

# LUNAR EVOLUTION: HOW WELL DO WE KNOW IT NOW?\*

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Of course I was not there when the solar system originated and I do not know how it originated. I am only a student of the subject and I modify my ideas as new evidence appears or as new ideas occur to me.

Harold Urey

**Abstract.** The currently known astronomical, chemical and magnetic data are not uniquely indicative of an extensively and globally molten Moon. We argue here for an accretional layering in the Moon, but at temperatures below solidus. The excess mass in the near side of the Moon compatible with a 2 km displacement in the center of mass relative to the centre of figure and the moment of inertia data is considered to be due to Fe-FeS liquid formation and inhomogeneous segregation. These Fe-FeS bodies, termed 'fescons', are shown to be capable of accounting for the presently available magnetization data, by acting as small regenerative dynamos with a time-stability less than that of the terrestrial equivalent. The chemical characteristics of the highly differentiated materials (KREEP, 'granite' etc.) are considered to be due to small scale localized melting caused by collisional events, from sources in which accessory phases play a significant role. Mare basalts are considered to be melts in the overlying material produced at a later time by  $^{40}\text{K}$  radioactivity in the fescons. Some consequences of the present hypothesis are suggested.

We conclude that these and other characteristics of the lunar materials are reconcilable with a 'cold' Moon such as discussed by Urey over the past two decades.

## 1. Introduction

A fundamental limitation to our understanding of the Moon is aptly expressed in the above quotation by one whose contributions to the field are legion. The wealth of data obtained on the lunar samples from the Apollo missions has provided both new evidence and new ideas. A prominent feature of many of these seems to call for a Moon molten soon upon formation, either totally or in its outer several hundred kilometer shell.

We are amazed at how firmly this conclusion is embedded into Apollo lunar science, which after all, is in its state of infancy. Note that after several decades of dedicated observations and extensive sampling, this point has not yet been settled for the Earth; hence our title to this paper, which is really a question. The purpose of this paper is to examine the known astronomical, geochemical and geophysical data and the constraints provided by them in that context. Some of these data have been known for the past several decades; the others are a direct result of the study of lunar samples.

\* Paper dedicated to Professor Harold C. Urey on the occasion of his 80th birthday on 29 April, 1973.

## 2. Physical and Astronomical Considerations

It was long known that the three moments of inertia about the principal axes of the Moon are such that the lunar globe is definitely not in hydrostatic equilibrium.

Denoting

$$\alpha = \frac{C - B}{A}, \quad \beta = \frac{C - A}{B} \quad \text{and} \quad \gamma = \frac{B - A}{C},$$

where  $A$  and  $C$  are the moments of inertia about the Earth–Moon axis and the axis of rotation respectively and  $B$  is moment about an axis orthogonal to the other two, a recent evaluation of these parameters gives  $\alpha=0.000397$ ;  $\beta=0.000628$ ; and  $\gamma=0.000230$  (Kopal, 1972).

For hydrostatic equilibrium at any distance from the Earth, the ratio  $\alpha/\beta$  should be 0.2523 whereas the observed value is  $0.633 \pm 0.011$  (Kopal, 1972). The idea that this disparity is due to a tidal bulge on the Earth-facing side of the Moon (Jeffreys, 1924 and several workers following him) is negated by recent laser altimetry studies of the shape of the Moon by Wollenhaupt and Sjogren (1973) and Kaula *et al.* (1972). These works showed that there is no such bulge and showed further that the center of mass is displaced toward the Earth-facing side from the center of volume by 2 km.

These observations clearly point out two important aspects of the Moon: that there is an excess mass due to high density material at some depth within the Earth-facing side and more importantly, as shown by Kopal (1972), that the Moon cannot have been extensively melted at any time and at whatever distance from the Earth.

Considering the first point, some general statements are possible with regard to the maximum depth at which the excess mass must be located. An outer 200 km shell of a moon of uniform density carries about 30% of the lunar mass and about 45% of the total moment of inertia. Thus, in order to account for  $\alpha/\beta=0.633$ , an appropriate excess mass must effectively reside in the outer 200 km shell on the near side rather than deep in the interior of the Moon, because the deep interior is expected to be in hydrostatic equilibrium resulting from lithostatic pressures greater than 10 kbar. Convincing arguments along these lines have been presented by Kopal (1972).

The discovery of large positive gravitational anomalies over the ringed maria on the Moon's Earth-facing side (Muller and Sjogren, 1968) and the analyses of the Moon's gravitational field based on these data (e.g., Michael and Blackshear, 1972) have an important bearing on this point. These data show that in these areas of the Moon, regardless of the origin and the depth of location, there are in effect excess concentrations of mass (mascons). If the mascons are near-surface phenomena, Booker *et al.* (1970) showed that *apparent surface mascons* approximately proportional to the area of the ringed maria and of the order of  $\sim 10^{-5}$  lunar mass each are consistent with the satellite acceleration data and the ratio of the moments of inertia,  $(B - A)/(C - A)$ . Several workers had suggested earlier that mascons are thin surficial plates (e.g., Conel and Holstrom, 1968; Wood *et al.*, 1970) formed by upwelling of dense mare lava flows into depressions in a lighter crust.

Although apparent surface mascons in the ringed maria totalling about  $10^{-4}$  lunar mass can explain the observed gravity field as well as the relationship of mascon size to mare diameter and the proper moment of inertia ratio of  $(B-A)/(C-A)$  (Booker *et al.*, 1970), there remain some difficulties. If the observed gravity anomalies are essentially due to this mechanism, the gravity anomalies over Maria Orientale and Smythii pose some problems. Both of these are only partly filled with mare material, and yet show mascons proportional to their diameter. This suggests that surface thin plate mascons produced by mare filling can only be a *partial answer* to the observed moment of inertia ratios and the gravity anomalies.

The other difficulty with thin plate surface mascons of a total mass of  $\sim 10^{-4}$  lunar mass is that they alone are inadequate to satisfy the absolute values of  $C/Ma^2$ , the absolute values of  $B-A$  and  $C-A$ , or the center of mass displacement. The current best estimate of  $C/Ma^2$  due to Michael and Blackshear (1972) is  $0.4019_{-0.002}^{+0.004}$ . These authors show that for a 2 km difference between the center of figure and center of mass of the lunar globe, the correct value of  $C/Ma^2$  can be satisfied by an Earth-side slab 10–50 km thick, having a range in density from 4.29 to 3.53 g cm $^{-3}$  corresponding to the range in thickness. Material of such densities in terms of any known rock types is definitely ruled out in the outer 80 km of the near-side lunar shell as shown by seismic data (Toksöz *et al.*, 1972). Thus, we are constrained to believe that a proper match to the best value of  $C/Ma^2$  can only be obtained by the presence of some high density material in the Earth-facing side at depths possibly greater than 100 km. Burial at such depths would require a mass higher than  $10^{-4}$  lunar mass to account for the center of mass displacement as well as the observed moments of inertia.

This point of view is also supported by the fact that while surface mascons of about  $10^{-4}$  lunar mass can give the correct ratio for  $(B-A)/(C-A)$  they do not satisfy the absolute values of  $B-A$  and  $C-A$  obtained by satellite gravity data (Booker *et al.*, 1970). In fact, this mass is inadequate by an order of magnitude. Thus, we can deduce that a total mass excess consistent with both the magnitudes and the ratios of the moments of inertia is of the order  $10^{-3}$  lunar mass and further, that in order to satisfy the observed peak accelerations of the lunar satellites and the center of mass displacement, it must be buried at a depth of about 200 km on the Earth-facing side of the Moon (Booker *et al.*, 1970; Chase and Murthy, 1972 in preparation).

The straightforward conclusion that emerges from these physical and astronomical data is that the lunar globe was never extensively melted and contains an excess mass of about  $10^{-3}$  lunar mass at about 200 km below the surface on the near side of the Moon. Any model of geochemical differentiation or thermal history of the Moon should satisfy these constraints.

### 3. Geochemical and Thermal History Considerations

Petrological and geochemical studies on lunar samples have shown that the lunar surface contains material that is extremely differentiated with respect to the interior. The simplest explanation would be that on the Moon large scale melting (outer 200 km

to total Moon) and consequent igneous processes have given rise to these differentiated materials. Important additional characteristics of this differentiation have been deduced from the detailed considerations of isotopic systematics, trace-element abundances and petrological aspects of the lunar samples (e.g., Papanastassiou and Wasserburg, 1971; Wood, 1970; Gast, 1972 and many others). On the basis of these data, models of a partially or completely molten early Moon have come into vogue.

If we neglect interpretive conclusions, a set of definitive statements can be made regarding the lunar surficial materials.

(1) There is evidence for material on the surface of the Moon showing extremely high concentrations of radioactive elements and other lithophile elements. These materials are sampled as small igneous rocks and as ubiquitous but variable component in all lunar soils. Further, these materials formed at  $\approx 4.6$  b.y. ago. (Papanastassiou and Wasserburg, 1971; Hubbard *et al.*, 1971).

(2) The source regions of mare basalts are extremely low Rb/Sr reservoirs, also established at about 4.6 b.y. ago. From the Rb/Sr isotopic systematics, it can also be deduced that inhomogeneities exist with respect to this ratio (e.g., Wasserburg and Papanastassiou, 1971; Compston *et al.*, 1971).

(3) Basaltic lavas flooded and solidified in the maria between 3.8–3.1 b.y. ago. There is evidence that these lavas were produced by internal heat production in the Moon. At present, there is no evidence that such processes occurred later than this time span (Papanastassiou and Wasserburg, 1971; Turner, 1971; Compston *et al.*, 1971; Murthy *et al.*, 1971).

(4) Neither the mare basalts, nor the highly radioactive materials sampled so far can represent the average lunar composition because of constraints imposed by either the mean density of the Moon or thermal considerations of the present Moon, respectively (Wetherill, 1968; Papanastassiou and Wasserburg, 1971).

The first two of the above statements have been attributed to selenochemical differentiation processes in at least the outer regions (200–400 km) of the Moon molten at the time of accretion about 4.6 b.y. ago. Materials enriched in trace elements and radioactivities segregate to the surface leaving impoverished material in the deeper regions. These depleted materials at depth were the source material from which mare basalts were generated later at 3.8–3.1 b.y. ago. Thermal considerations such as these have led to the description of two 'hot' and 'cold' cycles in the early thermal regime of the Moon (e.g., Hinners, 1971). The terms 'hot' and 'cold' can be taken to indicate liquidus and solidus temperatures, respectively.

Recent thermal history models of the Moon (Mizutani *et al.*, 1972; Toksöz *et al.*, 1972; Hanks and Anderson, 1972) have employed the above as 'constraints'. Another constraint used in thermal models is the  $33 \text{ ergs cm}^{-2} \text{ s}^{-1}$  heat flow observed by Langseth *et al.* (1972). This value of the measured heat flow at one site is often applied globally to the entire Moon, in spite of the cautionary qualifications stated by the original investigators. Implicit in all these thermal models are two pivotal assumptions: one, that selenochemical differentiation at 4.6 b.y. ago is due to extensive melting of

the Moon; and two, that from the trace element fractionations observed in returned lunar sample we know how much of the Moon was molten.

The presence of mascons, which indicates high rigidity with viscosities of the order of  $10^{25}$ – $10^{27}$  P (Baldwin, 1971), the low lunar seismicity (Latham *et al.*, 1971), the estimates of the lunar interior temperatures from electrical conductivity (Dyal and Parkin, 1972), and the non-equilibrium figure of the Moon, all tend to indicate that the interior of the Moon is cold ( $<1000^{\circ}\text{C}$ ) at present time. Coupled with the observations in Section 2 which indicate that it is very difficult to explain the observed physical characteristics of the moment of inertia and shape of the Moon with an entirely or partly molten Moon (Kopal, 1972) we are bluntly faced with the dilemma posed by the interpretation of the geochemical data for a molten Moon. The question, therefore, remains: Is a partly or totally molten Moon at 4.6 b.y. ago essential to explain the geochemical requirements mentioned above?

At these early stages of Apollo-mission based lunar science, we would like to suggest that these thermal constraints, particularly those that require a totally or partially molten Moon at 4.6 b.y. ago, may have been somewhat overemphasized. For example, the magnitude of enrichments of the trace elements in surficial materials (such as granites, KREEP, Luny rock, magic component of the soils, etc.) at 4.6 b.y. ago tells nothing about whether or not a surficial shell of the Moon was molten at 4.6 b.y. ago and by how much. All such estimates are model dependent and strongly influenced by terrestrial analogies to partial melting systematics and the assumed compositions of the source material (chondritic, achondritic, accretional gradients in composition, etc.) in the outer regions of the Moon. This is also true with regard to the mare basalts generated between 3.8–3.1 b.y. ago, in that no indication of how large a zone in the Moon was melted and differentiated can be specifically given to us in a non-model dependent fashion.

A further difficulty arises from the fact that we do not at present unequivocally know the extent of these highly differentiated materials on the lunar surface (see, for example, Anderson and Smith, 1971). Present data show that such materials are extremely common at the Imbrium site and are found as components in the soils from the various Apollo sites. Size fraction studies of the Apollo 15 soils (Murthy *et al.*, 1972), as well as the fact that radioactive and trace element enriched components progressively decreases with distance from the Imbrium site with little or none present in the Luna 16 soils, suggests that there is a distinct possibility that these highly differentiated materials are probably quite 'local' on the Moon, but are easily available for transport over hundreds of kilometers in cratering and other lunar surface erosional processes. As discussed later, if these highly differentiated materials are sporadic on the lunar surface, there may be no necessity to have a globally molten lunar layer at 4.6 b.y. ago.

Another powerful incentive for a molten Moon stems from the discovery of anorthositic material in Apollo 11 soils (Wood *et al.*, 1970) and subsequent findings of anorthositic material from all Apollo missions. The provenance of this material is considered to be the lunar highlands. Thus, the anorthositic 'crust' of the lunar highlands

has been postulated and has been attributed to separation of plagioclase crystals within a molten Moon (Wood *et al.*, 1970; Wood, 1970; Smith *et al.*, 1970).

It seems certain that the lunar highland regions are composed of rocks primarily dominated by plagioclase in their mineralogy as shown by the Al/Si ratios of lunar highland regions (Adler *et al.*, 1972) and their higher albedo. But the prime evidence for the postulated origin of the highland 'crust' by crystal fractionation from a melt comes from the europium anomalies observed in lunar materials. The data show that anorthositic fragments from lunar soils and breccias show positive Eu anomalies in contrast to the negative Eu anomalies observed in lunar basalts and the KREEP fragments (e.g., Laul *et al.*, 1972; Hubbard *et al.*, 1972).

Let us consider here only the positive Eu anomalies which have lent strong support to the concept of an anorthositic lunar 'crust'. The observed range of rare-Earth elements (REE) for anorthositic fragments from various Apollo missions normalized to chondritic abundances are shown in Figure 1, from the compilation by Laul *et al.* (1972). The peak at Eu shows the high enrichment of this element relative to adjacent REE. Extensive geological experience and laboratory studies show that plagioclase effectively enriches  $\text{Eu}^{+2}$  from a silicate melt with which it is in equilibrium. Therefore, it appears reasonable or even certain that crystallization from a melt must be invoked for the observed positive anomalies in lunar anorthositic samples.

But is a molten lunar surface shell the only possible locale for this process? It is instructive to compare, in this context, the REE patterns exhibited by achondritic meteorites whose composition closely resembles these materials from the moon (Duke and Silver, 1967). The achondritic meteorites which are products of melting and differentiation of solid matter in the solar system in locales other than the Moon show Eu anomalies of both positive and negative nature (Philpotts and Schnetzler, 1970; Boynton and Schmitt, 1972, personal communication). Clearly, these data suggest that solar system fractionations can induce Eu anomalies of various types in solid matter that may be available for planetary accretion. Note that these data are for bulk achondrites composed of various mineral phases in addition to plagioclase. Taking Moore County and Sierra de Mage as examples, the positive Eu anomalies are not as striking as in the lunar anorthosites. In order to make meaningful comparisons, however, we should investigate the REE abundances in plagioclase-rich fractions of achondritic meteorites comparable to lunar anorthosites (>90% plagioclase).

So far only one plagioclase separate from the achondrite Juvinas has been analyzed for its REE content (Philpotts and Schnetzler, 1968). The chondrite normalized pattern for this sample is shown in Figure 1, and falls in the range observed for lunar anorthositic fragments of comparable modal plagioclase content. The absolute enrichments relative to chondritic material and the  $\text{Eu}^{+2}$  anomaly in this achondritic plagioclase are virtually identical to Apollo 11 anorthosite, 10085 and are similar to those in Apollo 14 sample, 14161. Certainly this one sample falls within the range of absolute enrichments and Eu anomalies observed in lunar samples.

Several workers, notably Gast and his coworkers (e.g., Gast, 1972; Gast *et al.*, 1970) argued that the outer regions of the Moon may represent material enriched in Ca, Al,

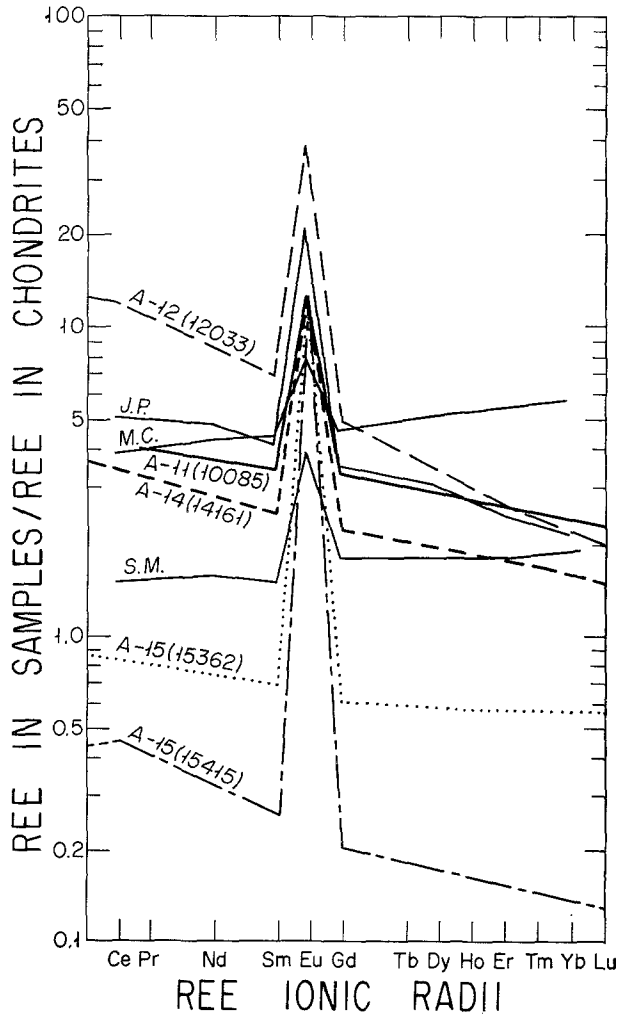


Fig. 1. Comparative rare Earth abundance patterns in lunar anorthosites, some Ca-rich achondrites and plagioclase rich fraction of achondrites. MC - Moore County; JP - Juvinas Plagioclase; SM - Sierra de Mage. Achondritic data from Philpotts and Schneltzer (1968; 1970). Lunar data from the compilation of Laul *et al.* (1972).

REE, U and Th and other trace elements by processes that occurred before the accretion of the Moon. We merely wish to note that if this is the case, the Eu anomalies could just as well represent such solar system fractionation processes. There appears no need to invoke the origin of highland 'crust' by plagioclase flotation from a largely molten Moon at 4.6 b.y. ago.

We suggest here that a permissible model can be explored in which the surface regions of the Moon have undergone major geochemical differentiation without attendant major melting, and that a single and local thermal cycle culminating in the

production of mare basalts may be sufficient to explain the geochemical characteristics. Our obvious aim in this is to satisfy the physical and astronomical constraints discussed in Section 2. Before we do that however, we wish to discuss the data on the magnetism of the Moon, which has also been widely used to postulate a largely molten Moon.

#### 4. Remanent Magnetization and Electrical Conductivity Considerations

The hypothesis of a molten Moon postulated from geochemical and petrological arguments has gained some support from the interpretations of the magnetic characteristics of returned lunar material. From these studies it has been suggested that an intrinsic stable dipole field prevailed in the Moon at some early time in its history and, from terrestrial analogy, the inference is drawn that a lunar convecting core had at one time acted as the dynamo (Runcorn *et al.*, 1970, 1971). This implies that the Moon had once been sufficiently molten to segregate a core.

Once again, let us consider the data and statements that can be made from the studies of the lunar magnetic field(s). These data come from three sources: The Explorer 35 and the Apollo 15 subsatellite observations, the surface magnetometer data obtained in Apollo missions, and the natural remanent magnetization (NRM) observed in returned lunar samples. The subsatellite data indicate some global aspects of the lunar magnetic fields while the surface magnetometer and the NRM data on lunar samples provide local details from which possible interpretations of the past lunar magnetic field(s) can be made.

A downward continuation of the observed data from the Explorer 35 magnetometer indicated a lunar surface field of  $\leq 2\gamma$  ( $1\gamma = 10^{-5}$  emu) leading one to conclude that if it is due to a central dipole field in the Moon, such a dipole should carry a magnetic moment of  $\leq 10^{20}$  emu (less than  $10^{-5}$  of the Earth's dipole moment). This was a conclusion fairly easy to accept in view of the previously mentioned physical and astronomical data.

The Apollo 15 subsatellite magnetometer has mapped a large number of magnetic anomalies of the order of  $1\gamma$  from a height of 10 km. This experiment convincingly showed that the field on the lunar surface is not due to some local and transitory mechanisms like a solar flare or a shock compression of the general interplanetary field which conveniently took place while the magnetic carriers, chiefly iron, were cooling through  $770^\circ\text{C}$ , their Curie point, at one or two places on the lunar surface. The anomalies observed by the Apollo 15 subsatellite fluxgate magnetometer indicate more frequent and extensive sources. For example, a typical anomaly could be due to a body with 10 km diam at the surface with a high value of NRM like  $10^{-2}$  emu  $\text{cm}^{-3}$  or a 100 km body with a more reasonable NRM value of  $10^{-5}$  emu  $\text{cm}^{-3}$ . In either case, large size lunar blocks with uniform magnetization are called for in these models, suggesting that the magnetizing field is due to intrinsic sources.

A further point of importance (Coleman *et al.*, 1972; Dyal and Parkin, 1972) is that the observed anomaly data indicate irregular vector directions, an observation some-



what at odds with a single central dynamo. Pearce *et al.* (1972) and Runcorn *et al.* (1971) have attributed these localized anomalies to edge effects near craters in an otherwise lunar-wide uniformly magnetized basaltic layer. Purely to serve a heuristic function, we submit that the irregular vectorial directions and the localized variability of the fields could equally well be primarily due to heterogeneous and multiple sources of fields in the lunar interior. We return to this point at a later time.

The surface magnetometers placed on the Moon during Apollo missions 12, 14, 15 and 16 have recorded local magnetic fields ranging from 6–300 $\gamma$ . Associated lack of steep gradients and the large variability in anomaly amplitudes at stations apart by about one kilometer lead to the simplest conclusion that these fields were produced by the NRM of magnetized surface layers of considerable extent. If the observed fields were due to a central dipole, it would have to carry a magnetic moment as large as  $10^{22}$  emu, which is not consistent with the Explorer 35 data.

The actual measurements of the NRM in returned lunar samples have been reported to be between  $10^{-3}$ – $10^{-6}$  emu g $^{-1}$ . From alternating field (AF) demagnetization studies which showed an order of magnitude decrease in the original NRM, several workers (Pearce and Strangway, 1972) have deduced that a secondary isothermal remanent magnetization (IRM) has been acquired by the samples during transitory exposure to weak (1–10 Oe) DC fields in the spacecraft. An intermediate stability (say, to 150 Oe or 300°C) component has been found and attributed to shock-induced piezo-remanent magnetization, an IRM due to sporadic and intense ( $\geq 10\gamma$ ) fields of solar activity, or a remanence acquired by magnetic chromites due to thermal cycling on the lunar surface (Banerjee, 1972). This intermediate stability component could well be due to a primary thermoremanent magnetization (TRM) carried by small iron particles with remanence-blocking temperatures of  $\sim 300^\circ\text{C}$ . Of main concern for us here will be the magnitude of the stable component of NRM in lunar rocks, with the qualification that part of this could be a chemical remanence produced by the slow grain growth over millions of years producing an apparent ‘younger’ NRM when compared to the absolute age of the rocks.

The stable fraction of NRM observed in lunar rocks is given in Table I, clearly showing that primary magnetizing fields and a primary magnetizing mechanism(s) existed. We note that the stable magnetization observed in these rocks is consistent with fields observed by the surface magnetometers. However, what has not been emphasized to date is the extreme variability in the size of the stable NRM in lunar rocks. This variability is not related to the iron content (Nagata *et al.*, 1972) which varies from 0.7 to 0.1 wt percent only. Also from Table I, it is apparent that there is no correlation with the absolute age. This specific point of information we consider to be particularly significant.

Runcorn *et al.* (1971) and following him, several others (e.g., Pearce *et al.*, 1972) have interpreted the stable NRM of lunar rocks as due to a TRM in an intrinsic lunar field of a central dynamo in a lunar core. We would expect however, that the postulated birth and death of such a dynamo requires that the dipole field it produces show a regular and continuous variation of intensity with time over a period of the order of

TABLE I  
Size of stable NRM of lunar rocks

Age (b. y.)	Lunar sample No.	Stable magnetization (emu/gm) $\times 10^{-6}$	Reference <sup>a</sup>
<i>A. Igneous rocks</i>			
Apollo 12	3.1-3.3	12002-78	1 [16]
		12002-84	< 1 [16]
		12002-86	2-3 [16]
		12017-4	1 [16]
		12017-5	2 [16]
		12021-66	3-5 [16]
		12021-105	3 [16]
		12038-29	3 [11]
		12038-32	4 [11]
		12038-75	2 [16]
		12051-17	1 [16]
		12051-24	~ 1 [16]
		12053	0.3 [4]
		12053-47	0.4 [11]
		12063	0.2 [4]
		12063-57	1 [16]
		12063-90	2 [16]
12063-98	0.3 [16]		
12065	0.8 [4]		
Apollo 15	3.3	15058-52	1 [17]
		15065-26	0.6 [6]
		15495-46	< 0.3 [17]
		15535-28	1 [17]
		15555-108	0.5 [6]
		15597-28	40 [6]
		15662-2	4.5 [6]
Apollo 11	3.7-3.9	10017-64	< 1 [15]
		10022	2.5 [7]
		10024-22	6 [9]
		10024-22	2 [10]
		10047	5 [4]
		10058	4.5 [4]
		10062	< 3 [4]
		10069	25 [7]
		10085,16-2	10 [7]
		10085,16-3	20 [7]
		10085,16-4	< 9 [7]
Apollo 14	3.9	14310-57	1 [5]
		14310-136	1.6 [17]
<i>B. Breccias</i>			
		10021-32	< 500 [10]
		10048-55	500 [11]
		10085-13	150 [9]
		12073-24	20 [16]
		14047-47	1 [12]
		14063-44	< 0.7 [5]
		14063-47	13 [12]
		14303-28	2 [5]
		14321-194A1	0.4 [5]
		14321-194A2	< 1.5 [5]
		15405-19	< 4 [17]

<sup>a</sup> For references see Table II.

TABLE II  
Paleointensities from returned samples

Age (b.y.)	Lunar sample No.	Paleo-intensity (1000 $\gamma$ )	Comments	Ref.	
<i>A. Igneous rocks</i>					
Apollo 12	3.1-3.3	12002	4.5	Thellier's Method; $T \leq 150^\circ\text{C}$	[8]
		12020-23	10	Comparison of PTRM's with total NRM	[14]
		12022	4.8	Thellier's Method; 8 independent determinations giving 2800-10060 $\gamma$ range; $T < 300^\circ\text{C}$	[8]
		12052-32	31-32	Thellier's method; $T = 195^\circ\text{C}$ ; Calculated from published data by present authors.	[2]
		12063-55, -98	< 4.8	Comparison of total NRM with total TRM	[13]
		12063-55, -98	0.4	Comparison of stable component of NRM (Chip,98) obtained by AF demagnetizing to a peak of 50 Oe with TRM (chip, 55) similarly demagnetized. Calculated from published data by present authors.	[13]
		12065-67	8-13	Thellier's method; $T = 195^\circ\text{C}$ ; Calculated from published data by present authors.	[2]
Apollo 15	3.3	15058-52	72-75	Thellier's method: (a) For $T = 75^\circ\text{C}$ using AF-stable Component. (b) For $150^\circ\text{C} \leq T \leq 350^\circ\text{C}$	[17]
Apollo 11	3.7-3.9	10022	1.57-1.88	Thellier's method; $150^\circ\text{C} \leq T \leq 350^\circ\text{C}$	[7]
		10024-22(A)	6	Comparison of total NRM with total TRM	[9]
		10024-22(B)	3	Comparison of total NRM with total TRM	[10]
		10069	1-170	Thellier's method; $T < 470^\circ\text{C}$	[7]
Apollo 14	3.9	14053-48	'Several tenths of an oersted'	Comparison of total NRM with total TRM	[1]
		14310-136	10-20	Thellier's method using AF-stable component only; $T \leq 225^\circ\text{C}$ .	[7]
<i>B. Breccias</i>					
		10048-55	1.2	Comparison of stable NRM with total TRM.	[11]
		14321-194	3.3-5.1	Thellier's method; $T \leq 160^\circ\text{C}$ .	[3]

For references see p. 170.

1 b.y. as seen, for example, for the Earth over the last 0.5 b.y. (Stacey, 1969). The irregular variation with age of the stable NRM of lunar rocks does not seem to reflect this.

It may be argued that a more appropriate technique for studying the time variation in the strength of a dipole source (core dynamo) would be by the conventional paleointensity determination methods on lunar rocks of different ages. As can be seen in Table II, the paleointensity values for the crystalline rocks range from  $700\gamma$  to  $25000\gamma$ , with little consistency among rocks of similar ages, even when we consider the influence of variable opaque mineralogy on paleointensity measurements as shown by Hoffman and Banerjee (1972).

In view of the above arguments on the lunar magnetic field history, we are led to conclude, that a single central lunar core dynamo is somewhat unsatisfactory. Other and potentially more serious difficulties with a core dynamo have been pointed out recently by Levy (1972). A case can be made on arguments of equal or better plausibility for a number of small field sources (dynamos) not too far from the surface which produced the observed irregular site-variation of NRM and equally irregular time variation of the paleointensities. The point we wish to stress, of course, is that if there is no compelling reason for a lunar core dynamo, there remains no compelling argument from the magnetization data for a molten moon with a core (e.g., Runcorn *et al.*, 1971).

Studies of electrical conductivity data deduced from induction response of the Moon to solar wind magnetic transients, both from the Explorer 35 and Apollo 12 surface magnetometer data have been used to infer the interior structure as well as the temperature of the Moon (Sonett *et al.*, 1971; Dyal and Parkin, 1972). Using the amplification ratio (of induced over inducing field strength) variation with frequency, Sonnett *et al.* (1971) postulated a three-layered structure for the Moon. A special feature of this model was a thin high conductivity layer about 250 km deep which has been suggested

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References for Tables I and II:

- [1] Dunn and Fuller (1972)
- [2] Gromme and Doell (1971)
- [3] Hargraves and Dorety (1972a)
- [4] Hargraves and Dorety (1971)
- [5] Hargraves and Dorety (1972b)
- [6] Hargraves and Dorety (1972c)
- [7] Helsley (1970)
- [8] Helsley (1971)
- [9] Nagata *et al.* (1970)
- [10] Nagata and Carleton (1970)
- [11] Nagata *et al.* (1971)
- [12] Nagata *et al.* (1972)
- [13] Pearce *et al.* (1971)
- [14] Runcorn *et al.* (1971)
- [15] Runcorn *et al.* (1970)
- [16] Strangway *et al.* (1971)
- [17] Banerjee and Hoffman (unpublished data, 1972)

to be an Fe metal or Fe-FeS layer (Urey *et al.*, 1971; Murthy *et al.*, 1971). However, the existence of this global layer of high electrical conductivity has been questioned (Kuckes, 1971).

We note that none of the lunar structure models deduced from electrical conductivity considerations show evidence for a small lunar iron core of  $\lesssim 0.2 R_1$  ( $R_1$  = lunar radius), such as that envisaged by Runcorn *et al.* (1971). Both the models of Sonett *et al.* (1971) and Dyal and Parkin (1972) show large cores of  $\gtrsim 0.6 R_1$  which cannot be iron, as constrained by the conductivity and by the moment of inertia of the Moon.

Another lunar interior property deduced from these considerations is the temperature distribution. The critical parameters here are the composition of material in the lunar interior and its conductivity variation as a function of temperature (Schwerer *et al.*, 1972). For various assumed compositions and temperature versus conductivity characteristics, lunar central temperatures in the vicinity of  $\sim 1250$  K have been calculated (Sonett *et al.*, 1971; Dyal and Parkin, 1972). We note the objections expressed toward these estimates by thermal history investigators (e.g., Hays, 1971; Toksöz *et al.*, 1972; Hanks and Anderson, 1972) but merely wish to extend it to inferred lunar interior temperatures in thermal models as well.

## 5. A Model of Early Lunar Evolution

From the discussion in the preceding sections, we feel that at this stage of development of lunar science it is too early to think exclusively of a Moon molten at accretion and deduce its further evolution. Alternative models still seem equally viable and should not be abandoned too soon. In fact, if the constraints provided by the data and discussion in Section 2 have any general validity at all, these alternative models may actually turn out to be the only viable models. We wish, therefore, to explore here a 'cold' Moon model that satisfies the geochemical characteristics of the lunar samples, their magnetization history as well as a number of other aspects alluded to in the previous sections.

In the present model, we employ the following constraints:

(1) Surficial regions of the Moon were never extensively molten, nor was the Moon ever totally melted to differentiate an iron core (Sections 2 and 4).

(2) Evidence exists for the presence of extremely differentiated materials on the lunar surface of  $\sim 4.6$  b.y. age (Section 3).

(3) Generation of mare basalts with near chondritic relative abundance patterns of several lithophile elements but with high absolute enrichments relative to the chondritic norm occurred during 3.8–3.1 b.y. ago (Section 3).

(4) Mass concentrations occur on the near Earth-side, somewhere in the outer 200 km of the Moon, such that the center of mass is displaced  $\sim 2$  km to the Earth-side relative to the center of figure (Section 2).

(5) Lunar intrinsic magnetic fields of highly variable nature existed during at least the first 1.5 b.y. of lunar history (Section 4).

From a conventional geochemical point of view, constraints 1 and 2 seem apparent-

ly irreconcilable. How can the highly differentiated 'granites', KREEP and other enriched components be generated at 4.6 b.y. ago without a molten Moon? Current interpretations for the origin of these materials strongly lean toward partial melting phenomena. Gast (1972) has discussed in detail the difficulties encountered in whole Moon melting and fractional crystallization models and suggested that KREEP materials are very small scale partial melts of a layer in the Moon initially enriched in U, REE and other such incompatible trace elements by 5–10 times the chondritic abundances. It is assumed that equilibrium prevails between the liquid and the solid mineral phases with which it coexists. This assumption probably is more valid for the extensive volcanism of the mare lavas or basalt generation in the Earth where sustained thermal activity in the planet leads to melting of the multiphase solid source regions. The extent or the nature of heat source required to produce KREEP basalts and 'granitic' differentiates on the Moon is not known. If the surficial radioactivity of the Moon is due to such materials, present indications are that its distribution is quite local (Metzger *et al.*, 1972). Lunar sampling so far shows that this material is present in small fragments in the soil from various Apollo mission sites and is abundant at the Apollo 14 site.

If these differentiates are indeed sporadic on the lunar surface, it would indicate that their generation locally need not be related to a Moon-wide thermal episode. Nor is there a need to postulate a global layer of material highly enriched in various lithophile trace elements.

The high enrichments observed in these materials may be due to partial melting, but in our model, this is a very local process, presumably initiated by deposition of thermal energy by incoming planetesimals, on to a lunar layer originally well below the solidus temperature. The role of accessory minerals such as apatite or whitlockite originally present in the lunar materials may be significant in this connection. In lunar materials, as well as in stony meteorites and the Earth's mantle, there is sufficient documentation to show that a significant fraction of the incompatible elements such as K, Ba, REE, U and Th reside in accessory phases and grain boundary films (e.g., Albee and Chodos, 1970; Brown *et al.*, 1971; Lovering and Kleeman, 1970; Mason and Graham, 1970; Taylor *et al.*, 1971; Griffin and Murthy, 1969; Condie *et al.*, 1969 and several others). Batch melting of these phases can produce small amounts of liquids characterized by high enrichments in these trace elements. Contamination of the local melts produced during some collisional events by the batch melts of accessory phases can lead to high abundances of various trace elements in the final liquids. A proposal somewhat similar to this has been advanced for the origin of mare basalts by Graham and Ringwood (1971); however Gast (1972) has shown that the low abundance of phosphorous in the mare basalts and their source regions may be a serious objection to this hypothesis as applied to mare basalt origin. KREEP basalts are rich in phosphorous and presumably their source region contained phosphate as a stable phase.

The chemical composition of KREEP basalts and laboratory experiments on melting experiments show that plagioclase is a liquidus phase for these rocks (Meyer *et al.*, 1971; Ringwood *et al.*, 1972). A transient and local heating such as in a collisional

event could induce complete melting of accessory phases plus small quantities of melts produced by surface melting of plagioclase, from the original lunar layer. The low relative abundance of Eu and Sr in KREEP materials is consistent with this idea. If KREEP materials are formed in the manner suggested above by local collisional events, we expect that high resolution age dating of these materials will show a scatter of ages confined to the first few hundred million years of lunar history and a range of  $(^{87}\text{Sr}/^{86}\text{Sr})_1$  values.

We have discussed earlier the europium anomalies in lunar anorthosites. In contrast to the lunar anorthosites, KREEP and similar components in the Moon show remarkable Eu deficiencies with gross enrichments of all REE, U and Th, relative to chondrites. This has been interpreted to mean that these materials were produced within regions in the Moon from which plagioclase has been extensively removed, presumably to form the highlands. However, we note that similar elemental fractionations have occurred elsewhere in the early solar system, outside the Moon. Boynton and Schmitt (1972, personal communication) report negative Eu anomalies similar to those observed in KREEP components from several enstatite and hypersthene achondrites. Schnetzler and Philpotts (1968) have earlier reported similar patterns for some Ca-rich achondrites as well. These data are for bulk samples, but it is clear that we can entertain the idea that accreting lunar matter may in fact merely represent an extreme of the fractionation trend noted in these meteorites. Incipient melting of such material after accretion on the Moon, as we discussed above, can lead locally to the production of small amount of liquids with gross REE enrichment and Eu depletion, such as that seen in the lunar differentiates. Settling and intergranular adjustments of the accreting lunar surface can squeeze these liquids in part to near-surface environments from which they can be widely dispersed on the Moon by collisional and cratering processes.

The point we wish to emphasize here is that whether these highly differentiated materials are partial melts or melts whose composition is dominated by mixing of batch melts of accessory minerals and small quantities of plagioclase in the source regions such as might happen in localized collisions, there seems no need for the constraint of Moon-wide melting at  $\sim 4.6$  b.y. ago, as is commonly employed in thermal history models.

In a Moon at its terminal stages of accretion, we thus envisage temperatures well below the liquidus of the accreting material. The actual temperature cannot be estimated unambiguously. Roedder and Weiblen (1971) have suggested that the highly differentiated materials on the Moon may have been produced by local impact melting and partial to complete homogenization of late stage immiscible liquids present as inclusions and mesostasis in lunar materials plus some plagioclase and pyroxene. Laboratory heating runs have shown that the first appearance of liquid and melting of plagioclase from inclusions in olivine can occur in a temperature range of  $< 1065$ – $1100^\circ\text{C}$ . The temperatures required for melting of glass inclusions and mesostasis would be of course, lower. Since these temperatures were obtained locally by impacts, we suggest that the ambient temperatures in the lunar outer regions at time of accre-

tion were  $\lesssim 1000^\circ\text{C}$  as an upper limit. Such a temperature is well below the liquidus temperature of anhydrous basalt (Cohen *et al.*, 1967). Thus it would seem possible to maintain a solid lunar outer shell during accretion and yet generate the highly differentiated materials known at  $\sim 4.6$  b.y. ago on the lunar surface.

The temperature deduced above satisfies our constraints 1 and 2 in this model, and bears importantly upon constraints 2, 3 and 4.

We have suggested elsewhere that one of the major chemical differentiation processes that occurs in planetary bodies such as the Earth and Moon will be that of an Fe-FeS eutectic liquid segregation (Murthy and Hall, 1970; Murthy *et al.*, 1971; Murthy and Hall, 1972). The melting temperature of the binary eutectic system Fe-FeS at zero pressure is  $988^\circ\text{C}$  (Brett and Bell, 1969). Addition of more components such as FeO, C, Ni will almost certainly reduce this melting temperature. Compared to the Fe-FeS eutectic melting point of  $988^\circ\text{C}$ , the system Fe-FeO-FeS has a ternary eutectic melting temperature of  $915^\circ\text{C}$  (Naldrett, 1968). Thus, the earlier deduced temperature in the outer shell of the Moon will lead to melting of these components at the time of accretion. Once these liquids are produced in the otherwise solid shell of silicates, liquid segregation into the interior will commence due to the higher density of these liquids.

We suggest that this is an unavoidable event in the temperature history deduced in this model of the Moon and even more emphatically so, in models of the Moon where significant melting has been proposed. There is some evidence from the lunar samples themselves that supports the idea of Fe-FeS liquid segregation in the Moon early in its history. The depletions of siderophile and chalcophile elements in lunar materials is well known (Anders *et al.*, 1971; Ganapathy *et al.*, 1970; Laul *et al.*, 1972). In particular, from a consideration of U-Pb isotopic systematics and the composition of initial Pb in some lunar rocks Tera and Wasserburg (1972) have shown that extreme differentiation between U and Pb occurred, after the formation of the Moon, possibly in a time interval of  $\sim 100$  m.y. These known fractionations are eminently consistent with the postulate of Fe-FeS liquid segregation in the Moon, early in its history.

It is the ultimate residence sites of these Fe-FeS liquid segregations and the physical and chemical perturbation they can induce in the lunar globe that concern us here.

Several recent thermal histories of the Moon have employed rapid accretion models in which gravitational energy of accretion is peaked at the outer regions of the Moon (for example, see Toksöz *et al.*, 1972). The absolute temperatures so reached are dependent on the model, and appropriate time scales of accretion from  $10^2$ – $10^4$  yr have been postulated. But a particular characteristic of all these models, even the most extreme ones, is that a large fraction of the radius of the Moon is at temperatures below  $1000^\circ\text{C}$  upon final accretion (Mizutani *et al.*, 1972; Toksöz *et al.*, 1972; Hanks and Anderson, 1972). Thus, we cannot escape the conclusion that even in these high temperature accretion models the Fe-FeS liquids must be trapped at some depth in the colder interior. We see no way that these would lead to a core of Fe-FeS as postulated by Brett (1972).

It is not necessary that the above process of Fe-FeS liquid segregation lead to the formation of an annular shell of Fe-FeS in the Moon. An original radial symmetry if it



existed at all can be easily disturbed. If the Moon was an Earth satellite at this time as suggested by Őpik (1961) (no matter how it originated), the then rapid rotation of the Moon would preferentially concentrate the Fe-FeS mass in the equatorial plane. Instabilities of the type described by Elsasser (1963) can produce inhomogeneities in mass distribution in this layer. These inhomogeneities may be unrelated to the present Earth-Moon direction, but once they are created, the lunar globe will align itself in such a manner that the axis of the least moment of inertia coincides with the line of centers of the Earth and Moon (Jeffreys, 1962). The net effect may well lead to localized concentrations of Fe-FeS preferentially on the front side of the Moon at a depth of  $\sim 200$  km. These, in keeping with contemporary lunar science parlance, we have dubbed as 'fescons' (Murthy *et al.*, 1971).

In Section 2 we have presented data and discussion, chiefly from Kopal (1972) which show that the moment of inertia data call for excess mass located in the outer 200 or so kilometers of the Moon and further that thin plate surficial mascons are not adequate to explain the magnitudes of the moments of inertia. Assuming fescons are the explanation, we show detailed calculations elsewhere (Chase and Murthy, 1972, in preparation) that call for an excess mass  $\sim 10^{-3}$  lunar mass, to be in accord with the moment of inertia data and the center of mass displacement in the Moon. For the mass in the outer 200 km of the Moon, this would correspond to about 0.3% FeS content, which appears eminently reasonable if the outer regions of the Moon accreted from silicate matter exposed to fractionations in the solar system similar to those in various achondritic types of meteorites which contain about 0.5% FeS (Wood, 1963). The well-documented lack of volatile elements in lunar materials is also consistent with extensive solar system fractionations of accreting lunar matter, as first noted by Gast *et al.*, 1970; Ganapathy *et al.*, 1970 and others.

A question to which we would like to address ourselves here is the possible role of fescons as multiple lunar dynamos as against the central core dynamo postulated by Runcorn *et al.* (1971) and Pearce *et al.* (1972).

Earlier we have suggested that an excess mass of fescons corresponding to  $10^{-3}$  lunar mass resides on the near side of the Moon, buried at about a depth of 200 km. We do not expect this to be a single coherent body; most likely several fescons exist as discrete bodies under the ringed maria where gravitational anomalies have been observed. Thus the mass of each of these would be approximately  $10^{-4}$  lunar mass. Assuming a density of 5.6, such a body will have a diameter of about 130 km for a spherical shape, and a thickness less than 100 km for a lens-shaped body. The question then is whether a 100 km fescon, composed of eutectic Fe-FeS, at a temperature of, say 1000°C, is capable of becoming a self-regenerating lunar dynamo in much the same way as the terrestrial core dynamo.

The critical parameter for calculating the magnetic Reynold's number ( $R_m$ ), which in turn will determine the efficiency of self-regeneration for such a body, is its electrical resistivity. Unless the resistivity is low enough the electromagnetic time constant  $\tau$  will be smaller than the mechanical time constant, resulting in non-regeneration. A conservative way to treat the fescon resistivity would be to assume a uniform distri-

bution of a few percent sulphur throughout the body which should produce the highest possible electrical resistivity. But as Gardiner and Stacey (1971) have recently pointed out, molten iron alloyed with silicon exhibits little or no difference in the electrical resistivity, it being nearly the same as that for molten pure iron. This is in agreement with the experimental observation of Busch *et al.* (1971) who find that in liquid transition metal alloys in general, addition of one percent of a polyvalent alloying agent like germanium changes, the resistivity also by only one percent. Sulfur is a polyvalent alloying agent and thus at the worst, we would expect changes of only about 25% in the resistivity of a fesccon from that of pure molten iron. In reality, of course, the situation promises to be even better. The effective resistivity in a heterogeneous body can be regarded as that due to a system of parallel resistors, one set (FeS) being consistently higher in resistivity than the other set (Fe). The net resistivity is, therefore, closer to the lower of the two resistivities. Thus in a Fe-FeS eutectic body like a fesccon, the effective resistivity can be taken as that due to pure iron alone, which is  $3 \times 10^5$  emu at its melting point. It is easy to show now that a 100 km fesccon sphere will indeed be self-regenerating. The  $R_m$ -value is given by  $(l\phi)(4\pi\mu)/\rho$  where  $l$  = characteristic length (130 km),  $\phi$  is the required velocity inside the fesccon, a quantity which has to be deduced and to be shown as reasonably small,  $\mu$  = magnetic permeability (unity) and  $\rho$  = electrical resistivity ( $3 \times 10^5$  emu). In order to be barely self-regenerating,  $R_m$  should be equal to unity but let us assume here a conservatively large  $R_m$  of 100. Then the above relation requires the internal velocity to be  $\geq 0.2$  cm  $s^{-1}$ , a value which is sure to be exceeded in a molten fesccon body which has internal  $^{40}\text{K}$  heat sources and additionally not in chemical equilibrium with its surrounding silicate environment. The velocity corresponding to an  $R_m$  value of 1 is, of course, 0.002 cm  $s^{-1}$ . Incidentally, the electromagnetic time constant  $\tau$  for a 100 km fesccon turns out to be about 230 yr, large enough for selfregeneration yet small enough to be responsible for the observation of wide variations in paleointensity in the NRM of surface rocks which will be near enough to the fesccon sources for recording these time variations. Reducing  $l$  to 20 km, as for example in a lens-shaped body, results in a critical velocity of  $\geq 0.01$  cm  $s^{-1}$  for  $R_m$  of 1, a velocity which is quite reasonable.

Thus, a few fesccons buried at shallow depths, which are capable of acting as sources of internal magnetic fields, can remove the requirement of an early whole Moon melting which the proponents of a single core dynamo perforce need. Our multiple lunar dynamos will have their individual life-cycles, thus producing the observed wide variation in lunar paleointensity as deduced from rocks from different missions.

Levy (1972) has recently shown that if a lunar dynamo is postulated to have a predominantly dipole field with time stability similar to that of the Earth's field (0.7––1.0 m.y.), lunar rotational velocities exceeding the break-up velocity are needed. From the presently available lunar data, very little is known about either the nature of the field (i.e., dipole / non-dipole ratio) or its stability over time. In fact, the discussion in Section 4 argues for a large variability of the lunar field in space and time. Relaxation of the time-stability criterion for the lunar field, of course, would allow for a

rotation period greater than the breakup rotation period of about 80 min (Levy, 1972). We therefore suggest that fescos dynamos can satisfy both the magnetic Reynolds number criterion as well as the rotation period criterion, if we accept that the lunar dynamos could have been less stable than the terrestrial core dynamo. A corollary of our hypothesis would be the observation of both normal and reversely magnetized bedrocks on lunar surface. Secondly, the finite time required by the fescos dynamos to become active will mean that very low paleointensity values should be observed in undisputed, ancient highland materials far from the maria.

Lewis (1971) and Hall and Murthy (1971) have suggested that in Fe-FeS segregation under appropriate conditions in a planetary body, a substantial amount of potassium is incorporated into the Fe-FeS melts. These predictions have been experimentally verified (Goettel, 1972). Ample metallurgical data show that K can be partitioned into sulfide melts and meteoritic analogies of this situation are well known. However, Oversby and Ringwood (1972) have recently reported experimental data which they argue bears against such K extraction into sulfide melts. A number of criticisms can be raised against their experiments as well as interpretation (see, for example, Goettel and Lewis, 1972) but for the time being, we merely wish to say that their experiments have little relevance to the question of K partition between *solid* silicates and *liquid* Fe-FeS melts such as we proposed.

The presence of  $^{40}\text{K}$  with a half life of  $\sim 1$  b.y. in the fescos has a bearing both on the magnetic field history and the generation of mare basalts about a billion years after the formation of the Moon. For example, if the fescos dynamos are driven by  $^{40}\text{K}$  heat generation, we should expect little or no signs of a large lunar paleointensity in rocks older than 4.1 b.y. and younger than, say, 2.6 b.y., if these are ever found. Incidentally, such a short life-time as we expect from the fescos dynamos (as against a core dynamo which, because of its location and thermal insulation, will be active longer) must mean that if in future the lunar soil is found to have a net NRM (being a depositional remanent magnetism, DRM,) the paleointensity deduced from the oldest (0.8 b.y.) of lunar soils should be no greater than the present interplanetary field (1 to  $10\gamma$ ).

The production of mare basalts about a billion years after the formation of the moon, but over a relatively short period between 3.8-3.1 b.y. ago has posed considerable problems in most thermal history models. In the present model this follows as a natural consequence of the heat generation due to  $^{40}\text{K}$  radioactivity in fescos at depths of  $\sim 200$  km (Murthy *et al.*, 1971). Assuming a uniform thermal gradient from 200 km depth to the surface and a conductivity of  $0.01 \text{ cal cm}^{-2} \text{ s}^{-1} \text{ }^\circ\text{C}$ , to produce and maintain silicate melting temperatures of  $1200^\circ\text{C}$  during the first billion years of the Moon's history would require a K concentration of approximately 0.2% in the fescos. This would correspond to the extraction of only about an amount equivalent to 10-20 ppm of K from the 200 km shell into the fescos at the time of Fe-S liquid segregation, if the FeS content of the outer lunar shell is about 0.5 to 1.0%. There seems no need to segregate excessive amounts of K from the outer silicate shell of the Moon into the fescos to cause melting of silicate materials at a later time.

From a consideration of the Rb/Sr systematics of the source regions of mare basalts and the coherence between Sr and  $\text{Eu}^{+2}$  behavior, it was observed that the negative Eu anomaly characteristic of the source regions of mare basalts was established nearly at 4.6 b.y. ago (Papanastassiou and Wasserburg, 1971). In our cold Moon model, this would be a characteristic of the material inherited from fractionations elsewhere in the solar system, such as those shown by various achondritic materials. From the known patterns of REE abundance, we expect that this accretional layer is pyroxene-rich, with minor plagioclase and accessory phases that contain a significant fraction of the bulk trivalent REE and the Eu depletion patterns observed in the mare basalts.

In conclusion, we suggest that a model of the Moon which has never been globally molten, such as has been championed by Harold Urey over the last two decades is still consistent with the known astronomical, geochemical and magnetization data. The model envisages accretional layering in the Moon such as proposed by Gast (1972) but at temperatures well below the solidus upon accretion. Local collisional events raised the temperature at points to produce melts of accessory phases and mesostasis from the near surface material, to produce the highly differentiated materials rich in trace and radioactive elements. The ambient temperature of the outer lunar shell was high enough to melt Fe-FeS binary or related ternary eutectic. Segregation of these liquids into 'fescons' at a depth of  $\sim 200$  km can satisfactorily account for the moments of inertia and the satellite acceleration data.

These fescons can also act as small regenerative dynamos to produce the highly variable magnetization observed in lunar rocks. Because of heat generation by  $^{40}\text{K}$  inside them, these fescons can produce silicate melts in the areas where they occur, up to practically 1.5 b.y. after the formation of the Moon. We suggest that the evidence obtained from the study of lunar samples so far only suggests localized melting in the outer lunar shell, and there is nothing in the data that warrants extrapolation to global melting such as has been used in current thermal models of the Moon.

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### References

- Adler, L., Trombka, J., Gerald, J., Lowman, P., Schmadebeck, R., Blodget, H., Eller, Y., Yin, L., Lamothe, R., Gorenstein, P., and Bjorkholm, P.: 1972, *Science* **175**, 436.  
 Albee, A. L. and Chodos, A. A.: 1970, *Proc. Apollo 11 Lunar Sci. Conf.* **1**, 135.  
 Anders, E., Ganapathy, R., Keays, R. R., Laul, J. C., and Morgan, J. W.: 1971, *Proc. Second Lunar Sci. Conf.* **2**, 1021.

- Anderson, A. J., Jr. and Smith, J. V.: 1971, *Proc. Second Lunar Sci. Conf.* **1**, 431.
- Baldwin, R. B.: 1971, *Phys. Earth Planetary Int.* **4**, 167.
- Banerjee, S. K.: 1972, *Proc. Third Lunar Sci. Conf.* **3** (in press).
- Booker, J. R., Kovach, R. L., and Lu, L.: 1970, *J. Geophys. Res.* **75**, 6558.
- Boynton, W. V. and Schmitt, R. A.: 1972, personal communication.
- Brett, P. R.: 1972, *Geochim. Cosmochim. Acta*, in press.
- Brett, P. R. and Bell, P. M.: 1969, *Ann. Rept. Geophys. Lab. Yearbook* **67**, 198.
- Brown, G. M., Emelens, C. H., Holland, J. G., Peckett, A., and Phillips, R.: 1971 *Proc. Second Lunar Sci. Conf.* **1**, 583.
- Busch, G., Güntherodt, H. J., and Künzi, H. U.: 1971, *Phys. Letters* **34A**, 309.
- Cohen, L. H., Ito, K., and Kennedy, G. C.: 1967, *Am. J. Sci.* **265**, 475.
- Coleman, P. J., Schubert, G., Russell, C. T., and Sharp, L. R.: 1972, *The Moon* **4**, 419.
- Compton, W., Vernon, M. J., Berry, H., and Rudowski, R.: 1971, *Earth Planetary Sci. Letters* **12**, 55.
- Condie, K. C., Kuo, C. S., Walker, R. M., and Murthy, V. R.: 1969 *Science* **165**, 57.
- Conel, J. E. and Holstrom, G. B.: 1968, *Science* **162**, 3860.
- Duke, M. B. and Silver, L. T.: 1967, *Geochim. Cosmochim. Acta* **31**, 1637.
- Dunn, J. R. and Fuller, M.: 1972, *The Moon* **4**, 49.
- Dyal, P. and Parkin, C. W.: 1972, *The Moon* **4**, 63.
- Elsasser, W. M.: 1963, in J. Geiss and E. D. Goldberg (eds.), *Earth Science and Meteorites*, p. 1.
- Ganapathy, R., Keays, R. R., Laur, J. C., and Anders, E.: 1970, *Proc. Apollo 11 Lunar Sci. Conf.* **2**, 1117.
- Gardiner, R. B. and Stacey, F. D.: 1971, *Phys. Earth Planetary Interiors* **4**, 406.
- Gast, P. W.: 1972, *The Moon* **5**, 121.
- Gast, P. W., Hubbard, N. J., and Wisemann, H.: 1970, *Proc. Apollo 11 Lunar Sci. Conf.* **2**, 1143.
- Goettel, K. A.: 1972, *Phys. Earth Planetary Int.* (in press).
- Goettel, K. A. and Lewis, J. S.: 1972, *Earth Planetary Sci. Letters* (in press).
- Graham, A. L. and Ringwood, A. E.: 1971, *Earth Planetary Sci. Letters* **13**, 105.
- Griffin, W. L. and Murthy, V. R.: 1969, *Geochim. Cosmochim. Acta* **33**, 1399.
- Gromme, C. S. and Doell, R. R.: 1971, *Proc. Second Lunar Sci. Conf.* **3**, 2491.
- Hall, H. T. and Murthy, V. R.: 1971, *Earth Planetary Sci. Letters* **11**, 239.
- Hanks, T. C. and Anderson, D. L.: 1972, preprint.
- Hargraves, R. B. and Dorety, N.: 1971, *Proc. Second Lunar Sci. Conf.* **3**, 2477.
- Hargraves, R. B. and Dorety, N.: 1972a, *Proc. Third Lunar Sci. Conf.*, in press.
- Hargraves, R. B. and Dorety, N.: 1972b, *Lunar Science III*. Revised Abstracts of papers presented at the *Third Lunar Sci. Conf.* (ed. by C. Watkins), p. 357.
- Hargraves, R. B. and Dorety, N.: 1972c, *The Apollo 15 Samples*, Lunar Science Institute, Houston, in press.
- Hays, F. J.: 1972, *Phys. Earth Planetary Int.* **5**, 77.
- Helsley, C. E.: 1970, *Proc. Apollo 11 Lunar Sci. Conf.* **3**, 2213.
- Helsley, C. E.: 1971, *Proc. Second Lunar Sci. Conf.* **3**, 2485.
- Hinners, N. W.: 1971, *Rev. Geophys. Space Phys.* **9**, 447.
- Hoffman, K. and Banerjee, S. K.: 1972, Abstract of talk presented at the 1972 National Fall Meeting of A.G.U.
- Hubbard, N. J., Meyer, C. Jr., Gast, P. W., and Wisemann, H.: 1971, *Earth Planetary Sci. Letters* **10**, 341
- Hubbard, N. J., Gast, P. W., Rhodes, M., and Wisemann, H.: 1972, *Proc. Third Lunar Sci. Conf.* in press.
- Jeffreys, H.: 1924, in *The Earth*, Chapter 4, Cambridge University Press.
- Jeffreys, H.: 1962, in *The Earth*, Cambridge University Press.
- Kaula, W. M., Schubert, G., Sjogren, W. L., and Wollenhaupt, H. D.: 1972, *Lunar Science III*, Revised Abstracts of papers presented at the *Third Lunar Sci. Conf.* (ed. by C. Watkins), p. 445.
- Kopal, Z.: 1972, *The Moon* **4**, 28.
- Kuckes, A. F.: 1971, *Nature* **232**, 249.
- Langseth, M. G., Clark, S. P., Jr., Chute, J. C., Keihm, S., and Wechsler, A. E.: 1972, *The Moon* **4**, 390.
- Latham, G., Ewing, M., Dorman, J., Lammlein, D., Press, F., Toksöz, M. N., Sulton, G., Dunnebie, F., and Nakamura, Y.: 1971, *Science* **174**, 687.

- Laul, J. C., Wakita, H., and Schmitt, R. A.: 1972, *The Apollo 15 Samples*, Lunar Science Institute, Houston, in press.
- Levy, E. H.: 1972, *Science* **178**, 52.
- Lewis, J. S.: 1971, *Earth Planetary Sci. Letters* **11**, 130.
- Lovering, J. F. and Kleeman, J. D.: 1970, *Proc. Apollo 11 Lunar Sci. Conf.* **2**, 627.
- Mason, B. and Graham, A. L.: 1970, *Smithsonian Contrib. Earth Sci.* **3**, 627.
- Metzger, A. E., Trombka, J. I., Peterson, L. E., Reedy, R. C., and Arnold, J. R.: 1972, *Lunar Science III*, Revised Abstracts of papers presented at the *Third Lunar Sci. Conf.* (ed. by C. Watkins), p. 540.
- Meyer, C., Brett, R., Hubbard, N., Morrison, D. A., McKay, D. S., Aitken, F. K., Takeda, H., and Schonfeld, E.: 1971, *Proc. Second Lunar Sci. Conf.* **1**, 393.
- Michael, W. H. and Blackshear, W. T.: 1972, *The Moon* **4**, 388.
- Mizutani, H., Takahumi, M., and Takeuchi, H.: 1972, *The Moon* **4**, 476.
- Muller, P. M. and Sjogren, W. L.: 1968, *Science* **161**, 680.
- Murthy, V. R. and Hall, H. T.: 1970, *Phys. Earth Planetary Int.* **2**, 276.
- Murthy, V. R. and Hall, H. T.: 1972, *Phys. Earth Planetary Int.* (in press).
- Murthy, V. R., Evensen, N. M., and Hall, H. T.: 1971a, *Nature* **234**, 267 and 290.
- Murthy, V. R., Evensen, N. M., Jahn, B., and Coscio, M. R., Jr.: 1971b, *Geochim. Cosmochim. Acta* **35**, 1139.
- Murthy, V. R., Evensen, N. M., Jahn, B., and Coscio, M. R.: 1972, *Proc. Third Lunar Sci. Conf.* in press, and unpublished data.
- Nagata, J. and Carleton, B. J.: 1970, *J. Geomag. Geoelectr.* **22**, 491.
- Nagata, T., Fisher, R. M., and Schwerer, F. C.: 1972, *The Moon* **4**, 60.
- Nagata, T., Ishikawa, Y., Kinoshita, H., Kono, M., Syono, Y., and Fisher, R. M.: 1970, *Proc. Apollo 11 Lunar Sci. Conf.* **3**, 2325.
- Nagata, T., Fisher, R. M., Schwerer, F. C., Fuller, M. D., and Dunn, J. R.: 1971, *Proc. Second Lunar Sci. Conf.* **3**, 2461.
- Naldrett, A. J.: 1968, *Ann. Rept. Geophys. Lab. Yearbook* **66**, 419.
- Ópik, E. J.: 1961, *Astron. J.* **66**, 60.
- Oversby, V. M. and Ringwood, A. E.: 1972, *Earth Planetary Sci. Letters* **14**, 345.
- Papanastassiou, D. A. and Wasserburg, G. J.: 1971, *Earth Planetary Sci. Letters* **11**, 37.
- Pearce, G. W. and Strangway, D. W.: 1972, *Apollo 16 Preliminary Science Report*, in press.
- Pearce, G. W., Strangway, D. W., and Larson, E. E.: 1971, *Proc. Second Lunar Sci. Conf.* **3**, 2451.
- Pearce, G. W., Strangway, D. W., and Gose, W. A.: 1972, *Proc. Third Lunar Sci. Conf.* **3**, in press.
- Philpotts, J. A. and Schnetzler, C. C.: 1970, *Proc. Apollo 11 Lunar Sci. Conf.* **2**, 1471.
- Philpotts, J. A. and Schnetzler, C. C.: 1968, *Chem. Geol.* **3**, 5.
- Ringwood, A. E., Green, D. H., and Ware, N. G.: 1972, *Lunar Science III*, Revised abstracts of papers presented at *Third Lunar Sci. Conf.* (ed. by C. Watkins), p. 654.
- Roedder, E. and Weiblen, P. W.: 1971, *Proc. Second Lunar Sci. Conf.* **1**, 507.
- Runcorn, S. K., Collinson, D. W., O'Reilly, W., Battey, M. H., Stephenson, A. A., Jones, J. M., Manson, A. J., and Readman, P. W.: 1970, *Proc. Apollo 11 Lunar Sci. Conf.* **3**, 2369.
- Runcorn, S. K., Collinson, D. W., O'Reilly, W., Stephenson, A., Greenwood, N. N., and Batley, M. H.: 1971a, Abstract of talk presented at the *Second Lunar Sci. Conf.*, Houston.
- Runcorn, S. K., Collinson, D. W., O'Reilly, W., Stephenson, A., Battey, M. H., Manson, A. J., and Readman, P. W.: 1971b, *Proc. Roy. Soc. London* **A325**, 157.
- Schnetzler, C. C. and Philpotts, J. A.: 1969 in P. M. Millman (ed.), *Meteorite Research*, D. Reidel, pp. 206-216.
- Schwerer, F. C., Huffman, G. P., Fisher, R. M., and Nagata, T.: 1972, *The Moon* **4**, 187.
- Smith, J. V., Anderson, A. T., Newton, R. C., Olsen, E. J., and Wyllie, P. J.: 1970, *J. Geol.* **78**, 381.
- Sonnett, C. P., Colburn, D. S., Dyal, P., Parkin, C. W., Smith, B. F., Schubert, G., and Schwartz, K.: 1971, *Nature* **230**, 359.
- Stacey, F. D.: 1969, in *Physics of the Earth*, Wiley, New York.
- Strangway, D. W., Pearce, G. W., Gose, W. A., and Timme, R. W.: 1971, *Earth Planetary Sci. Letters* **13**, 43.
- Taylor, S. R., Rudowski, R., Muir, P., Graham, A., and Kaye, M.: 1971, *Proc. Second Lunar Sci. Conf.*, **2**, 1083.
- Tera, F. and Wasserburg, G. J.: 1972, *Earth Planetary Sci. Letters* **14**, 281.
- Toksöz, M. N., Solomon, S. C., Minear, J. W., and Johnston, D. H.: 1972, *The Moon* **4**, 190.

- Toksóz, M. N., Press, F., Anderson, K., Dainty, A., Latham, G., Ewing, M., Dorman, J., Lammlein, D., Nakamura, Y., Sutton, G., and Duennenbier, F.: 1972b, *The Moon* **4**, 490.
- Turner, G.: 1971, *Earth Planetary Sci. Letters* **11**, 169.
- Urey, H. C., Marti, K. Hawkins, J. W., and Liu, M. K.: 1971, *Proc. Second Lunar Sci. Conf.* **2**, 987.
- Wasserburg, G. J. and Papanastassiou, D. A.: 1971, *Earth Planetary Sci. Letters* **13**, 97.
- Wetherill, G. W.: 1968, *Science* **160**, 1256.
- Wollenhaupt, W. R. and Sjogren, W. L.: 1972, *The Moon* **4**, 337.
- Wood, J. A.: 1963, in Middlehurst and Kuiper (eds.), *The Solar System*, vol. **4**, p. 337.
- Wood, J. A.: 1970, *J. Geophys. Res.* **75**, 6497.
- Wood, J. A., Dickey, J. S., Jr., Marvin, U. B., and Powell, B. N.: 1970 *Proc. Apollo 11 Lunar Sci. Conf.* **1**, 965.