THE MOON: COMPOSITION DETERMINED BY NEBULAR PROCESSES*

JOHN W. MORGAN

U.S. Geological Survey, National Center, Reston, Virginia, U.S.A.

and

JAN HERTOGEN** and EDWARD ANDERS

Enrico Fermi Institute and Department of Chemistry, University of Chicago, Illinois, U.S.A.

Abstract. The bulk composition of the Moon was determined by the conditions in the solar nebula during its formation, and may be quantitatively estimated from the premise that the terrestrial planets were formed by cosmochemical processes similar to those recorded in the chondrites. The calculations are based on the Ganapathy-Anders 7-component model using trace element indicators, but incorporate improved geophysical data and petrological constraints.

A model Moon with 40 ppb U, a core 2% by weight (1.8% metal with $\sim$ 35% Ni and 0.2% FeS) and $Mg/(Fe^{2+} + Mg)^{\ddagger} \sim 0.75$ meets the trace element restrictions, and has acceptable density, heat flow and moment of inertia ratio. The high Ni content of the core permits low-Ti mare basalts to equilibrate with metal, yet still retain substantial Ni. The silicate resembles the Taylor-Jakeš composition (and in some respects the waif Ganapathy-Anders Model 2a), but has lower SiO_2.

Minor modifications of the model composition (U = 30–35 ppb) yield a 50% melt approximating Apollo 15 green glass and a residuum of olivine plus 3 to 4% spinel; the low SiO_2 favors spinel formation, and, contrary to expectation, Cr is not depleted in the liquid. There may no longer be any inconsistency between the cosmochemical approach and arguments based on experimental petrology.

1. Introduction

The bulk composition of the Moon remains controversial. Some authors (Taylor and Jakeš, 1974; Ganapathy and Anders, 1974; Wänke *et al.*, 1974a) contend that the Moon differs markedly from the Earth, which is consistent with simultaneous formation (Öpik, 1972) or capture (Alfvén and Arrhenius, 1969, 1976; Kaula, 1977; Mitler, 1975; Singer and Bandermann, 1970; Smith, 1977; Urey, 1972). Others (Ringwood, 1977a; Binder, 1974; Wänke and Dreibus, 1977) maintain that the Moon matches the composition of the Earth's mantle, and consequently is of fission origin.

As a contribution to this problem, we shall attempt to determine the Moon's composition by a cosmochemical model, first developed for the Earth and Moon (Ganapathy and Anders, 1974), and later extended to Mars (Anders and Owen, 1977) and the eucrite parent body (Hertogen *et al.*, 1977; Morgan *et al.*, 1978). The model rests on the reasonable premise that solar system material has undergone a limited number of fractionation

---

* Paper dedicated to Professor Hannes Alfvén on the occasion of his 70th birthday, 30 May, 1978.
** Present address: Instituut voor Nucleaire Wetenschappen, Rijksuniversiteit Gent, Proeftuinstraat, 86, B-9000 Gent, Belgium.
‡ Unless otherwise stated, this ratio refers to the *molar* ratio throughout this paper.

processes in the solar nebula; the chondrites, as primitive nebular condensates, record these processes. The chemical elements show considerable internal coherence in these fractionations, so that in principle abundances for 83 elements can be derived from few geochemical and geophysical constraints. Bulk compositions derived for the Earth and eucrite parent body seem consistent with petrological and geophysical data (Stolper, 1975, 1977; Smith, 1977; Ringwood, 1977b). This work presents calculations for the bulk composition of the Moon, based upon improved geophysical and petrological data, and using more explicit algebra.

2. The Model

2.1. PLANETARY COMPONENTS

Equilibrium condensation from solar gas produces three types of primordial dust: a refractory-rich early condensate, metallic Ni–Fe, and Mg-silicates (Grossman and Larimer, 1974). Below about 700K, volatiles are collected; metal reacts with H_2S to give FeS (troilite) and with H_2O to give FeO (Larimer and Anders, 1967). The former becomes an additional component, the latter combines with Mg-silicates. These 4 components may not remain in original cosmic proportions, however. Early condensate may be gained or lost by settling to the median plane before precipitation of the other components; the E-chondrites, for example, show clear evidence of loss of refractory elements (Larimer and Anders, 1970). Metallic Fe–Ni may be fractionated from other components, possibly by ferromagnetic effects. This fractionation is most clearly seen in the depletion of total Fe in the L- and LL-chondrites relative to H-chondrites, but all chondrite groups appear to have lost metal relative to Cl-type composition (Larimer and Anders, 1970). Millimeter-sized chondrules are formed just before or during accretion, perhaps by impact melting of some portion of the original components (Whipple, 1972); volatiles are depleted and FeS is reduced to metal. Chondrule formation is assumed to affect the same fraction of all 3 components and provides 2 further components: remelted silicate and remelted Fe–Ni. Elements condensing below 600K may be only partially condensed to a degree which is unpredictable *a priori* but can be inferred from volatile patterns in planetary rocks. A seventh volatile-rich component is therefore postulated, containing these elements in proportions similar to those in C3V carbonaceous chondrites (Ganapathy and Anders, 1974; Anders and Owen, 1977).

  For a differentiated planet, the proportions of the various components can be esti-mated from whole-planet abundances of a few "index elements". The early condensate, the sole source of U, can be directly calculated from the bulk U, which is in turn derived from surface heat-flow measurements. The K abundance is a measure for the unremelted silicates and the Tl abundance for the volatile-rich component. Given a value for bulk U, planetary K and Tl contents can be determined from the K/U and Tl/U ratios in surface rocks. These ratios are nearly constant in igneous rocks on a given planet and presumably reflect the ratios in the bulk planet. The metal content is constrained by the oxidation

state of iron [Mg/(Fe^{2+} + Mg) ratio in silicates] and the bulk Fe abundance, estimated from geophysical measurements (e.g. seismic data) and planetary densities.

2.2. ABUNDANCE DATA

As a generally acceptable starting point we have used the cosmic abundance table of Cameron (1973), to which the following modifications have been made: Ca/Al has been adjusted to 1.08 (Ahrens, 1970); the abundance of Fe was raised to 8.9×10^5 (Mason, 1962); K was lowered from 4.2×10^3 to 3.79×10^3 (Nichiporuk and Moore, 1974); Th/U was taken to be 3.6, B was lowered to 54 (Weller et al., 1977).

2.3. COMPOSITION OF COMPONENTS

The Si and Mg content of the early condensate (EC) was adjusted so as to satisfy two constraints: an average enrichment factor of 17.5 × Cl for the refractory elements (Ca, Al, U, etc.) and a Si/Mg mass ratio equal to 2, closely resembling the Ca–Al rich inclusions in the Allende meteorite (Clarke et al., 1970; Wänke et al., 1974b; Grossman et al., 1977).

The remelted silicate component (RS) comprises the remaining Si and Mg, and those lithophile elements that are not appreciably volatilized upon remelting. Addition of lithophile elements condensing between 1300 and 600°K (Na, K, etc.) to the remelted silicates gives the unremelted silicate component (URS). And finally, the unremelted silicates plus the highly volatile elements (C3V chondrite proportions, relative to Tl, for H, C, N, and noble gases; cosmic proportions for all others) make up the volatile-rich component (VRS).

The composition of the silicates was calculated on a FeO-, MnO- and Cr$_2$O$_3$-free basis. The formal treatment of these metal oxides as separate entities simplifies the calculations, and obviates the need to tailor the composition of the silicate components to the oxidation state and iron content of specific planets.

2.4. CALCULATIONS

Let us use the following symbols for the mass fraction of the various components: c = early condensate, s = silicate, m = metal, t = troilite. The remelted, unremelted, and volatile-rich silicate components are designated by the subscripts r, u and v; subscript p refers to bulk planetary abundances. Greek letters are constants, whose numerical values depend upon the input data.

Mass balance requires that

$$c + s_r + s_u + s_v + m + t + \text{FeO} + \text{MnO} + \text{Cr}_2\text{O}_3 = 1. \tag{1}$$

The values of three of the parameters immediately follow from the index elements U, K, and Tl as

$$c = \text{U}_p/\text{U}_c,$$

$$s_v = \text{Tl}_p/\text{Tl}_v,$$

$$s_u = (\text{K}_p - s_v\text{K}_v)/\text{K}_u;$$

K_p and Tl_p are found from $(K/U)_p$, $(Tl/U)_p$ and U_p.

The $Mg/(Fe^{2+} + Mg)$ ratio, recast as $FeO = \alpha Mg_p$, gives

$$FeO = \alpha' + \beta's_r, \tag{2}$$

since $Mg_p = cMg_c + s_rMg_r + s_uMg_u + s_vMg_v$.

It is unlikely that Mn was completely lost upon remelting; therefore, MnO is tied to FeO, assuming that the FeO/MnO ratio remained approximately constant at 85 in lunar igneous processes. (Fe/Mn)-solid-liquid distribution coefficients actually vary from 0.8 to 1.3 (Stolper, 1977); but since Mn is not a critical element in the present calculations, FeO/MnO may serve as an acceptable index of the Mn abundance.

Chromium requires a more elaborate treatment. We assume that 80% of cosmic Cr condensed with metal, the remainder with Mg-silicates (Grossman and Olsen, 1974). Presumably, metallic Cr will react with H_2O to give chromite, which finds its way into the silicates. Then

$$Cr_p = 0.20 \times 1.086(Cr/Si)_{cos}(s_rSi_r + s_uSi_u + s_vSi_v)$$
$$+ 0.80(Cr/Fe)_{cos}Fe_p.$$

The factor 1.086 arises from the correction for Si present in the early condensate.

Equation (1) then reduces to

$$\alpha'' + \beta''s_r + \gamma''Fe_p + m + t = 1. \tag{3}$$

The metal and troilite mass fractions are fixed by the oxidation state of iron and the bulk Fe content

$$Fe_p[1 + (Ni + Co + P)_{cos}/Fe_{cos}] = 0.7773FeO + m + 0.6353t.$$

The two numerical coefficients are stoichiometric factors. With Equation (2) substituted for FeO, we find that

$$1.065Fe_p = \delta + \epsilon s_r + m + 0.6353t. \tag{4}$$

For a given total Fe content, Equations (3) and (4) comprise a system of two equations with three unknowns: s_r, m and t. We obviously need a third equation or an additional constraint to solve the problem in a straightforward way.

For want of a suitable index element for either troilite or metal, we will follow the Ganapathy-Anders approach, and suppose that for a Fe-poor body, such as the Moon, oxidation of primary metal went to completion prior to the remelting event, and that all lunar metal was produced by remelting of troilite. (This is a postulate and does not directly follow from the model.) The remelting process is assumed to be non-selective, and the relative proportions of metal and troilite thus have to satisfy the constraint

$$m/(0.6353t) = f/(1 - f), \tag{5}$$

where f is the remelted fraction, calculated from s_r, s_u and s_v after normalization to an involatile element, such as Si, by

$$f = s_r \text{Si}_r / (s_r \text{Si}_r + s_u \text{Si}_u + s_v \text{Si}_v).$$

Obviously. the troilite fraction is also limited by the available sulfur. Since it is rather unlikely that S would be more abundant in the Moon than less volatile K, the troilite fraction may be written as

$$t = K_p (\text{S/K})_{\text{cos}} \times 2.7417 \times \rho, \tag{6}$$

where 2.7417 is the stoichiometric factor to convert S to FeS, and ρ is an adjustable parameter (fudge factor) such that $(0 < \rho \leqslant 1)$.

For a given total Fe content and a chosen value of ρ, Equations (3), (4) and (6) reduce to a system of two equations with two unknowns, s_r and m. However, the actual value of ρ is not directly available, because we do not have any *a priori* knowledge about the planetary S/K ratio, or about the fraction of total S present as troilite. In practice, the appropriate value of ρ (and of s_r and m) was found by iterating the calculations with different values of ρ, until the constraint expressed in Equation (5) was met.

The calculations involve 9 components, but it needs to be stressed that these components are not completely independent. The bulk U content (derived from heat-flow measurements) determines the EC contribution. This in turn fixes s_u and s_v, via K/U and Tl/U respectively. FeO, MnO and Cr_2O_3, though formally treated as separate components, are determined by the silicate components [(via Mg/(Fe^{2+} + Mg)] and the total Fe content. They finally end up in the silicates, and the *actual* mass fractions of remelted, unremelted and volatile-rich silicate components can readily be calculated from s_r, s_u, s_v and the FeO, MnO and Cr_2O_3 abundances. Thus, the assertion that the model has 7 (or more) degrees of freedom (Ringwood and Kesson, 1976) is clearly incorrect. In fact, once planetary ratios are established there is very little flexibility. Accordingly, the model makes rather specific predictions and is subject to rigorous testing; it is *generally* applicable to terrestrial planets and does not invoke different *ad hoc* assumptions (i.e., volatile loss in magmatic processes, unlikely geochemical behaviour, improbable dynamic heating) to explain observed abundances.

3. Bulk Composition of the Moon: Model 4

3.1. INPUT PARAMETERS

There are two important reasons for recomputing the bulk composition of the Moon by the Ganapathy-Anders model. Firstly, values of such geophysical parameters as heat-flow and moment of inertia have been substantially revised and improved since 1974 (Langseth *et al.*, 1976; Blackshear and Gapcynski, 1977). Secondly, the determination of oxidized metal, which was formerly based on FeO/MnO and total lunar Fe (Parkin *et al.*, 1973; Reynolds and Summers, 1969) has been reconsidered. The new input data are summarized in Table I, and are explained in more detail below.

TABLE I

Constraints on lunar composition

Refractory fractionation	Bulk U = 40 ppb
Oxidized Fe	FeO/(FeO + MgO) = 0.20
Remelting	K/U = 1900
Voltatiles	Tl/U = 0.0023
Metal + Troilite	Geophysical data

3.2. HEAT FLOW AND BULK U CONTENT

Ganapathy and Anders (1974) adopted a lunar bulk U content of 59 ppb, derived from Apollo 15 and 17 heat flow data (Langseth *et al.*, 1973). More recently, Keihm and Langseth (1977) combined *in situ* geophysical data, orbiter geochemical measurements, and chemical and physical sample measurements to derive an improved model for the lunar thermal regime to 500 km. They conclude that bulk U content lies between 35 and 46 ppb; we have adopted 40 ppb U as a probable value for our calculations. Ringwood (1976) suggests that even the lower limit of 35 ppb is still too high and considers an estimate of 30 ppb U to be more acceptable. We will consider the relations between petrology and the bulk U content in greater detail in a later section.

3.3. IRON IN THE MOON

Ganapathy and Anders (1974) used a value of $9 \pm 4\%$ Fe derived from magnetometer data (Parkin *et al.*, 1973); this was preferred to, but not incompatible with, an earlier value of 13% (Reynolds and Summers, 1969). More recently, Sonett (private communication) has pointed out some shortcomings of the Parkin *et al.* estimate; in particular, that the contribution of ferromagnetic sub-curie point native Fe in the outer part of the Moon may completely swamp paramagnetic Fe in orthopyroxene and/or olivine, or super-curie point Fe in a possible lunar core. Since native Fe observed in lunar rocks is either meteoritic in origin (in breccias) or derived from surface desulfurization of sulfide (in basalts; Brett, 1976) the content of native Fe in the lunar mantle may not be enough to significantly affect bulk lunar Fe estimates. Nevertheless, in the present work we have chosen to determine FeO, troilite and metallic Ni–Fe independently.

The abundance of FeO is derived from estimates of "Mg number" (Ringwood, 1976) {i.e., $[(MgO/(FeO + MgO)] \times 100$} as determined for the lunar mantle from experimental petrology; the work of a number of authors suggests that values between 75 and 80 may be appropriate, and we have therefore adopted a value of 80, which corresponds to $(FeO/MgO)_{weight} = 0.446$.

The metal and troilite present a less accessible problem. We have assumed that these two components occur predominantly in a possible lunar core. Since the size of the core (or even its existence) is still hotly debated, three possible values have been adopted.

(1) 0 wt. %; this is clearly the absolute minimum value.

(2) 2 wt. %; this corresponds to a core radius of ~ 350 km, the maximum size acceptable to Nakamura *et al.* (1976), based on seismic data.

(3) 3.9%; this value is based on a new determination of the lunar moment of inertia ratio, 0.391 ± 0.002 (Blackshear and Gapcynski, 1977). The core size was derived from the curves of Dainty *et al.* (1974), assuming that the core contains 10% troilite.

3.4. RESULTS

Lunar compositions in terms of the "chondritic" components are listed in Table II for three possible lunar core sizes: 0, 2 and 3.9%. These shall be referred to as models 4a, b and c. The major element composition for Model 4b is given in Table III.

The abundances of silicate components are comparable with those of the Ganapathy-Anders Model 2a (Anders, 1977). Model 2a was originally abjured by its creators, because of an error in the U value for the early condensate, but has been partially rehabilitated since this error fortuitously anticipated the new, lower U abundance of the Moon. Moreover, it is petrologically more acceptable than the previously favored Model 3a. The new metal and troilite contents are lower and the FeO content is much higher. These were independently derived from petrological and geophysical information, whereas those of Ganapathy and Anders were based on the FeO/MnO ratio in lunar surface rocks. The calculation of Fe°/FeO from FeO/MnO critically depended upon the constancy of this ratio in igneous processes and upon the ill-defined fraction of Mn lost during remelting. Our approach does not require any *a priori* estimate of Mn loss. The resulting Mn abundance corresponds to about 30% retention, similar to the value observed in chondrules from carbonaceous chondrites (Schmitt *et al.*, 1965, 1967).

The total Fe content in our new estimates agree well with independent geophysical estimates of 13% (Reynolds and Summers, 1969) and 9 ± 4% (Parkin *et al.*, 1973).

The model predicts *ca.* 11 wt.% troilite in the putative core, corresponding to ρ-values – see Equation (6) – of 0.1 and 0.2 for a 2 and 3.9% core respectively. Taken at face value, these low ρ-numbers suggest either that the bulk lunar S/K ratio is significantly lower than the cosmic ratio, or that a major part of the total S is present in the mantle and crust. The maximum allowed troilite content in our model ($\rho = 1$) is 2.25% by weight. For most acceptable values of core size, this would imply that the lunar core contains no metal, only FeS. Therefore, the low ρ-values do not appear to be implausible.

The high Ni content of the metal in the assumed core is a consequence of the low Fe°/FeO ratio (as already pointed out by Kimura *et al.*, 1974), and is an obvious analogy to Prior's rules. This may have an important corollary. The Ni content of lunar low-Ti basalts is quite high; 50 ppm for the mean of 43 Apollo 12 and 15 basalts. The value of 150 ppm given by Ringwood (1976) is clearly too high, since only 2 of the 43 basalts listed in the references cited by Ringwood are as high as 150 ppm, and none exceed it. Ringwood has argued that the Ni content of the low-Ti basalts shows categorically that neither the primary low-Ti basalts nor their source material have been in equilibrium with metallic Fe or Fe-rich alloys. This argument is not germane, because by Prior's rules, the lunar core should be Ni-rich, not Fe-rich (Kimura *et al.*, 1974). From our model the Ni-content of the lunar core (2% by weight) would be ~ 35%. Kimura *et al.* (1974) have equilibrated a Ni-free, synthetic lunar basalt with metal of 30% Ni, and found Ni contents

TABLE II

Model compositions of the Moon

Component	4a	4b	4c
		Core, wt. %	
	0	2.0	3.9
Early condensate	0.252	0.252	0.252
Silicate, remelted	0.678	0.667	0.649
Silicate, unremelted	0.061	0.061	0.061
Volatile-rich material	0.0003	0.0003	0.0003
Metal	0	0.018	0.034
Troilite	0	0.002	0.005
Fraction remelted	0.919	0.918	0.915
FeO	0.129	0.127	0.124
Total Fe	0.100	0.111	0.125

TABLE III

Major element composition for Moon model 4b, assuming a core of 2% by weight

	Bulk Moon		Silicate only	
	Elements %	Oxides %	Oxides %	Norm %
Silicate				
Si	19.83	42.41	43.29	Ab 0.88 ⎫ 21.10
Ti	0.23	0.38	0.39	An 20.22 ⎭
Al	3.93	7.43	7.58	Di 8.16
Cr	0.20	0.29	0.30	Hy 13.10
Fe(oxide)	9.87	12.70	12.96	Ol 56.46
Mg	17.18	28.49	29.09	Ilm 0.74
Mn	0.12	0.15	0.15	Chr 0.44
Ca	4.29	6.01	6.13	
Na	0.071	0.095	0.097	
K	0.0076	0.009	0.009	
O	42.24			
Metal				
Fe	1.09			
Co	0.03			
Ni	0.63			
Troilite				
Fe	0.14			
S	0.08			

of 39 and 50 ppm after $2\frac{1}{2}$ hr at 1500°C. These values happen to agree with the mean for mare basalts, 50 ppm, and since the true equilibrium values may be even higher (Kimura *et al.*, 1974), mare basalts or their source region may well have equilibrated with Ni-rich metal. Ringwood's claim that they did not equilibrate with Fe-rich metal is irrelevant, even if true.

The norm for the silicate composition is also shown in Table III. For the bulk Moon, olivine (Fa_{19}) and plagioclase (An_{95}) are the predominant minerals. The 1 bar density for the silicates calculated from the normative composition is 3.27, and the mean density at STP of the model Moon is 3.31 g cm^{-3}. This value is about 1% too low, since the actual STP density is 3.343 ± 0.003 g cm^{-3} (Dainty *et al.*, 1974). Increasing the core size to 3.9 wt.% gives an almost perfect match to the bulk density and moment of inertia. This may not be a completely satisfactory solution, however, since the seismic properties of the mantle must also be satisfied. If we assume a 60 km anorthositic crust (0.100 total lunar volume; density, 2.95 g cm^{-3}) the STP density of the mantle becomes 3.31 g cm^{-3}; a value which is within the range acceptable for a Moon with a core, but only just. An increase of about 1% in the density would provide a somewhat more acceptable figure, and corresponds approximately to $MgO/MgO + FeO = 0.75$. Whether we choose to increase the core size or decrease $MgO/MgO + FeO$ in our model to achieve a satisfactory density, the total Fe content of the Moon will increase to correspond rather closely to the value of 13% Fe estimated by Reynolds and Summers (1969).

A complete set of abundances for Model 4b is given in Table IV. It supersedes a similar table for Model 3a (Ganapathy and Anders, 1974).

4. Two Alternative Compositions

The model lunar composition obtained here resembles that of Ganapathy and Anders (1974) Model 2, except for the higher FeO, as well as that derived by Taylor and Jakeš (1974) from geochemical rather than cosmochemical arguments (Table V). Melting equilibria experiments by Ringwood (1976) on material of the Taylor-Jakeš model suggest that such a composition is too high in (Ca + Al) relative to (Mg + Fe + Si) to be compatible with the petrology of lunar surface rocks, though other studies (Walter *et al.*, 1975; Weill and McKay, 1975; Longhi, 1977) show no such inconsistency. It is the purpose of this section to lessen the apparent inconsistency between Ringwood's petrological results and the cosmochemical approach (Ganapathy and Anders, 1974; Anders, 1977).

The major discrepancy lies in the Ca and Al contents. These in turn depend upon the bulk U content of the Moon; since this is derived from heat-flow measurements and is to some extent model dependent, it is perhaps the major uncertainty in the composition estimate. We feel that our adopted value of 40 ppb cannot be greatly in error because Taylor (1976) arrived at the same value from geochemical arguments based on orbital gamma-ray data. However, interpretation of orbital data in terms of bulk K and U is also model dependent, and a reanalysis of the orbital data (Metzger *et al.*, 1977) has apparently relaxed the lower limit constraint on the bulk U abundance.

TABLE IV

Elemental abundances[a] in Moon model 4b, with a 2 wt. % core

H	0.8	Ti	2300	Ru	3.3	Dy	1.01
He^4	300	V	230	Rh	0.71	Ho	0.22
Li	2.4	Cr	1700	Pd	0.32	Er	0.65
Be, ppb	126	Mn	1140	Ag, ppb	8.9	Tm	0.098
B, ppb	1.5	Fe%	11.48	Cd, ppb	0.39	Yb	0.64
C	11	Co	300	In, ppb	0.051	Lu	1.08
N	0.10	Ni%	0.65	Sn, ppb	79	Hf	0.64
O%	42.0	Cu	6.4	Sb, ppb	7.0	Ta, ppb	65
F	24	Zn	3.1	Te, ppb	31	W	0.51
Ne^{20}	1.0	Ga	0.61	I, ppb	0.32	Re, ppb	169
Na	710	Ge	1.53	Xe^{132}	0.042	Os	2.4
Mg%	17.79	As	1.15	Cs, ppb	26	Ir	2.4
Al%	3.92	Se	0.20	Ba	11.4	Pt	4.7
Si%	19.22	Br, ppb	2.6	La	1.06	Au, ppb	92
P	690	Kr^{84}	0.057	Ce	2.8	Hg, ppb	0.19
S	572	Rb	0.26	Pr	0.36	Tl, ppb	0.092
Cl	0.47	Sr	41	Nd	2.0	Pb^{204}, ppb	0.037
Ar^{36}	5.3	Y	7.4	Sm	0.58	Bi, ppb	0.070
K	76	Zr	44	Eu	0.22	Th, ppb	142
Ca%	28	Nb	2.2	Gd	0.80	U, ppb	40
Sc	27	Mo	6.6	Tb	0.15		

[a] Abundances in ppm unless otherwise marked; noble gases 10^{-10} cc STP/g.

TABLE V

Comparison of model lunar compositions

	wt. % (silicate only)				
	This work Model 4b 2% core	G & A (1974) Model 2	Taylor and Jakeš (1974)	Taylor and Bence (1975)	Ringwood (1976)
SiO$_2$	43.29	42.7	44.0	45.9	44.6
TiO$_2$	0.39	0.42	0.3	0.3	0.3
Al$_2$O$_3$	7.58	8.22	8.2	6.0	3.7
Cr$_2$O$_3$	0.30	0.16	–	–	0.4
FeO	12.96	3.21	10.5	10.5	13.9
MgO	29.09	30.7	31.0	32.4	33.4
MnO	0.15	0.04	0.1	–	–
CaO	6.13	7.68	6.0	4.9	3.4
Na$_2$O	0.1	0.11	0.3	0.1	–
K$_2$O	0.01	0.012	0.01	0.01	0.05

It would therefore add credence to our model if it could be tested independently of the bulk lunar U content.

The Apollo 15 green glass may provide an independent check of our compositional model if we accept provisionally the contention that it was derived by a large degree of

TABLE VI

Large degree partial melts and Apollo 15 green glass

	T-J comp. 20 kb		Lunar composition 5[a]		Lunar composition 6[c]		
	Olivine	Spinel	Bulk	50% melt[b]	Bulk	50% melt[d]	Green Glass
SiO_2	40.6	1.2	42.34	46.4	42.86	48.3	45.5
TiO_2		0.1	0.29	0.58	0.34	0.67	0.3
Al_2O_3		61.9	5.68	7.7	6.63	8.3	7.6
Cr_2O_3		4.9	0.35	0.41	0.32	0.25	0.4
FeO	14.8	9.8	17.30	20.1	15.08	15.8	19.8
MnO	0.2		0.20	0.21	0.18	0.18	0.2
MgO	43.5	21.7	29.11	16.0	29.13	16.5	17.9
CaO	0.4	0.2	4.60	8.8	5.36	10.3	8.6
Na_2O			0.07	0.14	0.09	0.18	0.1

[a] Lunar U = 30 ppb; MgO/(MgO + FeO) = 0.75; 2% core; total Fe = 14.2%.
[b] Removal of 47% olivine and 3% spinel.
[c] Lunar U = 35 ppb; MgO/(MgO + FeO) = 0.775; 2% core; total Fe = 12.6%.
[d] Removal of 46% olivine and 4% spinel.

partial melting of primitive lunar material. The fractionated REE pattern in at least some samples (Ridley *et al.*, 1972) and slight siderophile enrichment of probably predominantly meteoritic origin (Ganapathy *et al.*, 1973; Chou *et al.*, 1975) suggest that the actual history may be more checkered. The major assumption made is that 50% partial melting of the primitive material takes place, leaving behind a residue of mainly olivine plus a few per cent of spinel. The compositions of the residual phases are those determined by Ringwood (1977a) for Taylor-Jakeš composition at 20 kb.

The test will be to produce green glass composition with minimal changes to our model. Results for two possible compositions are given in Table VI. Both compositions correspond to core sizes of 2% by weight.

4.1. COMPOSITION 5

The bulk U content was lowered to 30 ppb, the upper limit acceptable to Ringwood (1976) and somewhat lower than the lowest value given by Keihm and Langseth (1977). A value of 0.75 was taken for $Mg/(Mg + Fe^{2+})$. Removal of 47% olivine and 3% spinel results in a composition which matches green glass well in the two troublesome elements Ca and Al; MgO is lower but only a slight change in $(Mg/Mg + Fe^{2+})$ would be required. It has been suggested that Cr depletion would severely limit the amount of spinel removal; yet in fact, for the chosen composition Cr is virtually identical in our 50% melt and in green glass. The rather low SiO_2 content of our bulk composition would also favor the formation of spinel (Longhi, 1977).

The total Fe content (14.2% for a 2% core) is on the high side, though probably still acceptable, given the uncertainties in the estimates of total lunar Fe. A smaller core size would obviously bring down the total Fe.

4.2. COMPOSITION 6

We now assume 30 ppb U and $(Mg/Mg + Fe^{2+}) = 0.775$. The composition of a 50% melt, in which Al_2O_3 is close to that in green glass and the match for Cr still acceptable, corresponds to the removal of 46% olivine and 4% spinel. The agreement for CaO, FeO and other elements is less good than for Composition 5. MgO remains low, but an increase in $Mg/(Mg + Fe^{2+})$ would aggravate the situation for FeO.

Acknowledgements

References

Ahrens, L. H.: 1970, 'The Composition of Stony Meteorites – IX. Abundance Trends of the Refractory Elements in Chondrites, Basaltic Achondrites and Apollo 11 Fines', *Earth Planet. Sci. Lett.* **10**, 1–6.

Alfvén, H. and Arrhenius, G.: 1969, 'Two Alternatives for the History of the Moon', *Science* **165**, 11–17.

Alfvén, H. and Arrhenius, G.: 1976, *Evolution of the Solar System*, NASA SP-345, National Aeronautics and Space Administration, Washington, D.C.

Anders, E.: 1977, 'Chemical Compositions of the Moon, Earth and Eucrite Parent Body'. *Phil. Trans. R. Soc. Lond.* A285, 23–40.

Anders, E. and Owen, T.: 1977, 'Mars and Earth: Origin and Abundance of Volatiles'. *Science* **198**, 453–465.

Binder, A. B.: 1974, 'On the Origin of the Moon by Rotational Fission', *The Moon* **11**, 53–76.

Blackshear, W. T. and Gapcynski, J. P.: 1977, 'An Improved Value of the Lunar Moment of Inertia', *J. Geophys. Res.* **82**, 1699–1701.

Brett, R.: 1976, 'Reduction of Mare Basalts by Sulfur Loss', *Geochim. Cosmochim. Acta* **40**, 997–1004.

Cameron, A. G. W.: 1973, 'Abundances of the Elements in the Solar System'. *Space Sci. Rev.* **15**, 121–146.

Chou, C. L., Boynton, W. V., Sundberg, L. L., and Wasson, J. T.: 1975, 'Volatiles on the Surface of Apollo 15 Green Glass and Trace-Element Distributions among Apollo 15 Soils'. *Proc. Sixth Lunar Sci. Conf.* 1701–1727, Pergamon Press.

Clarke, R. S., Jr., Jarosewich, E., Mason, B., Nelen, J., Gomez, M., and Hyde, J. R.: 1970, 'The Allende, Mexico, Meteorite Shower'. *Smithson. Contrib. Earth Sci.* **5**, 1–53.

Dainty, A. M., Toksöz, M. N., Solomon, S. C., Anderson, K. R., and Goins, N. R.: 1974, 'Constraints on Lunar Structure'. *Proc. Fifth Lunar Sci. Conf.* 3091–3114, Pergamon Press.

Ganapathy, R. and Anders, E.: 1974, 'Bulk Compositions of the Moon and Earth Estimated from Meteorites'. *Proc. Fifth Lunar Sci. Conf.* 1181–1206, Pergamon Press.

Ganapathy, R., Morgan, J. W., Krähenbühl, U., and Anders, E.: 1973, 'Ancient Meteoritic Components in Lunar Highland Rocks: Clues from Trace Elements in Apollo 15 and 16 Samples'. *Proc. Fourth Lunar Sci. Conf.* 1239–1261, Pergamon Press.

Grossman, L. and Olsen, E.: 1974, 'Origin of the High Temperature Fraction of C2 Chondrites'. *Geochim. Cosmochim. Acta* **38**, 173–187.

Grossman, L. and Larimer, J. W.: 1974, 'Early Chemical History of the Solar System'. *Rev. Geophys. Space Phys.* **12**, 71–101.

Grossman, L., Ganapathy, R., and Davis, A. M.: 1977, 'Trace Elements in the Allende Meteorite – III. Coarse-Grained Inclusions Revisited'. *Geochim. Cosmochim. Acta* **41**, 1647–1664.

Hertogen, J., Vizgirda, J., and Anders, E.: 1977, 'Composition of the Parent Body of Eucritic Meteorites'. *Bull. Am. Astron. Soc.* **9**, 458–9.

Kaula, W. M.: 1977, 'On the Origin of the Moon, with Emphasis on Bulk Composition'. *Proc. Eighth Lunar Sci. Conf.* 321–331.

Keihm, S. J. and Langseth, M. G.: 1977, 'Lunar Thermal Regime to 500 km'. *Lunar Science VIII*, 531–533. The Lunar Science Institute, Houston.

Kimura, K., Lewis, R. S., and Anders, E.: 1974, 'Distribution of Gold and Rhenium between Nickel-Iron and Silicate Melts: Implications for the Abundance of Siderophile Elements on the Earth and Moon'. *Geochim. Cosmochim. Acta* **38**, 683–701.

Langseth, M. G., Keihm, S. J., and Chute, J. C., Jr.: 1973, 'Heat Flow Experiment'. *Apollo 17 Preliminary Science Report*, NASA SP-330, 9–1 through 9–24.

Langseth, M. G., Keihm, S. J., and Peters, K.: 1976, 'Revised Lunar Heat-Flow Values'. *Proc. Seventh Lunar Sci. Conf.* 3143–3171, Pergamon Press.

Larimer, J. W. and Anders, E.: 1967, 'Chemical Fractionations in Meteorites – II. Abundance Patterns and their Interpretation'. *Geochim. Cosmochim. Acta* **31**, 1239–1270.

Larimer, J. W. and Anders, E.: 1970, 'Chemical Fractionations in Meteorites – III. Major Element Fractionations in Chondrites'. *Geochim. Cosmochim. Acta* **34**, 367–387.

Longhi, J.: 1977, 'Magma Oceanography 2: Chemical Evolution and Crustal Formation'. *Proc. Eighth Lunar Sci. Conf.* 601–621, Pergamon Press.

Mason, B.: 1962, *Meteorites*. Wiley, New York.

Metzger, A. E., Haines, E. L., Parker, R. E., and Radocinski, R. G.: 1977, 'Thorium Concentrations in the Lunar Surface. I: Regional Values and Crustal Content'. *Proc. Eighth Lunar Sci. Conf.* 949–999, Pergamon Press.

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.

Morgan, J. W., Higuchi, H., Takahashi, H., and Hertogen, J.: 1978, 'A "Chondritic" Eucrite Parent Body: Inference from Trace Elements'. *Geochim. Cosmochim. Acta* **42**, 27–38.

Nakamura, Y., Latham, G. V., Dorman, H. J., and Duennebier, F.: 1976, 'Seismic Structure of the Moon: A Summary of Current Status'. *Proc. Seventh Lunar Sci. Conf.* 3113–3122, Pergamon Press.

Nichiporuk, W. and Moore, C. B.: 1974, 'Lithium, Sodium and Potassium Abundances in Carbonaceous Chondrites'. *Geochim. Cosmochim. Acta* **38**, 1691–1701.

Öpik, E. J.: 1972, 'Comments on Lunar Origin'. *Irish Astron. J.* **10**, 190–238.

Parkin, C. W., Dyal, P., and Daily, W. D.: 1973, 'Iron Abundance in the Moon from Magnetometer Measurements'. *Proc. Fourth Lunar Sci. Conf.* 2947–2961, Pergamon Press.

Reynolds, R. T. and Summers, A.: 1969, 'Calculations on the Compositions of the Terrestrial Planets'. *J. Geophys. Res.* **74**, 2494–2511.

Ridley, W. I., Reid, A. M., Warner, J. L., and Brown, R. W.: 1972, 'Apollo 15 Green Glass'. *Phys. Earth Planet. Interiors* **7**, 133–136.

Ringwood, A. E.: 1976, 'Basaltic Magmatism and the Composition of the Moon. Part I, Major and Heat-Producing Elements'. *Publication 1221, Research School of Earth Sciences*, Australian National University, Canberra.

Ringwood, A. E.: 1977a, 'Limits on the Bulk Composition of the Moon'. *Phil. Trans. R. Soc. Lond.* **A285**, 23–40.

Ringwood, A. E.: 1977b, 'Composition of the Core and Implications for Origin of the Earth'. *Publication 1227, Research School of Earth Sciences*, Australian National University, Canberra.

Schmitt, R. A., Smith, R. H., and Goles, G. G.: 1965, 'Abundances of Na, Sc, Cr, Mn, Fe, Co and Cu in 218 Individual Meteoritic Chondrules via Activation Analysis 1. *J. Geophys. Res.* **70**, 2419–2444.

Schmitt, R. A., Smith, R. H., Ehmann, W. D., and McKeown, D.: 1967, 'Silicon in Meteoritic Chondrules'. *Geochim. Cosmochim. Acta* **31**, 1975–1985.

Singer, S. F. and Bandermann, L. W.: 1970, 'Where was the Moon Formed?' *Science* **170**, 438–439.

Smith, J. V.: 1977, 'Possible Controls on the Bulk Composition on the Earth: Implications for the Origin of the Earth and Moon'. *Proc. Eighth Lunar Sci. Conf.* 333–369, Pergamon Press.

Stolper, E.: 1975, 'Petrogenesis of Eucrite, Howardite and Diogenite Meteorites'. *Nature* **258**, 220–222.

Stolper, E.: 1977, 'Experimental Petrology of Eucritic Meteorites'. *Geochim. Cosmochim. Acta.* **41**, 587–611.

Taylor, S. R.: 1976, 'Geochemical Constraints on the Composition of the Moon'. *Proc. Seventh Lunar Sci. Conf.* 3461–3477, Pergamon Press.

Taylor, S. R. and Jakeš, P.: 1974, 'The Geochemical Evolution of the Moon'. *Proc. Fifth Lunar Sci. Conf.* 1287–1305, Pergamon Press.

Taylor, S. R. and Bence, A. E.: 1975, 'Evolution of the Lunar Highland Crust'. *Proc. Sixth Lunar Sci. Conf.* 1121–1142, Pergamon Press.

Urey, H. C.: 1972, 'Evidence for Objects of Lunar Mass in the Early Solar System'. *The Moon* **4**, 383–389.

Walker, D., Longhi, J., and Hays, J. F.: 1975, 'Differentiation of a very Thick Magma Body and Implications for the Source Regions of Mare Basalts'. *Proc. Sixth Lunar Sci. Conf.* 1102–1120, Pergamon Press.

Wänke, H. and Dreibus, G.: 1977, 'The Earth–Moon-System. Chemistry and Origin'. *Proc. 2nd Symposium Origin and Distribution of the Elements*, Paris, in press.

Wänke, H., Palme, H., Baddenhausen, H., Dreibus, G., Jagoutz, E., Kruse, H., Spettel, B., Teschke, F., and Thacker, R.: 1974a, 'Chemistry of Apollo 16 and 17 Samples. Bulk Composition, Late Stage Accumulation and Early Differentiation of the Moon'. *Proc. Fifth Lunar Sci. Conf.* 1307–1335, Pergamon Press.

Wänke, H., Baddenhausen, H., Palme, H., and Spettel, B.: 1974b, 'On the Chemistry of the Allende Inclusions and Their Origin as High Temperature Condensates'. *Earth Planet. Sci. Lett.* **23**, 1–7.

Weill, D. F. and McKay, G. A.: 1976, 'The Partitioning of Mg, Fe, Sr, Ce, Sm, Eu and Yb in Lunar Igneous Systems and a Possible Origin of KREEP by Equilibrium Partial Melting'. *Proc. Sixth Lunar Sci. Conf.* 1143–1158, Pergamon Press.

Weller, M. R., Furst, M., Tombrello, T. A., and Burnett, D. S.: 1977, 'The Solar System Boron Abundance'. *Astrophys. J.* **214**, L39–L43.

Whipple, F. L.: 1972, 'On Certain Aerodynamic Processes for Asteroids and Comets'. *Nobel Symposium 21, from Plasma to Planet* (editor, A. Elvius), pp. 211–232, Almqvist and Wiksell, Stockholm.