The Origin of the Moon: Geochemical Considerations

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"Of course I was not there when the solar system originated."
(Harold C. Urey, letter to NASA Planetology Committee, September 20, 1963)

Capture models for the origin of the Moon do not meet chemical, isotopic, or dynamical constraints and have effectively been abandoned. The composition of the bulk Moon differs in several important ways from that of the terrestrial mantle, ruling out fission hypotheses unless substantial element fractionation occurs during or following fission. Large Mars-sized impactor models cut several Gordian knots, since the lunar material is derived mostly from the impactor, whose chemical and isotopic composition become free parameters. Both volatile and siderophile element depletions, observed in the Moon, are common in meteorites (e.g., eucrites) formed at 4.5 aeons, so that the lunar composition is nonunique, and a population of precursor planetesimals of appropriate composition existed. Of the five hypotheses for lunar origin, either the large impactor models or the double-planet models remain as viable candidates. Establishment of the time of volatile depletion, and whether one or more episodes of volatile loss were involved, might distinguish between these two hypotheses. Whole Moon melting follows accretion. Core formation in the Moon depletes the mantle in siderophile elements. Additional Ni and Co are trapped in deep olivine and orthopyroxene phases. Crystallisation results in an anorthositic crust and a differentiated lunar mantle while the final residual liquid (KREEP) invades the crust at 4.3–4.4 aeons.

1. The Apollo Sample Return: Demise of Capture Models

The chemical and isotopic information from the samples returned by the Apollo manned missions immediately provided new limits on theories of lunar origin. Prior to 1969, the composition of the Moon was constrained only by density measurements.

Hence, it was known to be low in iron relative to bulk Earth compositions. From the evidence of massive cratering in the highlands, the age of that surface was supposed to be older than that of any exposed terrestrial surface. Its probable ancient age and the low density of the Moon led to hypotheses that the Moon might represent an essentially primitive object (Urey, 1962). Estimates of the cratering flux by perceptive students gave essentially correct ages of 3–4 aeons for the lunar maria (e.g., Baldwin, 1949; Hartmann, 1965). Although the material filling the maria was correctly identified as basaltic lava by acute observers (e.g., Baldwin, 1949), the anorthositic nature of the highlands eluded investigators until the first sample return.

The Apollo sample results immediately demonstrated that the Moon was formed at least 4.4 aeons ago, that it was highly depleted in volatile elements, and that a large fraction, if not all, had been extensively differentiated (e.g., Taylor, 1975). The similarity of oxygen isotope ratios to those in the Earth indicated a nearby origin, rather than in some remote part of the solar nebula. These observations reinforced the dynamical difficulties inherent in capture models, so that capture of a primitive object became less attractive. Since the hypothesis now possessed few advantages over double planet scenarios, it was effectively abandoned by most workers.

Attention then became focused on double-planet and fission scenarios. Double-planet hypotheses immediately encounter the difficulty of accounting for the density difference, and selective accretion of silicate rather than metallic iron became a feature of these models (Ruskol, 1977; Wood and Mitler, 1974; Smith, 1979, 1982). All such models presuppose prior fractionation of silicate-metal-sulfide phases, in addition to volatile element depletion in precursor planetesimals, before accretion of the Earth and Moon. This important point will be addressed later, but it should be recalled that the Earth is depleted in volatile elements (e.g., low Rb/Sr, K/U relative to CI meteorites), although not to the same extent as the Moon.

2. The Fission Hypothesis

The principal philosophical support for this hypothesis was the similarity between the density of the Moon (3.34 gm cm^{-3}) and the uncompressed density of the terrestrial mantle (3.32 gm cm^{-3}). Given the Apollo data, it became the most readily testable of all hypotheses. A necessary corollary is that core formation in the Earth should precede fission, and hence the Earth has to be close to its present size before fission occurs. Wetherill-Safronov scenarios (e.g., Wetherill, 1980) for the accretion of the terrestrial planets apparently demand sweep-up time scales of 10^8 years. If we assume that core formation was coeval or immediately followed accretion, then the origin of the Moon occurs at about 4.45 aeons, measurably later than the 4.56 aeon age for the formation of the meteorites (and presumably planetesimals). This point will be discussed later, along with the complexity introduced by the difference between lunar and terrestrial mantle FeO contents.

Serious dynamical constraints beset simple fission models. The angular momentum of the Earth-Moon system is insufficient by a factor exceeding three to form a bulge during rotation, from which the Moon might form (Weidenschilling, 1984). Catastrophic core formation might speed up the rotation rate (Wise, 1963), thereby facilitating fission. The precipitation hypothesis (Ringwood, 1970) was an attempt both to overcome the angular momentum problem and to provide for depletion in volatile elements.

2.1. The lunar orbit

The lunar orbit is inclined at 5.1° to the plane of the ecliptic, whereas the axial plane of the Earth is inclined at 23.4°. The inclination of the lunar orbit to the equatorial plane of the Earth thus varies between 18.3° and 28.5°. Simple versions of the fission hypothesis place the orbit of the Moon in the equatorial plane of the Earth. Although most planetary orbits lie close to the plane of the ecliptic, high axial inclinations are common and are generally attributed to the effects of large impacts during accretion consistent with the planetesimal hypothesis (Wetherill, 1980). Thus, the present Earth-Moon orbital arrangement could be due to a late large terrestrial impact following the formation of the Moon.

A more serious constraint comes from orbital calculations showing that the Moon has never been closer to the Earth than about 240,000 km (Hansen, 1982). This is consistent with a double-planet scenario, but not with fission models, or with disintegrative capture models in which a ring of debris enriched in silicates accumulates just beyond the Roche limit (2.89 earth radii or 18,400 km; Wood and Mitler, 1974; Smith, 1979). If Hansen's calculations are correct, all hypotheses for lunar origin except double-planet models are excluded.

2.2. Volatile element depletion

If the fission hypothesis is correct, then the mantle of the Earth should bear a close and identifiable relationship to the composition of the bulk Moon. The composition of the upper mantle of the Earth is constrained by experimental petrology, mantle derived xenoliths, and by basalts derived therefrom by partial melting. The composition of the lower mantle remains uncertain, but serious cosmochemical problems arise if it is the same as the upper mantle, whose Mg/Si ratio is far above that of any meteorite group. The lower mantle is probably dominated by perovskite ($MgSiO_3$) phases (Liu and Bassett, 1985) and hence may be different in composition to the upper mantle. Accordingly, there is no unambiguous evidence that the Earth has other than chondritic ratios for the major lithophile elements.

The volatile trace elements, however, are clearly depleted relative to CI abundances. The terrestrial Rb/Sr ratio is 0.031 compared to CI values of 0.30. The lunar Rb/

Sr ratio of 0.009 is depleted relative to the Earth. This volatile/refractory element depletion is likewise shown by K/U ratios (60,000 for CI, 10,000 for the Earth, and 2500 for the Moon) (Fig. 1) and for many other elements. The most volatile elements (e.g., Bi, Tl) are depleted in the Moon relative to the Earth by factors of about 50 (Wolf and Anders, 1980). This extensive depletion (shown also by total absence of lunar water) is a first-order fact requiring explanation in all hypotheses. However, Earth-Moon differences in volatile element abundances are not always simply related to volatility. Drake (1986) has pointed out that the lunar Cs/Rb ratio is higher than that of the Earth, contrary to predictions based on relative volatility.
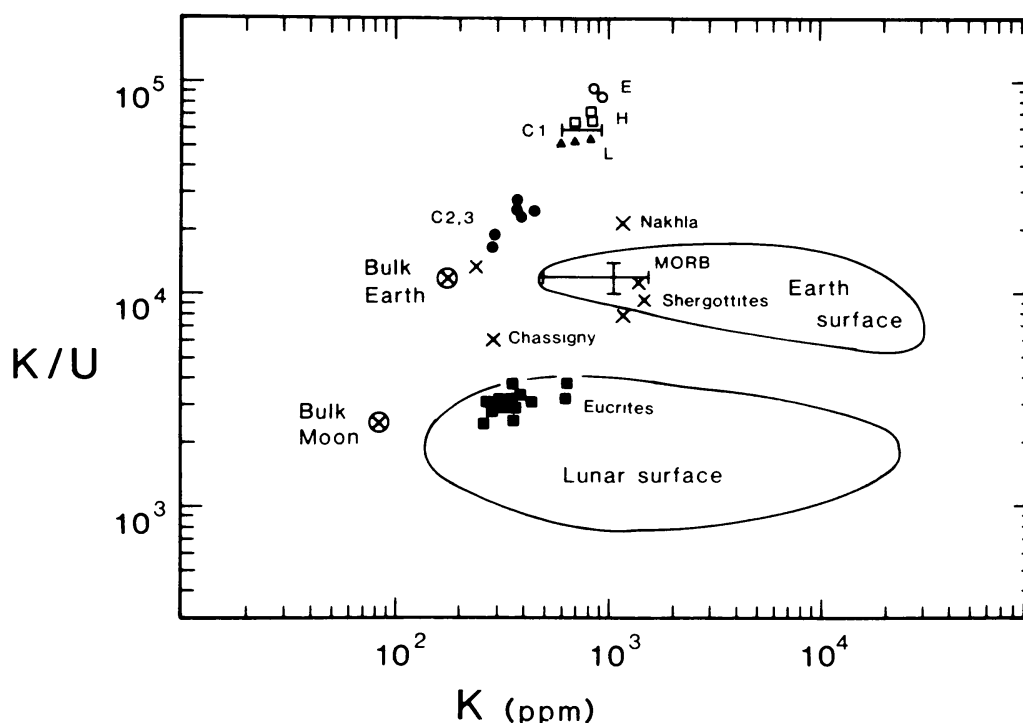
Fig. 1. Volatile/refractory element fractionation illustrated by variations in the K/U ratio, plotted against potassium abundance, in meteorites, Earth, and Moon. The CI data are taken as being representative of the original solar nebula, on account of their similarity to solar values. Potassium is volatile at about 1000 K and uranium is highly refractory. Widespread depletion in potassium, relative to uranium, is a feature of the inner solar system. Both elements concentrate in residual melts during intraplanetary igneous fractionation processes, so that the K/U ratio in surface rocks approximates to the bulk planetary ratio. If the SNC meteorites (Shergotty, Nakhla, Chassigny) come from Mars, then that body is at least as volatile-rich as the Earth. The K/U value for Nakhla, and the K/La data (Fig. 2) indicate higher K values for the SNC parent body. Uranium is mobile, under oxidising conditions as U^{6+}. This possibility explains the wide range in K/U ratios in the SNC meteorites. Both the Moon and the eucrites are the most depleted in K. (Adapted from Taylor, 1982, Fig. 8.3. Additional data from Smith et al., 1984.)

2.3. Lunar cores and siderophile element depletions

The $Mg/(Mg + Fe)$ ratio of the terrestrial mantle and the Moon should be similar, if the fission hypothesis is correct, since the ratio should not be altered by the fission process. The FeO content of the terrestrial mantle is usually estimated at about 8%. Density considerations limit higher values. The bulk Moon FeO content is estimated by most workers to be about 13%, making the bulk lunar $Mg/(Mg + Fe)$ value = 0.81. If about 2% Fe is removed in a lunar core, the $Mg/(Mg + Fe)$ value for the lunar mantle becomes about 0.84, consistent with the values required for the mare basalt source regions. The $Mg/(Mg + Fe)$ value for the terrestrial upper mantle is 0.90. Warren (1984) argued on the basis of estimates from highland samples that the lunar $Mg/(Mg + Fe)$ ratio could be similar to that of the terrestrial mantle. However, for bulk lunar FeO values of 13%, this would lead to bulk lunar MgO values exceeding 60%. These seem unrealistic. Lunar mare basalts possess very high Fe/Ni ratios and are depleted in Ni by a factor of four and trace siderophile elements by larger factors relative to terrestrial basalts (Taylor, 1979).

The various indirect arguments for the existence of a small (400-km radius) metallic core comprising about 2% of lunar volume have become stronger (Newsom, 1984). The existence of a lunar core has several consequences.

1. It provides a sink to deplete the lunar mantle in Ni and the other trace siderophile elements. The extreme depletion in these elements in the lunar mantle contrasts strongly with their higher abundances in the terrestrial mantle, indicating a lack of core-mantle equilibrium in the Earth in contrast to the Moon.

2. It reduces the lunar mantle Fe content to provide an acceptable $Mg/(Mg + Fe)$ value for the source region of lunar basalts.

3. It implies whole Moon differentiation. Whether the Moon was wholly or partially melted is less of an issue than the evidence that extensive planetary-wide differentiation occurred.

4. It raises a paradox for the fission hypothesis. The extreme depletion of the lunar mantle in siderophile elements is consistent with their removal into a lunar core (Newsom, 1984, 1986). If the Moon was derived from the terrestrial mantle, both would have the same FeO content at that time. Separation of a lunar core should then deplete the mantle of iron. However, the converse is true. The lunar mantle is enriched in FeO compared to the Earth's mantle, although Ni and other trace siderophile elements are severely depleted. One answer to this dilemma for the fission model is to remove the Moon from the terrestrial mantle before core separation is complete, requiring an ad hoc adjustment to the hypothesis.

2.4. Whole Moon differentiation and magma oceans

It is generally agreed that large scale melting and differentiation of the Moon occurred before 4.4 aeons. However, estimates of the depth to which the Moon

was affected have varied from shallow (200 km) to whole Moon involvement. In this section, it is concluded that evidence for the latter case is strong. Although alternative scenarios such as serial melting (Walker, 1983) have been proposed, the following lines of evidence support the concept of a magma ocean. These include the presence of a thick plagioclase-rich crust, complementary Eu anomalies in the crust and the mare basalts, the extensive near-surface concentrations of REE, U, Th, and other incompatible elements, the uniform REE and isotopic systematics of KREEP, and the isotopic evidence that this extensive lunar differentiation was completed very early, by about 4.4 aeons. Deep melting is consistent with geochemical balance arguments and by the most recent assessment of the lunar internal structure (Hood and Jones, 1985). This favors models in which Mg/(Mg + Fe) increases at depth, consistent with a deep lunar interior dominated by early Mg-rich olivine and orthopyroxene cumulates to depths of over 1000 km. Such models also require lunar cores (350–500-km radius) to account for the lunar coefficent of moment of inertia (0.3905 ± 0.0023), which needs a density increase in the deep interior, in addition to a low density crust, to offset the low density Mg-rich interior. Such scenarios remove the possibility of retention of unmelted primordial lunar compositions deep in the interior.

Such regions have appealed to some geochemists wishing to derive mare basalts from unfractionated lunar material, which accordingly would provide unique information on its origin (Delano and Ringwood, 1978). However, the most primitive lunar basaltic material available (green glass—15425–15427) comes of course from a source region already fractionated, shown by the presence of a negative europium anomaly. The evidence for a primitive volatile-rich interior deduced by Delano and Lindsley (1983) from volatile-rich rims on volcanic glass spheres is readily explained by coating of such spheres in fire-fountain-type eruptions, with the volatiles (e.g., Zn, Pb) being concentrated in such an environment, following scavenging from their miniscule lunar budget. The presence of primitive Pb isotope ratios, for example, is well explained by the segregation of Pb in sulfide phases in a low μ ($^{238}U/^{204}Pb$) environment during the initial lunar differentiation at about 4.4 aeons. The case for whole Moon differentiation seems reasonably consistent with the geochemical data and assists in explaining the siderophile element depletion, by sequestering these in a metallic core. An additional depletion of Ni will occur by segregation of early crystallising olivine and orthopyroxene in the deep interior. The lack of radioactive heat sources in those deep cumulates, and their depth and refractory nature, will prevent magma formation.

2.5. Refractory element abundances and the lunar crust

A basic question for the fission hypothesis is whether the refractory element abundances in the Moon and the Earth are similar or different. The refractory element abundances in the Earth are constrained by an interlocking set of elemental and

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isotopic ratios, so that a value of 18 ppb U represents a consensus among geochemists. In this case, heat flow is not equal to heat production and the Earth is losing heat (McKenzie and Richter, 1981). The abundance of uranium is tied to that of other refractory elements (e.g., Al, Ti) on the assumption that refractory elements are not substantially fractionated on a planetary scale in cosmochemical processes. Assuming no fractionation between U and Al, a uranium abundance of 18 ppb indicates an Al_2O_3 content of 3.60% for the bulk Earth, consistent with other estimates.

The lunar uranium abundance is more disputed. However, both Al and U are concentrated in the highland crust of the Moon. The average thickness of this crust is 73.4 km, comprising 12.3% of lunar volume. The Al content of this crust is thus a significant fraction of the lunar Al budget. The variation in crustal thickness (nearside 64 km; farside 86 km) is conventionally held responsible for the centre of figure (CF)/centre of mass (CM) offset, while a low density crust, coupled with an increase in density in the deep interior, accounts for the value of the coefficient of moment of inertia (see above). Thinner crusts must be of lower density (higher Al_2O_3), while a lower Al_2O_3 content for the bulk crust may be accommodated by a thicker crust. It is accordingly difficult to lower the bulk Al_2O_3 content of the crust because of these geophysical constraints. Alternative explanations for the CF/CM offset suggest core or mantle asymmetries. Such explanations seem unlikely for a fully differentiated moon. On the contrary, variations in thickness of a floating aluminous crust are readily explicable in terms of lunar magma oceans. It is difficult to escape from the conclusion that the Moon possesses a thick aluminous crust. Taylor (1982) concluded on geochemical grounds that the average crust contained 24.6% Al_2O_3. Spudis (1984), working from stratigraphic constraints, estimates 25% Al_2O_3. If it averages 24.6% Al_2O_3, this contributes 3.0% Al_2O_3 to the bulk lunar composition. Even if the crustal composition is based on the lowest observable Al_2O_3 content (about 20%), this contributes 2.5% Al_2O_3 to the bulk Moon composition. This exceeds the total Al_2O_3 content (2.4%) for a moon of chondritic composition and comprises a large proportion of a moon with a terrestrial mantle Al_2O_3 abundance of 3.6%.

A further petrological constraint favors high-alumina moons. It is a requirement that plagioclase crystallise early, during magma ocean cooling, not only to account for the thick plagioclase-rich highland crust, but also to account for the universal depletion of the lunar interior in the REE europium. Mare basalts, which sample to depths of 200–400 km (50% of lunar volume), provide a lower limit on the amount of prior plagioclase (and Eu) removal. Nevertheless, sufficient Al_2O_3 is retained in the source regions to produce basalts by partial melting that typically contain 8–14% Al_2O_3 (Taylor, 1982, Tables 6.3a,b). Terrestrial mantle compositions (3.6% Al_2O_3) do not meet these requirements, crystallising pyroxene before plagioclase, and so the bulk Moon must be enriched in Al_2O_3 relative to the terrestrial mantle. A bulk lunar Al_2O_3 content of 6.0% implies a uranium abundance of 30 ppb, a little below the lower heat flow limit. The heat flow estimates provide a range

from 33 to 46 ppb (Langseth *et al.*, 1976). This assumes that heat flow is equal to heat production, a reasonable view for the Moon on account of the near-surface concentration of heat-producing elements, its small size, and cessation of volcanic activity about 3 aeons ago. Both heat flow values were obtained at mare basin margins. Rasmussen and Warren (1985) suggest that focusing effects at such locations enhance the heat flow and estimate a bulk lunar uranium abundance of 19 ppb. Several factors suggest that this value is too low. The present basins filled with mare basalt (only a few tens of meters thick at the edges) represent only the final impacts. Modest estimates of cratering rates (Wilhelms, 1985) since crustal solidification at 4.4 aeons allow for at least 80 basins (>300 km diameter) and 10,000 craters in the range 30–300 km diameter. Spudis (1984) suggests that this impact history has created a megaregolith down to the 25-km seismic discontinuity. If the crustal structure is effectively that of a megaregolith 25 km thick, due to the superimposed effects of over 80 basins (plus several thousand craters), then it seems unlikely that any simple focusing mechanism will operate or that heat transport will be affected only by the most recent basin margin effects.

Further constraints may be placed by the abundance of potassium. The bulk lunar K/U ratio is about 2500. The highland crustal average is estimated at 500 ppm, which contributes 60 ppm to the bulk lunar budget. Mare basalts typically contain 500 ppm K (Taylor, 1982, Tables 6.4a,b), which for $D_{L/R}$ values of 10 indicate source region contents of at least 50 ppm. Thus, the bulk lunar K abundance is at least 85 ppm, which gives a bulk lunar U value of 34 ppb for K/U = 2500. The Rasmussen-Warren value of 19 ppb U yields a bulk lunar K value of 48 ppb, less than is apparently concentrated in the highland crust. It is thus difficult to escape from the conclusion that the lunar uranium value is greater than 30 ppb, consistent with the estimate by Drake (1985). In this case, the Moon is enriched in refractory elements relative to the Earth by a factor of greater than 1.5. Such an enrichment cannot, of course, be caused merely by loss of volatile elements during fission, but could be consistent with condensation from a vapor phase, as in the large impactor hypothesis.

2.6. Summary of Earth-Moon differences

Philosophical difficulties beset attempts to compare the chemistry of the Earth and the Moon. Even if a perfect compositional match was obtained, this implies similarity of process, rather than a genetic connection. Thus, the shergottite class of meteorites shows a very close correspondence in trace element chemistry to terrestrial basalts (Stolper and McSween, 1979). The correspondence is much closer than between lunar and terrestrial basalts. However, no one supposes that Mars, the probable source of the shergottites, is derived from the Earth. Likewise, the eucrites are much closer to lunar basalts than are terrestrial basalts, but a genetic connection is not suggested. The overall view of lunar and terrestrial mantle geochemistry reveals more

differences than similarities and refutes simple versions of the fission hypothesis (Table 1). The depletion of the volatile elements, the differences in FeO content, siderophile element and refractory element abundances all suggest irreconcilable differences. The high lunar FeO and low siderophile abundances present a paradox.

TABLE 1.   *Element Abundances in CI Chondrites (Volatile-free), Primitive Earth Mantle (= Mantle Plus Crust) and Bulk Moon.*

	CI	Earth Mantle +Crust	Bulk Moon
SiO_2	34.2	49.9	43.4
TiO_2	0.11	0.16	0.3
Al_2O_3	2.44	3.64	6.0
FeO	35.8	8.0	13.0
MgO	23.7	35.1	32.0
CaO	1.89	2.89	4.5
Na_2O	0.98	0.34	0.09
K_2O	0.10	0.02	0.01
ϵ	99.2	100.1	99.3
Volatile elements			
K (ppm)	854	180	83
Rb (ppm)	3.45	0.55	0.28
Cs (ppb)	279	18	12
Moderately volatile element			
Mn (ppm)	2940	1000	1200
Moderately refractory element			
Cr (ppm)	3975	3000	4200
Refractory elements			
Sr (ppm)	11.9	17.8	30
U (ppb)	12.2	18	33
La (ppb)	367	551	900
Eu (ppb)	87	131	210
V (ppm)	85	128	150
Siderophile elements			
Ni (ppm)	16500	2000	400
Ir (ppb)	710	3.2	0.01
Mo (ppb)	1380	59	1.4
Ge (ppm)	48.3	1.2	0.0035

Data sources: CI and terrestrial data from Taylor and McLennan (1985, Tables 11.1 and 11.3), lunar data from Taylor (1982, Tables 8.1 and 8.2), except for Ni, Ir, and Mo data calculated from CI/Moon depletion factors given by Newsom (1986), and Ge data from Dickinson and Newsom (1984).

Most attempts to find similarities between the chemistry of the terrestrial mantle and the Moon result in taking the extreme limits of the measured data or in the least plausible explanations. These models require the thinnest lunar crust (with the lowest Al_2O_3 content), that the heat flow data be ignored, that siderophile element abundances be based on highland rock data (particularly liable to meteoritic contributions), that the cause of the CF/CM offset lies deep in the inaccessible interior rather than in the crust, and that fission occur at a time earlier than the establishment of the present mantle-core relationship. Such attempts to match Earth-Moon chemistry have been labeled Procrustean, after the mythological inn-keeper (Anders, 1978). A final assessment is that "if the Moon was derived from the terrestrial mantle by fission, then the chemical evidence for such an event has been destroyed" (Taylor, 1982, p. 425).

3. The Large Impactor Hypothesis

This model states that the Moon was derived from material splashed during a glancing collision of a Mars-sized body with the Earth (Cameron and Ward, 1976; Hartmann and Davis, 1975). In the latest versions of this model (Cameron, 1984; Stevenson, 1984; Kaula and Beachy, 1984), two constraints have appeared. In order for the material to be placed in lunar orbit, it must be derived mostly from the Mars-sized impactor, removing the geochemical problems to another part of the wood. In addition, it must be vaporised, since solid ejecta will return to the Earth. This scenario cuts a number of Gordian knots. It removes the angular momentum problem. A grazing impact might place material in a nonequatorial orbit, accounting for the orbital dilemma. Several provisos may be made:

1. The impactor must form in the general vicinity of the Earth, to account for the similarity in oxygen isotopic signatures between Earth and Moon.

2. Mantle-core separation had already occurred in both Earth- and Mars-sized impactors before impact.

3. Time scales of 10^8 years are required to allow for accretion of Earth and impactor and for mantle-core separation in both bodies.

4. A seductive feature of the model is that it allows for an ad hoc composition for the impactor, within the general constraints imposed by inner planetary compositions. Accordingly, appropriate Fe/Mg ratios, siderophile element contents and Al and U abundances become free parameters, thus bypassing the imbroglio of lunar-terrestrial compositional comparisons.

5. It is apparently a requirement of the model that the material that finished up in the Moon, in addition to being derived mainly from the impactor, is vaporised at temperatures of about 2000 K (Stevenson, 1984). This could account for the extensive depletion of the Moon in volatile elements (e.g., alkalies, Bi, Tl, H_2O, etc.). However, recondensation temperature limits of 1500 K (Boynton, 1983) might be placed by the absence of combined Eu and Yb anomalies in bulk lunar REE

patterns, in contrast to those observed in some Allende inclusions. This constraint could be overcome by postulating simple evaporation and total recondensation for the REE. Since, however, the model wishes to use this mechanism to deplete the Moon in the other volatile elements, special pleading must be invoked and the shadow of William of Ockham appears. Evaporation of MgO and SiO_2 might also be expected at these temperatures, but such effects are not obvious. Extensive processing of material through a vapor phase might also produce some isotopic anomalies, although none have been detected. Depletion of volatile elements in the Moon appears to be limited to those elements with condensation temperatures below about 1100 K, which limits thermal processing of protolunar material to such temperatures or to short time intervals.

4. Early Volatile and Siderophile Element Depletion

Volatile element depletion is not restricted to the Moon but is common in the solar system, as is shown by the widespread variations in K/U and K/La ratios in meteorites and planets (Figs. 1 and 2). The terrestrial planets exhibit variations in volatile-refractory element ratios and in density consistent with metal-silicate fractionation. Since planets of Earth's size cannot lose elements of high atomic weight, three basic questions arise: When does the volatile-refractory element separation occur, where is the location, and what is the mechanism responsible? The meteorites provide some basic clues. They display much variation in metal-silicate and volatile-refractory ratios. The production of metal is apparently separate from and unrelated to the volatile depletion. The type 1 enstatite chondrites provide this interesting information, since they contain all their complement of iron as metal but retain the primitive levels of volatile element abundances. The H, L, and LL chondrites display both variations in metal and volatile elements.

4.1. The eucrite evidence

The lunar composition is clearly highly fractionated, relative to primitive solar nebula abundances. Does it represent an exceptional case among planetary compositions? Although there is a strong temptation to regard the Moon as a special case, volatile depletion is not restricted to that body, for very similar depletions (in K/U, Bi, Tl, H_2O, lack of Fe^{3+}, etc.) are observed in the eucrites, leading to the comment that "the lunar deficiency of alkalies and volatiles is not unique, and does not necessarily imply special processes in its evolution as a satellite" (Morgan et al., 1978, p. 35).

The eucrites are derived by partial melting from the mantle of a small asteroid. In addition to the volatile element depletion (K/U = 3100), they are depleted in Ni (~10 ppm) and the other siderophile elements. Metal was absent in the source regions (Stolper, 1977). Samarium-neodymium and $^{207}Pb/^{206}Pb$ data for eucrites

indicate crystallisation ages of 4.54–4.56 aeons. Two possibilities exist: (1) The eucrite parent body was accreted from metal-free volatile-depleted precursors. (2) The eucrites were derived from the mantle of an asteroid that had already formed a metallic core. In either alternative, K and Rb had already been depleted relative to CI abundances of U and Sr. The refractory element abundances in the parent body
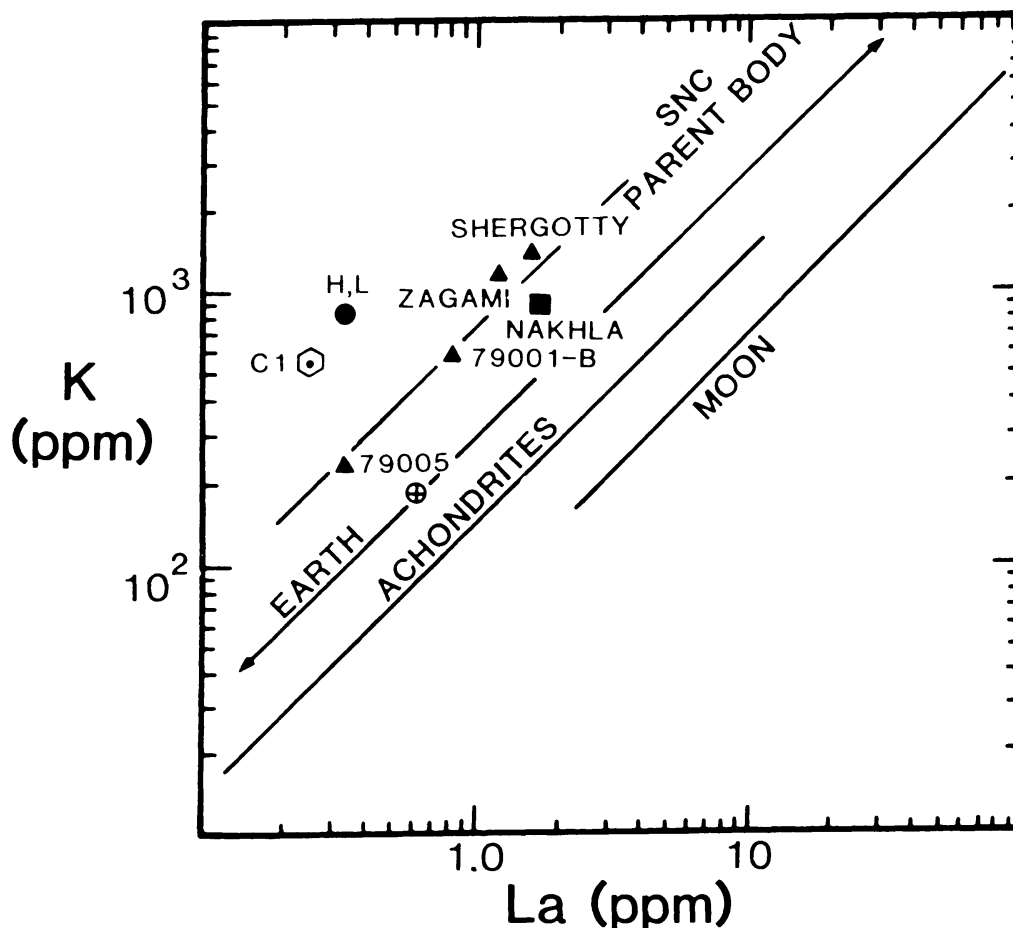

Fig. 2. *Relationship between volatile element potassium and refractory rare earth element lanthanum for CI, H, and L chondrites, Earth, Moon, achondrites, and SNC parent body (= Mars?). All planetary bodies are depleted in potassium (volatile at about 1000 K), relative to La, illustrating the large scale of depletion of volatile elements in the inner portions of the solar nebula. Both K and La are incompatible elements in intraplanetary igneous differentiation processes, concentrating in residual melts during crystallisation of silicate melts. The K/La ratio of surface rocks thus approximates to the bulk planetary ratio. Mars, if it is the source of the SNC meteorites, is thus more volatile-rich than the Earth. The K/U (Fig. 1) and K/La plots show slight differences, attributable to differences in behaviour of La and U in intraplanetary differentiation processes. (Based on Smith et al., 1984.)*

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are parallel to those of CI chondrites, showing no anomalies in the abundances of the more volatile REE, Eu, and Yb.

The similarity between eucrites and lunar mare basalts is remarkable. Were it not for the differences in age and oxygen isotope signature, it might be difficult to distinguish them on petrological or geochemical grounds. The similar levels of depletion in the highly volatile elements (e.g., Bi, Tl) and the absence of water is particularly significant. It demonstrates that it is unnecessary to appeal to high-temperature processing of terrestrial material following fission to cause the high lunar depletion in volatiles.

The fact that both volatile and siderophile element depletion occurred at 4.5 aeons and that basalts were generated by partial melting of a planetesimal mantle similar to that of the lunar mantle must give pause to those who advocate unique conditions for the origin of the Moon.

4.2. Timing of the lunar volatile depletion

Volatile depletion, in the fission and large impactor models for lunar origin, occurs during the fission or impact event. In the double-planet scenario, the Moon accretes from planetesimals in which volatile depletion has already occurred. Can we establish the time at which the lunar depletion in volatile elements occurred? The eucrites, and other meteorites, provide direct evidence of volatile loss at about T_0 (4.56 aeons). We also know that the Earth accreted from volatile-depleted planetesimals that had average K/U ratios of 10,000 and Rb/Sr = 0.031 and that volatile/refractory element fractionation was widespread in the early solar system (Figs. 1 and 2).

In principle the Rb-Sr isotopic systematics could provide evidence for the time of separation of volatile Rb from refractory Sr (and hence also K from U). The best constraints on the lunar initial $^{87}Sr/^{86}Sr$ (LUNI) come from measurements on the low Rb/Sr anorthosites. For example, sample 60025 has a measured $^{87}Sr/^{86}Sr$ ratio of 0.69895 ± 3, which constrains LUNI to be equal to or less than that value (Nyquist, 1977). This is very primitive and close to BABI (0.69898 ± 3), implying an early origin for the Moon. We do not know the terrestrial initial $^{87}Sr/^{86}Sr$, but if the Earth formed in 10^8 years by accretion from a suite of planetesimals with bulk Rb/Sr = 0.031, then the terrestrial initial $^{87}Sr/^{86}Sr$ ratio must be very low, closer to ALL (0.69877 ± 2) than to BABI, if we wish to derive the Moon from the terrestrial mantle. This evidence is consistent with models that establish the basic differences in volatile/refractory element ratios between the Earth and the Moon in precursor phases prior to their formation and accrete the two bodies from a distinct set of planetesimals, as in the double-planet scenario.

In the impact-induced fission model, the absence of knowledge of the terrestrial initial $^{87}Sr/^{86}Sr$ ratio is not critical since terrestrial mantle involvement in lunar composition is minimal. The impactor $^{87}Sr/^{86}Sr$ initial ratio is a free parameter.

However, in the 10^8 years or so required to accrete both Earth and impactor, a measurable increase in $^{87}Sr/^{86}Sr$ will occur so that the impactor initial $^{87}Sr/^{86}Sr$ must be very low. The very primitive $^{87}Sr/^{86}Sr$ ratio of the Moon presents some difficulties for these scenarios.

In summary, either the Earth or the impactor have initial $^{87}Sr/^{86}Sr$ much lower than BABI, or the Wetherill-Safronov times for planetesimal sweep-up of 10^8 years are too long, or the fission and impact-induced fission models are incorrect. Further constraints are derived from the I-Pu-Xe dates (Swindle *et al.*, 1984, 1986) that indicate that the Moon appears to be somewhat older than the Earth. A full discussion is given in Swindle *et al.* (1986).

4.3. Mechanisms for volatile depletion and element fractionation

Depletion of the volatile elements occurs both in the early solar nebula and in later collisional episodes. The mechanisms responsible for the early volatile depletion in the nebula are speculative, but some constraints may be noted. Temperatures of the order of 1000–1100 K seem required to account for the observed depletions. The separation of volatile (e.g., K) from refractory (e.g., U) elements occurs on a large scale, so that the Earth, Moon, meteorites, the eucrite parent body, and other planetesimals acquired differing K/U ratios (Figs. 1 and 2). However, heavy elements cannot be lost by heating from condensed bodies of planetesimal size, and the separation must take place in dispersed phases before accretion or during collisions between small bodies. Could this separation have occurred during condensation of an initially hot solar nebula? There is little evidence that the primordial solar nebula was hot, and there is considerable evidence from the isotopic heterogeneities (Clayton, 1978), and supported by astrophysical observations and theory, that it was initially cold (Gehrels, 1978). However, heating of the inner nebula (in the region now occupied by the terrestrial planets), either during collapse of the central region to form the sun or due to early solar activity, has occurred. A preferred scenario is heating of the inner nebula by early intense solar activity.

If volatile elements, along with the rare gases, are removed from the inner solar system by early strong solar winds and flares, then the astrophysical evidence provides stringent time constraints. Nebulae surrounding young stars clear on time scales of 10^6 years. Strong T Tauri type solar winds occur on similarly short time scales (Gehrels, 1978). In order to sweep out gases and volatiles by such a mechanism, then the mechanism operates on times one or two orders of magnitude shorter than the planetesimal sweep-up times of 10^8 years. Accordingly, volatile-refractory element separation, if accomplished by such means, can only occur within a few million years of the sun arriving on the main sequence. The evidence from the meteorites that such fractionation occurs on short time scales at about 4.5 aeons may thus be consistent with such scenarios.

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The timing is crucial, since not all volatile elements are removed from the inner solar system. Enough volatile material remains to account for the volatile components in the inner planets. Accordingly, by the time the early sun is radiating, some material must already be in planetesimals capable of surviving heating and dispersion by early strong solar activity.

Large impact events also provide a mechanism for depleting the parent material in volatile elements. Small-scale examples of depletion of very volatile elements (e.g., Bi, Cs, Tl, Pb) occur in tektite-forming events. Impact glasses (e.g., Henbury, Zhamanshin), in contrast, are subjected to less heating and retain their complement of such elements (Taylor and McLennan, 1979). Such mechanisms can operate either during planetesimal breakup, in disintegrative capture processes, in processing through a circumterrestrial disk, or during large-scale impactor events. In the latter case, the material that goes into the Moon is postulated to have condensed from a vapor phase. There are accordingly several possible mechanisms for depleting the lunar material in volatile elements. Temperatures of recondensation seem to have been about 1000–1100 K, since only elements whose condensation temperatures are less than this are depleted.

5. Implications for Terrestrial Planetary Formation

Current theories for the origin of the inner planets call for their assembly from a diverse suite of planetesimals, probably in a gas-free environment (Wetherill, 1980). The time scale for the accretion of the inner planets from a large population of planetesimals is of the order of 10^8 years, although much of the accretion is completed in about half that time (Wetherill, 1980).

If the terrestrial planets accrete from planetesimals, with sweep-up times of about 10^8 years, then the inner planets were not complete until accretion ended about 4.45 aeons, about 100 m.y. later than the time scales for element fractionation indicated by the meteorites. It is a consequence of such models that the inner planets thus accrete from a heterogeneous swarm of already differentiated planetesimals, with varying contents of metal and volatile elements.

The implications of preaccretion metal-silicate and refractory-volatile element fractionation for the compositions of the Earth, Moon, and the inner planets are considerable. The metal now in the Earth's core, and the silicates in the mantle are accreted as separate phases (but not necessarily in separate planetesimals) whose equilibration was established in precursor events. Thus, there is no necessary close relationship between, for example, the Ni, Co, and siderophile element contents of the mantle and the core. Melting and segregation of metal, sulfide, and silicate (i.e., core formation) most likely occurred during or shortly after accretion, but new metal-silicate equilibria was not necessarily established, and the bulk Mg/(Mg + Fe) ratio of the mantle is only indirectly related to core compositions. Partitioning

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of K, U, or Th into the core is thus unlikely. The light element in the core is probably mainly sulfur (Ahrens, 1979). If all element fractionation processes occurred at low pressures in small bodies or dispersed phases in the inner solar system close to 4.56 aeons, then the meteorites provide the classical Goldschmidtian evidence of siderophile-chalcophile-lithophile (plus volatile-refractory) element fractionation, which points to sulfur accreted as troilite or oldhamite as the most viable candidate. Limited high-pressure reaction between metal and silicate is not precluded but is not extensive in this scenario, in which the preexisting metal and sulfide fall out to form the core during melting concomitant with accretion.

Some further consequences follow. The satellites of the outer planets have low densities, although Ganymede, Callisto, Titan, and Triton are about the size of Mercury. Accordingly, it appears that free metal was not available in large amounts beyond the asteroid belt. Probably volatile-refractory element fractionation, which may be linked to early solar flare-ups in the inner solar system, was not effective at and beyond 5 AU (as is shown, inter alia, by the large amounts of condensed water/ ice in the Jovian and Saturnian satellites). Accordingly, it may be predicted that the volatile-refractory element ratios (e.g., K/U ratios) in these satellites are about the same as those in CI meteorites. This would have the consequence that radioactive heat generation is generally higher in these bodies than for the inner planets. The decrease in density of the Galilean satellites with distance from Jupiter is probably due to mild warming of a proto-Jupiter disk, changing the rock/ice ratio. This scenario provides a low-energy analogue of events close to the sun in the early solar nebula.

6. Conclusions

The current state of geochemical knowledge for the Moon, Earth, and meteorites clarifies several points with respect to the origin of the Moon.

1. Classical capture models can be excluded.

2. The composition of the bulk Moon differs significantly from that of the terrestrial mantle. Fission models can be ruled out, unless very complex fractionation processes occur following fission but preceding accretion of the Moon.

3. Large Mars-sized impactor models, in which the material making up the Moon is both derived from the impactor and is vaporised during the event, become viable. They cut the Gordian knot of Earth-Moon comparisons at the cost of making the chemical and isotopic composition of the impactor free parameters.

4. Double-planet scenarios remain viable (Taylor, 1982) since the meteorite data predict a population of precursor planetesimals depleted in siderophile and volatile elements. The problems of accreting these to the Moon remains an outstanding problem, possibly answered by Ruskol-type scenarios.

5. The origin of the Moon is complex and elements of several theories are needed to account for its composition. This, however, is not unique, and geochemical processes

operating at the beginning of the solar system produced similar depletions in siderophile and volatile elements to those observed in the Moon.

6. Hypotheses in science should be testable. A test to distinguish between the two viable hypotheses of lunar origin, large impactor and double planet, can be made in principle by dating the volatile depletion event. In the simplest case, this depletion occurs at T_o for the double-planet hypothesis, with the Moon then accreting in a circumterrestrial disk from fractionated planetesimals. The Mars-sized impactor hypothesis predicts that major volatile depletion occurs in a vapor phase as a consequence of the impact. However, this scenario is complicated by the documented meteoritic evidence of volatile depletion at T_o, so that we need to identify the signature of a second volatile depletion event.

7. References

Anders E. (1978) Procrustean science: Indigenous siderophiles in the lunar highlands, according to Delano and Ringwood. *Proc. Lunar Planet. Sci. Conf. 9th,* pp. 161–184.

Ahrens T. J. (1979) Equation of state of iron sulfides and constraints on the sulfur content of the Earth. *J. Geophys. Res., 84,* 985–998.

Baldwin R. B. (1949) *The Face of the Moon.* University of Chicago Press, Chicago. 239 pp.

Boynton W. V. (1983) Cosmochemistry of the rare earth elements: meteorite studies. In *Rare Earth Element Geochemistry* (P. Henderson, ed.), pp. 63–114. Elsevier, Dordrecht.

Cameron A. G. W. (1984) Formation of the prelunar accretion disk (abstract). In *Papers Presented to the Conference on the Origin of the Moon,* p. 58. Lunar and Planetary Institute, Houston.

Cameron A. G. W. and Ward W. R. (1976) The origin of the moon (abstract). In *Lunar Science VII,* pp. 120–122. The Lunar Science Institute, Houston.

Clayton R. N. (1978) Isotopic anomalies in the early solar system. *Annu. Rev. Nuclear Sci., 28,* 501–522.

Delano J. W. and Lindsley D. H. (1983) Mare glasses from Apollo 17: Constraints on the Moon's bulk composition. *Proc. Lunar Planet. Sci. Conf. 14th,* in *J. Geophys. Res., 88,* B3–B16.

Delano J. W. and Ringwood A. E. (1978) Siderophile elements in the lunar highlands: nature of the indigenous component and implications for the origin of the moon. *Proc. Lunar Planet. Sci. Conf. 9th,* pp. 111–159.

Dickinson T. and Newsom H. (1984) Ge abundances in the lunar mantle and implications for the origin of the moon (abstract). In *Papers Presented to the Conference on the Origin of the Moon,* p. 16. Lunar and Planetary Institute, Houston.

Drake M. J. (1986) Is lunar bulk material similar to Earth's mantle?, this volume.

Gehrels T. (ed.) (1978) *Protostars and Planets.* University of Arizona Press, Tucson. 756 pp.

Haines E. L. and Metzger A. E. (1980) Lunar highland crustal models based on iron concentrations: Isostasy and center-of-mass displacement. *Proc. Lunar Planet. Sci. Conf. 11th,* pp. 689–718.

Hansen K. S. (1982) Secular effects of oceanic tidal dissipation on the moon's orbit and the Earth's rotation. *Rev. Geophys. Space Phys., 20,* 457–480.

Hartmann W. K. (1965) Terrestrial and lunar flux of large meteorites in the last two billion years. *Icarus, 4*, 157–165.

Hartmann W. K. and Davis D. R. (1975) Satellite-sized planetesimals and lunar origin. *Icarus, 24*, 504–515.

Hood L. L. and Jones J. (1985) Lunar density models consistent with mantle seismic velocities and other geophysical constraints (abstract). In *Lunar and Planetary Science XVI*, pp. 360–361. Lunar and Planetary Institute, Houston.

Kaula W. M. and Beachy A. E. (1984) Mechanical models of close approaches and collisions of large protoplanets (abstract). In *Papers Presented to the Conference on the Origin of the Moon*, p. 59. Lunar and Planetary Institute, Houston.

Langseth M. G., Keihm S. J., and Peters K. (1976) Revised lunar heat flow values. *Proc. Lunar Planet. Sci. Conf. 7th*, pp. 3143–3171.

Liu L. and Bassett W. A. (1985) *Elements, oxides and silicates: High pressure phases.* Oxford University Press, in press.

Mason B. and Taylor S. R. (1982) Inclusions in the Allende meteorite. *Smithson. Contrib. Earth Sci., 25*, 1–30.

McKenzie D. and Richter F. M. (1981) Parameterized thermal convection in a layered region and thermal history of the earth. *J. Geophys. Res., 86*, 11667–11680.

Morgan J. W., Higuchi H., Takahashi H., and Hertogen J. (1978) A "chondritic" eucrite parent body: Inference from trace elements. *Geochim. Cosmochim. Acta, 42*, 27–38.

Newsom H. E. (1984) The lunar core and the origin of the moon. *EOS (Trans. Am. Geophys. Union,) 65*, 369–370.

Newsom H. E. (1986) Constraints on the origin of the moon from the abundance of molybdenum and other siderophile elements, this volume.

Nyquist L. E. (1977) Lunar Rb-Sr chronology. *Phys. Chem. Earth, 10*, 103–142.

Rasmussen K. L. and Warren P. H. (1985) Megaregolith thickness, heat flow and the bulk composition of the Moon. *Nature, 313*, 121–124.

Ringwood A. E. (1970) Origin of the moon: The precipitation hypothesis. *Earth Planet. Sci. Lett., 8*, 131–140.

Ruskol El. Y. (1977) The origin of the moon. *NASA SP 370*, p. 815–822.

Smith J. V. (1979). Mineralogy of the planets: A voyage in space and time. *Mineral. Mag., 43*, 1–89.

Smith J. V. (1982) Heterogeneous growth of meteorites and planets, especially the earth and moon. *J. Geol., 90*, 1–125.

Smith M. R., Laul J. C., Ma M. -S., Huston T., Verkouteren R. M., Lipschutz M. E., and Schmitt R. A. (1984) Petrogenesis of the SNC meteorites: Implications for their origin from a large dynamic planet, possibly Mars. *Proc. Lunar Planet. Sci. Conf. 14th*, in *J. Geophys. Res., 89*, B612–B630.

Spudis P. D. (1984) Apollo 16 site geology and impact melts: Implications for the geologic history of the lunar highlands. *Proc. Lunar Planet. Sci. Conf. 15th*, in *J. Geophys. Res., 89*, C95–C107.

Stevenson D. J. (1984) Lunar origin from impact on the Earth (abstract). In *Papers Presented to the Conference on the Origin of the Moon*, p. 60. Lunar and Planetary Institute, Houston.

Stolper E. M. (1977) Experimental petrology of the eucritic meteorites. *Geochim. Cosmochim. Acta, 41*, 587–611.

Stolper E. M. and McSween H. Y. (1979) Petrology and origin of the shergottite meteorites. *Geochim. Cosmochim. Acta, 43*, 1475–1498.

Swindle T. D., Caffee M. W., Hohenberg C. M., Hudson G. B., Laul J. C., Simon S. B., and Papike J. J. (1984) Noble gas component organisation in Apollo 14 breccia 14318; [129]I and [244]Pu regolith chronology. *J. Geophys. Res.*, in press.

Swindle T. D., Caffee M. W., Hohenberg C. M., and Taylor, S. R. (1986) I-Pu-Xe dating and the relative ages of the Earth and Moon, this volume.

Taylor S. R. (1975) *Lunar Science: A Post-Apollo View*. Pergamon, New York. 372 pp.

Taylor S. R. (1979) Nickel abundances in lunar mare basalts (abstract). In *Lunar and Planetary Science X*, pp. 1215–1216. Lunar and Planetary Institute, Houston.

Taylor S. R. (1982) *Planetary Science: A Lunar Perspective*. Lunar and Planetary Institute, Houston. 481 pp.

Taylor S. R. and McLennan S. M. (1979) Chemical relationships among irghizites, zhamanshinites, Australasian tektites and Henbury impact glasses. *Geochim. Cosmochim. Acta, 43*, 1551–1565.

Taylor S. R. and McLennan S. M. (1985) *The Continental Crust: Its Composition and Evolution*. Blackwell, Oxford. 312 pp.

Urey H. C. (1962) Origin and history of the moon. In *Physics and Astronomy of the Moon* (Z. Kopal, ed.), pp. 481–523. Academic, New York.

Walker D. (1983) Lunar and terrestrial crust formation. *Proc. Lunar Planet. Sci. Conf. 14th*, in *J. Geophys. Res, 88*, B17–B25.

Warren P. H. (1984a) Megaregolith thickness, heat flow and the bulk composition of the moon (abstract). In *Papers Presented to the Conference on the Origin of the Moon*, p. 18. Lunar and Planetary Institute, Houston.

Warren P. H. (1984b) The bulk-moon MgO/FeO ratio: A highlands perspective (abstract). In *Papers Presented to the Conference on the Origin of the Moon*, p. 19. Lunar and Planetary Institute, Houston.

Weidenschilling S. J. (1984) The lunar angular momentum problem (abstract). In *Papers Presented to the Conference on the Origin of the Moon*, p. 55. Lunar and Planetary Institute, Houston.

Wetherill G. W. (1980) Formation of the terrestrial planets. *Annu. Rev. Astron. Astrophys., 18*, 77–113.

Wilhelms D. E. (1985) Lunar impact rates reconsidered (abstract). In *Lunar and Planetary Science XVI*, pp. 904–905. Lunar and Planetary Institute, Houston.

Wise D. U. (1963) An origin of the moon by rotational fission during formation of the earth's core. *J. Geophys. Res., 68*, 1547–1554.

Wolf R. and Anders E. (1980). Moon and Earth: Compositional differences inferred from siderophiles, volatiles and alkalis in basalts. *Geochim. Cosmochim. Acta, 44*, 2111–2124.

Wood J. A. and Mitler H. E. (1974) Origin of the moon by a modified capture mechanism (abstract). In *Lunar Science V*, pp. 851–853. The Lunar Science Institute, Houston.