Moon Over Mauna Loa: A Review of Hypotheses of Formation of Earth's Moon

JOHN A. WOOD

Harvard-Smithsonian Center for Astrophysics, 60 Garden Street, Cambridge MA 02138

Important constraints that a model of lunar formation must satisfy are first reviewed: the large mass of the Moon and the substantial prograde angular momentum of the Earth-Moon system; the Moon's depletion in volatile elements and iron (the Moon contains about one-fourth of its cosmic complement of Fe); the correspondence of oxygen isotope signatures in Earth and Moon; and the lunar magma ocean. Some similarities in elemental abundance patterns between the mantles of Earth and Moon have been noted, but there is disagreement as to how and whether this should be applied as a constraint.

Five major models of lunar formation are then reviewed: (1) capture from an independent heliocentric orbit; (2) coaccretion from a swarm of planetesimals in geocentric orbit; (3) fission from a rapidly rotating Earth; (4) collisional ejection, in which a Mars-size planetesimal impacts the Earth, resulting in an orbiting debris disk from which the Moon accretes; and (5) disintegrative capture, wherein a large planetesimal passing through Earth's Roche zone is disrupted by tidal forces and the debris forms an orbiting disk from which the Moon accumulates. The coaccretion model, which has probably had the widest (if most passive) following in the past, is found to have a severe problem: It seems impossible to account for the angular momentum of the Earth-Moon system if it formed by this means. In the light of our present knowledge, the collisional ejection model appears to be the most plausible alternative.

1. Introduction

Considering how conspicuous and starkly beautiful Earth's Moon is, and what a unique object it is in the solar system, it is surprising how little scientific thought was given to its origin before the present epoch. The Moon appears only as a minor detail in the grand World Systems of authors such as Immanuel Kant (1755): Typically these writers considered that satellite systems, including Earth's Moon, are small-scale analogues of the planetary system, and that they formed about their primary planets just as the planets formed about the sun. The only widely cited reference on lunar origin from this period is George Darwin's (1879) treatise on fission of the Moon from the Earth. Early discussions of lunar origin are reviewed in this volume by Brush (1986).

It appears that the first person who really cared about the origin of the Moon was Harold Urey (1952). The Moon was a crucial element of Urey's World System, not a detail in it. He believed the Moon to be a "primary object," probably the only survivor in the inner solar system of a primordial population of bodies that accreted from presolar interstellar dust when the solar system formed. Collisions among the other primary objects, and complex fractionation processes that subsequently affected the debris, produced planets and meteorites with evolved, noncosmic chemical compositions. Urey's great interest in the Moon and his scientific prestige made him a powerful advocate of lunar exploration in the period when NASA was founded and potential space missions were being defined.

By 1964, with four different U.S. programs of lunar exploration at various stages of execution, scientific interest in the Moon had become intense. During January 20-21 of that year, just five months before the first successful Ranger flight to the Moon, a conference on "The Dynamics of the Earth-Moon System" was held at the Institute for Space Studies (New York City) of the NASA Goddard Space Flight Center. Discussions centered on the origin of the Moon via Earth fission or capture, and the subsequent orbital evolution of the Moon. This was (to my knowledge) the first conference that focused on the question of the origin of the Moon. A proceedings volume was published (*The Earth-Moon System*, B. G. Marsden and A. G. W. Cameron, eds., Plenum, New York, 1966).

There was widespread expectation that the Apollo exploration of the Moon would settle the question of its origin; this had been cited frequently as one of the scientific goals of the Apollo program. However, the Moon turned out to be a highly differentiated body, one that preserved a priceless record of the earliest igneous activity and geochemical fractionations in a small planet, but not of how the body formed. The only clear reading on lunar origin to come from Apollo data was that Urey's concept of a cold, primitive Moon had been wrong. As Apollo scientists set to work studying the geological evolution of the Moon, the question of its origin receded into the background.

Recently the question has been reopened by another conference on "The Origin of the Moon," this one held at Kona, Hawaii (October 14-16, 1984) and cosponsored by the Division of Planetary Seiences of the AAS, the Lunar and Planetary Institute (Houston), and NASA. It was the first conference on lunar origin since Apollo; the first, in fact, since the Institute for Space Studies meeting 20 years earlier. The conference revealed that more progress has been made in understanding lunar origin than most of us suspected. Apollo science contributed to the advance, but in relatively indirect ways. Most of the gain in understanding (if it is not illusory) has come from more sophisticated dynamical studies and from a more mature appreciation of the context of the problem, the larger question of the origin of the planetary system. A major share of credit probably goes to digital computers, which have come into widespread use in the years since the Goddard Institute for Space Studies meeting.

The present paper describes and critiques five models of lunar formation in the light of the discussions at Kona. I begin by reviewing those properties of the Moon that seem most likely to preserve information about its mode of origin. A successful model of lunar origin must account for these properties, so they can be thought of as constraints as well as clues.

2. *Clues and Constraints*

2.1. *Mass of the Moon*

The Moon is far more massive, relative to its primary, than is the satellite or satellite system of any other planet, with the probable exception of Charon (Table 1). This indicates rather clearly that the event or process that created Earth's Moon was an unusual one, and suggests that some relaxation of the normal aversion to invoking *ad hoc* circumstances might be permissible in this problem.

TABLE 1. Mass Ratios, Satellite(s)/ Primary Planet

Mercury	0.
Venus	0.
Earth	0.0122
Mars	0.00000002
Jupiter	0.00021
Saturn	0.00025
Uranus	0.00017
Neptune	0.0013
Pluto	~0.04

2.2. *Angular momentum of the Earth-Moon system*

The Moon moves in its geocentric orbit in the same prograde sense that the Earth rotates. The amount of angular momentum in the Earth-Moon system is substantial, 3.45×10^{41} rad g cm^2/sec. This is the sum of the rotational angular momentum of the Earth and the orbital angular momentum of the Moon. Figure 1 shows a relationship between angular momentum density and planetary mass, noted by MacDonald (1966); this suggests that the Earth-Moon system contains an anomalously large amount of angular momentum. However, it is questionable whether the curve of Fig. 1, which is established largely by the momenta of the Jovian planets, can be validly applied to the terrestrial planets, which may well

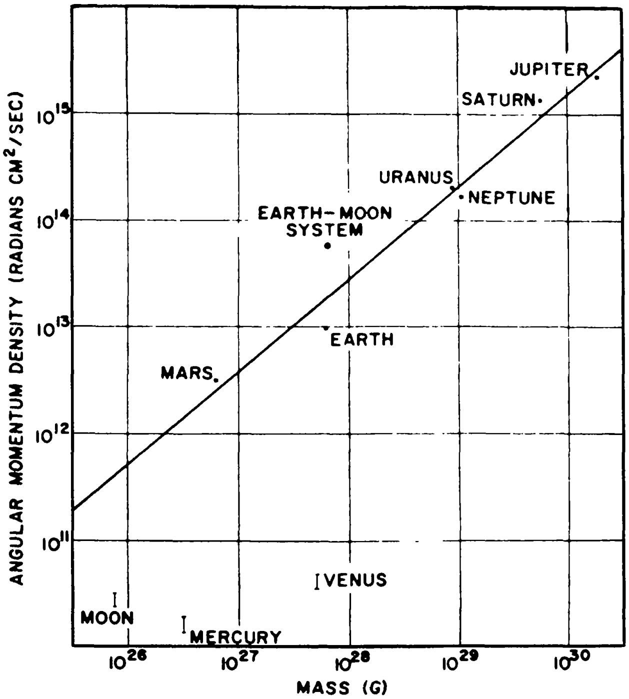


Fig. 1. The angular momentum density of planets (and the Earth-Moon system) as a function of their masses. Figure from MacDonald (1966).

have formed by a fundamentally different mechanism than the Jovian planets (accretion of solids vs. gravitational instability of nebular gases). Only one of the four terrestrial planets, Mars, falls on the curve. Hartmann and Larson (1967) and Burns (1975) have redefined the curve by including data for a number of asteroids at the low-mass end of the plot. This makes the Earth-Moon system fall on the curve, and leaves Mars somewhat deficient in angular momentum. But the value of this generalization is also questionable, since the data embrace objects formed by not two but three mechanisms (most of the asteroids are collisional debris).

That the Earth-Moon system is turning may seem a trivial point to use as a constraint, but as will be seen one of the long-established models of lunar formation has trouble accounting for this basic property. There has, of course, been major exchange of angular momentum between the Earth's rotation and the Moon's orbit. The Moon was once much closer to the Earth than it is now, possibly only a few Earth radii away from it, and tidal interactions have decelerated the rotation of the Earth while they accelerated the Moon and expanded its orbit. There is an extensive literature of theoretical studies of lunar orbital evolution (e.g., Burns, 1977). These considerations, and in particular the angular difference between the inclination of the lunar orbit and the obliquity of Earth's rotation axis, which figures importantly in orbital evolution, are often cited as constraints on lunar orgin (e.g., MacDonald, 1966). This article does not include them among constraints: In the first place, uncertainties in the orbital evolution calculations make it impossible to reliably project back to the earliest epochs of orbital history (e.g., Boss and Peale, 1986), and in the second place, the crucial angular difference between Earth's equator and the lunar orbital plane can have been arbitrarily changed at an early time by the impact of a single relatively small planetesimal at a high latitude on Earth, or on the Moon.

2.3. *Volatile element depletion*

One of the first observations made in the study of Apollo samples was that the lunar rocks and soils are markedly depleted in volatile elements. The degree of the volatile depletion is indicated in Fig. 2, which compares elemental abundances in the lunar mantle (as deduced from abundances in low-Ti mare basalts) with abundances in undifferentiated planetary material (C1 carbonaceous chondrites).

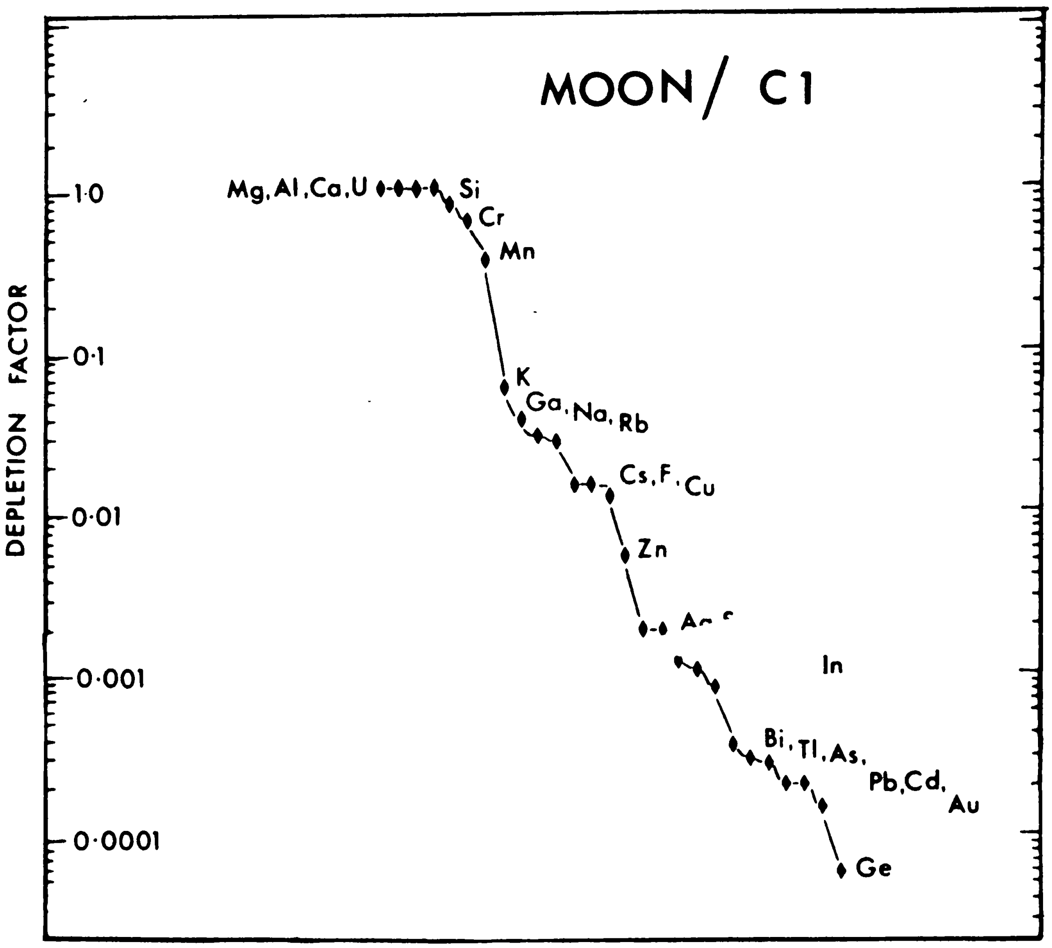


Fig. 2. Comparison of elemental abundances in the mantle source region of low-Ti mare basalts with abundances in C1 chondrites. Volatility of the elements tends to increase to the right. Figure from Ringwood and Kesson (1977).

It should be kept in mind that the Earth is also depleted in volatile elements, and so may be all the terrestrial planets. However, the Moon is much more severely depleted in volatiles than the Earth: For example, the ratio K/U in lunar materials is only about one-fifth as great as in terrestrial rocks. (Potassium and U are volatile and involatile elements, respectively, which have little tendency to be fractionated from one another by igneous processes.)

It is widely believed that the missing volatiles were lost from the substance of the Moon during some highly energetic process of lunar formation that heated the material to very high temperatures. The volatile elements were vaporized at that time, for the most part, and dispersed in space. All the models of lunar formation to be discussed are high-energy processes with the potential for volatile loss.

I-Pu-Xe dating shows that the last lunar volatile element fractionation occurred 20–100 m.y. after the formation of the most primitive meteorites (Swindle *et al*, 1986), a time consistent with any of the models of lunar formation reviewed in this paper. However, the similarity of initial $^{87}Sr/^{86}Sr$ values in the Moon (LUNI) and in basaltic achondrites (BABI) (Papanastassiou and Wasserburg, 1976) shows that the lunar material was depleted in the volatile element Rb very early, before the canonical $\sim 10^8$ yr of planetary accretion had elapsed (Taylor, 1986). Since all the models of lunar formation that are reviewed are tied to the time scale of planetary formation, this means none of them can be held fully responsible for the volatile depletion of lunar material, if the 10^8 yr figure is correct. The Moon would have to be made of material already depleted in volatiles (as the Earth was). This does not, of course, *exclude* a high-temperature process of lunar formation with additional volatile losses.

2.4. Elevated abundances of refractory elements

The Moon's depletion in volatile elements and the unexpected abundance of the mineral plagioclase (mostly $CaAl_2Si_2O_8$) in samples from the lunar crust also led to a conclusion, early in the Apollo program, that levels of refractory elements in the Moon are greatly enhanced. (Ca, Al, Ti, Ba, Sr, Sc, Y, the rare Earth elements, Zr, Hf, Th, V, Nb, Ta, Mo, W, U, Re, Ru, Os, Rh, Ir, and Pt are the refractory elements. They were believed to be concentrated in the Moon to levels far above the cosmic relative to Mg, Si, and Fe, the major elements of intermediate volatility.) This led to the proposal of several models of lunar origin (Anderson, 1972; Cameron, 1972) that involved incorporation in the Moon of large amounts of refractory Ca,Al-rich material of the type that has been studied extensively in the Allende meteorite.

However, as more geochemical and petrological evidence accumulated, the degree of the hypothetical refractory element enhancement diminished. Current estimates of refractory element abundances in the Moon (e.g., Ca and Al; expressed as oxides in Table 2) do not exceed the cosmic levels by factors large enough to affect ideas about lunar origin. Furthermore, the observations on which these estimates are based

*TABLE 2. Concentrations of the Refractory Oxides CaO and Al_2O_3 in the Moon, According to Several Models of the Lunar Bulk Composition.**

	CaO	Al_2O_3	MgO
Morgan *et al.* (1978)	6.1%	7.6%	29.1%
Taylor (1982)	4.5	6.0	32.0
Wänke *et al.* (1977)	3.8	4.6	32.4
Ringwood (1979)	3.7	4.2	32.7
Cosmic			
C1 chondrites	2.7	3.3	(32.0)

**C1 chondrite levels adjusted to MgO = 32.0.*

(Apollo heat-flow measurements at two sites, from which levels of the refractory element U are inferred, and measurements made by orbital geochemical sensors of the composition of the surface layer of the lunar highlands, from which the composition of the entire lunar crust has been inferred) have too much uncertainty associated with them to demonstrate conclusively that refractory elements are concentrated to any degree at all above cosmic in the Moon.

2.5. *Depletion of iron in the Moon*

The Moon's mass density (3.344 ± 0.002 g/cm^3; Bills and Ferrari, 1977) is much less than the uncompressed densities of the terrestrial planets (~3.7–5.4 g/cm^3), which means the Moon must contain a smaller proportion of some heavy major element than the planets do. Only Fe is heavy enough and abundant enough to make the difference. It was recognized before Apollo that the Moon must be deficient in Fe relative to Earth. What is generally pictured is that the dense metallic Fe core that contributes so importantly to the Earth's density is largely missing from the Moon; and indeed, geophysical measurements constrain the Moon's core, if it has one at all, to constitute no more than ~6% of the lunar mass (Hood, 1986), in contrast to a terrestrial core that makes up 32% of the Earth's mass. In fact, however, a broader statement than this can be made: The Moon must be deficient in *elemental* Fe—Fe in any form—not just metallic Fe. The low lunar density cannot be rationalized by postulating that the Moon contains Fe in the cosmic proportion, but all of it is oxidized to FeO and incorporated in silicate minerals.

This can be shown by calculating the mineral norm of a simple Moon consisting only of Si, Mg, and Fe in their cosmic proportions, with enough O to oxidize them fully (the Fe to FeO). The norm consists of 92.3 wt % olivine, 7.7% orthopyroxene; Mg/(Mg + Fe) = 0.56. Using the mineral density data of Deer *et al.* (1962), the net specific gravity of this assemblage is found to be 3.69 g/cm^3, much greater than the lunar density. To make this simple system reproduce the lunar density, its content of Fe must be reduced to ~0.25 of the cosmic abundance. In this case the norm works out to 34% olivine, 66% orthopyroxene, Mg/(Mg + Fe) = 0.84, density = 3.34 g/cm^3. This Mg/(Mg + Fe) ratio is close to the values indicated for the lunar mantle by mare basalt petrology (0.75–0.80; Ringwood, 1979) and geophysical constraints (0.75–0.85; Hood, 1986).

Thus the Moon failed to incorporate three-quarters of its cosmic complement of Fe when it formed. This is one of the strongest constraints that bear on the question of lunar origin. Most models of lunar origin dispose of the unwanted Fe mechanically: Much of the Fe is seen to exist as a metal phase apart from the silicate phases that will eventually join the Moon, and differences in physical properties (density, strength) of the metal and silicate phases cause them to be mechanically fractionated. However, there is another possibility that should also be considered. Hashimoto (1983) has experimentally evaporated oxide mixtures at high temperatures in a vacuum; his work shows that under these circumstances FeO is a relatively volatile component that is lost from the silicate melt more readily than are MgO, SiO_2, CaO, and Al_2O_3 (Fig. 3). It is possible that the Moon's Fe deficiency is simply an aspect of the general pattern of depletion of volatile elements, and that all these elements were distilled away at a time when the substance of the Moon was dispersed and very hot.

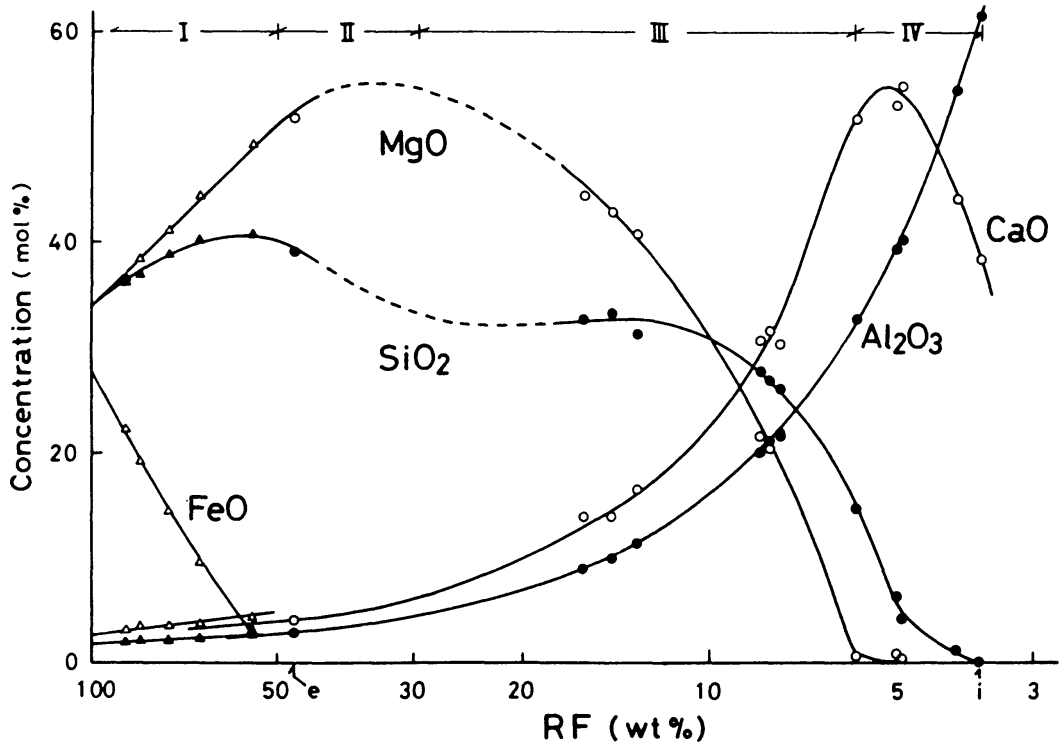


Fig. 3. Change of composition of residual molten oxide material as a charge is vaporized away into a vacuum at 2073 K. RF is the wt % of the original charge remaining. Initial charges contained solar proportions of the oxides shown. Figure from Hashimoto (1983).

2.6. *Oxygen isotopes*

The oxygen isotope signature of lunar materials collected by the Apollo missions is identical to that of the Earth to within measurement error (Clayton and Mayeda, 1975). Oxygen isotope compositions seem to vary with position in the solar system, since most types of meteorites differ in this property from the Earth and Moon (Clayton *et al.*, 1976). Therefore the similarity of oxygen in the Earth and Moon strongly suggests that both bodies were made from the same batch of protoplanetary material, at the same radial distance from the sun.

2.7. *Element abundance patterns in the lunar and terrestrial mantles*

Siderophile trace elements are highly depleted in the Earth's mantle. Presumably they were absorbed in drops of molten metal as the latter sank to the core during the primary differentiation of the Earth. Only minor aliquots of these elements, amounts defined by their respective metal-silicate melt partition coefficients, were left in the silicate minerals of the mantle. The mantle of the Moon is also depleted in these elements. The abundance patterns of many of the siderophile elements residual in the mantles of Earth and Moon are similar. Does this mean the two mantles had a common origin, and therefore the Moon was created by some form of fission from the Earth?

A. E. Ringwood and coworkers (Ringwood and Kesson, 1977; Delano and Ringwood, 1978a,b; Ringwood, 1979, 1986) have argued with some conviction that it does. They point out that abundances of the siderophile elements Co, Ni, W, P, S, Se, and Te are the same to within a factor of about two in the terrestrial and lunar mantles, and argue that two separate core-forming fractionations operating in an Earth and Moon that were always independent of one another could not have produced such a correspondence: The partitioning of these elements depends upon pressure, temperature, fO_2, and fS_2 in the interiors of the fractionating bodies, and these conditions are known to have been very different in the early Earth and Moon.

H. Wänke and colleagues (Rammensee and Wänke, 1977; Wänke *et al.*, 1978; Wänke and Dreibus, 1986) share the conviction of Ringwood and his coworkers. They draw attention to the depleted but remarkably similar concentrations of the lithophile elements V, Cr, and Mn in the terrestrial and lunar mantles. Again, this could not be a coincidence; these workers propose that fractions of the elements named joined the terrestrial core as sulfides or oxides before the substance of the Moon was separated from the Earth's mantle, leaving the observed residual amounts in the two mantles.

Drake (1983) has pointed out that for the set of siderophile elements P, Mo, Ge, and Re, mantle concentrations are not similar in Earth and Moon: These elements are more depleted in the Moon, and the amount of depletion correlates with the degree of siderophilic behavior of the element (i.e., its metal/silicate partition coefficient; Fig. 4). Thus if the Moon did separate from the Earth's mantle, an additional amount of core formation had to occur in the Moon but not the Earth, after the separation, to additionally deplete these elements only in the Moon. This hypothetical secondary depletion was earlier proposed by Wänke *et al.* (1978), and more recently has been endorsed by Ringwood and Seifert (1986).

Drake (1983) and also Newsom (1986) conclude that the lunar siderophile pattern is consistent with formation either in the Earth or as an independent system. Morgan *et al.* (1978) and Anders (1978) argue that the fractionation did not occur in the Earth. Kreutzberger *et al.* (1985) show that the patterns of alkali volatile trace elements in Earth and Moon are inconsistent with derivation of all the substance of the Moon from the Earth. The topic of mantle trace element abundances as a key to the origin of the Moon is complex and arcane, and an area of little consensus. A detailed discussion is outside the scope of this paper; the interested reader is referred to reviews by Newsom (1984) and Drake (1986). The present author agrees with Taylor (1982) that too many factors cloud the picture to allow mantle compositions to be used as a convincing argument either for or against formation

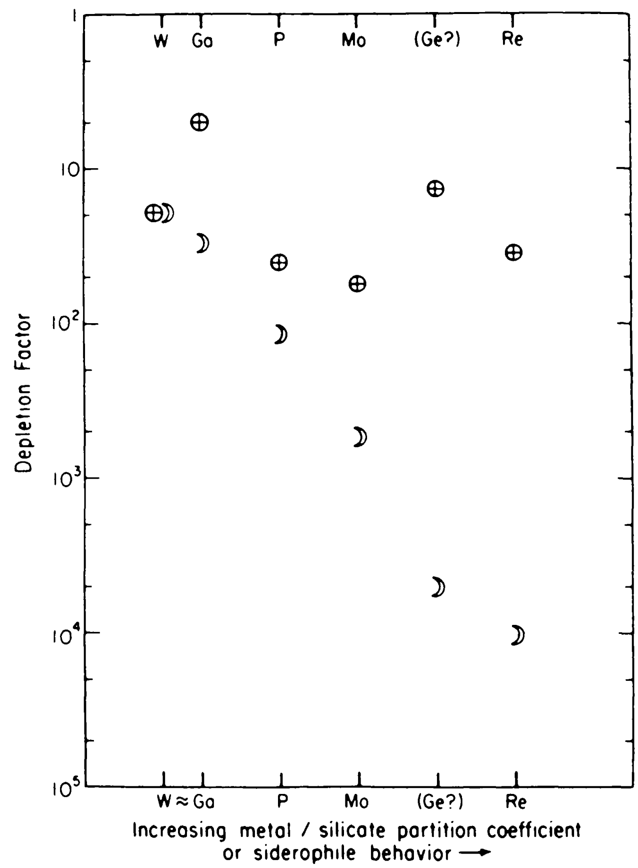


Fig. 4. The depletion of six siderophile elements in mantles of the Earth and Moon, as inferred from the compositions of basalts on both bodies. "Depletion factor" is the ratio (abundance in mantle)/(abundances in C1 chondrites). The "?" associated with Ge refers to a paucity of data and also the fact that the depletion is partly due to the volatile nature of Ge. These elements are seen to be much more depleted in the Moon than in the Earth, and the degree of depletion correlates with the metal/silicate partition coefficient. Figure from Drake (1983).

of the Moon from the Earth. Consider the number of ways in which the lunar and terrestrial trace element signatures would have been changed or obscured if the Moon *had* formed from the Earth's mantle:

1. The partial vaporization of protolunar material during formation of the Moon that probably contributed to the depletions of volatile elements observed in lunar materials (noted above) would also deplete siderophile elements selectively.

2. By far the most plausible mechanism of Earth fission involves the glancing impact of a large planetesimal that ejected material from the mantle of the Earth (as discussed below). A major amount of debris from the impacting planetesimal inevitably would be mixed with the mantle debris, imparting a composite trace element signature to the material that ultimately accreted into a Moon. Cameron (1985, 1986) concludes that *most* of the material placed in orbit by such an event would be derived from the impactor.

3. Additional core formation and fractionation of mantle siderophiles is likely to have occurred in the Earth after the fission event.

4. Additional core formation and fractionation is also likely to have occurred in the Moon after it accreted; indeed, this seems required. The nature and degree of fractionation is bound to have differed in the two bodies.

5. The Earth probably accreted additional material from heliocentric orbit after fission, and to some degree incorporated its siderophile elements in the terrestrial mantle.

6. The Moon also accreted additional material after its formation. The amounts of new material gained and the nature of its incorporation in the bodies of the Moon and Earth undoubtedly differed.

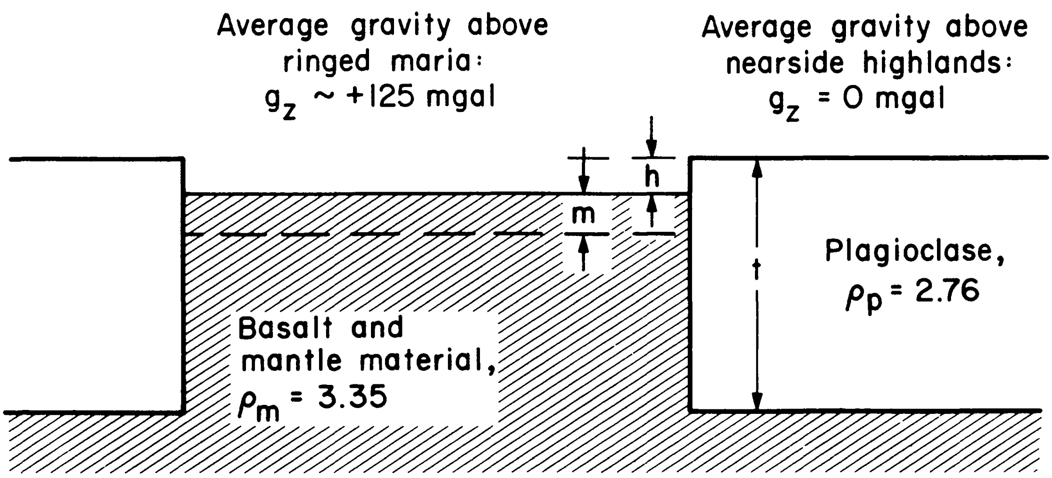
7. It is not straightforward to infer the composition of the lunar mantle from mare basalt and highlands rock samples. As Taylor (1982) points out, no lunar basalts can be identified as coming from primitive unfractionated material of bulk lunar composition.

8. Similarly, we have no assurance that present-day estimates of the terrestrial mantle composition (based on basalt compositions and xenoliths) are representative of the composition of the mantle at the time of the hypothetical lunar fission event.

2.8. The lunar magma ocean

Apollo samples from the highlands (terrae) of the Moon were found to be enriched in calcic plagioclase feldspar. Some of the samples are true anorthosites (~95% plagioclase); the mean composition of surface material in the highlands all the way around the Moon, as measured by orbiting geochemical sensors, corresponds to that of an anorthositic gabbro containing ~75% plagioclase (Warren, 1985). Magmas produced by the partial melting of planetary interiors are constrained by phase equilibria to contain less than ~55% plagioclase, so some other igneous process must be held responsible for the concentration of plagioclase at the lunar surface. Crystal fractionation is the universally accepted mechanism: Plagioclase is a relatively low-density phase, and would tend to float up to the top of lunar magma bodies and concentrate there.

The amount of plagioclase in the lunar crust is very great. Gravity and topography measurements provide a way of estimating the amount. The highlands are "high" because they are made of relatively low-density rock that "floats" in denser lunar mantle material. The highland rock is light because of its content of plagioclase, the only abundant low-density lunar mineral. For a given crustal density, the total thickness of floating crust needed to buoy the highlands to their observed mean altitude above the maria can be simply calculated. The formula appears in Fig. 5, along with gravity and topographic parameters applicable to the lunar near side (Kaula *et al.*, 1974); these are conservative values. For a hypothetical crust of pure anorthite plagioclase, the crustal thickness t works out to 20.4 km. To collect this much plagioclase from a Moon originally of chondritic composition (minus metal,



h = mean altitude difference, ~ 2.6 km

m = thickness of mare basalt that would have to be removed to eliminate +125 mgal gravity anomaly, leaving highlands and ringed maria in isostatic equilibrium, ~ 1 km

t = mean crustal thickness

$$\rho_p t = \rho_m \left[t - (h + m) \right]$$

Fig. 5. Summary of gravity and topography parameters on the lunar nearside, and the formula relating them to the mean crustal thickness t (Kaula et al., 1974). Gravity in the highlands is arbitrarily set at 0 mgal; the 125 mgal mean positive gravity anomalies (mascons) in the ringed maria are relative to this value.

sulfide, alkalis) requires that ~30% of its volume (i.e., to a depth of ~200 km) be partially melted, and the plagioclase components extracted. This assumes the plagioclase components are extracted and sent to the lunar crust with perfect efficiency; otherwise a larger fraction of the lunar volume would have to be extracted. Wood *et al.* (1970) and Smith *et al.* (1970) concluded that the outer layers of the Moon, to a depth of several hundred kilometers, were melted when the Moon formed. An integral magma ocean seemed required in order to give plagioclase crystals the freedom to float up from great depths and join the crust. This concept, modified in various ways, has gained wide acceptance. The topic is competently reviewed by Warren (1985).

The lunar crust does not actually consist of pure plagioclase, of course, but of rocks that contain various proportions of other minerals (mostly pyroxenes and olivine). The gravity/topography criterion informs us of the total amount of low-density plagioclase in the crust, but not its degree of dilution by denser minerals. The crust below the plagioclase-rich surface might consist entirely of gabbroic rock containing only ~50% plagioclase; of course, such a crust would have to be thicker than the 20.4 km calculated for pure plagioclase in order to buoy the highlands. These gabbros could have been produced in the Moon by partial melting over a protracted period of time and delivered to the surface little by little ("serial magmatism"); then there would be no need for an integral magma ocean to permit plagioclase crystal flotation. Wetherill (1975) and Walker (1983) have proposed effectively this, dispensing with the magma ocean. In their models the plagioclase-rich rocks characteristic of the near-surface highlands were produced by crystal flotation in gabbroic magma chambers of limited scale, and/or basaltic flow units.

However, these more prosaic models of lunar crust formation have a severe problem. Crystal fractionation in a gabbroic magma chamber would produce ultramafic cumulate rocks (by crystal sinking) as well as plagioclase cumulates (by crystal flotation). If the scale of the magma chambers was smaller than the depth of the impact craters that have excavated and scattered crustal material over the lunar surface, then fragments of ultramafic cumulate rock as well as anorthosite should be abundant on the lunar surface. However, ultramafic rocks are conspicuously absent from lunar highlands regions. Only one such sample, dunite 72415, was among the individual rock samples collected by the Apollo astronauts. The ejection depth from lunar mare basins is estimated to be 30–80 km (Grieve, 1980; Spudis, 1983), i.e., most of the thickness of the lunar crust. Therefore plagioclase fractionation must have consistently occurred in magma bodies of more than 50–120 km vertical extent in order to bury the ultramafic cumulates deeply enough to prevent them from being excavated during the subsequent cratering history of the Moon. Such bodies approach the vertical scale of the hypothetical magma ocean. It does not seem possible to avoid invoking very large-scale surface magmatism on the Moon, though it is arguable whether a surface shell was molten everywhere on the Moon simultaneously, as in the classic magma ocean.

The principal motivation for discounting evidence of a magma ocean has been the difficulty in accounting for a source of heat to melt it. Traditionally, accretional energy has been invoked for this purpose, but Ruskol (1973), Wetherill (1975), and Kaula (1979) have argued that the heat released during formation of the Moon by coaccretion in a geocentric planetesimal swarm is marginal to inadequate to melt a magma ocean. However, the problem may be with the coaccretion model rather than with the magma ocean. As will be seen, other models have no trouble melting a magma ocean during lunar formation.

Like the volatile-poor character of the lunar samples, the evidence for something like a magma ocean early in the history of the Moon requires the mode of origin of the Moon to be a high-energy process. However, where the volatile depletion seems to require that the substance of the Moon was hot while it was in a dispersed state (which facilitates devolatilization), the magma ocean requires high temperature after the Moon accreted.

3. *Models of Lunar Origin*

Traditionally, the debate about lunar origin has focused on three possibilities: (1) that the Moon was formed in heliocentric orbit elsewhere in the solar system, then captured intact into geocentric orbit; (2) that it accreted as a companion to the Earth from a disk of planetesimals and particles that had been captured into geocentric orbit; and (3) that it fissioned from an early Earth that was rotating so rapidly as to be unstable. More recently, two hybrid models have been added to the lineup: An off-center impact by a very large planetesimal (more properly a planet) threw debris into a geocentrically-orbiting disk from which the Moon accreted; or, a similarly large object that barely missed the Earth was tidally disrupted as it passed through the Roche zone; debris from the Earth-facing side of the object was retained in geocentric orbit, and the Moon accreted from it. These five models will be discussed in turn.

3.1. Intact capture

The intact capture model was much favored in the 1960s (Urey, 1966; MacDonald, 1966; Singer, 1968), but interest in it seems to have waned. Most recently it has been defended by Turcotte and Kellogg (1986) and Singer (1986).

Originally, the most attractive aspect of the capture model was its potential for rationalizing the major compositional differences (especially Fe) between Earth and Moon. If two objects were made in widely separated parts of the solar system and then joined by capture, a large compositional difference would not be hard to understand (e.g., Cameron, 1972). However, it has come to be appreciated that the Moon cannot have been captured from a remote region of the solar system. First, this entails motion of the proto-Moon in an eccentric orbit stretching from its place of origin to the Earth's orbit, which means the encounter would occur at relatively high velocity, and as will be seen this makes capture virtually impossible. Horedt (1976) estimates that a proto-Moon would have to be formed at a mean distance in the range 0.95–1.05 AU and with eccentricity very approximately 0.04 to be eligible for capture. Second, the essentially identical isotopic compositions of lunar and terrestrial oxygen appears to require that the substance of both bodies came from the same part of the solar system. Since the Moon cannot be "exotic," the capture model is deprived of its main selling point.

The dynamics of capture are best explored in the framework of the restricted three-body problem of celestial mechanics. Motions are considered in a reference frame that moves with the Earth and remains aligned with the sun-Earth line. In such a frame, the positions of critical boundary surfaces (zero relative velocity, or Hill, surfaces) can be calculated by use of Jacobi's integral. These surfaces define regions of space where the proto-Moon can and cannot move. The positions of the surfaces depend upon the amount of mechanical energy associated with the proto-Moon's motion relative to the Earth, and the masses of the sun and Earth. In a very low-energy situation, which corresponds to a small proto-Moon velocity relative to the Earth, a zero-velocity surface surrounds the Earth. An object moving around the Earth inside this boundary cannot cross it; the object is permanently locked in geocentric orbit. However, a proto-Moon from elsewhere in the solar system also cannot be captured into this situation, because if its energy is low enough for its Jacobi integral to predict a closed zero-velocity surface around the Earth, it cannot cross the intervening forbidden region and get inside this surface.

On the other hand, if the proto-Moon's energy is high, the zero-velocity surfaces expand, open out, and come to include the sun and a vast region of solar system space. An exotic proto-Moon has no trouble approaching the Earth, but there is nothing to hold it there; it invariably swings back out into distant reaches of space.

The situation is most interesting for intermediate values of proto-Moon energy. Then the Earth can be almost but not quite completely surrounded by a zero velocity surface: A door is left open at the l_1 libration point, on the side of the Earth toward the sun. The lower the energy, the smaller the door. An object from elsewhere in the solar system, having an amount of energy appropriate for this zero velocity surface, can wander in the door and become temporarily trapped. It circulates around and around the Earth until it finally finds its way out again (Fig. 6). Actually the zero velocity surface defined by the present Earth-Moon system has a tiny door; technically the Moon could escape the Earth and go into heliocentric orbit, and conversely there is a very small but nonzero dynamic probability that it was captured from heliocentric orbit without the assistance of energy-dissipating mechanisms (Kopal and Lyttleton, 1963; Szebehely and McKenzie, 1977).

Most capture models rely on the premise, whether explicitly stated or not, that the proto-Moon approached the Earth through a fairly large door in the zero velocity surface surrounding it; some dissipative effect then decreased the proto-Moon's energy,

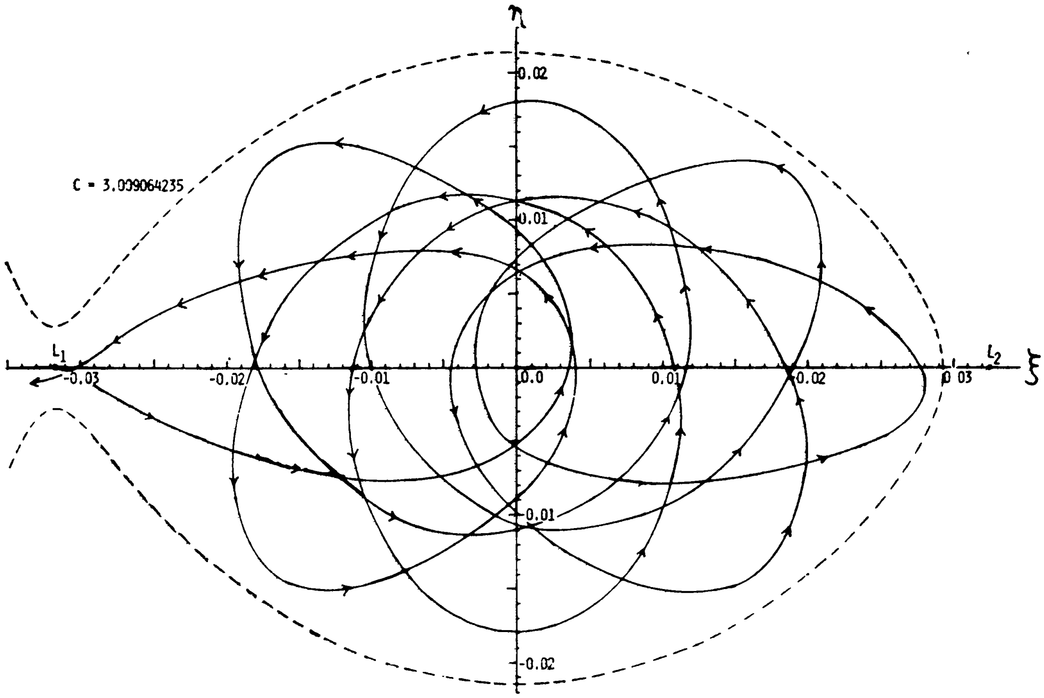


Fig. 6. Example of temporary capture of a satellite in the reference frame of the restricted three-body problem. The capturing planet is at the origin, the sun on the x-axis, far to the left out of the figure. The dashed line is the zero velocity surface for a satellite with a particular small amount of energy. Note that there is a door at the l_1 point to the left of the figure. The satellite for which this zero velocity surface is applicable enters the door and describes seven prograde orbits about the planet before it finds its way out the door again. Figure from Heppenheimer and Porco (1977).

which had the effect of pulling the zero velocity surface in and shrinking its door, so the Moon couldn't find its way out again. The mode of energy dissipation most often invoked is tidal interactions (MacDonald, 1966; Singer, 1968; Öpik, 1972; Winters and Malcuit, 1977), but there are several other possibilities. Kaula and Harris (1973) appeal to collisions with a circumterrestrial swarm of planetesimals; Nakazawa *et al.* (1983) appeal to gas drag in a massive primordial Earth's atmosphere that extended to the zero velocity surface. (In the latter case the timing must be exquisite: The atmosphere must be in the process of dissipation from the Earth at the time when the Moon begins to be decelerated by it, or the Moon will spiral into the Earth.) Clark *et al.* (1975) rely on a fortuitous sudden increase in the mass of the Earth (as by the accretion of a single major planetesimal), rather than energy dissipation, to close the zero velocity surface while the proto-Moon was passing the Earth.

The importance of a zero velocity surface with a small door is that once a proto-Moon is inside, it continues to circulate around the Earth in the same direction

(Fig. 6) and the energy-dissipating mechanism, whatever it may be, can operate repeatedly and with additive effect. The capture of an exotic object that approaches the Earth with high velocity and large energy is ruled out because its zero velocity surface would be too open to trap it, so it could make only one pass by the Earth; and it would have far too much energy to be dissipated in one pass.

Lunar capture is inherently improbable because the range of approach velocities and trajectories that corresponds to a zero velocity surface favorable for entrapment is extremely narrow. Most discouraging, however, is the absence of an explanation for the compositional difference between Earth and Moon (since exotic objects are ruled out). If the two bodies formed independently by similar processes of accretion at the same radial distance in the solar system and from the same reservoir of primordial material, it is reasonable to expect that they would end up with similar inventories of chemical elements. Other models involve more or less different processes of formation for the Earth and Moon, and their advocates take advantage of this to try to rationalize the compositional differences between the two bodies.

3.2. Coaccretion

The idea that the accreting Earth accumulated a disk or swarm of orbiting solid particles from which the Moon then accreted was first articulated by Schmidt (1959), and later developed by Ruskol (1960, 1963, 1972a, 1973), Harris and Kaula (1975), and Harris (1978). Most recently, the concept has been defended by Weidenschilling *et al.* (1986).

The great virtue of this model is that it does not invoke some special, low-probability event. The underlying premise is that the Earth accreted and the protolunar swarm accumulated in a natural way from untold numbers of small heliocentrically orbiting objects, and the dynamic and material properties of the Earth and Moon result inevitably from an averaging of the approach trajectories and behavior upon mutual interaction of these many objects. Coaccretion has probably gained wider acceptance than any of the other lunar origin hypotheses among lunar scientists, who were trained to shun *ad hoc* assumptions. However, the model has difficulty accounting for the compositional differences between Earth and Moon, the melting of the magma ocean, and the angular momentum content of the Earth-Moon system.

Heliocentrically orbiting planetesimals that approached the growing Earth but did not collide with it would have described quasihyperbolic trajectories about Earth's center of mass. Some planetesimals would have passed the Earth in a prograde direction, others in a retrograde sense. Collisions between these two populations would cancel some of the angular momentum of the planetesimals, relative to the Earth; a portion of the collision debris would fall to Earth, and some would be decelerated just enough to leave it in geocentric orbit. It has generally been understood (but see below) that most planetesimals approached the growing Earth-Moon system in a prograde sense—hence the Earth's prograde rotation—so a disk or protolunar swarm (PLS) of decelerated collision debris with net prograde rotation should have built up in orbit around the proto-Earth. Once the beginnings of a disk were established, it would have greatly accelerated the capture of additional disk material, since the collision probability for an incoming heliocentric planetesimal with accumulated disk material was far larger than the likelihood that it would encounter another heliocentric planetesimal. This mechanism of PLS formation is accepted by all the authors cited above. In addition, Weidenschilling *et al.* (1986) raise the possibility that a giant Earth impact of the sort that is central to the collisional ejection model (to be discussed) threw material into geocentric orbit, making a major contribution to the PLS.

Proponents of the coaccretion model explain the volatile element deficiency of the Moon as resulting from devolatilization of PLS material during the many high-energy collisions that established the disk, or as a property of the material that predated its incorporation in the PLS. Ruskol (1972b) proposed that the iron deficiency of the Moon is a consequence of the difference in strength and ductility between silicate minerals and metallic iron. Collisional fragmentation of protoplanetary material, both prior to and during encounter with the PLS, would reduce the relatively weak, brittle silicates to systematically smaller and lighter fragments than Fe metal. The silicate objects, with their lower momentum content, were more readily decelerated and captured by the PLS than the Fe fragments, so the PLS took on an Fe-depleted composition. Thus the PLS operated as a "compositional filter" (Weidenschilling *et al.*, 1986). The process would be particularly effective if, as is generally assumed, most of the protoplanetary material had already had an existence in small bodies that melted and differentiated (as the achondrite and iron meteorite parent bodies did), after which collisions among the bodies reduced many of them to debris fragments. Large Fe-metal core fragments would be especially difficult for the PLS to decelerate.

A crucial aspect of the process is the timing of the nucleation of the Moon from the PLS. Gravitational instability would cause an orbiting swarm to coagulate into one or more proto-Moons in a relatively short time, but Ruskol (1972a) argues that lunar embryos, which formed too early, while the space density of heliocentric planetesimals was still high, could not survive; they would be disrupted by high-velocity collisions with the abundant heliocentric planetesimals. She estimates that the embryos formed or were captured when the Earth had attained about half its present mass, and that there were probably two or three of them. They grew to ~1000 km radius by sweeping up low-velocity PLS material; thereafter they were large enough to accumulate high-velocity heliocentric planetesimal material, as well as additional PLS material. An advantage of several lunar embryos is that it increases the capture cross-section for material in heliocentric orbit, and hence the net accretion rate. Ultimately these moonlets merged to form the Moon.

Harris and Kaula (1975), assuming that once the lunar embryo formed all subsequent accretion was from heliocentric orbits, assessed the relationship between nucleation time and the final Moon/Earth mass ratio, and concluded that the lunar embryo formed relatively early, when the Earth had only ~0.1 its present mass. A difficulty with this picture and also Ruskol's, however, is that direct accretion of heliocentric material would not discriminate between silicate and iron objects; the compositional filter is turned off. Harris (1978), in a reanalysis of the situation, included the effect of accretion to the Moon of material that continued to join the PLS outside the orbit of the lunar embryo. This material is gradually decelerated by "accretion drag" (its encounter with heliocentric particles that have less angular momentum relative to Earth than it has), spirals inward, and is swept up by the Moon. For this model, Harris found that the lunar embryo can have nucleated when the Earth had attained ~0.4 of its final mass, and that one-third to one-half of its mass can have come from the PLS rather than directly from heliocentic orbit. (Even this does not rationalize the lunar composition, however, since, as noted, approximately three-quarters of the Moon's cosmic complement of Fe has to be withheld from it.)

The problem with melting a magma ocean during coaccretion of the Moon is that the heat source which is traditionally invoked, namely the gravitational potential energy that is converted to heat during the accretion of the Moon, may be released too gradually to be effective. Ruskol (1973) argues that the accretion of the Moon would not have ended until planetary accretion did, since protoplanetary material continued to be available for accretion until that time, and the duration of planetary accretion was ~10^8 yr. Another advantage of Ruskol's (1973) plan to make the Moon from several submoons is that a major fraction of the potential energy of accretion would be converted to heat in the ~1 hr during which two submoons of comparable size coalesced; this goes a long way toward providing the heat needed to melt the magma ocean. The melting of a magma ocean is also more plausible if lunar accretion was suppressed until late in the ~10^8 yr interval of planetary accretion, after which it proceeded very rapidly (Weidenschilling *et al.*, 1986), and/or if the act of accretion threw up a transient lunar atmosphere that retarded radiative heat loss (Matsui and Abe, 1986).

The most serious problem for the coaccretion model appears to be accounting for the angular momentum of the Earth-Moon system. Only as recently as the Kona Conference was this problem widely appreciated. It has always been understood, and occasionally proven in rather broad terms (e.g., Schmidt, 1959; Lyttleton, 1972), that the accretion of a planet from dispersed, heliocentrically orbiting material would naturally transform some of the prograde orbital angular momentum of the raw material into rotational angular momentum of the planet. As a corollary, material captured into Earth's PLS would also have net prograde angular momentum. Giuli (1968) attempted to confirm this by simulating the accretion of the Earth on a planetesimal-by-planetesimal basis, using a computer to follow the trajectories and assess the angular momentum inputs of discrete objects. (For the moment I will leave the PLS aside, and concentrate on the Earth.) Giuli investigated a number of sets of a, e for planetesimals, which define orbits in two dimensions that are capable of accreting the planetesimals to the proto-Earth, if the latter is in the right place at the right time in its orbit ($a = 1$ AU, $e = 0$). For each planetesimal a and e, there is a limited range of the angle θ (see Fig. 7) for which accretion can occur. Giuli refers to all the orbits in one of these ranges of θ, having a common a and e, as a "band." He found that for most bands investigated, one end of the angular range corresponds to a trajectory that barely skims the limb of the Earth in a prograde sense; the other end of the range corresponds to skimming the retrograde limb of the Earth, and trajectories in between impact symmetrically across the body of the Earth, half imparting prograde and half retrograde angular momentum (Fig. 8). Thus accretion of a large number of planetesimals from these bands would impart no net angular momentum to the Earth; their effects would cancel each other.

Some of Giuli's bands did not impact the Earth symmetrically. He found that bands originating in circular orbits contributed net negative angular momentum to the Earth. On the other hand, bands originating in eccentric orbits, for which Earth

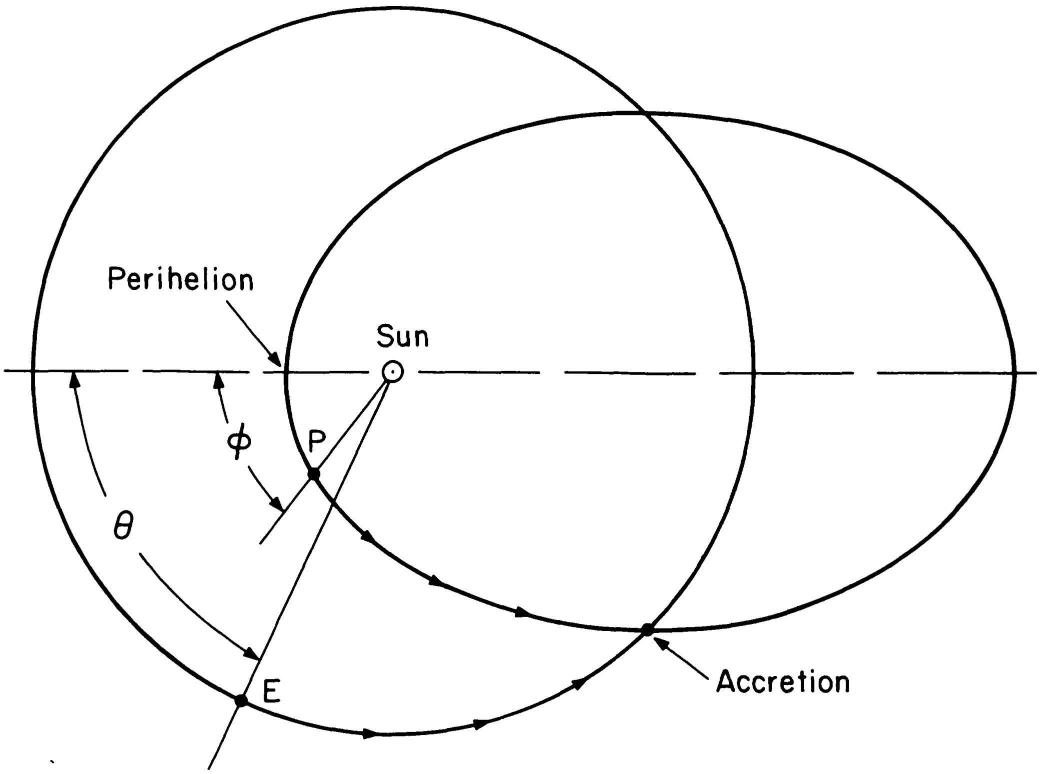


Fig. 7. Trajectories followed by Earth (started at E) and a planetesimal (started at P) during simulated Earth accretion. Variables in the situation are a *and* e *for the planetesimal orbit and also the angle* Θ*, which describes the orientation of the planetesimal orbit relative to the starting position of Earth. In general there is a small range of values of* Θ *(*ϕ *held constant) that result in Earth impact. Exactly where the planetesimal strikes the surface of the Earth, and how much angular momentum it imparts to the Earth, vary with the value of* Θ *in this range; see Fig. 8.*

impact occurred at the lowest possible velocity (the escape velocity), produced net positive angular momentum. These corresponded to a very narrow range of *a* and *e*. Giuli's conclusion was that an accretion scenario can be postulated that would produce an Earth with a prograde rotation period of 15 hours. However, to do this requires the planetesimal feedstock for the Earth to come from orbits with an implausibly limited range of orbital elements: *a*, 0.958–0.973 or 1.029–1.046 AU; *e*, 0.025–0.039. These values hold for an Earth with its present mass; for an earlier, smaller proto-Earth, values of *a* and *e* even closer to 1 and 0 are required.

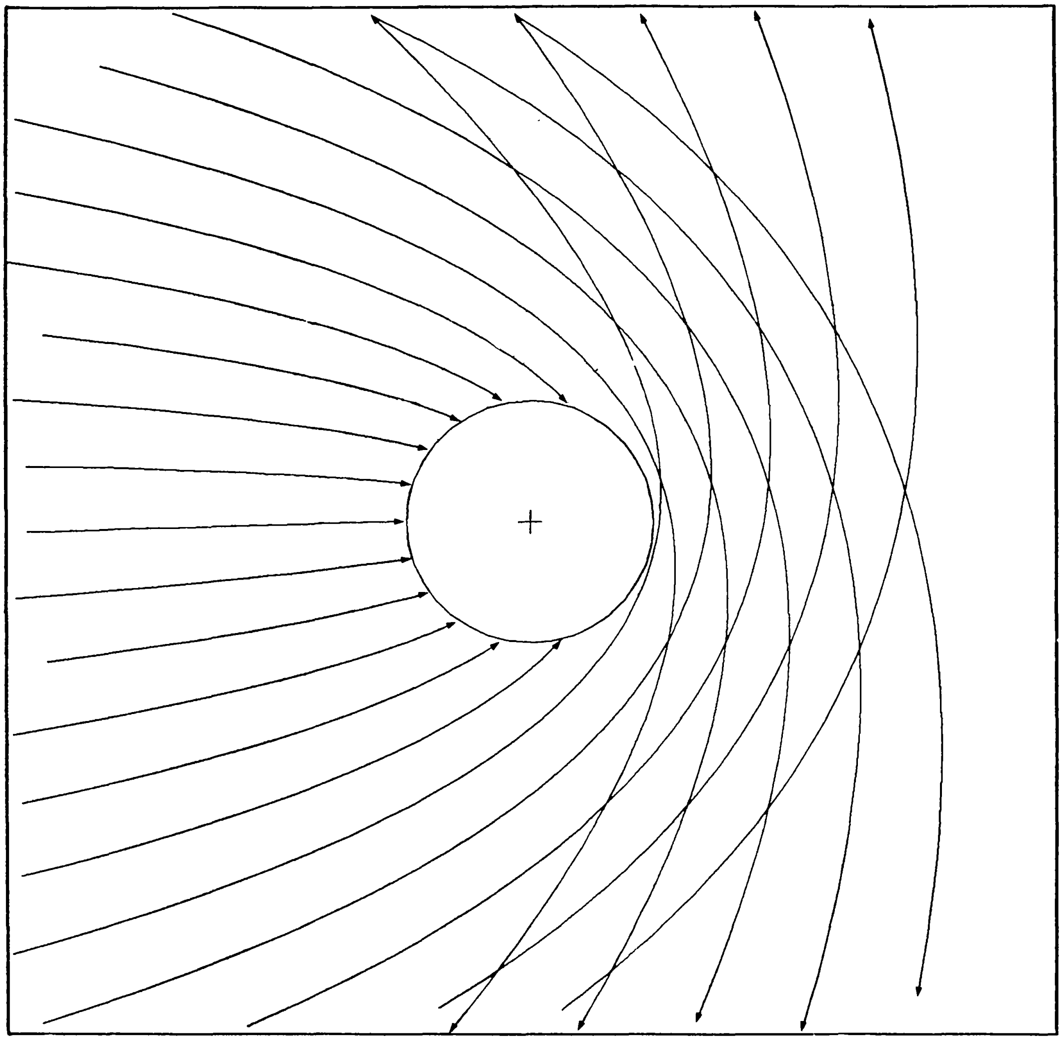


*Fig. 8. Trajectories of a "band" of planetesimals accreting to the Earth. Reference frame fixed in Earth; sun-Earth line maintains alignment. Earth has its present radius. Planetesimals shown originated in orbits having the same **a** (1 AU) and **e** (0.05), but with orientations of the line of apsides (Θ) rotated by equal small increments. Impacts that impart positive and negative angular momentum to the Earth are evenly balanced. Trajectories computed by the author (see Appendix).*

Harris (1977) noted that the bands of orbits that contribute positive angular momentum to Giuli's proto-Earth are just tangent to Earth's orbit at perihelion or aphelion. In fact, when Earth is not in the vicinity they don't quite cross Earth's orbit, but the presence of Earth's gravitational field enlarges their orbits to the point that the planetesimals can impact one (but only one) hemisphere of the Earth—the inner or sun-facing hemisphere if the planetesimal is at aphelion, the outer hemisphere for planetesimals at perihelion. Since eccentrically orbiting planetesimals at aphelion move more slowly than an object in circular orbit at the same radial distance, they are overtaken by the Earth; striking its sun-facing hemisphere, they impart prograde rotation to it. For analogous reasons the planetesimals at perihelion also impart prograde rotation to the Earth when they overtake it. In an analytical treatment of accretion that takes this effect into account, Harris (1977) confirmed that a suitably rotating Earth can be accreted from planetesimals if their range of orbital elements is sufficiently narrowly constrained. However, he observed that the value and range of e required are implausibly small, and an Earth accreted from planetesimals with a more credible range of orbital elements would have only ~10% of the required angular momentum.

Not surprisingly, it turns out that the same angular momentum problem is associated with the capture of a PLS from heliocentric orbits. Weidenschilling *et al.* (1986) studied the near-Earth trajectories of planetesimals started in a wide range of heliocentric orbits, and found it impossible to capture a PLS with an adequate amount of prograde angular momentum via planetesimal collisions in near-Earth space unless the source population of heliocentric planetesimals included very few objects in Earth-crossing orbit. This is equivalent to the orbital constraints Giuli (1968) invoked in order to explain the Earth's rotation. Weidenschilling *et al.* consider this to be a possible state of affairs, but their opinion is not widely shared. A commonly held view is that planetesimal eccentricities would have been large enough during the late stages of accretion (because of orbit perturbations by numerous growing subplanets in the system; e.g., Wetherill, 1986) to cause the great majority of objects entering near-Earth space to arrive in Earth-crossing orbits.

A simulation of Earth accretion and PLS accumulation (via disintegrative capture) carried out several years ago by the author (unpublished; reviewed in an Appendix to this article) confirmed the difficulty of accreting an Earth-Moon system with the observed content of angular momentum. This problem with the Earth's angular momentum forces a reexamination of the basic premise of coaccretion, that it consisted of the accumulation of a statistically large number of small planetesimals. The alternative one is driven to is a more hierarchical form of accumulation, involving a few very large planetesimals in the late stages. The fortuitous off-center impact of one large planetesimal could account for the rotation of the Earth, and perhaps also for the formation of the PLS. The absence of such an event might be the reason for the small, negative rotation rate of Venus.

3.3. *Earth fission*

In Darwin's (1879) original proposal of the fission origin of the Moon, rotational instability was not the cause of fission. Darwin assumed a prefission Earth with the same angular momentum as the present Earth-Moon system, rotating with a period of ~4 hours. He noted that the period of the slowest free oscillation of a fluid mass having the Earth's size and mean density is ~2 hours, and concluded that a resonance between this period and the period of tides raised on the Earth by the sun would cause the tidal bulges to grow higher and higher, until the tip of a highly elongated Earth broke off to become the Moon. The concept was refuted by Jeffreys (1930), who showed that frictional damping in the Earth, which increases as the cube of the tidal distortion, would prevent the amplitude of the distortion from growing large enough to result in fission.

The fission hypothesis was resurrected by Ringwood (1960) and Wise (1963), this time with rotational instability as the cause of the event. In this model the Earth was formed spinning at the verge of rotational instability, with a period of ~2.6 hours. Initially it had uniform density; segregation of its core decreased its moment of inertia, which caused it to spin faster (~2.1 hours), making it rotationally unstable. The Earth responded to this instability by adopting increasingly distorted geometrical forms (Fig. 9), until its most tenuously connected end broke off. The remaining Earth relaxed into a less distorted form and, rotating faster than the orbital period of the new Moon, began to tidally accelerate the motion and expand the orbit of the Moon. Jeffrey's objection does not apply in this case, because there is no differential rotation between the form and the substance of the prefission Earth. This model has been promoted by Wise (1969), O'Keefe (1966, 1969, 1972), O'Keefe and Sullivan (1978), and Binder (1974, 1978, 1986).

Recently, numerical hydrodynamic simulations of the fission of a rapidly spinning Earth have been carried out (Durisen and Scott, 1984; Durisen and Gingold, 1986; Boss and Peale, 1986) that represent a great advance over the crude understanding of the physical nature of the event that existed in the 1960s. These have shown that fission would not separate an intact Moon, as shown in Fig. 9, but instead dispersed material would be shed from the ends of the spinning elongated figure, building up a particulate disk or PLS. In some ways this makes the fission model more attractive, since hot, dispersed material shed by the Earth would be devolatilized more efficiently than a Moon fissioned intact, no matter how hot it was.

Advantages of the fission hypothesis are that it explains the Fe-poor character of the Moon, if the fission occurred after core formation in the Earth; and it also explains the common oxygen isotope signatures of the two bodies, the similarities that have been reported in trace element abundance patterns in the mantles of Earth and Moon, and the geochemical evidence that a large amount of metallic iron was once in contact with the substance of the Moon in spite of the small size of the

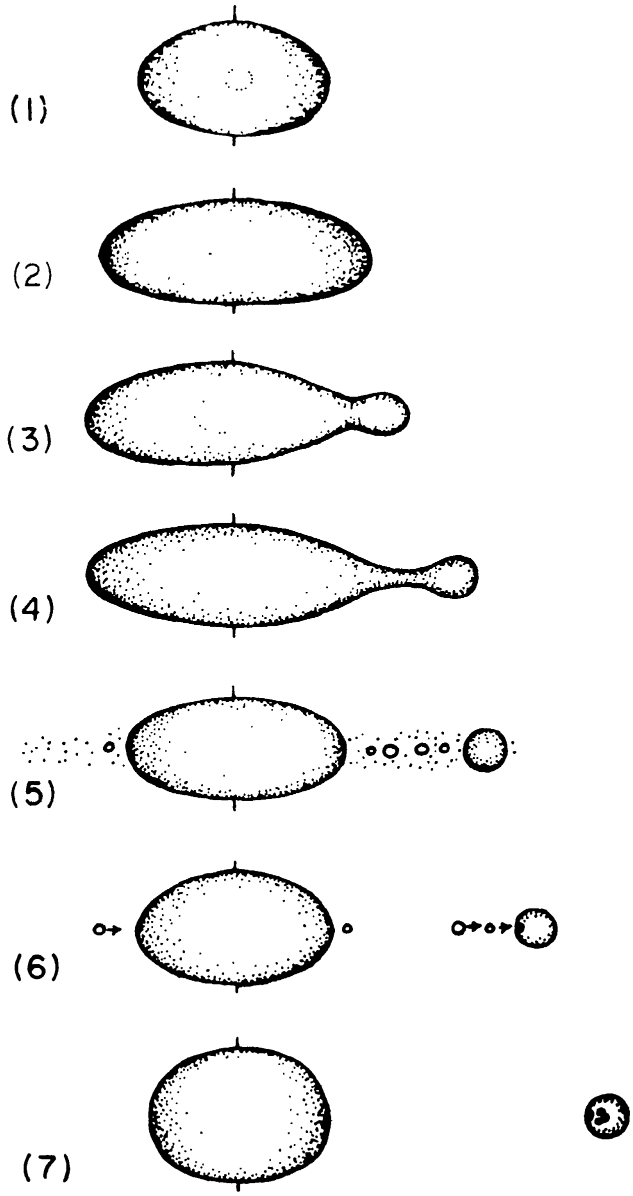


Fig. 9. Sequence of forms adopted by a rapidly spinning Earth after it becomes rotationally unstable. (1) Oblate Maclaurin spheroid; (2) Jacobian triaxial ellipsoid; (3) Poincare figure. Figure from Wise (1963).

present lunar core (Rammensee and Wänke, 1977; Ringwood, 1979). [But there is an uncomfortable aspect of this last triumph. Terrestrial core separation needs to be ~97% complete before fission occurs, to account for lunar siderophile element depletions. As O'Keefe and Sullivan (1978) point out, if fission is caused by core formation, it is unlikely that the process would go so nearly to completion before it caused destabilization.]

However, the fission model has very severe dynamic problems: In order to fission, the Earth had to have about four times as much angular momentum as the Earth-Moon system now has. There is no good explanation of why the Earth had such an excess of angular momentum in the first place, or where the surplus angular momentum went after fission occurred.

Why would a prefission Earth have so much angular momentum? Fission advocates rarely address this question. The previous section discussed at some length the evidence from a number of independent studies that accretion of the Earth from small heliocentrically orbiting planetesimals cannot even account for the amount of angular momentum the Earth-Moon system has, let alone four times this much. A series of major planetesimal impacts, all fortuitously off-center in a prograde sense, could do so, but lunar formation by this mechanism would have more in common with the collisional ejection model (discussed below) than with classic fission. Binder (1978, 1986) draws an analogy with contact binary stars, which have too much angular momentum to be stable as a single star. However, this has no relevance to the rotational state of planets and satellites. Stars are formed by the contraction of masses of interstellar gas to $\sim 10^{-7}$ of their original dimension. Conservation of even a fraction of the original angular momentum of the interstellar gas guarantees that the resulting stellar system will be rotating very rapidly, in many cases too rapidly to collapse, all the way to stellar dimensions as an integral object. No one has proposed that the terrestrial planets formed by an analogous process. G. P. Kuiper, H. C. Urey, and A. G. W. Cameron have advocated planet formation in gaseous protoplanets, but no element of shrinkage through many orders of magnitude of dimension with angular momentum conserved is involved. In these models the planets form by a rainout of refractory solids/liquids to the centers of the protoplanets, and it is to be expected that viscous interaction of the particles with protoplanet gases on the way down would remove most of their angular momentum, so that a planet which accumulated from them would be rotating at not much more than the moderate rate of the gaseous protoplanet itself.

Where could the excess angular momentum have gone after fission? Only very speculative answers have been offered. Wise (1969) suggests that the postfission Earth, at a very high temperature after the release of energy associated with accretion, core formation, rotational distortion, and tidal interaction with the fissioned Moon, boiled off a transient atmosphere of silicate vapors at ~5000K. If such an atmosphere extended to an altitude of one (additional) Earth radius, and rotated synchronously with the Earth beneath, then only 3.7% of the mass of the Earth would have to be lost by thermal escape from the top of the atmosphere to bring the angular momentum of the system down to its present value. (It is doubtful, however, that mechanical coupling between the Earth and its upper atmosphere would be good enough to brake the rotation of the Earth by the required amount.) O'Keefe (1972) proposes that the Moon that originally fissioned was much larger than the present Moon, ~0.1 Earth mass; the very high temperatures in that epoch boiled away 95% of the mass of the Moon. This mass and the angular momentum associated with it were lost to space, reducing the system's angular momentum by the required amount. (But the distillation away of such a large fraction of the substance of the

Moon should have left a residual body of very specialized composition: see Fig. 3. The Apollo samples do not bear this out.)

3.4. Collision ejection

The idea that the Moon was assembled from a disk of material that had been ejected from the Earth by a catastrophic collision was late in coming because (1) invoking the services of a sufficiently large planetesimal seemed uncomfortably *ad hoc*: Safronov (1966), always an advocate of a hierarchical process of accumulation of the planets, nonetheless estimated the size of the largest objects impacting the Earth toward the end of its accretion to be only $\sim 10^{-3}$ Earth mass, which is not large enough; and (2) it always seemed obvious that debris spalled off the Earth at less than the escape velocity would, after executing one geocentric orbit, return to its original geocentric radial distance, i.e., the surface of the Earth, and reaccrete.

Recently Hartmann and Davis (1975), Wetherill (1976, 1986), Greenberg (1979), and Hartmann (1986) have argued that the upper end of the size distribution of planetesimals during planetary accretion included much larger objects than those visualized by Safronov. According to this analysis the addition of an object of ~ 0.1 Earth mass to the Earth late in its accretion is to be expected, and is not *ad hoc* at all. Hartmann and Davis (1975) further pointed out that such a large collision might eject more than enough debris from the Earth to make a Moon, most of it Fe-depleted mantle material.

However, Hartmann and Davis did not address problem (2) above. Cameron and Ward (1976) first suggested a way out: An impact of the scale contemplated, between two bodies already hot and partially molten from the release of accretional and core-forming energy, would eject debris mostly in the form of vapor, not solids. The vapor would expand as it receded from the Earth, and gas pressure effects would continue to accelerate elements of the gas and also any solids entrained in them for some time after the material was initially dispatched from the point of impact. Because of this continued acceleration, the motion of the debris would not be strictly ballistic; it is to be expected that some of the debris would be accelerated into orbits whose perigees cleared the Earth. Cameron and Ward (1976) postulated an off-axis collision that would impart much of the early Earth's prograde angular momentum to it (Fig. 10), and most of the debris from such an impact would also move in a prograde sense. Thus after collisions and viscous interaction between elements of the ejecta placed in orbit had cancelled out most of its random motions, a disk of material would be left in prograde orbit about the Earth.

This theme has been developed further by Ward and Cameron (1978), Ringwood (1979), Cameron (1983, 1985, 1986), Thompson and Stevenson (1983), and Stevenson (1984, 1985). Major topics of discussion are the formation of the orbiting disk, its nature and evolution, and the source of observed similarities and differences in composition between Earth and Moon.

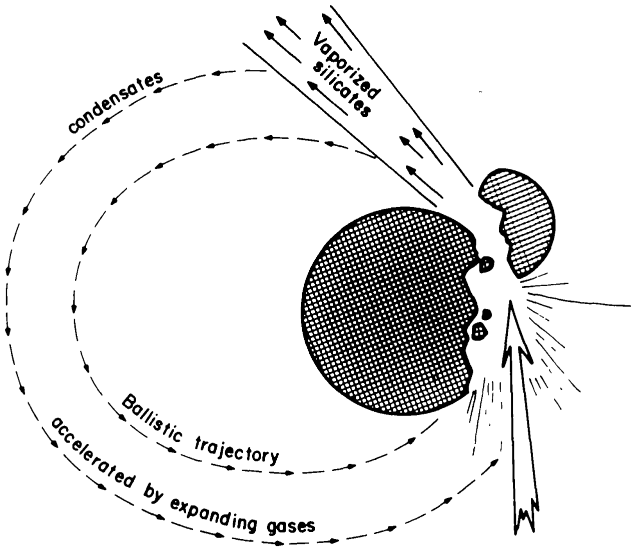


Fig. 10. The collisional ejection model: Expansion of ejected silicate vapor accelerates condensates into orbits with perigees that clear the Earth's surface (in contrast to the trajectories of solid debris knocked into simple ballistic orbits, which would reaccrete to the Earth).

Would the expanding-vapor model just sketched really place enough debris and enough angular momentum in geocentric orbit to make a Moon of? On the basis of a crude numerical simulation by Cameron (1985, 1986) and simple physical considerations cited by Stevenson (1984), the answer is yes. Success is favored by a large planetesimal (>0.1 Earth mass, i.e., Mars-size); a prograde off-axis impact point; and an impact velocity somewhat (~2 km/sec) above the escape velocity, which means a moderately eccentric (e ~0.2) preterrestrial orbit for the planetesimal. Cameron finds that most of the debris launched into geocentric orbit originated in the planetesimal, not the mantle of the Earth. Stevenson lists two physical effects in addition to vapor expansion that might have helped put a portion of the impact debris in orbit: the fact that the center of mass of the colliding system is significantly displaced from the center of mass of the Earth at the time of the collision; and the viscous transfer of energy and momentum in a broad jet of debris emanating from the impact.

Cameron and Ward (1976) assume the expanding vapor would begin to condense into solids soon after ejection, but Thompson and Stevenson (1983) argue that energy-dissipative processes (noted below) in the disk would keep it hot and largely vaporized during most of its existence. The disk is expected to behave qualitatively like the solar nebula: viscous interaction between differentially rotating annular elements of the disk cause it to spread, with angular momentum and some mass being transferred outward, the remaining mass inward. The spreading rate depends upon the effective viscosity of the disk, which controls the rate at which mechanical energy can be dissipated as heat. Ward and Cameron (1978) conclude that particulate disk material inside the Roche limit would try repeatedly to coalesce via gravitational instability, only to be torn apart by tidal forces. The "viscosity" of the disk would arise from the noncircular components of motion of particles as they moved into and out of these unsuccessful gravitational associations; this would determine the rate of energy dissipation in the disk. Normally a hot gaseous disk, such as Thompson and Stevenson (1983) envisage, would not be subject to such gravitational instability effects, because the tendency to instability is inversely related to the speed of sound in the gas, and the sound speed in hot gases is relatively high. However, Thompson and Stevenson point out that the sound speed in vapor-plus-solid or vapor-plus-liquid mixtures can be much lower than the value for pure vapors at the same temperature: As a result, the disk would have stayed hot, because wherever it became cool enough to condense extensively gravitational instability set in, the rate of viscous energy dissipation was turned up, and additional heat was generated. Consequently the disk could not have cooled and condensed fully until late in its spreading history.

As a result of this efficient energy-dissipation process and the high effective viscosity of the disk material, the disk is expected to expand beyond Earth's Roche limit and cool in ~100 yr. Once an important fraction of the disk material is outside the Roche limit, gravitational instability very promptly collects the condensed matter into Moonlets, and these soon coalesce to form the Moon. Because of the short time-scale of lunar accretion and the relatively high temperature of the solids being accreted, the Moon should be partly or wholly molten when it forms (Thompson and Stevenson, 1983).

This hypothesis is so new that its weaknesses have not yet become apparent. Much work needs to be done toward modelling the mechanical and thermal behavior of the hypothetical vapor/debris disk. No attempt has been made to understand the chemical fractionations that would result from formation of the Moon by collisional ejection. If Earth and the impactor had already separated cores, and if mostly mantle material was ejected and vaporized in the collision, this could account for the Fe-poor lunar composition. Also, Cameron (1985, 1986) argues that the Fe metal core of the impactor would be less decelerated during the collision than mantle material, because of its greater density; therefore the metal was less vaporized than silicates and was correspondingly underrepresented in the vapor disk. Another possibility is that the relatively high volatility of Fe (Fig. 3) prevented most of that element from being incorporated in the Moon along with the other volatile elements, even though it was present in the disk of vaporized debris. If the relatively eccentric planetesimal orbit favored by Cameron (1985, 1986) means that the planetesimal formed at a radial distance much different than 1 AU, we should worry about the close match in oxygen isotope compositions between Earth and Moon. Cameron postulates that the planetesimal was formed at ~1 AU and gravitationally perturbed into an eccentric orbit.

Undoubtedly collisional ejection would fractionate elements according to their volatility, which in principle would account for the volatile-depleted character of the Moon. However, there are limits on the degree to which volatility fractionation in the circumterrestrial disk can be invoked to rationalize the lunar composition, set by the fact that the rare Earth elements in the Moon have not been fractionated according to volatility (Taylor, 1986). Vapor fractionation appears to be limited to elements with condensation temperatures less than ~1100K. Kreutzberger *et al.* (1985) have shown that even at <1100 K the fractionation pattern is very difficult to understand.

3.5. Disintegrative capture

Öpik (1972) first showed that if an inviscid planetesimal in independent heliocentric orbit passed close enough to the Earth to be significantly decelerated by the tidal interaction (and hence potentially captured during a single passage), it would move inside Earth's Roche limit and would be tidally disrupted in the process. An interesting aspect of such an event, as Öpik pointed out, is that for a range of low-velocity encounters it would lead to geocentric capture of a portion of the debris of the disrupted planetesimal, even if tidal deceleration effects are ignored. Fragments derived from the inner (Earth-facing) side of the planetesimal would have been moving with the same velocity and energy as the center of mass of the planetesimal since (before disruption) they were bound together, yet the fact that they were closer to the center of the Earth would mean that the geocentric escape velocity applicable to them would be larger than the escape velocity for fragments derived from the center of mass or the outermost face of the planetesimal. For low-velocity encounters (asymptotic velocity <2 km/sec) this results in some debris from the inner side of the planetesimal being captured into geocentric orbits (Fig. 11). As much as 50% of the planetesimal debris can be captured, in the case of encounter at the parabolic velocity. It should be stressed that this material is captured because its position in the predisruption planetesimal constrains it to approach the Earth at an unnaturally low velocity, not because its motion is retarded by tides or any other dissipative process.

This concept was adopted by Wood and Mitler (1974) and Smith (1974), with the added observation that such a process would capture mostly crustal and mantle material from the inner face of a differentiated planetesimal; the bulk of the core material from such a system would be moving too fast to be captured, and would remain in heliocentric orbits. Thus the Fe-poor character of the Moon might be accounted for. Wood and Mitler (1974) proposed that many small planetesimals suffered this fate at the time when other such planetesimals were accreting to form the Earth, and that the integrated effect of many such disintegrative captures was to establish an Fe-poor PLS about the Earth. But this model suffers from the same angular momentum problem as coaccretion: There is no reason to expect substantially more planetesimals to be disrupted and partially captured while passing Earth in a prograde sense than in a retrograde sense.

Smith (1974) and Mitler (1975) postulated the disintegrative capture of just one planetesimal (again, Mars-size), which fortuitously encountered the Earth in a prograde direction. Clearly the conditions favorable for this disintegrative capture model and for collisional ejection (previous section) are almost identical. Smith (1974) and Mitler (1975) did not address the question of the rotation of the Earth itself. It may have been fortuitously the same as that of the PLS, but another possibility is that evolution of the PLS along the lines discussed in the last section caused

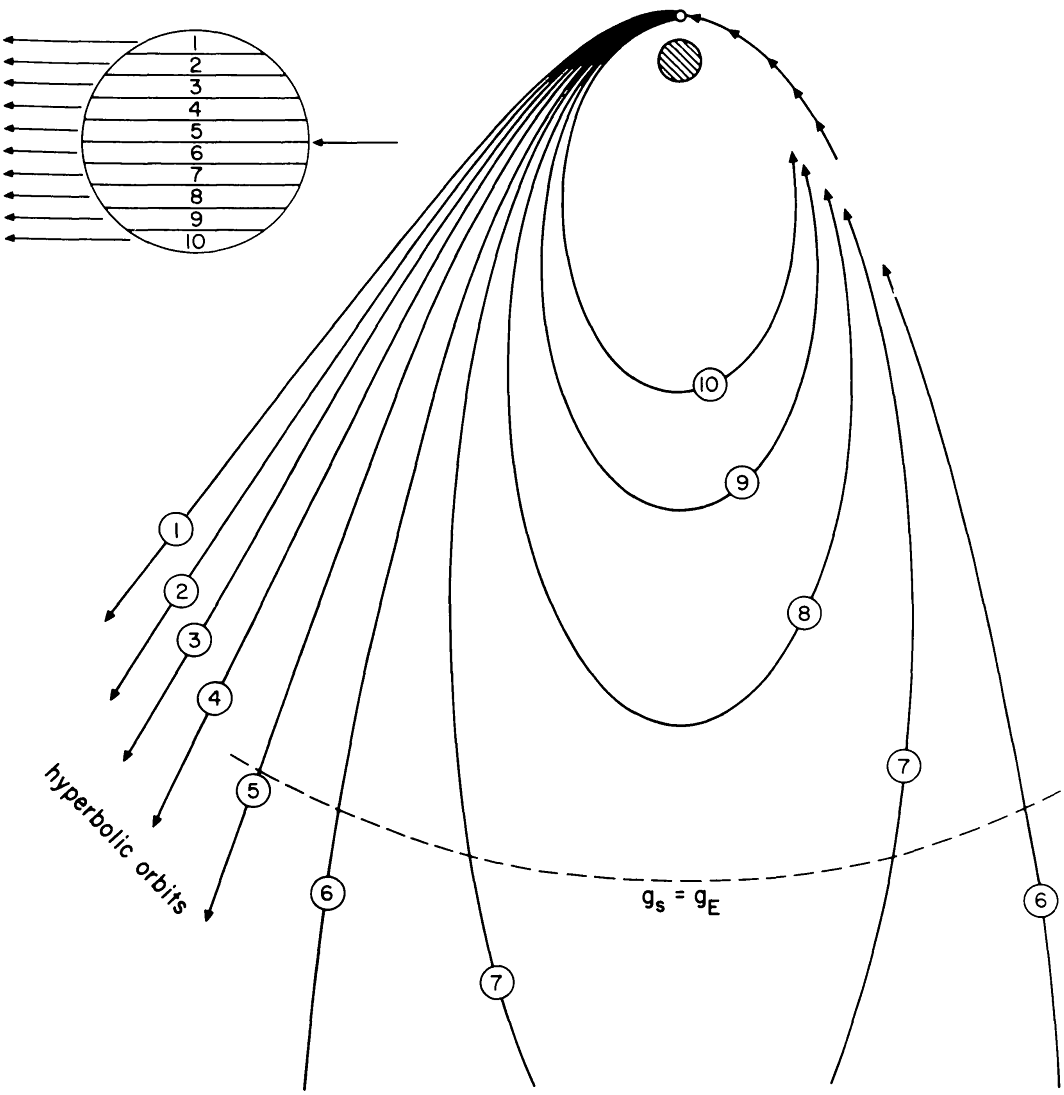


Fig. 11. An idealized example of disintegrative capture. A planetesimal at the top of the figure is tidally disrupted during passage at parabolic velocity through Earth's Roche zone. The subsequent trajectories of material from ten parallel slabs in the original planetesimal (inset, upper left) are followed. Debris from slabs 1–5 (farthest from Earth) is returned to heliocentric orbits; debris from slabs 6–10 is captured in geocentric orbits. In the case of encounters at greater than the parabolic velocity, a smaller number of slabs, those that pass closest to Earth, are retained in geocentric orbits. These would contain a disproportionately large amount of crustal and mantle material if a differentiated planetesimal was disrupted; most of the core material would return to heliocentric orbits. Figure from Wood and Mitler (1974).

a portion of it to lose angular momentum and migrate inward until it joined the Earth at Earth-skimming orbital velocity. This would spin up the Earth, and bring its rotation axis closer to that of the PLS. Whether it is possible for such a process to account for the early Earth's rotation rate is not known.

Advantages of the model are that it has the potential of accounting for the compositional differences between Earth and Moon, the Moon's prograde orbit, and the lunar magma ocean (if disintegrative capture occurred when planetary accretion was nearly completed, and the PLS promptly reconstituted itself as the Moon). The encounter postulated is as plausible as the one that would lead to collisional ejection; perhaps slightly more so, since the cross section for a disintegrative capture interaction (if it works at all) is probably larger than the cross section for collisional ejection.

The most serious liability of disintegrative capture is that it is not clear it would work. The process has been investigated in only a very approximate way. All the authors cited above except Öpik make the simplifying assumption that the passing planetesimal remains an integral sphere until it reaches perigee, whereupon it disrupts and fragments follow the trajectories dictated by their positions in the system at the time of disruption. In fact, the planetesimal would have been markedly distorted as it approached the Roche zone (Öpik, 1972; Kaula and Beachey, 1986), and disruption can have occurred before or after perigee passage, depending upon the circumstances of encounter. The above authors also ignore the continuing gravitational influence of the substance of the planetesimal after its disruption on the debris fragments as they recede from one another.

An apparently fatal objection to the model has been registered by Mizuno and Boss (1985), who numerically simulated the passage of viscous fluid planetesimals through Earth's Roche zone. These authors concluded that disruption would not occur at all, because the effective viscosity of the material of the passing planetesimal is so great and the tidal stresses are applied so briefly (< 2 hours) that the planetesimal would not have time to deform enough to disrupt.

4. A Summing Up

How the Earth's Moon was formed is still not known. Perhaps it will never be. But it became clear at the Kona conference that a major shift of confidence has occurred among lunar scientists toward the collisional ejection model. This did not occur because strong evidence was presented that the Moon was formed by this means, or even that it could have been: It happened because several independent investigators showed that coaccretion, the model that had been most widely accepted by lunar scientists (at least at a subconscious level), could not account for the angular momentum content of the Earth-Moon system. With this alternative seemingly removed, many in the lunar community turned to collisional ejection as the model that appeared most plausible among those remaining.

I have attempted to review the five major models of lunar formation in this paper. My personal editorial comments and judgments have crept in from time to time, as the reader will have noticed. By way of summarizing what I see as the strengths and weaknesses of the five models, I present a "Report Card" (Table 3). The line in parentheses symbolizes my doubt that similarities in trace element abundance patterns between the mantles of Earth and Moon are a reliable constraint. Perhaps the question of lunar origin should be systematically reexamined in another twenty years. An appropriate conference site would be the Moon.

TABLE 3. A "Report Card": The Author's Opinion of How Well Five Models of Lunar Origin Satisfy Seven Constraints.*

	Intact Capture	Coaccretion	Earth Fission	Collisional Ejection	Disintegrative Capture
Lunar mass	B	B	D	I	B
Earth-Moon angular momentum	C	F	F	B	C
Volatile element depletion	C	C	B	B	C
Fe depletion	F	D	A	I	B
Oxygen isotopes	B	A	A	B	B
(Similarity of mantle trace element patterns)	(C)	(D)	(A)	(C)	(C)
Magma ocean	D	C	A	A	B
Physical plausibility	D–	C	F	I	F

**For readers unfamiliar with the U.S. educational system: A is the best grade; F (failing) is the worst; I (incomplete) means all assignments have not been completed and a grade cannot yet be awarded.*

5. *Appendix*

Computer simulation of earth accretion and capture of a protolunar swarm

In 1981, the author developed a computer program to simulate the accretion of the Earth and the accumulation of a protolunar swarm from a population of heliocentrically orbiting planetesimals. The model was based on the premise that the systematic accretion of a large number of relatively small planetesimals was involved, and no one of them affected the outcome greatly.

A characteristic eccentricity for the planetesimals was entered as input, as well as a size distribution and maximum mass for the planetesimals An embryonic Earth some fraction of its present mass was assumed at the outset. A series of planetesimals were then allowed to interact with the Earth until it grew to its present mass. Planetesimals that missed the Earth but passed through its Roche zone were assumed to have been tidally disrupted if they were larger than the critical mass beyond which their material strength was comparable to their gravitational cohesion, and the fraction of their mass and angular momentum that would have been captured in this situation was added to the PLS. A record was kept of the proportions of core and mantle material captured from hypothetically differentiated planetesimals, in the forlorn hope that the disintegrative capture model of lunar formation would be found to work. When material was disintegratively captured into the PLS in a sense opposite to its net rotation, the new mass was assumed to collide with a like amount of PLS material, cancelling the angular momenta of both and adding their mass to the Earth.

The simulated accretion proceeded in a series of cycles consisting of two stages: First, values of e and a were chosen; then a number of planetesimals were released with this e and a, but different values of the angle θ (Fig. 7). The simulation was carried out in two dimensions, but the results were corrected for the effect of planetesimals whose inclinations caused them to approach the Earth-Moon system out of the plane of the ecliptic.

The eccentricity for each set (band) of planetesimals was randomly chosen from a range of values centered on and weighted in favor of the value of characteristic eccentricity that was entered as input. The range of values of mean distance for which an object with this e would cross Earth's orbit (a = 1 AU, e = 0) was reckoned, and a was randomly chosen from this range ±0.05 AU. The extra ±0.05 AU of range was added to include the effect of planetesimal orbits that do not normally intersect Earth's orbit, but which under the gravitational influence of the Earth might be perturbed into interaction with it. For orbits that intersect Earth's without perturbation, a random choice was made of whether the encounter would occur as the planetesimal was crossing into or out of Earth's orbit.

A series of test points were released suitably far away from the Earth in orbits having these values of a and e but different angles θ between their lines of apsides and the position of the Earth at the time of release, and by a process of trial and error the range in θ was found that would result in collision with the Earth or capture into its PLS. A set of planetesimals was then released with the established values of a and e, and with values of θ randomly chosen in this fertile range. The number of planetesimals in the set was proportional to the angular width of the allowed range in θ. The planetesimal size was randomly chosen within the weighted size distribution previously specified. Planetesimal trajectories were integrated in double precision, using an Adams-Bashforth-Moulton predictor/corrector technique.

Several dozen runs were carried out using different input parameters, experimental biasing factors, and increasingly complex program modifications that were intended to make the simulation more realistic. The results were surprising. None of the runs created an Earth and PLS with as much as half the angular momentum of the present Earth-Moon system, and about half of the runs produced Earth-Moon systems with retrograde rotatation.

The most successful run for which data survives had a characteristic eccentricity of 0.04. It accreted the Earth from a 0.25-Earth-mass nucleus to its present mass, producing an Earth and PLS with 1.37×10^{41} rad g cm^2/sec of prograde angular momentum, 40% of the value for the present Earth-Moon system. The mass of material in the PLS was 3.6×10^{25} g, 49% of the present lunar mass. The record of growth of the system is shown in Fig. A1. 1366 planetesimals, having masses 10^{-5}–10^{-3} the current mass of the system, joined the Earth and PLS in 388 sets. The simulation had to be biased to achieve even this much success: All planetesimals passing through the Roche zone, regardless of size, were assumed to be disrupted and partially captured; this maximizes the mass of the PLS and hence the angular momentum of the system.

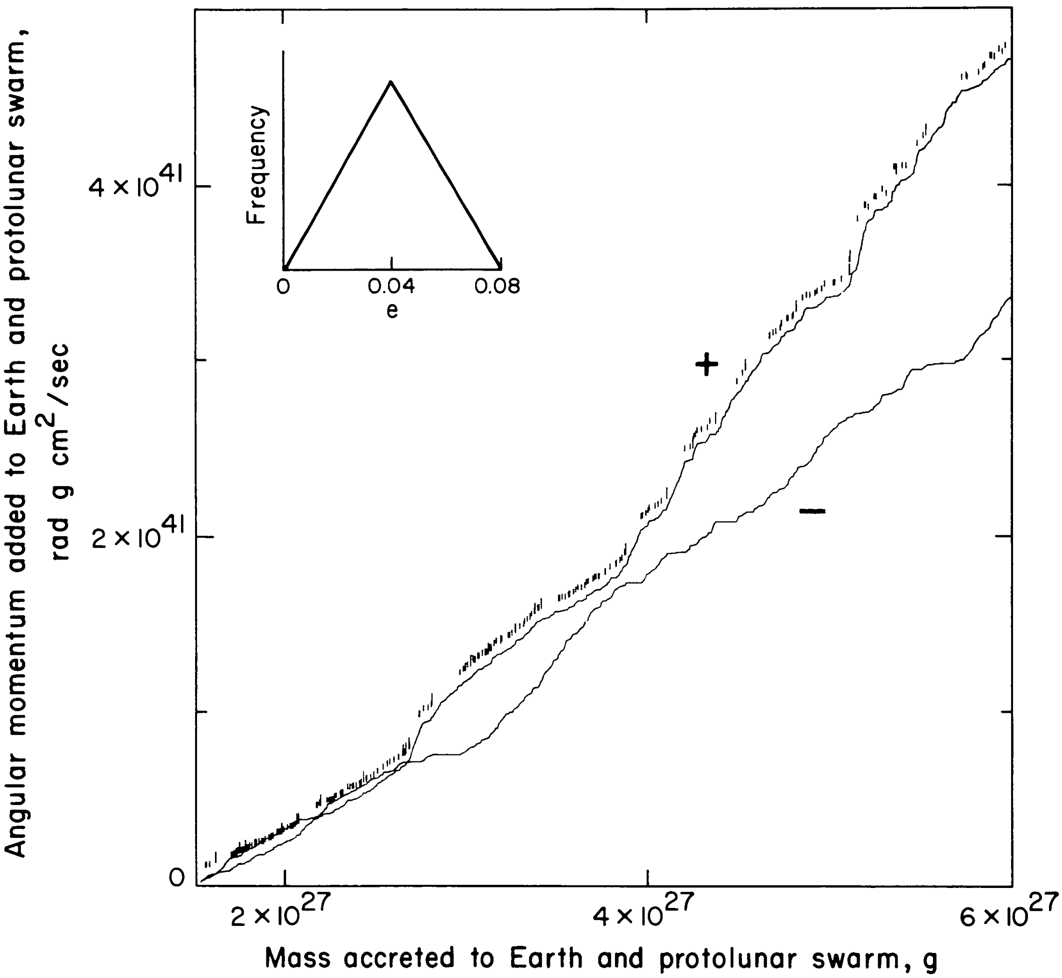


Fig. A1. Record of the accumulation of mass and angular momentum by the Earth and protolunar swarm in the simulation described in the text. Assumed distribution of eccentricities among arriving planetesimals is shown at upper left. Two curves are shown, one for positive and one for negative angular momentum accumulated; net angular momentum of the system is the vertical difference between the curves. Small tick marks divide the mass increments due to the 388 sets (bands) of planetesimals accreted and captured.

Inspection of Fig. A1 suggests that the Earth-Moon system in this run acquired its record amount of angular momentum because the roll of the dice favored it this time, not because some systematic physical effect inexorably converted the planetesimals' orbital angular momentum into rotational angular momentum of the Earth-Moon system. The curves of positive and negative angular momentum do not diverge steadily: For the first 20% of their lengths, slightly more negative than positive angular momentum is being acquired by the system; at 40% of their lengths, the curves almost touch or cross again; beyond this they diverge, but along paths so erratic as to suggest they might rejoin if the system continued to grow.

I abandoned this effort when it began to seem clear that, as Safronov (1966) has argued all along, the rotation (and obliquity) of the Earth must be due to the off-center accretion of a very small number of larger planetesimals, instead of some systematic effect associated with the accretion of a large number of small planetesimals.

Acknowledgments. I am grateful to a number of people for their critical reading of this paper, and the helpful suggestions they made: A. G. W. Cameron, W. Cameron, C. R. Chapman, M. J. Drake, R. J. Greenberg, A. W. Harris, L. L. Hood, W. M. Kaula, J. Longhi, H. E. Newsom, J. A. O'Keefe, G. Ryder, S. F. Singer, and S. R. Taylor.

6. *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.

Anderson D. L. (1972) The origin of the moon. *Nature, 239*, 263–265.

Bills B. G. and Ferrari A. J. (1977) A harmonic analysis of lunar topography. *Icarus, 31*, 244–259.

Binder A. B. (1974) On the origin of the moon by rotational fission. *The Moon, 11*, 53–76.

Binder A. B. (1978) On fission and the devolatilization of a moon of fission origin. *Earth Planet. Sci. Lett., 41*, 381–385.

Binder A. B. (1986) The binary fission origin of the Moon, this volume.

Boss A. P. and Peale S. J. (1986) Dynamical constraints on the origin of the Moon, this volume.

Brush S. G. (1986) The early history of selenogony, this volume.

Burns J. A. (1975) The angular momenta of solar system bodies: implications for asteroid strength. *Icarus, 25*, 545–554.

Burns J. A. (1977) Orbital evolution. In *Planetary Satellites* (J. A. Burns, ed.), pp. 113–156. Univ. of Arizona, Tucson.

Cameron A. G. W. (1972) Orbital eccentricity of Mercury and the origin of the moon. *Nature, 240*, 299–300.

Cameron A. G. W. (1983) Origin of the atmospheres of the terrestrial planets. *Icarus, 56*, 195–201.

Cameron A. G. W. (1985) Formation of the prelunar accretion disk. *Icarus*, in press.

Cameron A. G. W. (1986) The impact theory for origin of the Moon, this volume.

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.

Clark S. P., Jr., Turcotte D. L., and Nordmann J. C. (1975) Accretional capture of the moon. *Nature, 258*, 219–220.

Clayton R. N. and Mayeda T. K. (1975) Genetic relations between the moon and meteorites. *Proc. Lunar Sci. Conf. 6th*, pp. 1761–1769.

Clayton R. N., Onuma N., and Mayeda T. K. (1976) A classification of meteorites based on oxygen isotopes. *Earth Planet. Sci. Lett., 30*, 10–18.

Darwin G. H. (1879) On the precession of a viscous spheroid and on the remote history of the earth. *Phil. Trans. Roy. Soc. Part II, 170*, 447–530.

Deer W. A., Howie R. A., and Zussman J. (1962) *Rock-forming Minerals*, vols. 1 and 2. Longman, London. 333 pp. and 379 pp.

Delano J. W. and Ringwood A. E. (1978a) Indigenous abundances of siderophile elements in the lunar highlands: Implications for the origin of the moon. *Moon and Planets, 18*, 385–425.

Delano J. W. and Ringwood A. E. (1978b) 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.

Drake M. J. (1983) Geochemical constraints on the origin of the moon. *Geochim. Cosmochim. Acta, 47*, 1759–1767.

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

Durisen R. H. and Gingold R. A. (1986) Numerical simulations of fission, this volume.

Durisen R. H. and Scott E. H. (1984) Implications of recent numerical calculations for the fission theory of the origin of the moon. *Icarus, 58*, 153–158.

Giuli R. T. (1968) On the rotation of the earth produced by gravitational accretion of particles. *Icarus, 8*, 301–323.

Greenberg R. (1979) Growth of large, late-stage planetesimals. *Icarus, 39*, 141–150.

Grieve R. A. F. (1980) Cratering in the lunar highlands: Some problems with the process, record and effects. In *Proceedings of the Conference on the Lunar Highlands Crust* (J. J. Papike and R. B. Merrill, eds.), pp. 173–196. Lunar and Planetary Institute, Houston.

Harris A. W. (1977) An analytical theory of planetary rotation rates. *Icarus, 31*, 168–174.

Harris A. W. (1978) Satellite formation, II. *Icarus, 34*, 128–145.

Harris A. W. and Kaula W. M. (1975) A co-accretional model of satellite formation. *Icarus, 24*, 516–524.

Hartmann W. K. (1986) Moon origin: The impact-trigger hypothesis, this volume.

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

Hartmann W. K. and Larson S. M. (1967) Angular momenta of planetary bodies. *Icarus, 7*, 257–260.

Hashimoto A. (1983) Evaporation metamorphism in the early solar nebula—evaporation experiments on the melt FeO-MgO-SiO_2-CaO-Al_2O_3 and chemical fractionations of primitive materials. *Geochem. J., 17*, 111–145.

Heppenheimer T. A. and Porco C. (1977) New contributions to the theory of capture. *Icarus, 30*, 385–401.

Hood L. L. (1986) Geophysical constraints on the lunar interior, this volume.

Horedt Gp. (1976) Pre-capture orbits of the moon. *Moon and Planets, 15*, 439–443.

Jeffreys H. (1930) The resonance theory of the origin of the moon. *Mon. Not. Roy. Astron. Soc., 91*, 169–174.

Kant I. (1755) *Allgemeine Naturgeschichte und Theorie des Himmels.*

Kaula W. M. (1979) Thermal evolution of Earth and Moon growing by planetesimal impacts. *J. Geophys. Res., 84*, 999–1008.

Kaula W. M. and Beachey A. E. (1986) Mechanical models of close approaches and collisions of large protoplanets, this volume.

Kaula W. M. and Harris A. W. (1973) Dynamically plausible hypotheses of lunar origin. *Nature, 245*, 367–369.

Kaula W. M., Schubert G., Lingenfelter R. E., Sjogren W. L., and Wollenhaupt W. R. (1974) Apollo laser altimetry and inferences as to lunar structure. *Proc. Lunar Sci. Conf. 5th*, pp. 3049–3058.

Kopal Z. and Lyttleton R. A. (1963) On the elliptic case of the restricted problem of three bodies and the remote history of the earth-moon system. *Icarus, 1*, 455–458.

Kreutzberger M. E., Drake M. J., and Jones J. H. (1985) Origin of the earth's moon: constraints from alkali volatile trace elements. *Geochim. Cosmochim. Acta*, in press.

Lyttleton R. A. (1972) On the formation of planets from a solar nebula. *Mon. Not. Roy. Astron. Soc., 158*, 463–483.

MacDonald G. J. F. (1966) Origin of the moon: dynamical considerations. In *The Earth-Moon System* (B. G. Marsden and A. G. W. Cameron, eds.), pp. 165–209. Plenum, New York.

Matsui T. and Abe Y. (1986) Origin of the moon and its early thermal evolution, this volume.

Mitler H. E. (1975) Formation of an iron-poor moon by partial capture, or: Yet another exotic theory of lunar origin. *Icarus, 24*, 256–268.

Mizuno H. and Boss A. P. (1985) Tidal disruption of dissipative planetesimals. *Icarus, 63*, 109–133.

Morgan J. W., Hertogen J., and Anders E. (1978) The Moon: composition determined by nebular processes. *Moon and Planets, 18*, 465–478.

Nakazawa K., Komuro T., and Hayashi C. (1983) Origin of the moon—capture by gas drag of the earth's primordial atmosphere. *Moon and Planets, 28*, 311–327.

Newsom H. E. (1984) The lunar core and the origin of the moon. *EOS (Trans. Amer. 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.

O'Keefe J. A. (1966) The origin of the moon and the core of the earth. In *The Earth-Moon System* (B. G. Marsden and A. G. W. Cameron, eds.), pp. 224–233. Plenum, New York.

O'Keefe J. A. (1969) Origin of the moon. *J. Geophys. Res., 74*, 2758–2767.

O'Keefe J. A. (1972) The origin of the moon: theories involving joint formation with the earth. *Astrophys. Space Sci., 16*, 201–211.

O'Keefe J. A. and Sullivan E. C. (1978) Fission origin of the moon: cause and timing. *Icarus, 35*, 272–283.

Öpik E. J. (1972) Comments on lunar origin. *Irish Astron. J., 10*, 190–238.

Papanastassiou D. A. and Wasserburg G. J. (1976) Early lunar differentiates and lunar initial $^{87}Sr/^{86}Sr$ (abstract). In *Lunar Science VII*, pp. 665–667. The Lunar Science Institute, Houston.

Rammensee W. and Wänke H. (1977) On the partition coefficient of tungsten between metal and silicate and its bearing on the origin of the moon. *Proc. Lunar Sci. Conf. 8th*, pp. 399–409.

Ringwood A. E. (1960) Some aspects of the thermal evolution of the earth. *Geochim. Cosmochim. Acta, 20*, 241–259.

Ringwood A. E. (1979) *Origin of the Earth and Moon*. Springer-Verlag, New York. 295 pp.

Ringwood A. E. (1986) Composition and origin of the Moon, this volume.

Ringwood A. E. and Kesson S. E. (1977) Basaltic magmatism and the bulk composition of the moon, II. Siderophile and volatile elements in the moon, earth, and chondrites: implications for lunar origin. *The Moon, 16*, 425–464.

Ringwood A. E. and Seifert S. (1986) Nickel-cobalt abundance systematics and their bearing on lunar origin, this volume.

Ruskol E. L. (1960) The origin of the moon. I. Formation of a swarm of bodies around the earth. *Soviet Astronomy-AJ, 4*, 657–668.

Ruskol E. L. (1963) On the origin of the moon. II. The growth of the moon in the circumterrestrial swarm of satellites. *Soviet Astronomy-AJ, 7*, 221–227.

Ruskol E. L. (1972a) The origin of the moon. III. Some aspects of the dynamics of the circumterrestrial swarm. *Soviet Astronomy-AJ, 15*, 646–654.

Ruskol E. L. (1972b) On the possible differences in the bulk chemical composition of the earth and the moon forming in the circumterrestrial swarm. In *The Moon* (S. K. Runcorn and H. C. Urey, eds.), pp. 426–428. D. Reidel, Dordrecht.

Ruskol E. L. (1973) On the model of the accumulation of the moon compatible with the data on the composition and the age of lunar rocks. *The Moon, 6*, 190–201.

Safronov V. S. (1966) Sizes of the largest bodies falling onto the planets during their formation. *Soviet Astronomy–AJ, 9*, 987–991.

Schmidt O. Yu. (1959) *A Theory of the Origin of the Earth*. Lawrence and Wishart, London. 139 pp.

Singer S. F. (1968) The origin of the moon and geophysical consequences. *Geophys. J. Roy. Astron. Soc., 15*, 205–226.

Singer S. F. (1986) Origin of the Moon by capture, this volume.

Smith J. V. (1974) Origin of the moon by disintegrative capture with chemical differentiation followed by sequential accretion (abstract). In *Lunar Science V*, pp. 718–720. The Lunar Science Institute, Houston.

Smith J. V., Anderson A. T., Newton R. C., Olsen E. J., Wyllie P. J., Crewe A. V., Isaacson M. S., and Johnson D. (1970) Petrologic history of the moon inferred from petrography, mineralogy, and petrogenesis of Apollo 11 rocks. *Proc. Apollo 11 Lunar Sci. Conf.*, pp. 897–925.

Spudis P. D. (1983) The excavation of lunar multi-ringed basins: Additional results for nearside basins (abstract). In *Lunar and Planetary Science XIV*, pp. 735–736. Lunar and Planetary Institute, Houston.

Stevenson D. J. (1984) Lunar formation from impact on the earth: is it possible? (abstract). In *Papers Presented to the Conference on the Origin of the Moon*, p. 60. Lunar and Planetary Institute, Houston.

Stevenson D. J. (1985) Implications of very large impacts for Earth accretion and lunar formation (abstract). In *Lunar and Planetary Science XVI*, pp. 819–820. Lunar and Planetary Institute, Houston.

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.

Szebehely V. and McKenzie R. (1977) Stability of the sun-earth-moon system. *Astron. J., 82*, 303–305.

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

Taylor S. R. (1986) The origin of the Moon: Geochemical considerations, this volume.

Thompson A. C. and Stevenson D. J. (1983) Two-phase gravitational instabilities in thin disks with application to the origin of the moon (abstract). In *Lunar and Planetary Science XIV*, pp. 787–788. Lunar and Planetary Institute, Houston.

Turcotte D. L. and Kellogg L. H. (1986) Implications of isotope data for the origin of the Moon, this volume.

Urey H. C. (1952) *The Planets*. Yale University, New Haven. 245 pp.

Urey H. C. (1966) The capture hypothesis of the origin of the moon. In *The Earth-Moon System* (B. G. Marsden and A. G. W. Cameron, eds.), pp. 210–212. Plenum, New York.

Walker D. (1983) Lunar and terrestrial crust formation. *Proc. Lunar Planet. Sci. Conf. 14th*, in *J. Geophys. Res., 88*, B17–B26.

Ward W. R. and Cameron A. G. W. (1978) Disc evolution within the Roche limit (abstract). In *Lunar and Planetary Science IX*, pp. 1205–1207. Lunar and Planetary Institute, Houston.

Wänke H. and Dreibus G. (1986) Geochemical evidence for formation of the Moon by impact induced fission of the proto-Earth, this volume.

Wänke H., Baddenhausen H., Blum K., Cendales M., Dreibus G., Hofmeister H., Kruse H., Jagoutz E., Palme C., Spettel B., Thacker R., and Vilcsek E. (1977) On the chemistry of lunar samples and achondrites. Primary matter in the lunar highlands: a re-evaluation. *Proc. Lunar Sci. Conf. 8th*, pp. 2191–2213.

Wänke H., Dreibus G., and Palme H. (1978) Primary matter in the lunar highlands: The case of the siderophile elements. *Proc. Lunar Planet. Sci. Conf. 9th*, pp. 83–110.

Ward W. R. and Cameron A. G. W. (1978) Disc evolution within the Roche limit (abstract). In *Lunar and Planetary Science IX*, pp. 1205–1207. Lunar and Planetary Institute, Houston.

Warren P. H. (1985) The magma ocean concept and lunar evolution. *Ann. Rev. Earth Planet. Sci., 13*, 201–240.

Weidenschilling S. J., Chapman C. R., Herbert F., Davis D. R., Drake M. J., Jones J., and Hartmann W. K. (1986) Origin of the Moon from a circumterrestrial disk, this volume.

Wetherill G. W. (1975) Possible slow accretion of the moon and its thermal and petrological consequences (abstract). In *Papers Presented to the Conference on Origins of Mare Basalts and their Implications for Lunar Evolution*, pp. 184–188. The Lunar Science Institute, Houston.

Wetherill G. W. (1976) The role of large bodies in the formation of the earth. *Proc. Lunar Sci. Conf. 7th*, pp. 3245–3257.

Wetherill G. W. (1986) Accumulation of the terrestrial planets and implications concerning lunar origin, this volume.

Winters R. R. and Malcuit R. J. (1977) The lunar capture hypothesis revisited. *Moon and Planets, 17*, 353–358.

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.

Wise D. U. (1969) Origin of the moon from the earth: Some new mechanisms and comparisons. *J. Geophys. Res., 74*, 6034–6045.

Wood J. A. and Mitler H. E. (1974) Origin of the moon by a modified capture mechanism, or: Half a loaf is better than a whole one (abstract). In *Lunar Science V*, pp. 851–853. The Lunar Science Institute, Houston.

Wood J. A., Dickey J. S., Marvin U. B., and Powell B. N. (1970) Lunar anorthosites and a geophysical model of the moon. *Proc. Apollo 11 Lunar Sci. Conf.*, pp. 965–988.