Moon. The Earth-Moon system has an anomalously high angular momentum compared to other planets and this has been the rock on which most hypotheses of lunar origin have foundered. This excess cannot arise through random small multiple impacts since these must average out. Less than 10% excess angular momentum could arise by such means. However, one very large glancing impact could account for the observed excess (Cameron 1986, 2000; Canup and Righter 2000; Canup and Asphaug 2001). The model proposes that a body larger than Mars struck the Earth and a hot disk of material was ejected to form the Moon. This "Mars-sized impactor" hypothesis thus solves the angular momentum problem by definition. This involves collision with the Earth by the next largest body in the hierarchy and assumes prior core formation in the impactor.

A primary requirement is that such bodies were common in the early solar system. It is clearly established that melting and internal differentiation of planetesimals took place within a few million years after T_0. There is no shortage of potential impactors. Kortenkamp et al. (2000) consider how to assemble planetesimals into larger bodies, with results converging towards runaway accretion forming perhaps 20 Mars-sized bodies within timescales of 10^5 years after T_0. However, it appears "that the final number, masses and spacings of planets are stochastic, and not very dependent on the intermediate embryo formation stage of accretion" (Kortenkamp et al. 2000). Random impact events capable of supplying the angular momentum of the Earth-Moon system were probably common in the early solar system.

The model provides sufficient energy to melt at least half the Moon as required by the geochemical evidence that indicates that melting occurred more or less concurrently with accretion. Indeed the energetic nature of the event provides not only the lunar magma ocean, required by the geochemical evidence, but also induces terrestrial mantle melting (Stevenson 1987). Computer simulations indicate that the material that makes up the Moon comes primarily from the silicate mantle of the impacting body, not from the Earth (Canup 2004). In the giant impact hypothesis, the impactor is assumed to have differentiated into a metallic core and a silicate mantle. The low density and low bulk Fe content of the Moon, long-standing problems, are thus resolved in the hypothesis, as in the simulations, the metallic core of the impactor accretes to the Earth following the collision. The hypothesis also resolves the geochemical dilemma that the composition of the Moon does not match that of the Earth's mantle by cutting this particular Gordian Knot.

Cameron (2000) postulated that the impact occurred when Earth was 2/3 accreted. It seems difficult to add 1/3 of the Earth subsequently without affecting the Moon. The lack of evidence of such a late addition of presumably iron-rich material to the Moon, is a serious problem for this iteration. However, in revised models by Canup and Asphaug (2001) and Canup (2004), a Mars-sized impactor (10% Earth mass) impacts at a late stage in the accretion of the Earth (Fig. 7.16) thus returning to the original Cameron models. This avoids the problem of having to park the Moon somewhere out of the way while one third of the mass of the Earth is added. The revised models form the Moon as the last major collisional event during the hierarchical accretion of the Earth.

The impacting body was clearly derived from the inner solar system as shown by similarity between the Earth and Moon for the oxygen (Wiechert et al. 2001) and chromium isotopes (Carlson and Lugmair 2000) and by the dynamical constraints of the model. Although the Hf-W isotopic system has provided new information on the separation of silicate from iron metal (Halliday et al. 2000), this is difficult to interpret. It is possible that we may never be able to disentangle all the details of the accretion of the Earth and the impactor from a series of planetesimals that had already differentiated into iron cores and silicate mantles and that had already been involved in multiple collisional episodes.

The Giant Impact model has not been without its critics. Thus Jones and Palme (2000) assess the various geochemical parameters in an attempt to decide whether the giant impact remains as a viable hypothesis. Their main concern is the problems of a lunar core and of relating

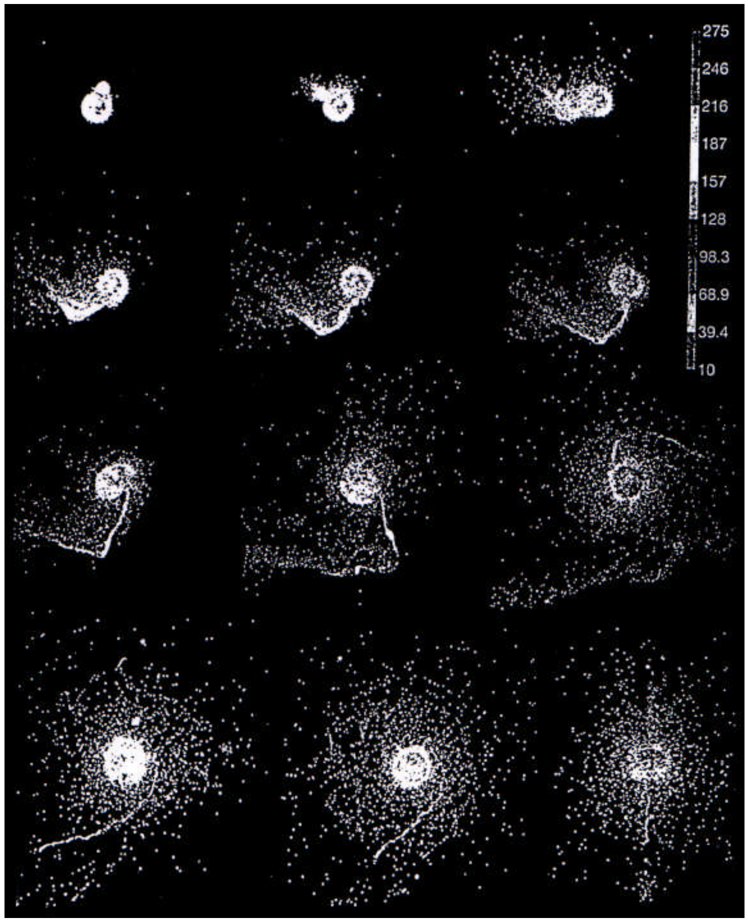

Figure 7.16. A computer simulation of the Moon-forming impact at times 0.3, 0.7, 1.4, 1.9, 3, 3.9, 5, 7.1,11.6, 17 and 23 hours after the impact. All views are in the plane of the impact except the last which is an edge on view. (courtesy Robin Canup and Erik Asphaug).

the composition of the Earth and Moon, but many of the problems disappear if the Moon is derived from the silicate mantle of the impactor rather than from the earth (Canup 2004). A small metallic lunar core might well be expected as the separation of silicate from metal during the giant impact is unlikely to be perfect. Thus not all geochemists would agree with their conclusions that the geochemical and geophysical evidence "casts doubt on the giant impact hypothesis." The stochastic nature of that event, the original depletion of volatile elements in the nebula, the unknown nature of the impactor, from which most of the Moon comes in the models, the great assemblage of precursor planetesimals and possible loss of volatiles during the impact make the task of geochemists attempting to reconstruct that distant event, model-dependent at best. It is also worth recalling that the Moon constitutes only about 5% of the mass of the impactor so perhaps is not even a uniform sample of the impactor mantle (e.g., Warren 1992).

Coupled with the very low value for the initial $^{87}Sr/^{86}Sr$ ratio (LUNI = BABI) (Nyquist 1977), the depletion of the Moon in many volatile elements might also be attributed to a volatile-depleted impactor rather than being due to the impact (Taylor 2001b).

Many complex factors need to be thought about in modeling the formation of a Moon following a giant impact. Stewart (2000) notes that the Moon "may have accreted slowly from the outer (cooler) edge of a protolunar disk, while the inner disk remained hot and molten. He points out that many of the apparent problems stem from "our primitive understanding of how the terrestrial planets formed", so that "idealized models....can also mislead the unwary" noting that "collisions between planetesimals... are inherently messy processes."

In summary, the peculiarities of the chemistry and dynamics of the Moon can be ascribed in one way or another to the extraordinary events surrounding a giant impact.

2.2.4. The nature of the impactor. A critical feature of the large-impactor hypothesis is that most of the material making the Moon comes from the silicate mantle of the impactor. Thus, the composition of the interloper is the most significant component for the geochemical constitution of the Moon.

The similarity in oxygen isotopes between the Earth and Moon indicate derivation of both the Earth and the impactor from the same region of the nebula, thus excluding models that derive the impactor from the outer reaches of the solar system. The similarity in $^{53}Cr/^{52}Cr$ ratios (^{53}Cr is derived in part from ^{53}Mn: $t_{1/2}$= 3.7 m.y.) between the Earth and the Moon and their contrast with higher meteoritic values (Lugmair and MacIsaac 1995) carries the same implication of derivation of lunar material from around 1 A.U. A third constraint is the relatively low collision velocity (Benz et al. 1989; Cameron and Benz 1991) required to produce a Moon-sized body, that once again restricts the impactor to have been formed in the neighborhood of Earth.

Current models (e.g., Canup 2004) assume that core mantle separation occurred before impact, to account for the lunar siderophile element abundances and the lunar depletion in iron (13% FeO) relative to primordial solar nebula volatile-free abundances (as shown by the CI meteorites) of 36%. The abundance of FeO in the mantle of the impactor must however have been greater than that of the terrestrial mantle (8% FeO), since the bulk Moon contains a much higher abundance (13–14.6%, Jones and Palme 2000). Such values are intermediate between the FeO abundance in the martian mantle (in which an estimate based on the SNC meteorites is 18%; Longhi et al. 1992) and that of the terrestrial mantle. Since Mars with a high mantle FeO content exists in the inner solar system, the postulate of an impactor with a high mantle FeO content is not unreasonable.

O'Neill (1991, p.1169) proposed that the impactor was an "exotic interloper.......from further out in the solar system" of CI composition, "fully oxidized and volatile rich." However, the factors noted above constrain the impactor, most likely to be volatile-depleted, to have originated in the neighborhood of the Earth. In addition, the rare gas and O isotopic evidence make CI chondrites an unsatisfactory component for building the terrestrial planets, so that there is no reason to suppose that planetesimals of CI composition were abundant in the inner nebula. The great distinction between the region now occupied by the terrestrial planets and that from which the CI chondrites are derived lies in the striking depletion of volatile elements. This depletion as shown by low K/U ratios, was common to Venus, Earth and Mars, and was thus probably due to a major event affecting the inner solar nebula (Taylor 2001a).

2.2.5. Resolution of the origin of the Moon. Of most significant immediate interest would be various geophysical data, specifically for heat flow which would resolve some geochemical questions about bulk composition, seismic data on the internal structure that will establish the nature of lunar deep structure and the presence or absence of a metallic core. These data will refine our ideas about initial thermal history and the initial depth of the magma ocean.

Another series of problems deals with events in the solar system before formation of the Moon and the history of the precursor material that is critical to the interpretation of the W-Hf isotopic systematics. Is the Moon a unique object or are there bodies (asteroids) with similar histories. What was the nature of the impactor? What was thermal history of precursor material and was the volatile-element depletion in the Moon due to the collision or inherited from the impactor, or is it a memory of events in the earliest nebula, as suggested by the low value of LUNI?

2.3. Highlights of recent discoveries

As discussed in previous chapters, a series of recent small missions to the Moon created an entirely new perspective of this intriguing planetary body, and the small amount of new data caused several paradigm shifts in our understanding of the character of the Moon. The fly-by of the Moon by the Galileo spacecraft on its way to Jupiter (Belton et al. 1992, 1994) was the first time a spacecraft had visited the Moon with modern sensors since Apollo. This was followed by the Department of Defense Clementine mission (Nozette et al. 1994), which was principally designed to test a suite of small sensors in space. NASA sponsored a science team shortly before launch to plan and implement science data. After several calibration issues (some still ongoing) the first global topography (Zuber et al. 1994) and global multispectral imagery data sets were produced for the Moon (e.g., Eliason et al. 1999). The only mission flown in this new wave of exploration whose principal objective was lunar science was the Lunar Prospector mission (Binder 1998), selected through NASA's competitive Discovery Program. This mission provided the first global assessment of elemental abundance (especially Th, K, Fe, Ti, H; see Chapter 2) (e.g., Lawrence et al. 1998, 2000, 2002; Feldman et al. 1998, 2000, 2001).

The input of data from this combined group of encounters and small missions has provided a valuable foundation upon which to reevaluate the evolution of silicate bodies. This is not to say important research had not continued via lunar sample studies and telescopic measurements of the lunar nearside during the dark decades. The origin of the Moon and the Earth/Moon system through a violent planetesimal encounter was well on the way to being understood (see Section 7.1.2). But our perception and understanding of the Moon changed with this minor amount of new data from relatively modern sensors. Three particularly important surprises are highlighted below.

2.3.1. SPA—the biggest basin in the solar system. Although several lunar scientists had long argued for a big basin on the lunar farside (e.g., Wilhelms 1987), the South Pole-Aitken basin (SPA) is missing from most commercial lunar maps produced using Apollo era data. This remarkable oversight reflects the limited character of the data available, changed, however by the first digital images from Galileo. Not only was the basin rim readily recognized, but the basin interior had a distinctively low albedo and exhibited spectral properties indicating that it was rich in mafic minerals (Belton et al. 1992). Subsequent data from Clementine illustrated the dramatic topography (up to 12 km deep!) (e.g., Zuber et al. 1994), enhanced mafic composition (e.g., Lawrence et al. 1998 2002; Lucey et al. 1998) and unusual mineralogy exposed in the interior (Pieters et al. 2001). It became apparent that SPA may be the largest documented basin in the solar system (Spudis et al. 1994; Zuber et al. 1994), and the deep-seated material of the interior represents a terrane on the Moon that is quite distinctive and unlike the lithologies documented with samples returned from the near side (Jolliff et al. 2000).

2.3.2. Thorium hot spot. The global elemental data from Lunar Prospector confirmed that the SPA interior is iron-rich (midway between feldspathic highlands and mafic mare basalt). One of the most unexpected results, however, was the very well-defined pattern of radiogenic materials, almost all of which are associated with the nearside Procellarum Basin, especially the Imbrium basin and its deposits. Since such elements (KREEP material) are expected to be concentrated in the residua of primordial lunar differentiation, it is not surprising that a large basin tapped such a concentration. The surprise is that the pattern is *only* associated with one part of the Moon and strongly associated with the Imbrium basin and circum-Imbrium deposits. Imbrium is not only smaller than SPA but also occurred near the end of the period of basin forming events. This highly localized concentration of radiogenic (incompatible) elements indicates there are enormous gaps in our understanding of the first few hundred million years of planetary evolution and differentiation.

2.3.3. Polar H (water?). The extremely low volatile content of lunar samples was one of the first characteristics noted for returned lunar samples (along with their ancient age). Thus, the contested hint of polar ice by Clementine (Nozette et al. 1996; Simpson and Tyler 1999) and the unquestioned discovery of hydrogen concentration at the lunar poles by Lunar Prospector (Feldman et al. 2000) has led to some optimism that water resources may indeed exist in the permanently shadowed regions of the Moon. However a reassessment by Campbell et al. (2003) indicates that any ice sheets in the regolith are probably only a few cm thick. This counter-intuitive concept, although reasonable in theory (Arnold 1979), is made more credible by the discovery of extensive volatile (perhaps ice) deposits in craters at Mercury's poles (Harmon et al. 1994).

2.4. Uniqueness of the Earth-Moon system

There are many peculiar features about the Moon and the Earth-Moon system that make this pair unique in the solar system. The orbit of the Moon about the Earth is neither in the equatorial plane of the Earth, nor in the plane of the ecliptic, but is inclined at 5.1° to the latter. The angular momentum of the Earth-Moon system is anomalously high compared to the other planets. The Moon has a low density relative to the terrestrial planets. The Earth-Moon mass ratio (1/81.3) is the largest in the solar system (excluding the Pluto-Charon system in the Edgeworth-Kuiper Belt).

Neither Venus, close in size to the Earth, nor Mercury have moons, while Phobos and Deimos, the two moons of Mars, may be tiny captured asteroids, or pieces of one, with what appear to be primitive compositions. The satellites of the outer planets are mostly ice-rock mixtures. Our satellite has an unusual composition. It is bone dry, depleted in elements such as K, Ga, Pb, Tl and Bi that are volatile below about 1100 K, and also in siderophile elements; and it is probably enriched in refractory elements such as Al and U relative to CI. However, in this respect the Moon is probably not unique as Vesta, at least, appears to possess similar depletions in volatile and siderophile elements (see Section 2.3 on Vesta).

The spin angular momentum of the Earth-Moon system is anomalously high compared with that of Mars, Venus or the Earth alone. Some event or process spun up the system, relative to the other terrestrial planets.

The Moon has a lower bulk density (3.34 g/cm^3) than the Earth (5.54 g/cm^3) (Fig. 7.17). However, the lunar density is close to the uncompressed density of the terrestrial mantle, implying that the Moon is composed largely of silicate minerals. Thus, it lacks the massive amounts of metallic iron that are responsible for the high densities of the inner planets.

2.5. Role of the Moon in making Earth a habitable planet

What were the implications for the Earth of the single impact origin of the Moon? It may be responsible for the 24 hour rotation period and the obliquity of the Earth, both significant parameters affecting biological evolution. An entertaining discussion by Williams and Pollard (2000) notes that the large mass and close proximity of the Moon stabilizes the obliquity of the Earth and has important implications for life and our climate, concluding that perhaps the Earth does owe its habitability and long-term sustainment thereof (and its suitability for humans) to the presence of the Moon.

According to Lascar and coworkers (1993), the obliquity of the Earth would be "chaotic with large variations reaching more than 50° in a few million years and even, in the long term, more than 85° in the absence of the Moon" and so "our satellite is a climate regulator for the Earth" (Fig. 7.18). Any primitive atmosphere was removed during the impact event. The Moon also plays a unique role by raising substantial tides in the Earth's oceans. In the context of this discussion, two questions without answers are how crucial is the presence of the Moon in producing this habitable planet and how likely are such events in other planetary systems.

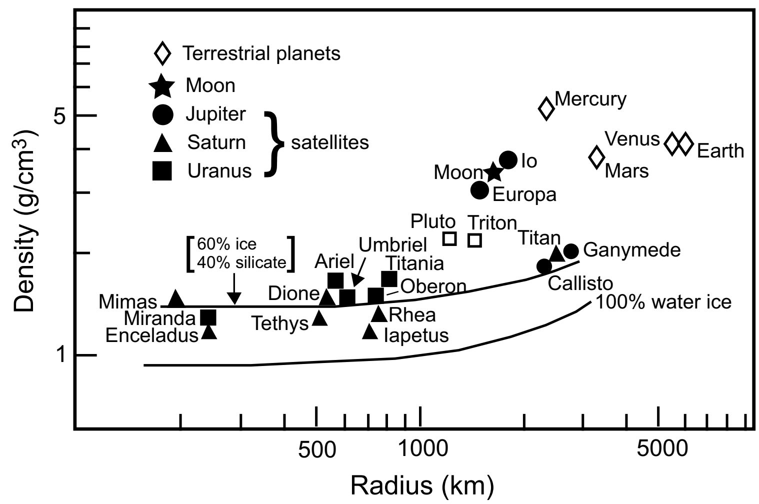

Figure 7.17. The densities (in g/cm^2) of the principal satellites of the giant planets, the Moon and the terrestrial planets (adapted from Taylor 2001).

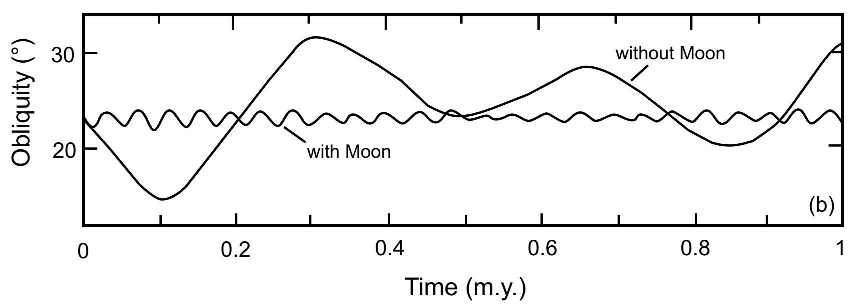

Figure 7.18. Obliquity variations for the Earth over one million years with and without a Moon (adapted from Williams and Pollard 2000, Fig 3).

3. PLACE OF THE MOON IN THE SOLAR SYSTEM

The Earth and Moon share the same part of the solar system. Since their birth they have both witnessed and experienced the same abundance of material and debris moving through our solar system. Without an atmosphere or magnetic field, however, the Moon has also constantly been subject to high-energy radiation from the sun and galactic sources as well as a constant dusting of small particles.

3.1. The Moon as a satellite

The density of the Moon is compared with that of the other satellites in the solar system in Fig. 7.17. Only Io and to a lesser degree Europa, are comparable with the Moon. The rest contain significant proportions of ice. Io owes its volcanic nature to its close proximity (six Jupiter radii) to that gas giant and so is a unique object. Ganymede, Callisto and Titan, are all larger, while Europa, Triton (and Pluto) are significantly smaller than the Moon. The similarity of Io to the Moon in density and radius has no fundamental significance, but must be due to stochastic processes. Size is not a very useful discriminator. Thus the planet Mercury is smaller than Ganymede and Titan, while seven satellites, including the Moon, outrank the largest Edgeworth-Kuiper belt object, Pluto (although larger Triton also has a case to be considered as a former resident of that belt). Most of the satellites in the solar system are mixtures of rock and water ice and have low densities. They formed in the early solar nebula, out beyond the "snow line" around 4 to 5 AU, where water-ice was stable. The captured satellites of the giant planets represent icy planetesimals left over following the formation of the giant planets, or captured bodies from the Edgeworth-Kuiper belt. The regular satellites of the giant planets, of which the Galilean satellites of Jupiter are the best known, appear to have formed from planetary subnebulae, and thus differ fundamentally in origin from the Moon.

In the inner solar system, only one rocky planet managed to grow to around the size of the Earth, but the Earth and Venus are similar only in a Jekyll-Hyde sense. Thus Venus is significant as it shows what happens when nature tried to duplicate our planet. Although Venus is close enough in mass and density to be regarded as a possible twin planet to the Earth, there are major differences between the two planets. Venus does not possess a satellite. The slow backward rotation and low obliquity of the planet are indicative of a very different collisional history.

3.2. Comparison with other airless differentiated bodies

3.1.1. Physical and compositional properties of Vesta. Vesta was the first asteroid studied with remote spectroscopic analysis of sufficient spectral resolution to characterize its surface mineralogy (McCord et al. 1970). Based on the inferred mineralogy, a strong association was made between Vesta and the basaltic achondrite meteorites (Howardites, Eucrites, Diogenites, or HEDs). This link was strengthened as new instruments and more sensitive measurements were made, especially when it became apparent that Vesta was the only large asteroid in the main belt with properties comparable to the HEDs (e.g., Tholen 1989; Gaffey et al. 1993; Pieters and McFadden 1994).

Dynamists were skeptical of the link between the HEDs and Vesta at first because it was believed to be extremely difficult to remove rocks from Vesta's surface and get them to a part of the solar system that favored perturbations into Earth-crossing orbits. All that changed when telescopic instrumentation improved sufficiently to accurately measure more and smaller asteroids. It was discovered (Binzel and Xu 1993), that a spray of small "Vestoids" exists between Vesta and the 3:1 resonance with Jupiter, an accessible "escape hatch." The case continued to grow stronger as additional spectroscopic observations were made (Gaffey 1997; Burbine et al. 2001) and HST images showed there to be a 460 km diameter central-peak crater, 13 km deep on Vesta's surface (Thomas et al. 1997), which may account for the Vestoids.

The basaltic eucrites derived from 4 Vesta display many similar compositional features to the lunar basalts (Fig. 7.19). Vesta produced basaltic material with similar element patterns to low-Ti lunar basalts, including strong depletions in volatile elements, at about 4557 m.y. ago (Lugmair and Carlson 2000) close to T_0. The angrites and the eucrites are the only meteorites that have such low K/U ratios that are less than those of the Earth and that overlap those of the Moon. The eucrites are also depleted in siderophile elements and have very similar high Cr abundances to those of mare basalts (Ruzicka et al. 1998, 2001).

Thus, at least one asteroid went through a sequence of volatile and siderophile element depletion similar to that experienced by the Moon. But the timing of the thermal event on the HED parent body was within only a few millions of years after planetesimal formation. The canonical view is that asteroids with surfaces of achondritic composition are rare (based on astronomical surveys such as Tholen 1989), so the processes that produced the HED parent body may have been uncommon in the early solar nebula, at least in the region of the asteroid belt. However, such processes producing "igneous" planetesimals may have been more common in the inner nebula that supplied planetesimals to form the inner planets.

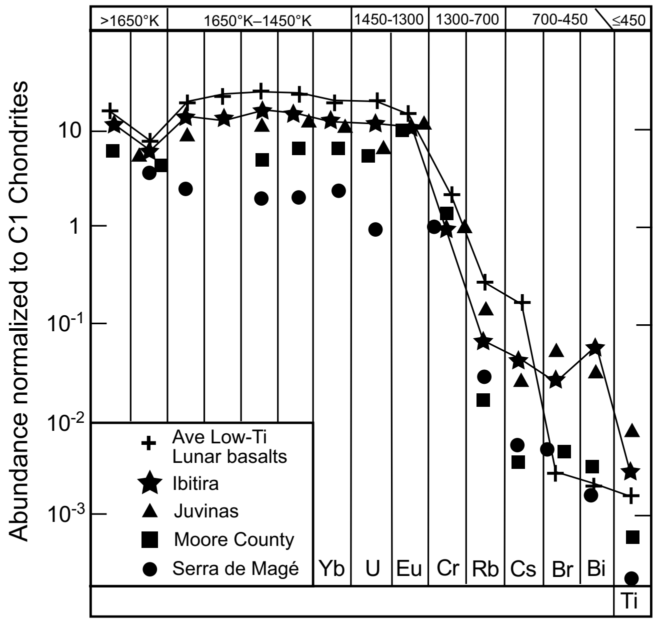

Figure 7.19. The close correspondence between the abundances of the lithophile elements in eucrites and low-Ti mare basalts (adapted from Taylor and Esat 1996, Fig. 2).

3.2.2. Physical and compositional properties of Mercury. The Moon and Mercury represent special cases even by the standards of the solar system. Mercury is unique due to its high density, with an iron-silicate ratio about twice that of the other inner planets. In contrast, the Moon is of interest because of its low density and low metal-silicate ratio. Mercury is of great interest as the scant evidence shows that the surface bears a close resemblance to that of the lunar highlands (Fig. 7.20). A comparison between the integrated spectrum for the mercurian surface and that of a laboratory spectrum for Apollo 16 lunar highlands samples shows that the mercurian data are consistent with a regolith surface. Spectral data from thermal infrared exclude both ultramafic and granitic soil compositions, and appear to be compatible with a feldspathic crust similar to that of the lunar highlands although the surface appears to be heterogeneous with feldspar compositions ranging from Ab_{50} up to anorthite (Mitchell and de Pater 1994; Sprague et al. 1997; Sprague and Roush 1998). More recent evaluations of the spectral reflectance data in the 0.4–0.65 μm region, coupled with the earlier data, indicate that the concentrations of iron and titanium appear to be near zero, while the spectral reflectance data are consistent with a surface mineralogy consisting of a 3:1 labradorite-enstatite mixture, with an average grain size a factor of two smaller than the lunar soils (Warell and Blewett 2004).

The presence of Na and K in the tenuous mercurian atmosphere is consistent with a feldspathic surface. Na and K ions are also observed up to 1200 km above the surface of the Moon. In the latter case, they are almost certainly derived by sputtering from the feldspathic surface.

Some information can be gleaned about composition from the surface morphology. There are two types of plains units on Mercury, the early intercrater plains of preTolstojan age and the later "smooth" plains of Calorian age (Spudis and Guest 1988). The origin of both sets of plains is disputed between those who equate them with the lunar Cayley-type plains (debris deposits or impact melt sheets, both produced by impacts) (e.g., Wilhelms 1976)) and those who propose that they are volcanic plains analogous to the lunar maria, and hence are formed

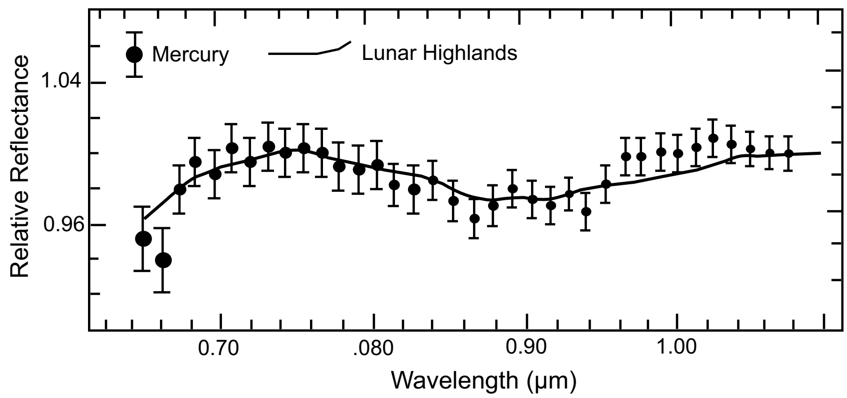

Figure 7.20. The reflectance spectrum of the surface of Mercury, covering both smooth and intercrater plains, shows a close resemblance to a laboratory spectrum of an Apollo 16 lunar highland soil sample. (adapted from Vilas et al. 1988, p 71, Fig. 6).

by rather fluid basaltic lava flows (Spudis and Guest 1988). It is interesting to note parallels with the history of the pre-Apollo interpretation of the lunar surface in this controversy.

If the intercrater plains are volcanic, the mercurian lavas must possess the same low viscosity as the lunar basalts. This assumption remains to be tested, but it would imply a similarity in mantle sources that, given the unique nature of lunar evolution, seems surprising. Their albedo is twice as high as that of the lunar maria, a fact not consistent with the presence of dark low viscosity iron-rich basaltic lavas. The wrinkle ridges on the floor of the Caloris Basin, cited in support of an analogy with lunar mare basalts, appear to be very large. Only more data will resolve this issue, but it is clear that if these features are volcanic, then their petrogenesis is distinct from that of the lunar mare basalts.

Additional light on this problem has been shed by microwave and mid-infrared data (Jeanloz et al. 1995; Mitchell and de Pater 1994; Blewett et al. 1997). The regolith of Mercury is two to three times more transparent to microwave radiation than the lunar mare regolith and 40% more transparent than that of the lunar highlands. This limits the amount of FeO and TiO_2 on the surface of Mercury to much less than that present in the lunar highlands and the concentrations of these elements may be close to zero (Warell and Blewett 2004). The apparent low content of iron on the mercurian surface is consistent with a lack of volcanic activity. It is possible that "volcanic heat piping has not played a major role in Mercury's thermal evolution" (Jeanloz et al. 1995) thus insulating the core. Perhaps the alkali and other incompatible elements were concentrated at an early stage in the crust of this one plate planet. In this scenario, the crust of Mercury may have had a similar origin (flotation of feldspar from a magma ocean) to that of the Moon. The difference between the two bodies appears to be a lack of subsequent basaltic volcanism on Mercury.

Some spectra may be compatible with that of nepheline-bearing alkali syenite (Jeanloz et al. 1995) and presence of such sodium-rich volcanic rocks could provide an additional source for the sodium ions detected in the atmosphere. Silicic volcanism, which might produce low-albedo deposits, seems an unlikely candidate on Mercury and may indeed be a mainly terrestrial phenomenon. Accordingly, magmas derived from the mercurian mantle may indeed be unique.

3.2.3. Implications of space weathering processes on airless body regolith. An understanding of regolith evolution may be key to interpreting remote observations of Mercury and other airless bodies in the Solar System. As detailed in Chapter 2, the harsh space environment alters planetary surfaces in a manner that makes remote compositional analysis difficult if regolith processes are not taken into account. Gradual physical and compositional transformations occur as a result of bombardment by micrometeorites and solar and galactic particles. Without the detailed information available from lunar soil samples, we would be ignorant about the myriad of processes that have a profound effect on observed properties of solar system surfaces. On the Moon as soil particles are exposed to the space environment, they accumulate thin (a few hundred angstroms) amorphous silicate rims that contain abundant sub-microscopic native iron, or nanophase metallic iron ($npFe^0$) (see Chapter 2 for detailed discussion). The longer a soil is exposed, the more particles develop such $npFe^0$-bearing rims and the total abundance of $npFe^0$ increases. Although $npFe^0$ is a minor component of lunar soils (<<1%), it is exceptionally optically active and absorbs radiation passing through particles in a very non-linear manner. Also, since deposits of $npFe^0$ originate as rims on individual particles, finer particles with more surface area are affected most strongly.

The effects of $npFe^0$ on optical properties depend on the size of these tiny metallic grains and their overall abundance, both of which depend on the host material and the environment in which $npFe^0$ forms. In all cases, the presence of $npFe^0$ causes the soil to darken and diagnostic mineral absorption bands to weaken. The effects are not uniform with wavelength. For lunar soils the size of $npFe^0$ ranges from about 3 to 20 nm (Keller and McKay 1993, 1997). This size of grains imparts a wavelength dependence to absorption of radiation, which is highly sensitive to the abundance of $npFe^0$ (Noble et al. 2001, 2003; Hapke 2001). For the Moon, more radiation is absorbed at shorter wavelength, creating a "red" continuum for most lunar materials. The curvature of this lunar continuum is controlled by the abundance of $npFe^0$, and to a lesser extent the size of grains (Hapke 2001; Noble et al. 2003; Noble 2004).

The origin of $npFe^0$ is linked to processes that release and mobilize iron from surface materials, the two most prominent of which are vaporization from micrometeorite impacts and solar wind sputtering (Keller and McKay 1997; Taylor et al. 2001; Hapke 2001). For the Moon most of the iron comes from FeO in regolith minerals, with minor amounts from meteoritic contaminants. Soils developed from feldspathic highland lithologies have lower amounts of FeO available to produce $npFe^0$ than soils from mare regions. Systematic differences in optical effects are thus observed as a soil matures in the space environment for mare and for highlands regoliths (Noble et al. 2001, 2003; Noble 2004). This understanding of the processes affecting space weathering allows predictions for other solar system bodies. The effects of space weathering and development of $npFe^0$ on soil particles is expected to be considerably reduced for asteroids and readily accounts for optical effects that transform ordinary chondrite parent bodies into "S-type" asteroids (Pieters et al. 2000; Hapke 2001). It should be noted that metallic grains larger than 100 nm (i.e., approaching micron scale) simply act as absorbers at all optical wavelengths (Noble et al. 2003; Noble 2004). The impact darkening observed in ordinary chondrites due to dispersed metal and sulfides is a good example of this effect (Britt and Pieters 1994). On the surface of Mercury, the effects of space weathering and development of $npFe^0$ should be enhanced (Noble and Pieters 2003), but is complicated by the extreme thermal cycling and higher flux of particles. Experimental evidence suggests the size of the $npFe^0$ is slightly larger on Mercury than on the Moon and that the total abundance of Mercurial regolith $npFe^0$ is somewhat more than typical lunar highlands (Noble et al. 2004, in prep).

4. HOW TO EXPLORE A SOLID BODY

An overview of lunar exploration and the historic sequence and context of its multiple aspects is provided in Chapter 1. Subsequent chapters provide details and depth of this incredible undertaking. It is constructive to reflect on what has been learned and some of the important lessons that might guide further detailed exploration of other solar system bodies.

4.1. What we learned (and changing perspectives)

4.1.1. From samples. In the 30 years since the last Apollo and Luna samples were brought to Earth, many review papers and books have documented the profound impact that the lunar rocks had in changing nearly all preconceived notions about the nature of the Moon (e.g., Taylor 1982; Wood 1986). Prior to Apollo, it is doubtful if anyone could have predicted some of the remarkable properties of the lunar samples. They serve still as an object lesson about the dangers of over-interpreting planetary bodies in the absence of sample-derived knowledge. Yet, if the single most important lesson to be learned from the Apollo samples is "expect the unexpected," that lesson seems not to have been universally learned. Great technological advances in robotic analytical instrumentation have led to some speculation that costly and difficult sample return might not be required to adequately explore Mars (e.g., Paige 2000). But, the fact is that there will always be major limitations to what a robotic lander or rover can do. Partly this is because of restrictions in power and mass, and partly it is because no robotic spacecraft instruments can ever conduct experiments under such controlled conditions or with sufficiently well prepared samples as is possible in an Earth laboratory. An additional very important reason is that robots cannot adapt in the way a human scientist can in response to unexpected results, bringing additional and unforeseen tests to bear.

There are of course over 37 known (as of this writing) lunar meteorites (~80 separate stones) that have been found in Antarctica and elsewhere. Even without the Apollo and Luna samples, we might by now hypothesize much about the formation and evolution of the Moon from the meteorites (albeit 10 years later; the first lunar meteorite was identified in 1981). Although their identity as such rests on comparison with the Apollo and Luna specimens, even in the absence of the latter it is reasonable to assume that a lunar origin for the meteorites would be suggested from such evidence as identity in oxygen isotope compositions with Earth rocks (recall, however that enstatite chondrites share this property), and the young ages of many mare basalts relative to most other meteorites. The meteorites are considerably less informative than mission-returned samples in the sense that the geologic context of the meteorites is completely unknown. The Apollo and Luna samples come from known localities and in some cases even from outcrops, so ages of specific sites, formations, and impact events can be determined. But the fact remains that meteorites are samples that are analyzed in terrestrial labs just as the mission-returned samples are. The only difference is the delivery mechanism and the lack of definitive information about their place of origin.

Previous reviews (e.g., Taylor 1975; Drake 1986; Wood 1986) have written in detail about what has been learned from the Apollo samples. There would be little point in restating here what has been so eloquently said elsewhere, were it not for the fact (alluded to above) that 25+ years (since Viking) of Mars exploration and exploration planning still have not resulted in a sample return mission. While admitting the high cost and technical difficulty of implementing a robotic martian sample return mission, there nonetheless remains a continuing lack of recognition of the critical importance of sample return to achieving Mars science goals. The emphasis is on building bigger and better robotic rovers with larger and more capable science payloads, in order to find "just the right place" to eventually collect samples for return to Earth. But this attitude forces sample return to be continually postponed into the almost indefinite future. Such planning either does not recognize or else ignores a single inescapable fact: The most important science goals (e.g., the existence of ancient or present life, crustal evolution, atmospheric evolution) of Mars exploration will likely not be achieved in the absence of sample return because robotic rovers will simply not be able to carry analytical instruments with the levels of precision and sensitivity of Earth-based instruments, nor will they have the flexibility to react to unknown findings with anything like the flexibility of human scientists in a lab.

Therefore, in the interest of using the lunar experience as an object lesson for the exploration of other solar system bodies—especially Mars—this section takes a somewhat different approach than previous reviews. Specifically, we highlight here a sampling (by no means exhaustive) of important lunar measurements that we could *not* have made, or make even now, with strictly robotic means; these findings derive uniquely from the study of the returned samples (including the meteorites) by humans in Earth laboratories. These measurements were necessary precursors to current concepts of lunar formation and evolution. For example, the oxygen isotopic similarity of lunar and terrestrial rocks plus the extreme depletion of volatiles in the lunar rocks and the Eu anomalies are fundamental underpinnings of the giant impact hypothesis. We do not attempt to review in detail the results themselves—that is done elsewhere in this volume—but rather to explain why laboratory analysis of returned samples plays such a critical role in understanding the evolution of the Moon—and, hence, any other planet. Not only would some of the most important lunar discoveries have been missed 25 years ago had we relied solely on telescopic/orbital/*in situ* robotic exploration, they would be missed *even today* using the much more sensitive and accurate remote sensing instruments that currently are available.

Lunar differentiation and igneous diversity. One of the most fundamental scientific properties of any rocky planet or satellite is its basic geologic structure: whether it is primitive (chondritic) or differentiated into a core, mantle, and crust. And, if it is differentiated, what are the basic crustal units and how did they evolve? Up until nearly the time of Apollo 11 some scientists believed the Moon to be not a differentiated planet but, rather, a chondritic aggregate (e.g., Urey 1965). Although remotely obtained chemical data from Surveyor missions suggested the presence of basalts in the Mare basins and hinted at feldspar-rich rocks elsewhere (e.g., Turkevich 1971), it was the preliminary examination of the Apollo 11 rocks that largely dispelled the possibility of a primitive undifferentiated Moon (LSPET 1969). Samples from the subsequent Apollo missions showed even more clearly, in their petrographic and geochemical characteristics, the strongly differentiated character of the Moon. Seismic experiments set up by the Apollo crews confirmed that the Moon has a stratified structure similar to that of Earth, with a crust, a mantle, and possibly a small core (Hood and Zuber 2000). From the samples themselves, three major igneous rocks suites were recognized—the highlands ferroan anorthosites, mare basalts, and magnesium-rich norites and gabbros. A fourth important "rock type," KREEP, is mainly a geochemical component contained to varying degrees in other suites, such as the Mg suite. Detailed geochemical and petrologic studies showed these suites to be interrelated by differentiation on a planetary scale, albeit not from a homogeneous reservoir. For example, trace-element studies showed ubiquitous negative Eu anomalies in the diverse mare basalts and complementary positive anomalies in the ferroan anorthosites. This observation suggests that the basalts originated by partial melting of a mantle source from which plagioclase of the crust was also removed, and supports the idea of formation of the anorthositic crust by plagioclase flotation and fractionation of a magma ocean.

Consider how the situation would now be different had there never been any Apollo and Luna samples, or even any lunar meteorites (or, at least, that they were not recognized as such). The Clementine and Lunar Prospector orbital data obviously reveal the chemical dichotomy between the feldspathic highlands and the mafic mare, the existence of a huge geochemical anomaly in the region of Procellarum basin (i.e., the Procellarum KREEP terrane), and the mafic South Pole-Aiken basin. What then would we learn by sending robotic landers to those regions of the Moon today for the first time, equipped with modern robotic analytical instrumentation? The Mars 2003 rovers, for example, carried an instrument package that included a rock-abrasion tool (for abrading below the surfaces of coated or weathered rocks), an alpha particle X-ray spectrometer for bulk chemical analysis, a Mössbauer spectrometer, a miniature thermal emission spectrometer, and a high-resolution imager. A similar package on the surface of the Moon could characterize the dominant mineralogy and major-element chemistry of, and thus identify, the dominant lunar igneous types such as basalts, anorthosites, and troctolites/norites. Thus we would undoubtedly recognize that the Moon is a differentiated body. But deciphering the details of that differentiation, and seeing through the devastating effects of more than 4 billion years of impact processing, requires tools that mostly are beyond the capabilities of any rovers now or in the foreseeable future. For example, recognizing and characterizing an impact melt breccia is straightforward for a petrologist in a lab using a thin section, a petrographic microscope, and an electron microprobe. The hypothetical robotic lander described above could, in fact, probably also recognize a coarse impact breccia through the presence of glass and large diverse clasts. However, barring the presence of xenoliths, recognizing a hybrid impact melt would be difficult at best for the rover. Distinguishing between a primary basalt, a fine grained basaltic breccia, and an impact-melt of basaltic composition, probably would be even more difficult (e.g., Fig. 7.21).

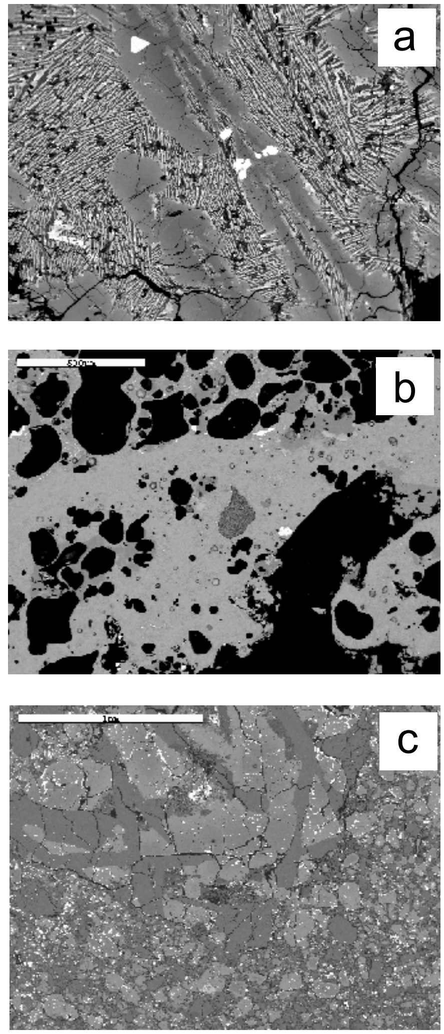

Figure 7.21. Back-scattered electron photomicrographs of three lunar basaltic rocks: (a) primary igneous basalt; (b) basaltic impact melt; (c) basaltic breccia. Photos courtesy of C. Shearer.

Petrogenetic modeling of the lunar igneous suites is critically dependent on trace elements, among which the rare earths are of course important because of how igneous processes fractionate the individual elements from one another. In terrestrial laboratories, bulk determination of REE in common igneous rocks is routinely done either by neutron activation analysis or inductively-coupled plasma mass spectrometry (ICP-MS); These elements generally are beyond the detection limit for X-ray fluorescence, and elemental peak overlaps in any case are severe. Consider that the REE abundances in lunar mare basalts range from several tens of ppm for the most abundant elements (Ce) down to less than one ppm for the less abundant (e.g., Lu), and abundances in highland anorthosites are less than one ppm for even Ce and Eu and much lower for the other REE. Even in KREEP, the abundances of Sm and heavier REE generally are several tens of ppm at most. No current generation robotic lander instrument is capable of precisely analyzing trace elements, and it is unlikely that such instruments will be capable any time soon of analyzing the REE at the levels indicated with sufficient precision to resolve intra-REE fractionation patterns. Thus without the returned Apollo and Luna samples and the lunar meteorites, there would be no accurate data for REE or other trace element abundances that are so important in constraining the petrogenesis of the dominant lunar igneous rock types and their possible relationships to one another. Without those samples, it is also unlikely that we would recognize KREEP at all beyond simply knowing that the geochemically-anomalous Procellarum region exists.

In summary, without having the Apollo/Luna samples and lunar meteorites we would not understand any of the details about lunar differentiation. This is because of the simple reason that we would not be able to bring to bear two of the most critical tools for constraining igneous petrogenesis: precise trace element determinations and classical petrographic analysis. There is no robotic lander yet that can even begin to do those things as well as a human scientist studying returned samples in a lab.

The antiquity of lunar rocks. The ages of major crustal units and the timing of global differentiation are essential facts for understanding the crustal evolution of a differentiated planet. On Earth, plate tectonic recycling and crustal reprocessing have erased most traces of the first 500 million years of Earth's history. Uplift and erosion have destroyed most traces of its early impact history as well. Therefore the Moon offers critical insight into Earth's earliest history.

Prior to Apollo, the ages of lunar surface units were estimated based on crater counts. Noting that pre-Apollo crater age estimates for the mare ranged from 3.5 Ga to a mere few million years, Taylor (1982, p 31; and refs. therein) wryly concluded that the understanding of lunar ages at that time "might best be described as chaotic." One of the fundamental discoveries from Apollo is that the lunar surface is uniformly very old. The youngest measured crystallization (igneous) age of any lunar rock is 3.1 Ga for a mare basalt, although the lunar volcanism that produced the mare basalts extended over a substantial time interval from 4.2 Ga. Ferroan anorthosite from the highlands dates back to at least 4.44 Ga (Lugmair 1987) or 4.46 Ga (Norman et al. 2003), meaning that the Moon had formed and began to solidify within about 100 my of solar system formation.

What was true 30 years ago remains equally true today: the extreme care required to make precise isotopic age determinations of ancient lunar (or any other) rocks is far beyond the capability of any robotic instrumentation and probably always will be. Epsilon-level (part per 10,000) precision in isotopic ratio measurements is achieved only through elaborate sample preparation (elimination of contamination, complete digestion of samples in ultra-pure reagents, separation of the isotopic systems of interest by ion exchange, etc.), low blank levels, the frequent measurement of known standards during unknown runs, and the use of stable, high-resolution, magnetic-sector mass spectrometers under carefully controlled and monitored conditions. Without the lunar samples, we would not have any knowledge about the absolute timing of lunar evolution. It is worth emphasizing in this regard, yet again (as other writers have done), that modern methods for using crater counts as a lunar surface dating tool have only been possible because the crater counts are calibrated against the radiometric ages of the lunar samples.

Radiometric age dating is sometimes cited as the one first-order scientific investigation that absolutely requires sample return. Thus it is very interesting that there have been recent suggestions (e.g., Swindle 2000) that it might be possible to measure radiometric ages robotically via landed rovers. This has been driven by the high cost of returning samples from other large and distant bodies, especially Mars, and the need to provide some reasonable calibration for crater counting as a means of dating planetary surfaces. It is estimated that the obtainable precision for robotic K-Ar ages might be on the order of ±10–20%. Even the most ardent proponents of robotic age dating are very careful to emphasize that such measurements are no substitute for terrestrial laboratory measurements. However, as Swindle (2000) points out, if the cratering history of some martian surfaces (for example) spans 3 Ga or more, then a robotic measurement with a precision of 10–20% to constrain those crater count ages would be useful.

Constraining rough ages of planetary surfaces is one thing, but deciphering the details of early planetary crustal evolution is another matter entirely. Current knowledge of the chronology of lunar crustal evolution required laboratory analysis of returned or meteoritic samples, and this situation will remain true until such time as we actually build manned laboratories on other worlds. The truth of this statement can be appreciated by realizing that a robotically-acquired rock age of 3.8 Ga with an uncertainty 20% could actually be anywhere from 3.0 Ga to 4.6 Ga; on Earth's Moon, that leaves its entire known igneous history unresolved. Moreover, a robotically measured age (on the Moon or anywhere else) will be very difficult to interpret correctly unless enough supplemental information is collected about the sample to reveal whether the age is meaningful, such as:

(1) Did the rock form approximately as an equilibrium system (i.e., is it igneous or metamorphic) or is it a clastic sediment?

(2) Has the rock been severely shocked or otherwise affected subsequent to initial crystallization?

(3) Is the rock an impact melt that might possibly contain (on, e.g., Mars) trapped atmospheric gas?

Answers to these questions require human scientists using detailed petrographic and electron microprobe studies of carefully prepared rock slices or thin sections; for the foreseeable future, it is difficult to envision the ability to do so via robotic means.

Identity of oxygen isotope compositions in the Earth-Moon system. The oxygen isotope compositions of rocks from the Earth and Moon lie within ±0.1–0.2‰ of the same mass-dependent fractionation line (Clayton and Mayeda 1975; Valley et al. 1995; Wiechert et al. 2001); very little else in the solar system lies on that same line (notably, enstatite chondrites and aubrites) (see Fig. 7.14). In a solar system where most solid bodies have their own characteristic oxygen isotopic fingerprints, the Earth and its Moon are alike. This observation is one of the most important constraints on the origin of the Moon, because it inextricably links together the matter from which the Earth and Moon were made. It renders unlikely whole classes of models for lunar origin; e.g., random capture of a passing body (Wood 1986). It also places severe constraints on the current best model, namely collision of a Mars-sized body with proto-Earth: either the colliding body had the same oxygen isotopic composition as Earth (again, aubrites and enstatite chondrites) or else it contributed a very small mass of material (< 3–5%; Wiechert et al. 2001) to the new moon.

Precise oxygen isotopic analysis is difficult under any circumstances. Conventional analysis of bulk material uses a highly reactive compound (BrF_5) to liberate oxygen from oxides and silicates (Clayton and Mayeda 1963). Isotopic fractionation during extraction is a significant problem unless >98% of the theoretical amount of oxygen is recovered. Microanalysis via laser ablation of small spots is done in an even more reactive environment (F_2), and because of the small amounts of gas liberated and the low abundance of ^{17}O, achieving low blanks in the extraction line are critical. Both of these techniques yield analytical precisions that generally are better than 0.2‰. Analyses of oxygen isotopes by secondary ionization mass spectrometry have (so far) significantly worse precision, generally ± ~1.0–1.5‰.

Whether twenty-five years ago or today, the prospect of designing a spacecraft to handle BrF_5 or F_2 robotically is not a concept that any spacecraft engineer would be likely to embrace enthusiastically. And setting aside such hazard issues, the blank and instrumental isotopic fractionation problems associated with bulk or laser fluorination techniques are beyond the capability of any current robotic instrumentation that we are aware of.

Without the laboratory analysis of oxygen isotopes in the Apollo, Luna, or lunar meteorite samples, either 25 years ago or today, we would not possess one of the most important existing constraints on lunar origin.

Depletion of volatiles. It was recognized almost immediately after Apollo 11 that lunar rocks are virtually anhydrous and are depleted in volatile trace elements relative to their terrestrial equivalents (e.g., LSPET 1969; Wasson 1971; Taylor 1975). Moreover, estimates of lunar and terrestrial bulk compositions suggest that the whole Moon is depleted in volatile elements relative to Earth (see review by Drake 1986). It should be noted that both Earth and the Moon are depleted in volatile elements relative to bulk solar (CI chondritic) values (Taylor and Jakes 1974; Ringwood and Kesson 1977), and the HED meteoritic basalts are equally as depleted in volatiles as the Moon. Nevertheless, lunar volatile depletion is commonly used as an argument that the lunar matter underwent high-temperature processing relative to that of the Earth even though some (e.g., Wasson 1971; Drake 1986) have pointed out that the Earth-Moon volatile differences are not entirely compatible with volatility considerations alone. For example, the Moon is less depleted in cesium relative to rubidium than the Earth is, even though cesium is the more volatile of the two elements.

Regardless, volatile elements useful for Earth-Moon comparisons are (with the exception of Na and possibly K) trace elements whose precise determination is beyond the capabilities of robotic analytical instruments even today, much less 25 years ago (e.g., Table 7.1). Even for K the analyses are difficult robotically in the case of the Moon, because K also essentially is a trace element in most lunar rocks. Even worse, Earth-Moon comparisons regarding volatile elements commonly are expressed as volatile/nonvolatile ratios such as K/U, and U in mare basalts is in the abundance range ~0.1–0.9 ppm.

Table 7.1. Typical Abundances of volatile elements in some lunar rock suites.

	Mare Basalts	Monomict Anorthosites	KREEP
Rb	0.28–6.2	0.12–<6	2.4–113
Cs	<0.1	0.015–<0.2	0.15–2.2
Zn	<2–30	0.26–93	2.6–5
Ag (ppb)	—	0.41–1.73	0.44–2.3
As	—	2–4.1 ppb	0.37
K_2O (%)	0.01–0.36	0.07–0.28	0.1–3.6

(Abundances in ppm except where indicated)
Data summarized from Lunar Sourcebook, BVSP

KREEP. Orbital gamma ray measurements by Apollo 15 and 16 demonstrated gross thorium heterogeneities in the lunar crust, with high concentrations in the Imbrium and Procellarum regions. More complete global coverage by the Lunar Prospector mission showed that the lunar surface has a profound geochemical asymmetry, with a huge geochemically-anomalous region centered on Mare Imbrium that shows up not only in Th but also in Sm, inferred from modeling neutron spectrometer data (e.g., Lawrence et al. 2000; Elphic et al. 2000). But it was laboratory studies of regolith fines and breccias brought back by Apollo 12 that first showed the presence of some mafic fragments that are highly enriched in a range of large ion lithophile elements (notably: K, REE, and P, hence the enduring acronym KREEP; Hubbard et al. 1971). It has taken 25+ years of detailed laboratory studies to reveal the nature of KREEP in ways that could never have been achieved from orbit or by *in situ* landers. Moreover, the orbital data cannot distinguish between KREEP and other lunar rocks that also have elevated abundances of some of the same trace elements (e.g., Th). In short, even though orbital measurements demonstrated the existence of a lunar geochemical anomaly, the methods required to decipher the detailed nature of the material making up that anomaly are another matter.

KREEP exists most commonly as a cryptic geochemical component in breccias and glasses. Voluminous trace-element determinations (by activation analysis, isotope dilution) showed that the characteristic KREEP “signature” is one of surprisingly constant elemental ratios despite large ranges in absolute trace element enrichments from rock to rock. Because of these constant elemental ratios, KREEP is interpreted (Taylor and Jakes 1974, Warren and Wasson 1979) to represent a late-stage residual liquid following the extensive fractional crystallization of an ultramafic liquid rather than the result of partial melting. Radiometric age determinations show that KREEP is very old. Relatively pristine KREEP basalts give ages of ~3.85–4.08 Ga (Shih et al. 1992), with formation of the KREEP source from bulk Moon being estimated at ~ 4.4 Ga (review by Nyquist and Shih 1992). For reasons already noted, none of these precise trace element and isotopic studies would be possible robotically. They were and are possible only in terrestrial laboratories and have yielded a set of KREEP properties that, taken together, suggest KREEP originated as a residue of the fractional crystallization that produced the lunar crust and mantle from a lunar magma ocean (e.g., Wood 1972; Lipin 1976; see discussion by Warren and Wasson 1979, and other references therein).

Character of the lunar regolith; effects of impacts and "space weathering." One of the great unknowns prior to Apollo was the nature of the lunar regolith: its origin, and the cause of the surficial darkening that is so apparent in comparison with ejecta near relatively young craters. The Apollo samples revealed a complexity far beyond the capability of any robotic instruments at the time to decipher: detailed studies including bulk chemistry, isotope chemistry, and electron microscopy showed that the regolith is not simply comminuted bed rock but is the result of eons of impact-induced brecciation and melting, vaporization and vapor re-deposition, and grain-surface changes induced by radiation and solar-wind-ion implantation. One early result following the return of the first Apollo samples was the recognition that the lunar regolith is not just highly comminuted lunar rocks, but has a significantly lower albedo and very unusual spectral properties (e.g., McCord and Adams 1973). This unusual regolith is produced from the uppermost lunar surface by processes that came to be loosely called "space weathering." Although it was recognized early that glass welded aggregates in regolith, or *agglutinates*, contain abundant extremely fine-grained metallic iron (e.g., Housley et al. 1973), the iron was believed to be widely dispersed and the complex agglutinate glass itself was responsible for the dominant optical properties of the lunar "soil." As laboratory spectrometers improved, it was later shown that it is actually the very fine-grained fraction of the lunar soil, not agglutinates, that dominates the optical properties and that the optically active component must occur on the surface of soil grains (Pieters et al. 1994). Recent work has taken advantage of technological advances in the field of transmission electron microscopy to show that the smallest iron metal particles in fact are concentrated within very thin rims on the surfaces of individual grains, even grains that intrinsically contain little iron (e.g., feldspar). The grain rims are generally ≤ ~200 nm thick, and the iron particles themselves are typically < 10 nm (Keller and McKay 1997).

These observations led in turn to the conclusion that the nanophase reduced iron is the result of impact vapor condensation or sputter deposition from nearby grains (Keller and McKay 1997), and contributed to the new understanding that the distinctive optical properties of the lunar regolith is in fact due mainly to the nanometer-sized iron metal on soil grains (reviews by Pieters et al. 2000; Hapke 2001). This model has now been extended to include alteration effects on S-asteroid regoliths as well (Pieters et al. 2000; Hapke 2001). The critical point, stated very clearly by Hapke (2001) in his review, was that even the transmission-electron-microscopy studies of 25 years ago showing the existence of tiny iron grains in the lunar agglutinate glass were inadequate to recognize that the iron is a ubiquitous surficial feature derived from deposition onto the grains. Only the modern analytical TEM methods (e.g., Keller and McKay 1997) sufficed to establish this fact. Studies of anticipated returned samples of martian, asteroidal, or other regoliths will require similar levels of detailed analysis to reveal their likely very different origin and evolution. For the foreseeable future, such analysis can occur only in the confines of a terrestrial laboratory.

Common themes. There are two common threads running through the science accomplishments unique to sample return. The first is the need not just for highly sensitive analytical equipment, operating without severe power and mass restrictions, but also that the analyses are able to be conducted under highly controlled conditions on samples that are ideally prepared, and with unlimited possibility for repeatability. Such conditions are necessary for valid science.

The continued importance of petrography to understanding planetary materials is manifest by the considerable effort that is currently underway to develop a robotic petrographic instrument, e.g., by using a laser Raman spectrometer with a small spot size on a rock surface (Haskin et al. 1997; Wang et al. 2003).

4.1.1. From remote sensing. Virtually all of our global information about the Moon comes from remote observations of gravity, topography, surface morphology, and composition. The only exception is the limited *in situ* seismic array emplaced during Apollo, which was able to obtain hints of the structure of the deep interior from random meteorite impacts and "moonquakes" induced by tidal stress. Even such fundamental lunar evolution concepts as the magma ocean would be meaningless without the recognition from remote data that vast areas of the exposed lunar crust consist of feldspathic dominated lithologies. The geologic context of the samples collected by Apollo and Luna have been critical to their scientific value.

That said, by today's standards of remote sensing data, exemplified by that being accomplished in the Mars Program, remote observations of the Moon are extremely primitive. Recall that the Apollo and Luna programs were accomplished in the late 1960's and early 1970's, an era without the benefit of even simple CCD detectors. With the advent of Clementine and Lunar Prospector decades later in the 1990's, the amount of geophysical and compositional data for the Moon improved significantly (any delta above zero is enormous), but the quality and detail of current lunar remotely acquired data is nevertheless embarrassingly below the standards expected of modern sensors.

This irony is mitigated by the fact that a *little* data goes a long way in a void when there is much to discover. We have learned some extremely fundamental information about the character of the Moon from this recent remotely acquired data. The data have allowed a host of additional interesting hypotheses of a more focused nature to be explored: diversity of mare basalts; impact processes; crustal evolution; local and regional geology; etc. Highlights of key lessons learned from remote sensing of the Moon of a global nature are summarized below. Much of the details, especially from Clementine and Lunar Prospector results, are discussed in Chapter 2.

Coverage is essential. If we have learned anything in the decades since Apollo, it is that ignorance is indeed bliss—and often wrong. All our models of lunar evolution were predicated on limited knowledge. Now we must rewrite aspects of early lunar evolution (and the implied consequences for the rest of the terrestrial planets) to accommodate the non-uniform distribution of heat-producing elements, the implications of the giant SPA basin, and the volatile deposits at the poles.

Diversity of approaches is fundamental to a balanced program of planetary exploration. There is no single experiment that will provide data to resolve fundamental problems like the origin of the Moon or even the origin of polar volatiles. Few interesting problems are that simple. But each step, each set of remote observations with different instruments, provides key boundary conditions to the unknown.

Capabilities (spectral resolution) cannot be compromised. Simple instruments provide simple data. We must bear in mind, however, that the worlds we explore are complex. If we measure the world in two "colors," that is all we will see. Technology advancements for most remote sensors have been astounding in expanding capabilities. Many modern developments, for example, are centered on vastly improved spectroscopic analyses. We are not providing the customer (the paying public) with the best value if we do not use the best that have been developed for remote-sensing purposes.

Spatial resolution sets the types of problems addressed. Every remote measurement made of the Moon spawns additional questions that require higher spatial resolution for resolution (the words are related). What is it, where is it, and how did it get to be that way? The cycle continues with each step.

Precision is essential. Bad data (poorly calibrated; low S/N; inadequate digitization, etc.) is usually worse than no data.

4.1.3. From combination of samples and remote sensing (1 + 1 + 1 >> 3). Our knowledge about the Moon and its evolution has progressed through an intimately interwoven series of events that have involved both remote sensing and analysis of returned samples. Much of this progression was driven by history, politics, and the technical tools available at the time: The sequence of programs started with orbital, *in situ*, and sample return at the dawn of our first venture into space: Ranger, Surveyor, Lunar Orbiter, Apollo, Luna, …. [the big gap]……. Galileo, Clementine, Lunar Prospector. Our perspectives today are intimately colored by the progression and it is difficult, if not impossible, to identify the single source of a specific insight or breakthrough. There is nevertheless no question that we would be ignorant about the body as a whole if we had only the remote data or only the sample data with which to work. The items below are just a few examples of how the whole is *much* greater than the sum of individual parts.

Assessment of principal global properties. The returned Apollo and Luna samples showed that material from the high albedo "highlands" is feldspathic and that material from the dark "maria" is basaltic in overall character. Even the rudimentary remote gamma-ray and X-ray measurements from the Apollo Command Module showed such a pattern to be universal across the equatorial path. This first-order assessment of lunar rock types is so ingrained in our thinking that it is difficult to recognize that it is actually a full marriage between remote measurements and sample return.

Validation of magma ocean concept. The hypothesis of a magma ocean early in lunar evolution came directly from analyses of returned lunar samples. For this hypothesis to be valid, it necessarily must be consistent with global compositional data. First order assessments (feldspathic highlands, basaltic maria) have kept the magma-ocean hypothesis intact, supported by later similar global data. A detailed physical model, however, requires more complete and detailed measurements. The new remote data on asymmetric distribution of KREEP, for example, illustrates how additional data upsets the equilibrium achieved with limited data (both samples and remote sensing). Large issues such as understanding the magma ocean and early planetary evolution depend on a cyclical feedback between remote measurements and sample analyses.

Identification of the character of the lunar regolith. Before spacecraft landed on the Moon, there was definitely what could be called wild speculation about the physical and chemical properties of surface material. That is of course history now, but without the return of the lunar samples, it is likely we would still be formulating hypotheses about the unusual optical properties of lunar soils. Now that we know and understand the soil's unusual characteristics, we are able to use remote-sensing measurements with confidence.

Identification of the "dark mantling material:" a case study. The last Apollo mission was selected to be on the rim of a large basin near unusual terrain that appeared to be mantled by some mysterious "dark mantling material." This mantled area was distinct in remote measurements, but its origin was totally unknown. Even the age of this material relative to surrounding units was debatable. An unplanned discovery during Apollo 17 by geologist-astronaut Jack Schmitt led to a core sample that turned out to be the key. The core was filled with unusual particles of black and orange spheres that had the same properties as the dark mantle (Pieters et al. 1973, 1974). Subsequent detailed laboratory analyses of the samples identified them to be pyroclastic particles with volatile-enriched surfaces and a composition derived from a relatively primitive source (e.g., Heiken and McKay 1977, 1978). This direct link between the remote sensing measurements and the samples allowed a new type of lunar material to be recognized and several areas mapped across the surface.

4.2. Lessons learned

The Moon is the only extraterrestrial body in the solar system to which we have sent the two ultimate exploratory tools: humans, and landed sample return. As a result we understand more about the Moon's complex history than any body in the solar system other than Earth. Even that point can be disputed with some scientists holding the opinion that we understand the Moon better than the Earth. A reader might therefore expect that a concluding chapter in a book dedicated to our modern understanding of the Moon, especially a chapter with the title "Lessons Learned", would be a two-note symphony about the necessity of human exploration and sample return for other solar system bodies as well. In fact, the conclusion is not nearly so simple. Sample return (even robotic) and human exploration are difficult and very costly, even for the Moon, and implementing such programs for every rocky or icy body we choose to explore is simply not practical. It may be expected that the ultimate human destiny is to colonize space, but this destiny might be centuries away. In the short term we will need to rely largely on robotic exploration of our solar system. And landed robotic sample return missions (as opposed to missions such as Genesis or Stardust) will always be more difficult and costly than robotic *in situ* or orbital missions. So although human exploration coupled with sample return unquestionably results in the most profound scientific understanding of a planet, both options must be judiciously exercised and timed. But they *are* essential.

For the contributors to this volume, the "New View" of the Moon that has resulted from the entire range of exploratory tools and cross-disciplinary studies has been eye-opening. It points compellingly to a model for how future planetary exploration should be conducted, because we now know how to do it right: what kinds of missions, and (in hindsight) in what order. The lesson is timely: thirty years after the Viking landings on Mars, there still is not a firm commitment to a sample return mission. Global chemical and mineralogical mapping of the martian surface is only now beginning. And, only now is the kind of logical mission queue planning underway for Mars that the lunar experience suggests is needed.

This section, therefore, is about using the lunar experience as a model for exploration of other rocky bodies in the solar system: the "lessons learned." The first four "lessons" are programmatic ones. That is, they bear on how to explore a silicate body. The last two lessons are profound scientific ones that will apply to most future exploration of our solar system.

4.2.1. Lesson #1: Remote sensing, in situ, and sample return are ALL required. *A comprehensive and balanced program for exploring a terrestrial body requires orbital remote sensing, landed in situ measurements, and return of samples to Earth-based laboratories.* There is no single "silver bullet" for deciphering a planet. Orbital remote sensing in the absence of landed *in situ* exploration and laboratory analysis of returned samples can provide only the largest scale kinds of visual and spectral data, and which are difficult to interpret unambiguously without detailed knowledge acquired "on the ground." Conversely, returning samples without knowing in advance about the global chemistry and mineralogy runs the risk of those samples being systematically nonrepresentative in the way that the Apollo samples are. The Moon is certainly not unusual in this regard, considering that a landed "sample return" on Earth would miss entirely the most abundant rock on the planet—sea floor basalt. Mars' Tharsis Plateau seems to be a similar large scale anomaly on the surface of that planet. We have to assume that every other rocky body in the solar system is heterogeneous in its own way. Moreover, the need for multiple kinds of analysis holds true at every scale, from the largest (planetary) scale down to the scale of a single crystal. Consider that even in a terrestrial laboratory, no one analytical method, no matter how sensitive, and even if carried out under the most highly controlled conditions, is sufficient for understanding the genesis of a complex rock. Multiple techniques, accompanied by careful petrographic studies, are always required for meaningful laboratory analyses of geologic samples. The same holds true for orbital and landed *in situ* missions as well. The ability to conduct multiple experiments, each serving as checks on the others, is essential.

The proper model, or paradigm, for exploration of a planet or moon is field geologic mapping here on Earth. A first season to a new area consists of reconnaissance, usually based on aerial photography as well as surficial (topographic) maps, in which the main goal is to identify major distinct units and bring back representative samples of those units. Laboratory studies of the samples following the field season results in much better understanding of what the units are, their relationships to each other, and some sense of the geologic history of the field area. Work during the second field season involves detailed mapping of the principal units, determination of essential geologic structures, and detailed sampling of units to reveal any systematic variations (chemical, mineralogic, sedimentary, or otherwise). The process is iterative, and so too should be the exploration of another world. Although much can be learned by single remote sensing missions followed by a single sample return mission, the best information especially for a geologically-complex body requires multiple missions. The lunar experience also suggests that, in order to make most efficient use of sample return, global (orbital) surface chemistry/ mineralogy should precede sample return. In the case of Apollo, the difficulties and danger of landing on the far side of the Moon may have resulted in doing things the way we did even if a global chemical map had been available. Nevertheless, it was not known in advance that all of the Apollo landing sites would be within or close to the chemically anomalous Procellarum KREEP Terrane. For the most part, we have done all the right things during our exploration of the Moon, but we have not necessarily done them in the most logical order. Early exploration of the Moon, up to and including Apollo, was driven as much by national political goals as by science goals. This was especially true of the imperative to land humans on the Moon by 1970.

4.2.2. Lesson #2: Expect the unexpected. From start to finish, lunar exploration provided astonishing surprises. The geologic record on the Moon turned out to be nothing like that on Earth, and many predictions about the Moon turned out to be wrong. Immediately after the first Apollo missions, analyses of returned samples revealed that the lunar chemistry, complex regolith, oxidation state, mineralogy, and petrologic diversity were in general unexpected. Later, Clementine and Lunar Prospector revealed the Moon's global dichotomy. It is likely that every planet we explore will hold unique surprises. We already know that Mars differs from both Earth and the Moon in very macroscopic ways: like the Moon it retains the scars of a long bombardment history, suggesting an absence of global crustal recycling, yet is also has an oxidizing atmosphere and did—and perhaps still does—have a hydrosphere. Thus, it seems unlikely that Mars' regolith will retain many of the small-scale signatures of impact processes the way that the lunar regolith does. But, especially considering that no known samples of martian regolith are represented in the meteorite collections, we have little idea what to expect in that regolith other than its apparently highly-oxidized state. Similarly, much remains to be learned about how the martian crust has evolved or what igneous rock types commonly occur there other than basalt and ultramafic rocks. If these rocks are derived from a mantle that has undergone fractionation, it may be very difficult to see back through such events to a primitive martian composition. We know far less about other planets, moons, and asteroids in the solar system. Exploration programs should be planned with no preconceptions. In particular, utilizing only orbital global imaging and hyperspectral studies, even in conjunction with sophisticated *in situ* landers, is likely to provide a very incomplete or even misleading picture of the body. They must be followed up with sample return.

4.2.3. Lesson #3: There is no substitute for the ultimate mobile sensor: a human. Robotic landers are becoming increasingly sophisticated: they can "see" things in ways the human eye cannot. Thus there is no doubt that robotic exploration—being both less costly and less risky then human exploration—is a practical way of exploring more bodies in the solar system and more places on each individual body. Yet if the Apollo experience taught anything, it is that the human ability to recognize interesting features quickly and then independently act to follow up on that information can lead to important discoveries. The Apollo astronauts received extensive geology training prior to their missions, as a result of which they were able to recognize features that were unusual relative to the local surroundings.

One of the most famous examples occurred during the Apollo 15 mission, when a suspiciously white rock was spotted sitting perched atop a small pedestal of soil (Fig. 7.22). The "Genesis Rock", 15415, turned out to be nearly pure sample of the kind of anorthosite that Wood et al. (1970) had predicted might compose the lunar highlands. The astronauts were on the lookout for anorthosite, and immediately recognized 15415 for what it was and collected it. It is quite possible, even likely, that a robotic rover on the surface of the Moon would have collected this sample once the human controllers on earth spotted it from panoramic images; unquestionably, 15415 stands out from it surroundings on account of its white color. However, the near-instantaneous recognition and collection of the sample by the astronauts is nothing like the laborious process of finding it with a rover and then programming the rover to travel to and collect the sample. Humans are completely autonomous and process large amounts of complex information very quickly.

Another example, which illustrates a very different advantage of human explorers, is the story of the "seatbelt basalt," 15016. During a transit by the Apollo 15 crew back to the lunar module, astronaut Scott spotted an unusually vesicular rock along the way. Judging that they would not be given permission to make an unscheduled stop for the sole purpose of picking up a sample, Scott instead stopped to "fix a seatbelt." At the end of the stop, the sample coincidentally ended up inside the LEM (Scott's pocket, actually). Outside of science fiction, robotic rovers do not fib or otherwise evade instructions in order to take advantage of a scientific find. Humans are independent and creative.

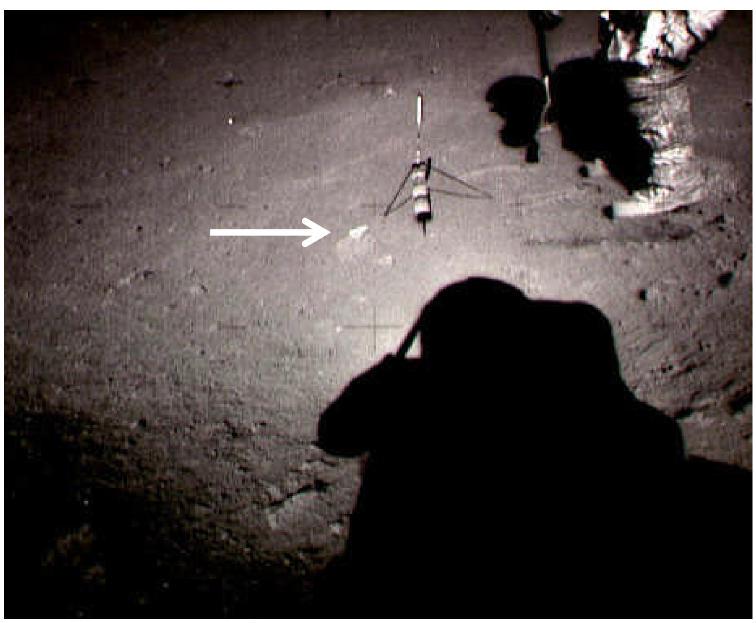

Figure 7.22. The Genesis Rock, 15415, sitting on its perch just prior to collection by the A15 crew. [NASA photo AS15-90-12227]

A third famous example is the collection of the orange glass by astronaut Schmitt on Apollo 17. In the process of investigating an interesting boulder, Schmitt noticed that his walking and scuffing of the ground had exposed orange soil underneath a centimeter-thick gray surface layer. Although orange soil subsequently was found to be exposed on the surface nearby, the lighting made it difficult to recognize; Schmitt's attention was drawn to it because of the footprint exposures. Pronounced colors are rare on the Moon, and orange in particular suggested (erroneously, as it turned out) the possibility of oxidized iron. Thus this famous discovery raised considerable excitement. It is debatable whether a robotic rover might have made the discovery at all. The clue was visual color alone, under difficult lighting conditions. A geologist's instinctive attraction to an unexpected and potentially significant color, and his ability to instantly act upon a serendipitous discovery, led to the collection of the orange glass soil.

The point of these anecdotes is not to advocate that all planetary exploration should be done by human missions—that is not realistic. Rather, it is to emphasize the kinds of discoveries that are possible when humans are present: ones that require quick decisions, astute judgment arising from intense training, and the ability and willingness to take quick advantage of serendipity (even if it means sometimes using subterfuge to do so!). Our knowledge of the Moon would be much poorer if we had not sent astronauts. There is no question that similar discoveries will be made when (hopefully) we send astronauts to Mars.

4.2.4. Lesson #4: Science requires exceptionally sensitive instrumentation. Achieving the best science requires the most precise and sensitive analytical instrumentation, whether terrestrial laboratory-based or spacecraft-based. One of the great successes of the Apollo program was NASA's far-sighted investment in building up a new generation of laboratory instrumentation in terrestrial labs prior to the return of the first samples. Given what turned out to be the very low volatile element inventory of lunar samples, the new instrumentation was essential in accurate determinations not only of elemental compositions but also for precise radiometric age determinations. Refined techniques for activation analysis also permitted trace element determinations for very small samples. Had NASA not implemented the instrumentation program ahead of time, much valuable sample and time would have been wasted. As an added bonus, the new laboratory instruments fortuitously were in place when two extremely important meteorites (Allende and Murchison) fell in the same year in which the first lunar samples came back. Because work on these chondritic meteorites entailed analyzing individual components (chondrules, calcium-aluminum-rich inclusions, matrix) rather than bulk meteorite samples, the ability to do highly sensitive and precise measurements on the minute masses of material was perhaps even more critical than for the Apollo samples. We are now at a similar critical juncture in space exploration. The first Mars surface-sample return is likely to bring back 0.5 kg or less of material that must satisfy the demands of the entire world scientific community. The only laboratories that will be able to compete successfully for sample allocations are those suitably equipped to work on very small amounts of material, and few are fortunate enough to be at the leading edge of analytical technology. For some foreseeable missions, <u>no</u> laboratories have suitable equipment for the simple reason that no current analytical technology is capable of making the critical measurements. For example, the Genesis (Discovery class) mission to collect implanted solar wind has as one of its most important goals the determination of the oxygen isotopic composition of the solar wind. Instruments to do this are (as of this writing) under development with NASA funding, but they do not yet exist and when they do only one or two labs will have them. The comet dust that will be collected by the Stardust mission (also Discovery class) can be analyzed by many currently available techniques, but relatively few laboratories are equipped with state-of-the-art equipment that can do so. For both of these missions, the amount of material that will be available to satisfy the world demand is tiny, and only the most capable laboratories will compete successfully for samples. A consortia approach employing the skills of several laboratories might be employed, as was successfully done with some Apollo samples. Fortunately a new funding program to correct this deficit has been put in place by NASA, but only recently and (arguably) just barely in time. The lesson is that studying astromaterials is intensely instrument-dependent, and there must always be an ongoing program to extend the analytical state-of-the-art and make it available to many investigators.

This lesson does not apply to laboratory instruments alone, although the lesson does not actually derive from past lunar experience. For some time there has been a disconnect between the design of spacecraft instrumentation and the actual scientific requirements for what those instruments need to do to make meaningful measurements. Worse, currently there is no requirement for proposed space-flight instruments to prove their performance capabilities—precision, accuracy, and sensitivity—prior to being selected for flight. The result is that instruments can fly without necessarily being able to return highly meaningful data. Robotic *in situ* landers are going to be one of the chief means of exploring extraterrestrial bodies, and it is critical that the analytical instruments they carry be capable of providing remote sensing data that seriously addresses important problems and can be interpreted confidently by the scientists back on Earth. The counterexample is the experience with the Mars Pathfinder Sojourner rover, whose major-element chemical data remain ambiguous to this day on account of poor spectral resolution and lack of standardization prior to flight. Space-flight analytical instruments must be developed to satisfy the precision and accuracy requirements of specific scientific problems. Perhaps there should be a programmatic requirement that all such instruments undergo double-blind tests on Earth using standard materials in order to establish performance characteristics, before they are selected for flight.

Although analysis of samples in terrestrial laboratories will almost always be superior to what is possible with analysis of samples by spacecraft *in situ* instruments, it must be stressed that such spacecraft instruments can and must be made much better. For example, the laboratory state-of-the-art at the time the Apollo samples were first returned included such capabilities as major elemental and mineralogical analysis at the scale of tens of microns; these capabilities are today feasible for robotic analysis *in situ* on planetary surfaces, but such instruments are not yet flight-certified. These and other analytical capabilities must be constantly improved for flight instruments. It is imperative that the scientific community—especially those who do actual laboratory analysis of planetary materials and thus understand the real state-of-the-art—work closely with instrument developers and communicate challenging but achievable science requirements and technical goals. Together with a more coordinated and vigorous program of instrument development funding and testing than currently exists, this will help close the gap between what is available and what is possible.

Finally, two profound scientific lessons were learned from the lunar experience that are worth repeating here, even though they have been emphasized by so many others in so many places that they are understood by even the most beginning students of planetary geology:

(1) Basaltic volcanism is a widespread phenomenon and a natural consequence of melting and differentiation of a body having chondritic (solar) bulk composition. The magnificent *Basaltic Volcanism Study Project* volume (BVSP 1984) is eloquent testimony to this fact.

(2) Impacts are profoundly important in the evolution of planetary surfaces. Even for planets (e.g., Earth and Venus) where much of the long record of bombardment has been obliterated, we now understand that impacts played an enormous role. For bodies where the record is preserved, notably the Moon, understanding the effects of eons of impacts is essential to interpreting geochemical data. Identifying primary lunar lithologies can be difficult even in the lab. Doing so with robots and remote-sensing instrumentation would be more so, and failing to recognize the hybrid nature of (e.g.) martian rocks on a robotic mission would render certain kinds of measurements, such as age determinations, nearly meaningless.

4. SUMMARY

As the stepping stone, cornerstone, capstone, and keystone of planetary science, exploration of the Moon has provided the foundation, inspiration, and bridge to understanding the planets of our solar system. These concepts are highlighted with countless examples throughout various chapters of this book.

Much of our detailed geochemical understanding of this extraterrestrial neighbor comes, of course, from the samples returned during the Apollo era. The more recent remote measurements of Clementine and Lunar Prospector nevertheless jolted a complacent community. The decades-long neglect of lunar exploration after Apollo stunted what had been an enormous growth of knowledge in the years immediately following the return of the initial samples. The small pulse of new data in the early 1990's made an enormous impact. The "New View" of possible accumulation of water ice at the lunar poles, the unique character of the enormous South Pole-Aitken basin on the farside, and the asymmetric concentration of heat-producing elements are fundamental aspects of lunar science whose importance was unrecognized a decade ago. All three of these paradigm shifts in our understanding beg explanations. They remind us in no uncertain terms, that our knowledge is terribly incomplete.

Clearly, remote observations, *in situ* measurements, and sample return all play a significant role in exploration of planetary bodies. There is ample evidence that to extend exploration beyond simple reconnaissance in fact requires both *in situ* and orbital measurements as well as—and integrated with—analysis of returned samples. There are arguments (and examples) about which sequence is the most efficient. On one hand, detailed remote analyses coupled with targeted *in situ* measurements allow assessment of regional properties and selection of sites that will optimize the value of samples returned from limited targets. On the other hand, the enormous amount of information inherent in the detailed analyses performed on samples in Earth-based laboratories has repeatedly identified broad-scale issues that were unimagined with the more limited remote measurements. A useful strategy could involve two stages. Firstly a simple (and less costly) grab sample from the surface (regolith) that would enable an interpretation of orbital data. The second, more costly phase would target specific localities identified from this combination of orbital and *in situ* measurements.

Like most areas of science, discovery and progress in understanding how planets work is cyclical in nature. As valuable data are first acquired in one area, the information derived lays the foundation for all subsequent questions and the definition of new steps. The danger, of course, is that our ignorance can never tell us what we have missed. We must make the best use of lessons learned.

5. REFERENCES

Adams JB, McCord TB (1971) Alteration of lunar optical properties: Age and composition effects. Science 171:567-571

Adams JB, McCord TB (1971) Optical properties of mineral separates, glass, and anorthositic fragments from Apollo mare samples. Proc Lunar Sci Conf 2:2183-2195

Adams JB, McCord TB (1973) Vitrification in the lunar highlands and identification of Descartes material at the Apollo 16 site. Proc Lunar Sci Conf 4:163-177

Anders E (1977) Chemical compositions of the Moon Earth and eucrite parent body. Phil Trans R Soc A 285:23-40

Anderson JD, Sjogren WL, Schubert G (1996) Galileo gravity results and the internal structure of Io. Science 272: 709-712

Arnold JR (1979) Ice in the lunar polar regions. J Geophys Res 84:5659-5668

Belton MJS, Greeley R, Greenberg R, McEwen AS, Klaasen KP, Head JW III, Pieters CM, Neukum G, Geissler PE, Heffernan C, Breneman H, Anger CD, Carr MH, Davies ME, Fanale FP, Gierasch PJ, Ingersoll AP, Johnson TV, Pilcher CB, Thompson R, Veverka J, Sagan C (1994) Galileo multispectral imaging of the North Polar and Eastern Limb regions of the Moon. Science 264:1112-1115

Belton MJS, Head JW III, Pieters CM, Greeley R, McEwen AS, Neukum G, Klaasen KP, Anger CD, Carr MH, Chapman CR, Davies ME, Fanale FP, Gierasch PJ, Greenberg R, Ingersoll AP, Johnson TB, Paczkowski B, Pilcher CB, Veverka J (1992) Lunar impact basins and crustal heterogeneity: New western limb and farside data from Galileo. Science 255(5044):570-576

Binder AB (1998) Lunar Prospector: Overview. Science 281:1475-1476

Binzel RP, Xu S (1993) Chips off Asteroid 4 Vesta: Evidence for the parent body of basaltic achondrite meteorites. Science 260:186-191

Blewett DT, Hawke BR, Lucey PG, Taylor GJ, Jaumann R, Spudis PD (1995) Remote sensing and geologic studies of the Schiller-Schickard region of the Moon. J Geophys Res 100:16,959-16,997

Blewett DT, Lucey PG, Hawke BR (1997) A comparison of mercurian reflectance and spectral quantities with those of the Moon. Icarus 129:217-231

Borg L, Norman M, Nyquist L, Bogard D, Snyder G, Taylor L, Lindstrom M (1999) Isotopic studies of ferroan anorthosite 62236: A young lunar crustal rock from a light-rare-earth-element-depleted source. Geochim Cosmochim Acta 63:2679-2691

Boss AP, Cameron AGW, Benz W (1990) Tidal disruption of inviscid protoplanets. Lunar Planet Sci XXI:117-118, Lunar and Planetary Institute, Houston

Boss AP, Peale SJ (1986) Dynamical constraints on the origin of the Moon. *In*: Origin of the Moon. Hartmann WK, Phillips RJ, Taylor GJ (eds) Lunar and Planetary Institute, p 59-101

Brearley A, Jones JH (1998) Chondritic meteorites. Rev Mineral 36:3-01–3-398

Britt DT, Pieters CM (1994) Darkening in black and gas-rich ordinary chondrites: The spectral effects of opaque morphology and distribution. Geochim Cosmochim Acta 58:3905-3919

Brophy JG, Basu A (1989) Europium anomalies in mare basalts as a consequence of mafic cumulate fractionation from an initial lunar magma. Proc Lunar Planet Sci Conf 20:25-30

Burbine TH, Buchanon PC, Binzel RP, Bus SJ, Hiroi T, Hinrichs JL, Meibom A, McCoy TJ (2001) Vesta, vestoids, and the howardite, eucrite, diogenite group: Relationships and the origin of spectral differences. Meteor Planet Sci 36:761-781

Bussey DBJ and Spudis PD (1997) Compositional analysis of the Orientale basin using full resolution Clementine data: Some preliminary results. Geophys Res Lett 24:445-448

BVSP (1981) Basaltic Volcanism on the Terrestrial Planets. Pergamon

Cameron AGW (1986) The impact theory for the origin of the Moon. *In*: Origin of the Moon. Hartmann WK, Phillips RJ, Taylor GJ (eds) Lunar and Planetary Institute, p 609-616

Cameron AGW (2000) Higher-resolution simulations of the giant impact. *In*: Origin of the Earth and Moon Canup RM, Righter K (eds) Univ Arizona Press, p 133-144

Cameron AGW (2001) From interstellar gas to the Earth-Moon system. Meteorit Planet Sci 36:9-22

Campbell BA, Campbell DB, Chandler JF, Hine AA, Nolan MC, Perillat PJ (2003) Radar mapping of the lunar poles. Nature 426:137-138

Canup R, Asphaug E (2001) Origin of the Moon in a giant impact near the end of the Earth's formation. Nature 412: 708-712

Canup RM (2004) Simulations of a late lunar-forming impact. Icarus 168:433-456

Canup RM, Righter K (2000) Accretion of the terrestrial planets. *In*: Origin of the Earth and Moon Canup RM, Righter K (eds) Univ Arizona Press, p 113-129

Canup RM, Righter K (eds) (2000) Origin of the Earth and Moon. Univ Arizona Press

Carlson RW, Lugmair GW (2000) Timescales of planetesimal formation and differentiation based on extinct and extant radioisotopes. *In*: Origin of the Earth and Moon Canup RM, Righter K (eds) Univ Arizona Press, p 25-44

Clayton RN, Mayeda TK (1963) The use of bromine pentafluoride in the extraction of oxygen from oxides and silicates for isotopic analysis. Geochim Cosmochim Acta 27:43-52

Clayton RN, Mayeda TK (1975) Genetic relations between the Moon and meteorites. Proc Lunar Sci Conf 6:1761-1769

Clayton RN, Mayeda TK, Rubin AE (1984) Oxygen isotopic compositions of enstatite chondrites and aubrites. J Geophys Res 8:C245-C249

Drake MJ (1986) Is lunar bulk material similar to Earth's mantle? *In*: Origin of the Moon. Hartmann WK, Phillips RJ, Taylor GJ (eds) Lunar and Planetary Institute, p 105-124

Drake MJ, Righter K (2002) Determining the composition of the Earth. Nature 416:39-43

Durisen RH, Gingold RA (1986) Numerical simulations of fission. *In*: Origin of the Moon. Hartmann WK, Phillips RJ, Taylor GJ (eds) Lunar and Planetary Institute, p 487-498

Elphic RC, Lawrence DJ, Feldman WC, Barraclough BL, Maurice S, Binder AB, Lucey PG (2000) Lunar rare earth element distribution and ramifications for FeO and TiO_2: Lunar Prospector neutron spectrometer observations. J Geophys Res 105:20,333-20,345

Eriksson KA (1977) Tidal deposits from the Archaean Moodies Group, Barberton Mountain Land, South Africa. Sediment Geol 18:257-281

Eriksson KA, Simpson EL (2000) Quantifying the oldest tidal record: The 3.2 Ga Moodies Group, Barberton Greenstone Belt, South Africa. Geology 28:831-834

Eshleman VR, Parks GA (1999) No ice on the Moon? Science 285:531
Feldman W, Maurice S, Lawrence DJ, Little RC, Lawson SL, Gasnault O, Wiens RC, Barraclough BL, Elphic RC, Prettyman TH, Steinberg JT, Binder AB (2001) Evidence for water ice near the lunar poles. J Geophys Res 106: 23,231-23,251
Feldman WC, Lawrence DJ, Elphic RC, Barraclough BL, Maurice S, Genetay I, Binder AB (2000) Polar hydrogen deposits on the Moon. J Geophys Res 105:4175-4195
Feldman WC, Maurice S, Binder AB, Barraclough BL, Elphic RC, Lawrence DJ (1998) Fluxes of Fast and Epithermal Neutrons from Lunar Prospector: Evidence for water ice at the lunar poles. Science 281:1496-1500
Gaffey MJ (1997) Surface lithologic heterogeneity of Asteroid 4 Vesta. Icarus 127:130-157
Gaffey MJ, Burbine TH, Binzel RP (1993) Asteroid spectroscopy, progress and perspectives, Meteoritics 28:161-187
Gast PW (1972) The chemical composition and structure of the Moon. Moon 5:121-148
Gonzalez G (1998) Spectroscopic analyses of the parent stars of extrasolar planetary system candidates. Astronom Astrophys 334:221-229
Halliday AN, Lee D-C, Jacobsen SB (2000) Tungsten isotopes, the timing of metal-silicate fractionation, and the origin of the Earth and Moon. *In*: Origin of the Earth and Moon. Canup RM, Righter K (eds) Univ Arizona Press, p 45-62
Hapke B (2001) Space weathering from Mercury to the asteroid belt. J Geophys Res 106:10,039-10,073
Hapke B, Cassidy W, Wells EN (1975) Effects of vapor-phase deposition processes on the optical, chemical and magnetic properties of the lunar regolith. The Moon 13:339-353
Harmon JK, Slade MA, Velez RA, Crespo A, Dryer MJ, Johnson JM (1994) Radar mapping of Mercury's polar anomalies. Nature 369:213-215
Hartmann WK, Ryder G, Dones L, Grinspoon D (2000) The time-dependent intense bombardment of the primordial Earth/Moon system. *In*: Origin of the Earth and Moon. Canup RM, Righter K (eds) Univ Arizona Press, p 493-512
Haskin LA (1998) The Imbrium impact event and the thorium distribution at the lunar highlands surface. J Geophys Res 103:1679-1689
Haskin LA, Gillis JJ, Korotev TL, Jolliff BL (2000) The materials of the lunar Procellarum KREEP Terrane: A synthesis of data from geomorphological mapping remote sensing and sample analyses. J Geophys Res 105:20403-20415
Haskin LA, Wang A, Rockow KM, Jolliff BL, Korotev RL, Viscupic KM (1997) Raman spectroscopy for mineral identification and quantification for *in situ* planetary surface analysis: A point-count method. Journal of Geophysical Research. J Geophys Res 103:19,293-19,306
Head JW (1976) Lunar volcanism in space and time. Rev Geophys Space Phys 14:265-300
Heiken G, McKay DS (1977) A model for eruption behavior of a volcanic vent in eastern Mare Serenitatis. Proc Lunar Sci Conf 8:3243-3255
Heiken G, McKay DS (1978) Petrology of a sequence of pyroclastic rocks from the valley of Taurus-Littrow (Apollo 17 landing site). Proc Lunar Sci Conf 9:1933-1943
Heiken G, Vaniman D, French BM (eds) (1991) Lunar Sourcebook, Cambridge University Press
Hood LL, Jones JH (1987) Geophysical constraints on lunar bulk composition and structure: A reassessment. J Geophys Res 92:E396-E410
Hood LL, Mitchell DL, Lin RP, Acuña MH, Binder AB (1999) Initial measurement of the lunar induced magnetic dipole moment using Lunar Prospector magnetometer data. Geophys Res Lett 26:2327
Hood LL, Zuber MT (2000) Recent refinements in geophysical constraints on lunar origin and evolution. I *In*: Origin of the Earth and Moon. Canup RM, Righter K (eds) Univ Arizona Press, p 397-412
Housley RM, Grant RW, Paton NE (1973) Origin and characteristics of excess Fe metal in lunar glass welded aggregates. Proc Lunar Sci Conf 4:2737-2749
Humayun M, Cassen P (2000) Processes determining the volatile abundances of the meteorites and terrestrial planets. *In*: Origin of the Earth and Moon. Canup RM, Righter K (eds) Univ Arizona Press, p 3-23
Humayun M, Clayton RN (1995) Potassium isotope chemistry: Implications of volatile element depletion. Geochim Cosmochim Acta 59:2131-2148
Javoy M (1995) The integral enstatite chondrite model of the Earth. Geophys Res Lett 22:2219-2222
Jeanloz R, Mitchell DL, Sprague AL, de Pater I (1995) Evidence for a basalt-free surface on Mercury and implications for internal heat. Science 268:1455-1457
Jolliff BL, Gillis JJ (2002) Lunar crustal and bulk composition. Workshop on The Moon Beyond 2002: Next Steps in Lunar Science and Exploration, #3056, Taos, New Mexico
Jolliff BL, Gillis JJ, Haskin LA (2000) Thorium mass balance for the Moon from Lunar Prospector and sample data: Implications for thermal evolution. Lunar Planet Sci 31:1763
Jolliff BL, Gillis JJ, Haskin LA, Korotev RL, Wieczorek MA (2000) Major lunar crustal terranes: Surface expressions and crust-mantle origins. J Geophys Res 105:4197-4216
Jones JH, Delano JW (1989) A three-component model for the bulk composition of the Moon. Geochim Cosmochim Acta 53:513-527
Jones JH, Palme H (2000) Geochemical constraints on the origin of the Earth and Moon. *In*: Origin of the Earth and Moon. Canup RM, Righter K (eds) Univ Arizona Press, p 197-216

Kahn A, Mosegaard K (2001) New information on the deep lunar interior from an inversion of lunar free oscillation periods. Geophys Res Lett 28:1791-1794

Keller L, McKay D (1997) The nature and origin of rims on lunar soil grains. Geochim Cosmochim Acta 61:2331-2340

Keller LP, McKay DS (1993) Discovery of vapor deposits in the lunar regolith. Science 261:1305-1307

Khan A, Mosegaard K, Rasmussen KL (2000) A new seismic velocity model for the Moon from a Monte Carlo inversion of the Apollo lunar seismic data. Geophys Res Lett 27:1591-1594

Kirk RL, Stevenson DJ (1989) The competition between thermal contraction and differentiation in the stress history of the Moon. J Geophys Res 94:12133-12144

Kokubo E, Canup RM, Ida S (2000) Lunar Accretion from an Impact-Generated Disk. *In*: Origin of the Earth and Moon. Canup RM, Righter K (eds) Univ Arizona Press, p 145-163

Konopliv AS, Binder AB, Hood LL, Kucinskas AB, Sjogren WL, Williams JG (1998) Improved gravity field of the Moon from Lunar Prospector. Science 281:1476-1480

Korotev RL (2000) The great lunar hot spot and the composition and origin of the Apollo mafic ("LKFM") impact-melt breccias. J Geophys Res 105:4317-4345.

Korotev RL, Haskin LA (1988) Europium mass balance in polymict samples and implications for plutonic rocks of the lunar crust. Geochim Cosmochim Acta 52:1795-1813

Kortenkamp SJ, Kobuko E, Weidenschilling SJ (2000) Formation of planetary embryos. *In*: Origin of the Earth and Moon. Canup RM, Righter K (eds) Univ Arizona Press, p 85-100

Kuskov OL (1997) Constitution of the Moon: 4. Composition of the mantle from seismic data. Phys Earth Planet Inter 102:239-257

Kuskov OL, Kronrod VA (1998) Constitution of the Moon: 5 Constraints on composition density temperature and radius of a core. Phys Earth Planet Inter 107:285-306

Kuskov OL, Kronrod VA (2000) Resemblance and difference between the constitution of the Moon and Io. Planet Space Sci 48:717-726

Laskar J, Joutel F, Robutel P (1993) Stabilization of the Earth's orbit by the Moon. Nature 361:615-617

Lawrence DJ, Feldman WC, Barraclough BL, Binder AB, Elphic RC, Maurice S, Thomsen DR (1998) Global elemental maps of the Moon: The Lunar Prospector gamma-ray spectrometer. Science 281:1484-1489

Lawrence DJ, Feldman WC, Barraclough BL, Binder AB, Elphic RC, Maurice S, Miller MC, Prettyman TH (2000) Thorium abundances on the lunar surface. J Geophys Res 105:20,307-20,331

Lawrence DJ, Feldman WC, Elphic RC, Little RC, Prettyman TH, Maurice S, Lucey PG, Binder AB (2002) Iron abundances on the lunar surface as measured by the Lunar Prospector gamma-ray and neutron spectrometers. J Geophys Res 107(E12):DOI:10.1029/2001JE001530

Lee D-C, Halliday AN, Snyder GA, Taylor LA (1997) Age and origin of the Moon. Science 278:1098-1103

Lipin BR (1976) The origin of Fra Mauro basalts. Lunar Sci VII:495-497

Longhi J (1992) Experimental petrology and petrogenesis of mare volcanics. Geochim Cosmochim Acta 56:2235-2251

Longhi J (2003) A new view of lunar ferroan anorthosites: Post magma ocean petrogenesis. J Geophys Res 108:5083, doi:10.1029/2002JE001941

LSPET (1969) Summary of the Apollo 11 Lunar Science Conference. Science 167:449-451

Lucey PG, Taylor GJ, Hawke BR, Spudis PD (1998) FeO and TiO_2 concentrations in the South Pole-Aitken basin: Implications for mantle composition and basin formation. J Geophys Res 103:3701-3708

Lucey PG, Taylor GJ, Malaret E (1995) Abundance and distribution of iron on the Moon. Science 268:1150-1153

Lugmair G (1987) The age of the lunar crust: 60025—Methuselah's legacy. Lunar Planet Sci XVIII:584-585

Lugmair GW, Shukolyukov A (1998) Early solar system timescales according to $^{53}Mn/^{53}Cr$ systematics. Geochim Cosmochim Acta 62:2863-2886

McCord TB, Adams JB (1973) Progress in remote optical analysis of lunar surface composition. The Moon 7:453-474

McCord TB, Adams JB, Johnson TB (1970) Asteroid Vesta: spectral reflectivity and compositional implications. Science 168:1445-1447

McKay GA (1982) Partitioning of REE between olivine, plagioclase, and synthetic melts: Implications for the origin of lunar anorthosite. Lunar Planet Sci XIII:493-494

McKay GA, Wagstaff J, Le L (1990) REE distribution coefficients for pigeonite: Constraints on the origin of the mare basalt europium anomaly. *In:* Workshop on Lunar Volcanic Glasses: Scientific and Resource Potential. Delano JW, Heiken GH (eds) LPI Tech Rpt. 90-02:48-49

Mitchell DL, de Pater I (1994) Microwave imaging of Mercury's thermal emission at wavelengths from 03-205 cm. Icarus 110:2-32

Morbidelli A, Chambers J, Lunine JI, Petit JM, Robert F, Valsecchi GB, Cyr KE (2000) Source regions and time scales for the delivery of water to the Earth. Meteor Planet Sci 35:1309-1320

Mueller S, Taylor GJ, Phillips R (1988) Lunar composition: A geophysical and petrological synthesis. J Geophys Res 93:6338-6352

Nakamura Y (1983) Seismic velocity structure of the lunar mantle. J Geophys Res 88:677-686

Neal CR (2001) Interior of the Moon: The presence of garnet in the primitive deep lunar mantle. J Geophys Res 106: 27,865-27,855
Neal CR, Taylor LA (1992) Petrogenesis of mare basalts: a record of lunar volcanism. Geochim Cosmochim Acta 56: 2177-2211
Neumann GA, Zuber MT, Smith DE, Lemoine FG (1996) The lunar crust: Global structure and signature of major basins. J Geophys Res101:16481-16863
Newsom HE (1986) Constraints on the origin of the Moon from the abundance of molybdenum and other siderophile elements. *In*: Origin of the Moon. Hartmann WK, Phillips RJ, Taylor GJ (eds) Lunar and Planet. Inst., p 203-229
Newsom HE, Taylor SR (1989) Geochemical implications of the formation of the Moon by a single giant impact. Nature 338:29-34
Noble SK (2004) Turning Rocks into Regolith: The Physical and Optical Consequences of Space Weathering in the Inner Solar System. Brown University, Providence, RI
Noble SK, Pieters CM (2003) Space Weathering on Mercury: Implications for Remote Sensing. Solar System Res 37: 31-35
Noble SK, Pieters CM, Keller LP (2003) The optical properties of nanophase iron: Investigation of a space weathering analog. Lunar Planet Sci 34:1172
Noble SK, Pieters CM, Taylor LA, Morris RV, Allen CC, McKay DS, Keller LP (2001) The optical properties of the finest fraction if lunar soil: Implications for space weathering,. Meteorit Planet Sci 36:31-42
Norman MD, Borg LE, Nyquist LE and Bogard DD (2003) Chronology, geochemistry, and petrology of a ferroan noritic anorthosite clast from Descartes breccia 67215: Clues to the age, origin, structure, and impact history of the lunar crust. Meteor Planet Sci 38:645–661
Nozette S, Lichtenberg CL, Spudis PD, Bonner R, Ort W, Malaret E, Robins M, Shoemaker EM (1996) The Clementine Bistatic Radar Experiment. Science 274:1495-1498
Nozette S, Rustan P, Pleasance LP, Horan D, Regeon MP, Shoemaker EM, Spudis PD, Acton CH, Baker DN, E. BJ, Buratti BJ, Corson MP, Davies ME, Duxbury TC, Eliason EM, Jakosky BM, Kordas JF, Lewis IT, Lichtenberg CL, Lucey PG, Malaret E, Massie MA, Resnick JH, Rollins CJ, Park HS, McEwen AS, Priest RE, Pieters CM, Reisse RA, Robinson MS, Smith DE, Sorenson TC, Vorder Breugge RW, Zuber MT (1994) The Clementine mission to the Moon: Scientific overview. Science 266:1835-1839
Nozette S, Spudis PD, Robinson MS, Bussey DBJ, Lichtenberg C, Bonner R (2001) Integration of lunar polar remote-sensing data sets: Evidence for ice at the lunar south pole. J Geophys Res 106:23,253-23266
Nyquist LE (1977) Lunar Rb-Sr chronology. Phys Chem Earth 10:103-142
Nyquist LE, Shih C-Y (1992) The isotopic record of lunar volcanism. Geochim Cosmochim Acta 56:2213-2234
O'Hara MJ (2000) Flood basalts basalt floods or topless bushvelds? Lunar petrogenesis revisited. J Pet 41:1545-1651
O'Neill HSC (1991) The origin of the Moon and the Earth-A chemical model. Geochim Cosmochim Acta 55:1135-1157
Paige DA (2000) Mars exploration strategies: Forget about sample return! *In*: Concepts and Approaches for Mars Exploration. Lunar Planet Inst Contrib 1062:243-244
Papanastassiou DP, Wasserburg GJ (1971) Lunar chronology and evolution from Rb-Sr studies of Apollo 11 and 12 samples. Earth Planet Sci Lett 11:37-62
Papike JJ (1998) Comparative planetary mineralogy: Chemistry of melt-derived pyroxene, feldspar, and olivine. Rev Mineral 36:7-1–7-11
Papike JJ, Bence AE (1978) Lunar mare basalts versus Mid-Ocean Ridge basalts: Planetary constraints. Geophys Res Lett 5:803-806
Papike JJ, Ryder G, Shearer CK (1998) Lunar Samples. Rev Mineral 36:5-1–5-234
Pieters C, Taylor L, Noble S, Keller L, Hapke B, Morris R, Allen C, McKay D, Wentworth S (2000) Space weathering on asteroids: Resolving a mystery with Lunar samples. Meteorit Planet Sci 35:1101-1107
Pieters CM, Fischer EM, Rode O, Basu A (1993) Optical effects of space weathering: the role of the finest fraction. J Geophys Res 98(E11):20,817-20,824
Pieters CM, Head JW, III, Gaddis L, Jolliff BL, Duke M (2001) Rock types of the South Pole-Aitken basin and extent of basaltic volcanism. J Geophys Res 106:28,001-28,022
Pieters CM, McCord TB, Charette MP, Adams JB (1974) Lunar surface: Identification of the mantling material in the Apollo 17 soil samples. Science 183:1191-1194
Pieters CM, McCord TB, Zisk SH, Adams JB (1973) Lunar black spots and the nature of Apollo 17 landing area. J Geophys Res 78:5867-5875
Pieters CM, McFadden LA (1994) Meteorite and asteroid reflectance spectroscopy: clues to early solar system processes. Ann Rev Earth Planet Sci 22:457-497
Pieters CM, Taylor LA (2003) Systematic global mixing and melting in lunar soil evolution. Geophys Res Lett 30:2048 doi:10.1029/2003GL018212
Pieters CM, Tompkins S (1999) Tsiolkovsky crater: A window into crustal processes on the lunar farside. J Geophys Res 104:21935-21949
Pieters CM, Tompkins S, Head JW, Hess PC (1997) Mineralogy of the mafic anomaly in the South Pole-Aitken Basin: Implications for excavation of the lunar mantle. Geophys Res Lett 24:1903-1906

Premo WR, Tatsumoto M, Misawa K, Nakamura N, Kita NI (1999) Pb-isotopic systematics of lunar highland rocks (>39 Ga): Constraints on early lunar evolution. *In*: Planetary Geology and Geochemistry. Snyder GA, Neal CR, Ernst WG (eds) Geological Society America, p 207-240

Pritchard ME, Stevenson DJ (2000) Thermal aspects of a lunar origin by giant impact. *In*: Origin of the Earth and Moon. Canup RM, Righter K (eds) Univ Arizona Press, p 179-196

Righter K (2002) Does the Moon have a metallic core? Constraints from giant-impact modeling and siderophile elements. Icarus 158:1-13

Righter K, Drake MJ (1996) Core formation in Earth's Moon, Mars and Vesta. Icarus 124:513-529

Righter K, Walker RJ, Warren PH (2000) The origin and significance of highly siderophile elements in the lunar and terrestrial mantles. *In*: Origin of the Earth and Moon. Canup RM, Righter K (eds) Univ Arizona Press, p 291-322

Ringwood AE (1986) Terrestrial origin of the Moon. Nature 322:323-328

Ringwood AE (1990) Earliest history of the Earth-Moon system. *In*: Origin of the Earth. Newsom HE, Jones JH (eds) Oxford Univ Press, p 101-134

Ringwood AE, Kesson SE (1977) Basaltic magmatism and the bulk composition of the Moon, II. Siderophile and volatile elements in Moon, Earth, and chondrites: Implications for lunar origin. The Moon 16:425-464

Ruzicka A, Snyder GA, Taylor LA (1998) Giant impact and fission hypotheses for the origin of the Moon: A critical review of some geochemical evidence. Int Geol Rev 40:851-864

Ruzicka A, Snyder GA, Taylor LA (2001) Comparative geochemistry of basalts from the Moon Earth HED asteroid and Mars: Implications for the origin of the Moon. Geochim Cosmochimica Acta 65:979-997

Ryder G (1994) Coincidence in time of the Imbrium basin impact and Apollo 15 KREEP volcanic flows: The case for impact-induced melting. Geol Soc Am Spec Paper 293:11-18

Ryder G. (1990) Lunar samples, lunar accretion and the early bombardment of the Moon. Eos, Transactions AGU 71: 313-323

Schultz PH, Spudis PD (1983) Beginning and end of lunar mare volcanism. Nature 302:233-236

Shearer CK, Floss C (2000) Evolution of the Moon's mantle and crust as reflected in trace-element microbeam studies of lunar magmatism. *In*: Origin of the Earth and Moon. Canup RM, Righter K (eds) Univ Arizona Press, p 339-359

Shearer CK, Papike JJ (1989) Is plagioclase removal responsible for the negative Eu anomaly in the source regions of mare basalts? Geochim Cosmochim Acta 53:3331-3336

Shih C-Y, Nyquist LE, Bansal BM, Wiesmann H (1992) Rb-Sr and Sm-Nd chronology of an Apollo 17 KREEP basalt. Earth Planet Sci Lett 108:203-215

Short NM (1975) Planetary Geology. Prentice Hall

Simpson RA, Tyler GL (1999) Reanalysis of Clementine bistatic radar from the lunar South Pole. J Geophys Res 104: 3845-3862

Snyder GA, Borg LE, Nyquist LE, Taylor LA (2000) Chronology and isotopic constraints on lunar evolution. *In*: Origin of the Earth and Moon. Canup RM, Righter K (eds) Univ Arizona Press, p 361-395

Solomon SC, Chaiken J (1976) Thermal expansion and thermal stress in the Moon and terrestrial planets: clues to early thermal history. Proc Lunar Sci Conf 7:3229-3243

Sprague AL, Nash DB, Witteborn FC, Cruikshank DP (1997) Mercury's feldspar connection - Mid-IR measurements suggest plagioclase. Adv Space Res 19:1507

Sprague L, Roush TL (1998) Comparison of laboratory emission spectra with Mercury telescopic data. Icarus 133: 174-183

Spudis PD, Guest JE (1988) Stratigraphy and geologic history of Mercury. *In*: Mercury. Vilas F, Chapman C, Matthews M (eds) Univ Arizona Press, p 152-164

Spudis PD, Reisse RA, Gillis JJ (1994) Ancient multi-ring basins on the Moon revealed by Clementine laser altimetry. Science 266:1848-1851

Stacey NJS et al. (1997) Arecibo radar mapping of the lunar poles: A search for ice deposits. Science 276:1527-1530

Stern SA (1999) The lunar atmosphere: History status current problems and context. Rev Geophys 37:453-491

Stevenson DJ (1987) Origin of the Moon-the collision hypothesis. Ann Rev Earth Planet Sci 15:271-315

Steward GR (2000) Outstanding questions for the giant impact hypothesis. *In*: Origin of the Earth and Moon. Canup RM, Righter K (eds) Univ Arizona Press, p 217-223

Swindle TD (2000) (2000) *In situ* noble-gas based chronology on Mars. Concepts And Approaches For Mars Exploration. Lunar and Planetary Institute Contribution 1062:294-295

Taylor LA, Pieters CM, Morris RV, Keller LP, McKay DS (2001) Lunar mare soils: Space weathering and the major effects of surface-correlated nanophase Fe. J Geophys Res 106:27,985-28,000

Taylor SR (1975) Lunar Science: A Post-Apollo View; Scientific Results and Insights from the Lunar Samples. Pergamon Press

Taylor SR (1982) Planetary Science: A Lunar Perspective. Lunar and Planetary Institute

Taylor SR (1987) The unique lunar composition and its bearing on the origin of the Moon. Geochim Cosmochim Acta 51:1297-1309

Taylor SR (1998) Destiny or Chance: Our Solar System and its Place in the Cosmos. Cambridge University Press

Taylor SR (1999) On the difficulties of forming Earth-like planets. Meteor Planet Sci 34 317-329

Taylor SR (2001a) Solar System Evolution: A New Perspective (2nd ed). Cambridge University Press
Taylor SR (2001b) Does the lunar composition retain a memory from the early solar nebula? Meteor Planet Sci 36: 1567-1569
Taylor SR (2001c) Flood basalts basalt floods or topless Bushvelds? Lunar petrogenesis revisited: A critical comment. J Pet 42:1219-1220
Taylor SR, Esat TM (1996) Geochemical constraints on the origin and evolution of the Moon. *In*: Earth Processes: Reading the Isotopic Code. Basu A, Hart SR (eds) AGU Geophysical Monograph 95:33-46
Taylor SR, Jakes P (1974) The geochemical evolution of the Moon. Proc Lunar Sci Conf 5:1287-1305
Tholen DJ (1989) Asteroid taxonomic classifications. *In*: Asteroids II. Binzel RP, Gehrels T, Matthews MD (eds) Univ. of Arizona Press, p 1139-1150
Thomas PC, Binzel RP, Gaffey MJ, Storrs AD, Wells EN, Zellner BH (1997) Impact excavation on asteroid 4 Vesta. Hubble Space Telescope Results:1492-1495
Tompkins S, Pieters CM (1999) Mineralogy of the lunar crust: Results from Clementine. Meteorit Planet Sci 34:25-41
Turkevich A (1971) Comparison of the analytical results from the Surveyor, Apollo, and Luna missions. Proc Lunar Sci Conf 2:1209-1215
Urey HC (1965) Meteorites and the Moon. Science 147:1262-1265
Vilas F, Chapman C, Matthews M (eds) (1988) Mercury. University of Arizona Press
Walker D (1983) Lunar and terrestrial crust formation. Proc Lunar Planet Sci Conf 14, in J Geophys Res 88:B17-B25
Walter MJ, Newsom HE, Ertel W, Holzheid A (2000) Siderophile elements in the Earth and Moon: Metal/silicate partitioning and implications for core formation. *In*: Origin of the Earth and Moon. Canup RM, Righter K (eds) Univ Arizona Press, p 265-289
Wang A, Haskin LA, Lane AL, Wdowiak TJ, Squyres SW, Wilson RJ, Hovland LE, Manatt KS, Raouf N and Smith CD (2003) Development of the Mars microbeam Raman Spectrometer. J Geophys Res 108(E1):5005, doi:10.1029/2002JE001902
Wänke H, Dreibus G (1986) Geochemical evidence for the formation of the Moon by impact-induced fission of the proto-Earth. *In*: Origin of the Moon. Hartmann WK, Phillips RJ, Taylor GJ (eds) Lunar and Planetary Institute, p 649-672
Ward PD, Brownlee D (2000) Rare Earth: Why Complex Life Is Uncommon in the Universe. Springer-Verlag
Warell J, Blewett DT (2004) Properties of the Hermean regolith: V. New optical reflectance spectra, comparison with lunar anorthosites, and mineralogical modelling. Icarus 168:257-276
Warren PH (1992) Inheritance of silicate differentiation during lunar origin by giant impact. Earth Planet Sci Lett 112: 101-116.
Warren, PH (2001) Early lunar crustal genesis: The ferroan anorthosite epsilon-neodymium paradox. Meteor Planet Sci 36:A219
Warren PH, Wasson JT (1979) The origin of KREEP. Rev Geophys Space Phys 17:73-88
Wasson JT (1971) Volatile elements on the Earth and Moon. Earth Planet Sci Lett 11:219-225
Wiechert U, Halliday AN, Lee D-C, Snyder GA, Taylor LA, Rumble D (2001) Oxygen isotopes and the Moon-forming giant impact. Science 294:345-348
Wieczorek MA, Phillips RJ (1997) The structure and compensation of the lunar highland crust. J Geophys Res 102: 10,933-10,943
Wieczorek MA, Phillips RJ (1999) Lunar multiring basins and the cratering process. Icarus 139:246-259
Wieczorek MA, Phillips RJ (2000) The Procellarum KREEP terrane: Implications for mare volcanism and lunar evolution. J Geophys Res 105:20,417-20,430
Wieczorek MA, Zuber MT (2001) A Serenitatis origin for the Imbrium grooves and South Pole-Aitken thorium anomaly. J Geophys Res 106:27,825-27,840
Wieczorek MA, Zuber MT (2001) The composition and origin of the lunar crust: Constraints from central peaks and crustal thickness modeling. Geophys Res Lett 28:4023-4026
Wieczorek MA, Zuber MT, Phillips RJ (2001) The role of magma buoyancy on the eruption of lunar basalts. Earth Planet Sci Lett 185:71-83
Wilhelms DE (1976) Mercurian volcanism questioned. Icarus 28:551-558
Wilhelms DE (1987) The Geologic History of the Moon. US Geol Surv Prof Paper 1348, Washington, DC
Williams DE, Pollard D (2000) Earth-Moon interactions: Implications for terrestrial climate and life. *In*: Origin of the Earth and Moon. Canup RM, Righter K (eds) Univ Arizona Press, p 513-525
Williams GE (1989) Late Precambrian tidal rhythmites in South Australia and the history of the Earth's rotation. J Geol Soc London 146:97-111
Williams GE (2000) Geological constraints on the Precambrian history of Earth's rotation and the Moon's orbit. Rev Geophys 38:37-59
Wood JA (1972) Fragments of terra rock in the Apollo 12 soil samples and a structural model of the Moon. Icarus 16: 462-501
Wood JA (1986) Moon over Mauna Loa. *In*: Origin of the Moon. Hartmann WK, Phillips RJ, Taylor GJ (eds) Lunar and Planetary Institute, p 17-55