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PETROGENESIS OF APOLLO 11 BASALTS AND IMPLICATIONS

FOR LUNAR ORIGIN

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ABSTRACT

The origin of Apollo 11 basalts is discussed in terms of two hypotheses: (i) formation by a small degree of partial melting in the lunar interior and (ii) formation by prolonged near-surface crystallization differentiation in a lava lake. The second hypothesis is rejected on the following grounds: Most Apollo 11 magmas are far removed from the plagioclase-pyroxene-ilmenite cotectic; fractional crystallization cannot explain the large variations in concentrations of incompatible trace elements in conjunction with the small variations in major element compositions, particularly, $\frac{Mg}{Fe}$ ratios; experimentally determined partition coefficients show that the high abundances of Cr and V in Apollo 11 *rocks* cannot be reconciled

with the previous separation of large quantities of ore minerals and pyroxenes. On the other hand, the major element and trace element contents of Apollo 11 rocks can be _readily_ explained by partial melting of sourꞔe material which buffers the major element compositions and causes enrichments of incompatible elements according to the degree of partial melting (first hypothesis)

Two alternative sources have been suggested for Apollo 11 basalts formed by partial melting: (i) unfractionated pyroxenite source region at depths of 200-600 km, (ii) fractionated source region with incompatible elements (e.g. Ba, U, rare earths) strongly enriched over chondritic abundances and containing plagioclase (approximate eucritic composition). Mass balance calculations and plagioclase stability conditions show that the second hypothesis requires Apollo 11 basalts to be generated by partial melting in the outer 150 km of the moon. This is very difficult to achieve one billion years after the moon's formation, since the outer 200 km will have cooled well below the solidus by conduction. Furthermore, magmas generated by partial melting of a plagioclase-bearing source region should have plagioclase on the liquidus, which is contrary to observation. The second hypothesis accordingly appears improbable. The first hypothesis is capable of explaining the major element chemistry

and the trace element abundances (Eu-see below) in terms of a simple, single stage model which is consistent with the moon's density, moment of inertia and inferred thermal history. A possible explanation of the europium anomaly is suggested on the basis of the first hypothesis. It will be necessary to determine the appropriate partition coefficients in order to test this explanation.

If the lunar highlands are anorthositic, extensive differentiation of the outer 150 km of the moon is required. This may have been caused by heating arising from partial conservation of gravitational potential energy during the final stage of accretion. Formation of Apollo 11 basalts by partial melting 3.7 billion years ago was probably the result of radioactive heating (U, Th) in the deep interior of the moon. A two-stage magmatic history for the moon is thus required.

Similarities between compositions of Apollo 11 and terrestrial basalts and between their respective source regions are suggestive of a genetic relationship between moon and earth. Nevertheless, important differences in trace element abundances, major element compositions and oxidation states exist. These abundance patterns are unfavourable to the traditional fission, binary planet and capture hypotheses of lunar origin. However, they may be explicable

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in terms of the precipitation hypothesis proposed by the author.
This maintains that during the later stages of accretion of the earth,
a massive primitive atmosphere developed which was hot enough to
evaporate selectively a substantial proportion of the silicates
which were accreting upon the earth. Subsequently, the atmosphere
was dissipated and the relatively non-volatile silicate components
were precipitated to form a swarm of planetesimals or moonlets, from
which the moon accreted.

PETROGENESIS OF APOLLO 11 BASALTS AND IMPLICATIONS FOR LUNAR ORIGIN

1. INTRODUCTION

It can scarcely be denied that the revelations from the Apollo 11 Lunar Science Conference in Houston, January 1969, were something of a disappointment to the world at large. Instead of finding a consensus among experts as to the broader significance of the Apollo 11 rocks and their bearing upon the nature of the lunar interior and on the origin and evolution of the moon, nearly all interpretative aspects (as distinct from the descriptive) were submerged in controversy and the overall impression obtained by non-scientists was one of considerable confusion. This was unfortunate in view of the intense worldwide public interest shown in the Conference.

In retrospect, it is difficult to see how this situation could have been avoided. With over 130 scientific teams working on specialized problems and ignorant of the findings being made by their colleagues (and rivals), it is not surprising that broad interpretations which were frequently based upon investigations with narrow scientific perspectives should have been so often in conflict. Now that the detailed scientific results on Apollo 11 rocks have been

published, there is no longer any excuse for this degree of confusion. It is important that the broad scientific conclusions which can reasonably be drawn from the study of Apollo 11 rocks with acceptable degrees of probability should be defined so that they can be used as boundary conditions for further investigations. It is equally important that unsolved problems which remain the subject of legitimate scientific controversy should be clearly defined, so that the investigations and data needed to secure solutions can be characterized.

In this paper, certain aspects of the petrogenesis and chemistry of Apollo 11 basalts are discussed with the above considerations in mind, and an attempt is made to critically assess existing information and hypotheses relative to the origin and significance of the Apollo 11 basalts. The bearing of data and interpretations derived from the basalts upon theories of lunar origin is then considered.

2. PETROGENESIS OF LUNAR BASALTS

The first order discovery from the Apollo 11 mission was that rocks from the Sea of Tranquility were clearly recognisable as members of a class of basaltic-type rocks. This is also true of Apollo 12 rocks and is inferred from compositions elsewhere on maria

sites obtained from Surveyors 5 and 6. Furthermore, detailed studies of lunar soil containing material of much wider provenance than the sites studied so far reveal that about 95% of recognisable rock fragments are composed of the same mineral assemblage as occurs in the rocks of local origin. These observations justify the conclusion that the maria generally are almost certainly composed of basalt-like rocks.

High-pressure phase relationships of Apollo 11 rocks and basalts show that they cannot be representative of the lunar interior. (Figure 3). At a relatively shallow depth, they would transform to eclogite with a density of 3.74 g/cm^3, compared to the moon's density of 3.36 g/cm^3. It follows that Apollo 11 rocks and the maria generally are differentiates of a less dense (lower Fe/Mg), more primitive material which comprises the bulk of the moon. (Ringwood and Essene 1970 a, b; O'Hara et. al., 1970 a,b).

This is confirmed by the very high concentrations (up to 100 times the chondritic abundances) of incompatible elements, found in Apollo 11 basalts. These elements (eg, U, Th, Ba, rare earths) possess ionic radii and charges which inhibit their ready entry into the major rock-forming silicates. Accordingly they are very strongly partitioned into the liquid phase during fractionation processes involving common silicates and magma. The observed concentrations of incompatible elements imply the operation of highly efficient crystal-liquid fractionation processes.

Two mechanisms have been proposed. The first maintains that Apollo 11 magma was primarily derived by a small degree (1 to 5%) of partial melting in the lunar interior during which the incompatible elements were almost totally partitioned into the liquid phase (eg. Ringwood and Essene (1970 a,b). According to the second hypothesis, the Apollo 11 magma is a highly residual liquid resulting from extensive crystallization differentiation in near-surface regions, probably in large lava lakes (eg. O'Hara et. al., 1970 a,b; Smith et. al., 1970 a,b). It is of the utmost importance to many broader problems of lunar origin and evolution to determine which of these two hypotheses is correct. I believe that sufficient, evidence is now available to permit a firm conclusion to be drawn.

(a) Fractional Crystallization versus Partial Melting

O'Hara et. al., (1970b) state: "Tranquility Base rocks have compositions similar to those of liquids in equilibrium simultaneously with crystals of olivine, two clinopyroxenes, feldspar and iron-titanium oxide. This remarkable situation must be attributed to coincidence if the Tranquility samples are primary magma (i.e. formed directly by partial melting). However, if the samples are the product of extensive low pressure fractional crystallization, their cotectic character is not only unremarkable: it is mandatory". These authors also cite the small liquidus-solidus temperature interval (60-140°C) and the depletion of Apollo 11 magma in europium

as evidence that Apollo 11 basalts are the "residual liquids of

advanced near-surface crystal fractionation, most probably in a

vast lava lake".

The experimental phase equilibrium data on which these con-

clusions are based are very limited — 3 runs at atmospheric

pressure above the solidus on an Apollo" rock and some near-

liquidus runs in simplified analogue systems lacking Na_2O and Cr_2O_3.

Half of these latter runs were carried out at an oxygen fugacity

of 10^{-8} bar and are not directly relevant to the crystallization

of Apollo 11 basalts which occurred at an oxygen fugacity of

$10^{-13.5}$ bar ($1200^{\circ}C$). When one examines the remaining data, it

becomes evident that ~~it is~~ they are much too sparse to justify the conclusions

reached.

It is true that the temperature interval between solidus and

liquidus $(\ldots^{\circ}-\ldots^{\circ}C)$ for different Apollo 11 rock compositions is

not large and is in the same range as for terrestrial basalts.

This property, however, does not permit firm conclusions to be

drawn concerning the closeness of the liquid to a multiphase cotectic.

It could also arise in situations where the liquidus temperatures

in primary phase volumes were not strongly dependent upon composi-

tion of the melt. Examples of this kind of behavior are not

uncommon in complex systems.

The closeness of a crystallizing liquid to a cotectic of major phases is to be measured not in terms of crystallization temperature intervals, but by the amount of crystallization necessary to bring the liquid from its observed composition to the cotectic composition. One of the most important observations made upon Apollo 11 basalts at atmospheric pressure was the late crystallization of plagioclese which did not appear as a primary phase until 30 to 50 percent of the liquid had crystallized as olivine, pyroxene and ore minerals (Ringwood and Essene, 1970 a,b; Roedder and Weiblen, 1970; Smith et. al., 1970 a,b; Weil et. al., 1970). This observation shows decisively that Apollo 11 basalts were not close to the cotectic composition and sharply contradicts the conclusions (above) of O'Hara et. al. It has been suggested by Smith et. al., (1970b) that at higher pressures in the lunar interior, Apollo 11 basalts may represent a cotectic composition in equilibrium with pyroxene, plagioclase and ilmenite. This is quite impossible, since pressure has exactly the opposite effect to that required, further depressing the crystallization field of plagioclase relative to pyroxene. (Green and Ringwood, 1967; Ringwood and Essene 1970b).

The fact that most Apollo 11 basalts are so far from the cotectic composition is a strong argument that the converse of O'Hara et. al.'s conclusion holds — namely that the parent magma

had <u>not</u> undergone extensive, near-surface fractionation. This would

lead to saturation with plagioclase which should, therefore, appear

on the liquidus in cotectic relationship to other major phases.

Further arguments showing that fractional crystallization

has <u>not</u> been a dominant factor in producing the abundance patterns

characteristic of Apollo 11 rocks, but to the contrary, these have

been produced dominantly by partial melting, are given below. See

also, Ringwood and Essene (1970b).

<u>Major element and trace element abundances</u>: An important measure

of the degree of crystallization undergone by a magma is given by

the $\frac{Mg}{Mg + Fe}$ ratios of derived rocks and minerals. This is a

particularly useful parameter (eg. Figure 2) in Apollo 11 rocks

which show the effects of very strong crystal fractionation on a

microscopic scale, with $\frac{Mg}{Mg + Fe}$ ratios varying from 0.74 to 0.02

(eg. Essene et. al., 1970 a). It is most significant that despite

very different cooling histories as indicated by textural varia-

tions from basaltic to gabbroic (LSPET, 1969), the $\frac{Mg}{Mg + Fe}$ ratios

of individual rocks are very similar and the $\frac{Mg}{Mg + Fe}$ ratios

of the earliest pyroxenes to crystallize are also similar and

close to 0.74 (eg. Essene et. al., 1970). Thus, the variations

of total rock compositions do not provide any evidence of the

operation of extensive fractional crystallization.

In contrast to the relative constancy in abundances of major

elements between individual rocks, the abundances of incompatible

trace elements vary systematically over a five-fold range among

different Apollo 11 rocks (eg. Compston et. al., 1970; Gast et.

al., 1970; Haskin et. al., 1970; Philpotts and Schnetzler, 1970).

If these five-fold variations in the abundances of incompatible

elements are to be explained in terms of fractional crystallization,

it would be necessary that the degree of fractional crystallization

experienced by different ~~Apollozok~~. Apollo 11 rocks has varied

by a factor of five or more. This would cause corresponding major

variations of $\frac{Mg}{Mg + Fe}$ ratios which are not observed. The relative

constancy of $\frac{Mg}{Mg + Fe}$ ratios of Apollo 11 rocks and the small varia-

tion in the abundances of other major elements cannot be reconciled

with an explanation of the variations of incompatible element

abundances in terms of fractional crystallization.

In contrast, this general pattern is exactly what might be

expected for small but variable degrees of partial melting where

the major elements in the liquid are buffered by equilibrium with

residual unmelted phases and are present in relatively constant

proportions, whereas the incompatible elements are almost totally

partitioned into the liquid phase and display wide variations in

amount according to the degree of partial melting. (Section 2d)

This conclusion is further supported by detailed studies of

the fine structure of rare earth abundances, including the europium

anomaly. In their latest papers, Haskin et. al., (1970 a,b), Philpotts

and Schnetzler (1970) and Gast and Hubbard (1970) all favour the

interpretation that rare earth patterns were caused by partial

melting rather than by fractional crystallization. Results of

Gast and Hubbard (1970) comparing rare earth abundances in basaltic

Fig. 1. Rare earth and barium abundances from Apollo 11 and 12

lunar samples and from some basaltic achondrites. The Hl-Rb

group and sample 10020 represent Apollo 11 rocks. Most other

Apollo 11 rocks fall between these samples as does the Apollo

11 soil, 10084. In contrast, the Apollo 12 rocks, 12051,

12053 and 12002 occupy a generally intermediate position

between the Apollo 11 rocks and the basaltic achondrites.

Apollo 12 soils 12070 and 12044 have much higher rare earth

abundances than the corresponding rocks and fall within the

Apollo 11 range.

(After Gast and Hubbard, 1970)
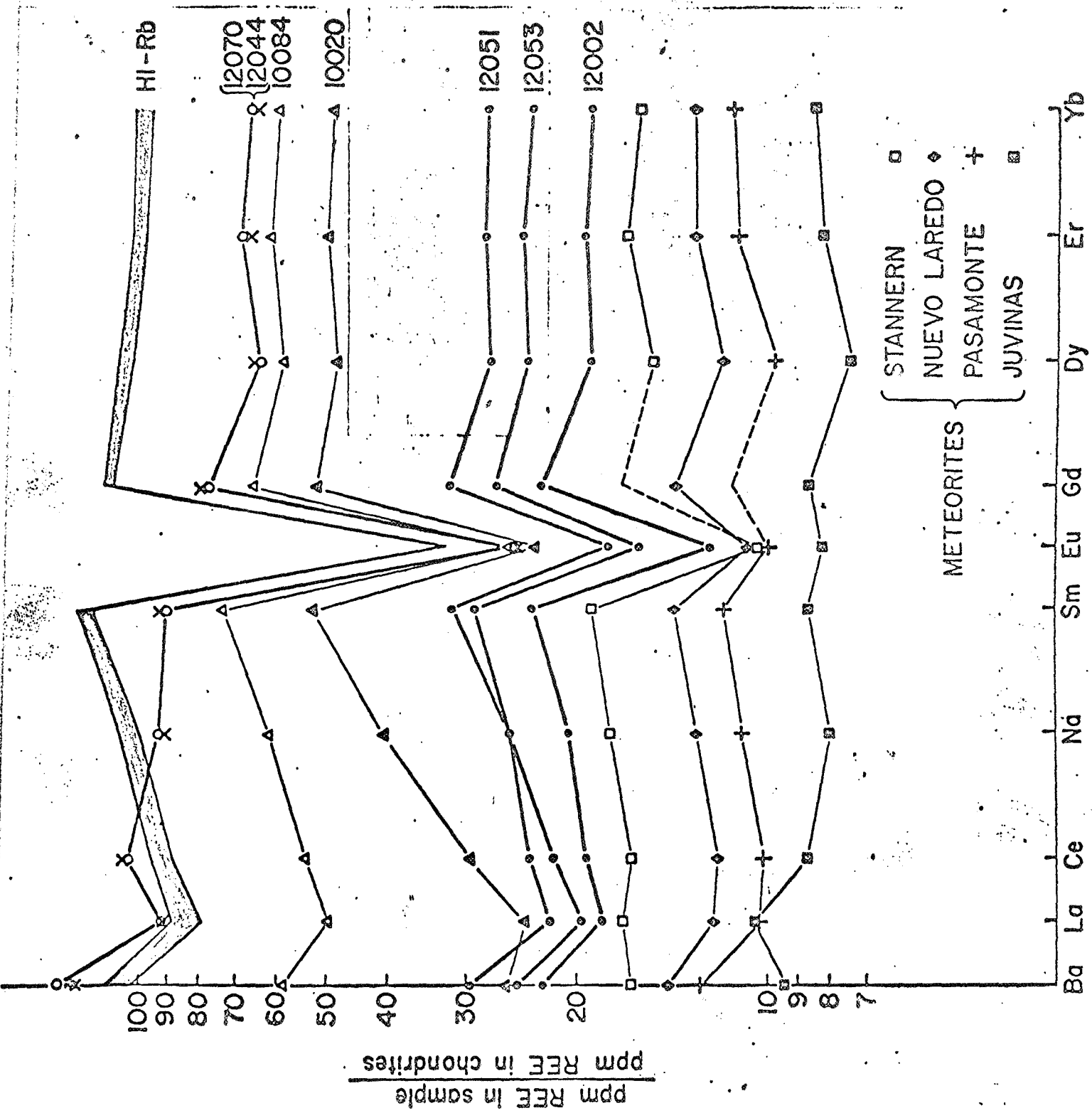
ppm REE in sample / ppm REE in chondrites

achondrites, and in Apollo 11 and 12 rocks are of particular interest

(Figure 1). The Apollo 12 rock abundances are generally intermediate

between those of Apollo 11 rocks and the achondrites, and a system-

atic relationship exists between the absolute abundances of the rare

earths and the magnitudes of the europium anomalies, strongly

suggesting a genetic relationship between the achondrites and the

lunar basalts. Despite variation in the absolute abundances of

rare earths by a factor of 15, the major element compositions of

the Apollo 11, 12 and basaltic achondrites are generally similar.

(Some Apollo 12 rocks display evidence of moderate major element

fractionation, but on a negligible scale compared to the trace element

variations). Gast and Hubbard (1970) point out that the patterns are

readily interpreted in terms of a model which has Apollo 11, 12

rocks and basaltic achondrites formed by increasing degrees of

partial melting of a common source material. This hypothesis had

been suggested earlier by Ringwood (1970 a) and Ringwood and Essene

(1970 a,b).

<u>Abundances of chromium and vanadium:</u> The high abundances of

chromium (av. 2000 ppm) and vanadium (av. 75ppm) in Apollo 11 basalts

are particularly notable. These elements are present at levels of

0.5 and 0.8 of the chondritic abundances. Essene et. al., (1970b)

have determined the crystal-liquid partition coefficients for Cr,

V and some other elements for the principal phases crystallizing from Apollo 11 magma. (Table 1) Note the very high partition coefficients of Cr and V in the ore minerals which are observed on the liquidus of Apollo 11 basalts at atmospheric pressure.

or near

These show that extensive near-surface fractional crystallization with accompanying precipitation of ore minerals and clinopyroxenes would cause drastic decreases in Cr and V abundances, leading to the effective removal of these elements from the magma. The high average abundances of chromium and vanadium remaining in Apollo 11 rocks severely limit the amount of previous low pressure fractionation which may have occurred, and are totally incompatible with the amount of fractionation which has to be postulated in order to explain the abundances of incompatible trace elements. They also contradict O'Hara et. al.'s (1970 a,b) hypothesis that mascons are caused by massive accumulations of ore minerals which have crystallized from the Apollo 11 lava near the surface of lava lakes and sunk to the bottom.

On the other hand, the magnitude of the chromium and vanadium abundances are readily explicable in terms of the experimental partition coefficients (Table 1) if Apollo 11 basalt had formed by direct partial melting from parental material possessing the chondritic Cr and V abundances, leaving behind a refractory residuum of pyroxenes and olivine. (Essene et. al., 1970b).

Table 1

Crystal - liquid partition coefficients (K) for phases on or

near liquidus of synthetic Apollo 11 basalts* containing (A)

0.1% V and (B) 25% additional normative anorthite

plus 0.5% each of Ba, Sr, Eu, and Yb.

(A) Phase	K Vanadium	K Chromium
Armalcolite	18	7
Spinel	38	77
Ilmenite	12	6
Sub-calcic clinopyroxene	3.4	3.5
Olivine	1.3	0.9

(K) Plagioclase - Liquid   (B)

Ba	0.16 ± 0.04
Sr	1.5 ± 0.1
Eu	1.0 ± 0.1
Yb	≤ 0.04

* The basalts were enclosed in sealed iron capsules and the

oxidation states were similar to those of natural Apollo 11

basalts. Composition of basalt is given by Ringwood et. al., (1970a,b).

Fig. 2. Cr_2O_3 contents plotted against $\dfrac{Mg}{Mg + Fe}$ ratios for pyroxenes from Apollo 11 basalt sample 10047. Data were obtained by electron microprobe analyses.
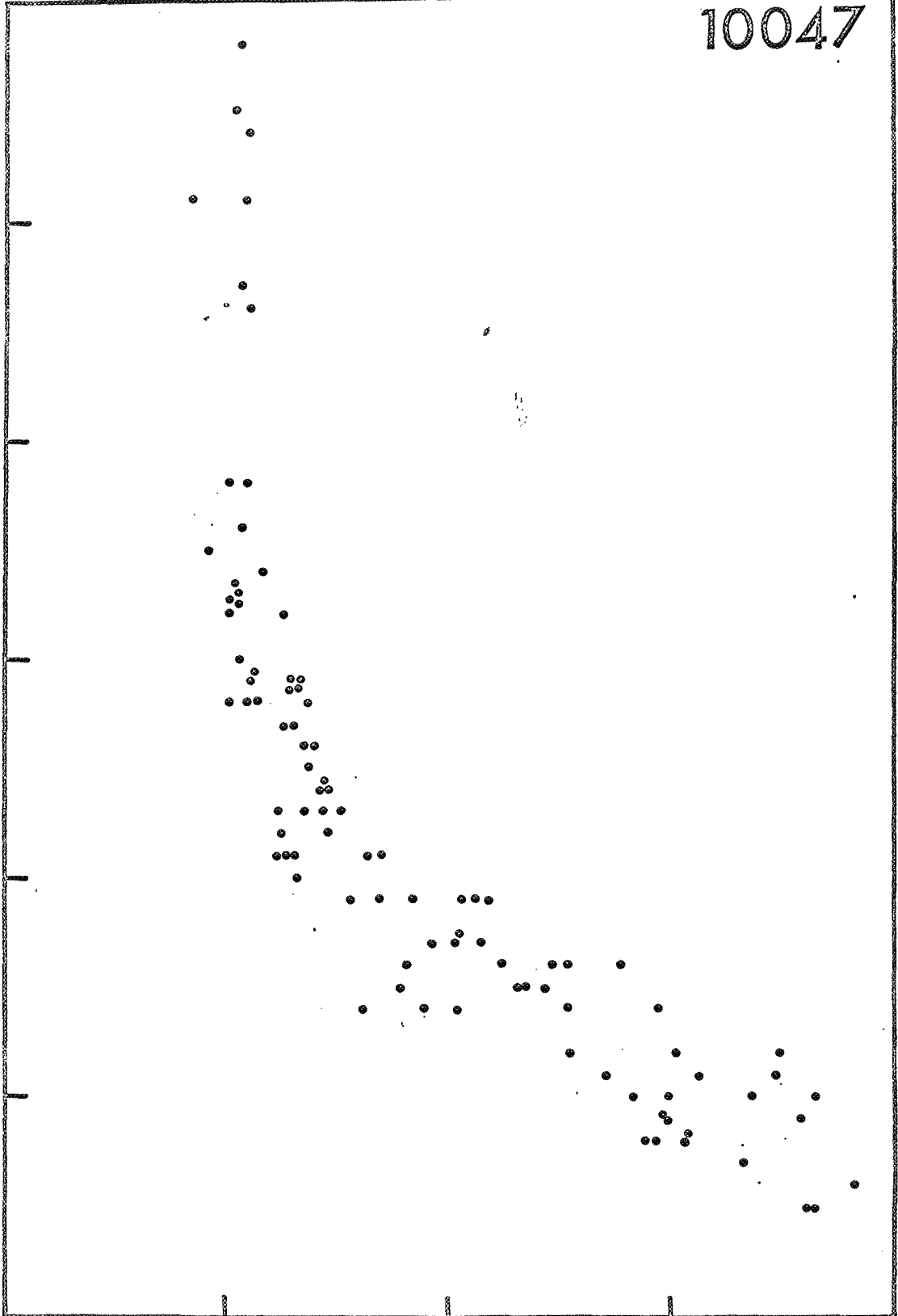
(After Essene et al, 1970a)

The chromium contents of most Apollo 11 rocks fall between

rather narrow limits (2000 $\pm$ 700 ppm Cr). In Figure 2 (Essene

et. al., 1970) the Cr_2O_3 contents of pyroxenes from rock 10047

are plotted against corresponding $\frac{Mg}{Fe + Mg}$ ratios. The strong decrease

of chromium with increasing degree of differentiation as indicated

by $\frac{Mg}{Mg + Fe}$ ratios is notable. The initial sharp decrease from 0.6

to 0.3% Cr_2O_3 is probably caused by crystallization of armalcolite,

spinel and ilmenite from the magma. (Table 1) These minerals

commenced crystallization before pyroxene. The continuous depletion

from Cr_2O_3 values of 0.3 to less than 0.1 is probably caused by

pyroxene fractionation (pyroxene-liquid partition coefficient 3.5)

combined with late separation of plagioclese.

These results, particularly the sharp decline between $\frac{Mg}{Mg + Fe}$

ratios of 0.74 and 0.70 apparently caused by separation of spinel

and armalcolite, demonstrate the large changes of chromium contents

which are caused by moderate degrees of crystal-liquid fractionation.

It is also significant that the chromium content of most Apollo 11

rocks falls within such a narrow range despite the fact that these

rocks have been subjected to very different cooling histories as

evidenced by textural variations from basaltic to gabbroic. (LSPET,

1969).

(b) Fractionation Mechanisms: Conclusions and Some Consequences

From evidence discussed earlier and also from additional considerations (Ringwood and Essene, 1970b) it is concluded that the characteristic abundance patterns of Apollo 11 rocks were not established primarily by near-surface crystallization differentiation processes. This does not rule out the possibility that a small amount of near-surface fractionation, probably less than 10 percent in Apollo 11 rocks (and perhaps somewhat greater in Apollo 12 rocks) might have occurred. This, however, would have a negligible effect on the abundances of incompatible elements. It seems clear that the high abundances of this group of elements is a primary character-istic of the Apollo 11 magma, and can be explained only if the magma were formed by a small degree of partial melting of appropriate source material. (Section 2d) Ringwood and Essene (1970b) demonstrated that the partial melting hypothesis explained the major element abundances of Apollo 11 basalts and furthermore, that the composition and minerology of the source region, as derived experimentally, provided an explanation of the moon's density and moment of inertia.

(i) Impact melting hypothesis

It has been suggested (eg. Urey, 1952; Öpik, 1967) that the maria were formed by impact melting caused by the collisions of

large planetesimals with the moon. It is difficult to reconcile

this process with the highly differentiated nature of Apollo 11

lavas and the conclusions that the differentiation was not the

result of near-surface fractionation, but on the contrary was

caused by partial melting within the lunar interior. (Sections

2a,d). If impact events led to shock melting, it would be expected

that the melting was confined to near-surface regions and would

be non-selective and total (where it occurred). It is difficult

to envisage impact leading directly to partial melting in the

interior. It follows that if impact melting is assumed, the surface

rocks must have attained their highly differentiated nature prior

to impact. The impact hypothesis does not, therefore, explain the

observed chemistry of Apollo 11 lavas; it relegates the problem to

an earlier episode of magma genesis, which as we have seen, almost

certainly involved partial melting.

Impact melting on the scale necessary to produce the maria

is a highly speculative process which has yet to receive an adequate

theoretical or experimental justification. The conditions which

are required if the kinetic energy of an impacting planetesimal

is to be transformed efficiently into thermal energy, resulting in

the relatively uniform and restricted degree of heating required

to cause melting throughout a large volume, are very difficult to fulfil. For this reason, advocates of impact melting, (eg. Öpik 1967, 1969) conceded that the process requires as a precondition that the surface regions of the moon were already very close to the melting point before the impacts occurred.

However, this requirement is contradicted by the age of the Apollo 11 basalts which are about one billion years younger than the age of the moon. Even if the moon were originally at high temperature, thermal calculations employing a wide range of boundary conditions (eg Urey, 1962) show that deep-seated cooling would occur to a depth of about 200 km, over a period of 10^9 years and that the mean temperature of the outer 200 km of the lithosphere would be about $500^\circ C$. This is not consistent with Öpik's requirement for impact melting. We conclude that this process appears most improbable as an explanation of the maria.

The above arguments are not directed against the suggestion (eg. Ronco, 1966) that impacts did not form the lavas directly, but "triggered off" the subsequent volcanic activity by some sort of tectonic action (eg. pressure release or thermal blanketing). Such models are entirely consistent with deriving the lavas by partial melting processes from a source region within the lunar interior.

(ii) Total melting of moon

Smith et. al., (1970 a,b) propose that the moon was formed from chondritic material (depleted in iron and volatile metals), after which it was completely melted. The entire interior was subjected to crystallization differentiation and the Apollo 11 basalts are regarded as the residuum of extensive and prolonged near-surface crystallization differentiation. Their hypothesis is the very antithesis of the one presented by Ringwood and Essene (1970b).

The first problem encountered by the Smith et. al., model is the mechanism which caused complete melting of the moon. This is dismissed in one sentence "There is no difficulty in finding sources of energy to melt the moon (radioactive heating, gravitational infall energy, compression)". It is a simple matter to show that the second factor is capable only of melting the outer regions of the moon, whilst the third factor is trivial. For, radioactive heating, the ad hoc assumption must be made that a short lived nuclide, eg. Al^{26} was present in sufficient amount. However, a chondritic moon heated internally by radio-activity is incapable of complete melting. The melting interval of chondritic material is about 500° C. Melting commences at about $1100-1200^{\circ}C$ and as soon as a significant amount of partial melting has occurred.

throughout the interior, probably less than 10 percent, convection will occur accompanied by segregation of basaltic liquid from its source region, forming a magma which rises to the surface. Since the heat source is strongly partitioned into the liquid, further heating of the depleted source region is greatly reduced. Transfer of heat sources to near-surface regions by this type of differentiation permits the radioactive heat to be removed by conduction.

Smith et. al., account for the undersaturation of the residual liquid with plagioclase by postulating that the parent material of Apollo 11 basalt consisted of a suspension of pyroxene and ilmenite crystals in a residual liquid which was at the plagioclase—pyroxene—ilmenite cotectic. The unmelted crystals had settled from an overlying body of crystallizing liquid. The suspension of crystals and liquid was then completely remelted by subsequent meteoritic impact to form the Apollo 11 magma.

Whilst this sequence of events may be capable in principle of explaining the delayed crystallization of plagioclase, it is highly contrived and entirely lacking in supporting evidence.

The model encounters a fatal difficulty in connection with the proposed time scale. If the moon were extensively melted 4.7 billion years ago, it would solidify from the core outwards in a few million years at most, depending upon the assumed heat source (e.g. Al^{26}).

Smith et. al., maintain that fractional crystallization ultimately led to the formation of a thin plagioclase-rich crust, 5 to 20 km thick underlain by residual liquid magma containing suspended pyroxene and ilmenite crystals. Meteoritic impacts 3 $\cdot$3$\cdot$7 billion years ago caused total melting of this magma-crystal mush to form Apollo 11 basalts.

The time constant for thermal conduction in a crust up to 20 km thick is much shorter than 10^8 years. According to Smith et. al.'s model, the liquid magma which was the parent of Apollo 11 lavas remained unfrozen at a depth of 20 km for 10^9 years waiting to be tapped by meteoritic impact. This is simply out of the question.

Smith et. al.'s hypothesis is subject to most of the earlier objections which have been raised against fractional crystallization. To account for the incompatible trace element concentrations in Apollo 11 basalts, from 95 to over 99 percent of a parental melt possessing chondritic abundances would need to have crystallized. At least half of the crystallizing phases would be pyroxenes, the remainder being *mostly* olivine. The partition coefficients for chromium and vanadium in pyroxene and olivine (Table 1) show that this amount of prior crystallization would have effectively removed Cr and V from the residual liquid. This does not agree with the observed abundances (2000 ppm Cr, 75 ppm V) in Apollo 11 basalts. Smith et. al.'s hypothesis fails entirely to explain the wide variations

in abundances of incompatible elements observed within Apollo 11

basalts (and also within Apollo 12 basalts — Figure 1) as contrasted

with the very restricted degree of major element fractionation

between these rocks as indicated by $\frac{Mg}{Mg + Fe}$ ratios.

Mascons were presumably formed at about the same time as the

maria, and their survival implies the existence of a strong cool,

lithosphere underlying the maria at the time they were formed

(Urey, 1969). According to Smith et. al.'s hypothesis, the outer

regions of the moon were still very hot and weak at the time of

mare formation. Accordingly they do not provide a framework for

understanding the continued existence of mascons.

(c) Depth of Origin

In this section, we adopt the hypothesis, following earlier

discussion, that Apollo 11 basalts have formed by partial melting.

The following lines of evidence indicate that the source region was

probably at a depth greater than 200 km.

(i) Apollo 11 basalts were generated one billion years after

formation of the moon. We have already noted (P -   ) that over

this period, extensive cooling would occur to depths of about 200 km

and the mean temperature of the outer 200 km is unlikely to exceed

500° C. Cooling of the moon is effective to greater depths than in

the earth because its surface area to volume ratio is nearly four times higher than the earth's.

It is difficult to understand how magmas might be formed by partial melting processes in this outer cool shell. On the other hand, radioactive heating at depths greater than 200 km where substantial conductive heat-loss had not occurred, might readily lead to partial melting (Section 3).

(ii) The mascons were presumably formed either before, during, or soon after the formation of the respective maria in which they occur. Their continued existence for 3.7 billion years implies the presence of an underlying lithosphere possessing substantial long-term strength (Urey, 1969; Urey and MacDonald, 1970). An origin for the lavas by partial melting within the lithosphere implies loss of strength and destruction of the lithosphere. Preservation of mascons would not be explicable.

(iii) Incompatible elements are present in Apollo 11 basalts at levels ranging from 30 to 100 times the chondritic abundances. If this range of compositions is characteristic of maria, it implies that maria have been derived by partial melting of volumes 30 to 100 times greater than their own, assuming chondritic abundances in the source material. The thicknesses of maria may range from

10 to 50 km (Baldwin, 1963; O'Keefe, 1968). These relationships imply extensive fractionation of much of the moon's deep interior and are not explained by near-surface melting processes. (Baldwin, 1970).

(iv) The assumption that the lunar highlands are anorthositic implies that complete differentiation of the outer 150 km of the moon has occurred (Section 3). This differentiation must have occurred long before the period of mare formation, and most probably during or soon after the formation of the moon. It is difficult to understand the high absolute abundances of incompatible elements and the chondritic-type abundance patterns (Eu excepted) in Apollo 11 basalts in terms of a second phase of partial melting of a highly differentiated outer crust (Figure 5) in which the incompatible elements are strongly depleted and fractionated. The source regions of Apollo 11 basalts must lie in relatively undifferentiated material, below the level reached by the early anorthositic phase of different-iation (Figure 5).

(v) Ringwood and Essene (1970 a,b) have determined the melting relationships of Apollo 11 basalts up to pressures equal to those reached at a depth of 700 km. They concluded that Apollo 11 basalts

were unlikely to have been generated at depths less than about
100 km.

(d)   Nature of Source Region

Two principal methods have been used to obtain information on
the chemical composition and mineralogy of the source region of
Apollo 11 basalts. One was based upon high pressure-high temperature
investigations of phase relationships, and conclusions were drawn on
the basis of major element chemistry. The other was based upon an
interpretation of the trace element geochemistry of Apollo 11 lavas
in terms of partition coefficients with possible source materials.
The conclusions reached by these two methods have been contradictory.

(i)   Partial melting of pyroxene ± olivine source region

Ringwood and Essene (1970 a,b) carried out an extensive
high pressure-high temperature investigation of melting equilibria
displayed by an average Apollo 11 basalt composition in which
compositions of near-liquidus phases were determined with the
electronprobe microanalyzer. (Figure 3). Three principal
pressure regimes were recognized: (i) a low pressure regime
in which the liquidus phases were olivine and armalcolite, (ii)
an intermediate pressure regime in which the liquidus phase was
a sub-calcic clinopyroxene,   (Table 2), and (iii) a high
pressure regime in which liquidus and near-liquidus phases

Fig. 3. Stability fields of mineral assemblages and melting

equilibria in average Apollo 11 basalt composition at high

pressures and temperatures. Each dot represents a separate

experiment. Note: Armalcolite is the approved mineral name

corresponding to the term "karooite" used on this figure.

MODEL LUNAR BASALT (APOLLO 11)
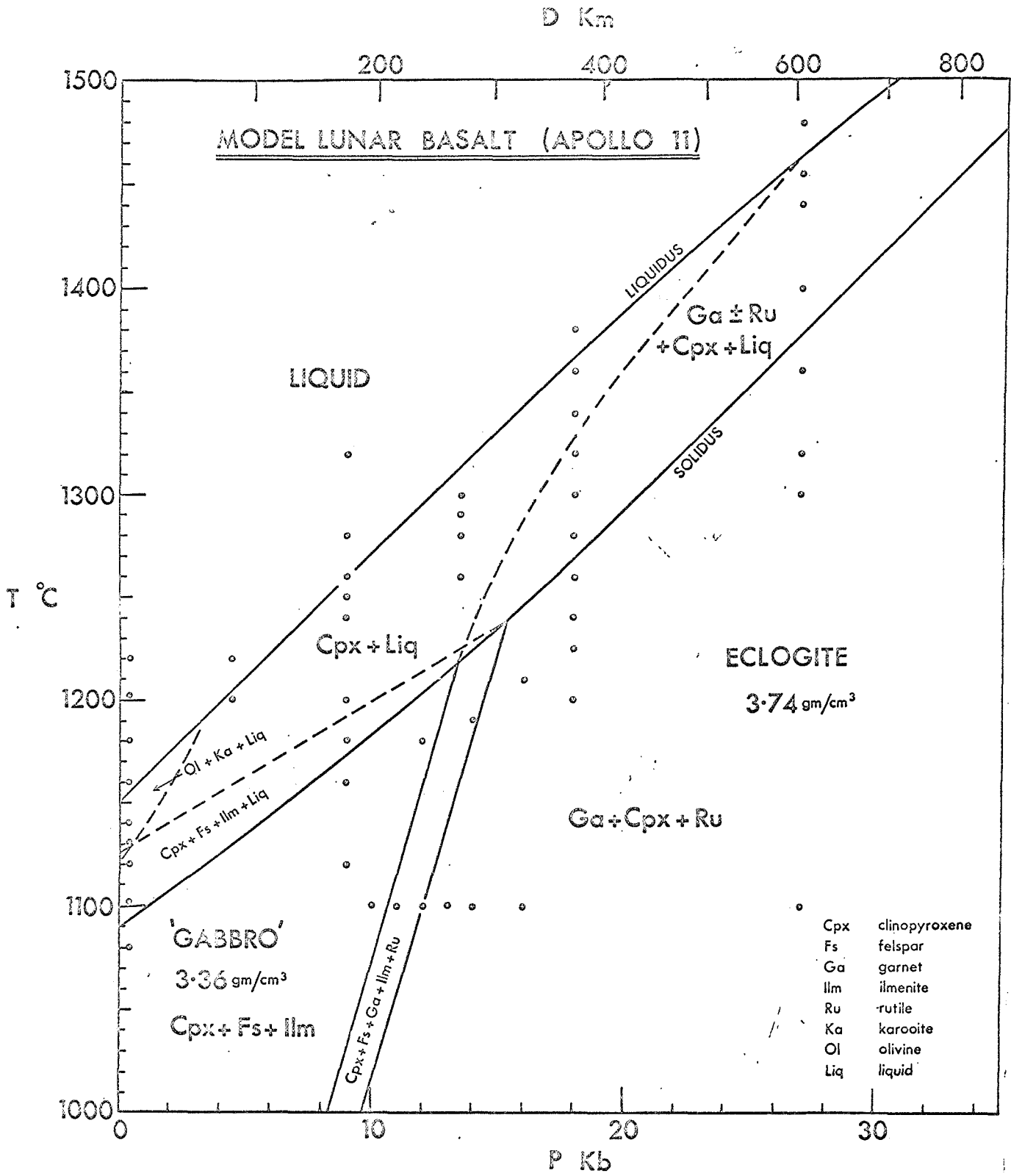
were clinopyroxene and garnet (eclogite). Since the equilibrium between crystals and magma is independent of the proportions of either phase present, the compositions of the liquidus and near-liquidus phases so determined represent possible bulk compositions for the source regions from which the magma was derived by partial melting. It was possible to test these derived source compositions for different depth intervals to see whether they were capable of explaining the moon's mean density and moment of inertia. This test eliminated the low pressure and high pressure regimes as likely sources; whereas the intermediate pyroxenite regime provided a reasonable match. Kushiro et. al., (1970) and O'Hara et. al. (1970 a,b) also concluded that the parent of the Apollo 11 magma might be derived by partial melting of a clinopyroxenite source material at depth, but did not specify the nature or composition of the clinopyroxene.

Ringwood and Essene (1970 a,b) also demonstrated that between 10 and 20 kb, Apollo 11 basalt was very nearly saturated with orthopyroxene (Table 2), suggesting that this phase may also have been present in the source region. Clinopyroxene alone did not appear to be a suitable source since its Al_2O_3/CaO ratio of 0.53 was substantially lower than the value of 1.2 ± 0.1 which was characteristic of all chondrites, eucrites and howardites (Ahrens and Michaelis, 1979) and probably also of the earth. Because of

Table 2

Compositions of model lunar pyroxenite and of Apollo 11 basalt

near-liquidus pyroxenes (After Ringwood and Essene 1970b)

	(1)	(2)	(3)
SiO_2	52.0	50.6	53.9
TiO_2	1.0	2.1	1.0
Al_2O_3	5.0	3.8	2.9
Cr_2O_3	0.4	0.9	0.6
FeO	13.5	12.8	12.2
MgO	22.5	22.4	26.4
CaO	4.0	7.2	2.8
Na_2O	0.1	0.1	—
$\dfrac{100\ Fe}{Fe\ +\ Mg}$	25	24	20

(1) Model lunar pyroxenite capable of yielding Apollo 11 basalt by small degree of partial melting.

(2) Av. sub-calcic clinopyroxene on liquidus of Apollo 11 basalt in 5-20 kb interval.

(3) Orthopyroxene 10kb, $1240^{O}C$, near liquidus of Apollo 11 basalt (modified by slight iron-loss).

the very low and similar volatiles of Al_2O_3 and CaO, and the wide-

spread belief (eg. Gast et. al., 1970, Haskin et. al., 1970) that the

relative abundances of other elements possessing oxides of low vola-

titility in the source regions of Apollo 11 basalts are similar to

their relative abundances in chondrites, it is reasonable to assume

that the overall Al_2O_3/CaO ratio in the moon is similar to that in

meteorites and in the earth. In the light of the experimental data

cited above, this suggests strongly that orthopyroxene possessing

a much higher Al_2O_3/CaO ratio than the clinopyroxene $(Table\ 2)$ was probably

also present in the source region. Using the compositions of sub-

calcic clinopyroxene and orthopyroxene demonstrated to be in

equilibrium or close to equilibrium with Apollo 11 basalts at 10-

20 kb, a model lunar pyroxenite representing a possible source rock

composition was synthesized. (Table 2). Mineral stability fields in

the possible source rock were determined over a wide range of P, T

conditions (Fig. 4). It was demonstrated that for the probable range

of temperature distributions existing in the lunar interior, this

composition, with only very slight modification, was capable of

explaining the moon's observed density and moment of inertia. The

composition of the source region thus derived was not unique, to

the extent that variable amounts of olivine could also be present

in addition to the pyroxenes. This, however, did not affect the

Fig. 4. Stability fields and densities of mineral assemblages

displayed by model lunar pyroxenite (Table 2) in relation

to probable range of lunar internal temperature distributions

(Urey, 1962; MacDonald, 1959; Urey and MacDonald, 1970;

Fricker et al 1967; Levin, 1960; McConnell et al, 1967). The

broken lines in the garnet pyroxenite field indicate the

Al_2O_3 contents (specified in weight percent) of pyroxenes in

equilibrium with garnet. They also indicate the respective

densities of the garnet pyroxenite assemblages as a function

of pressure and temperature.
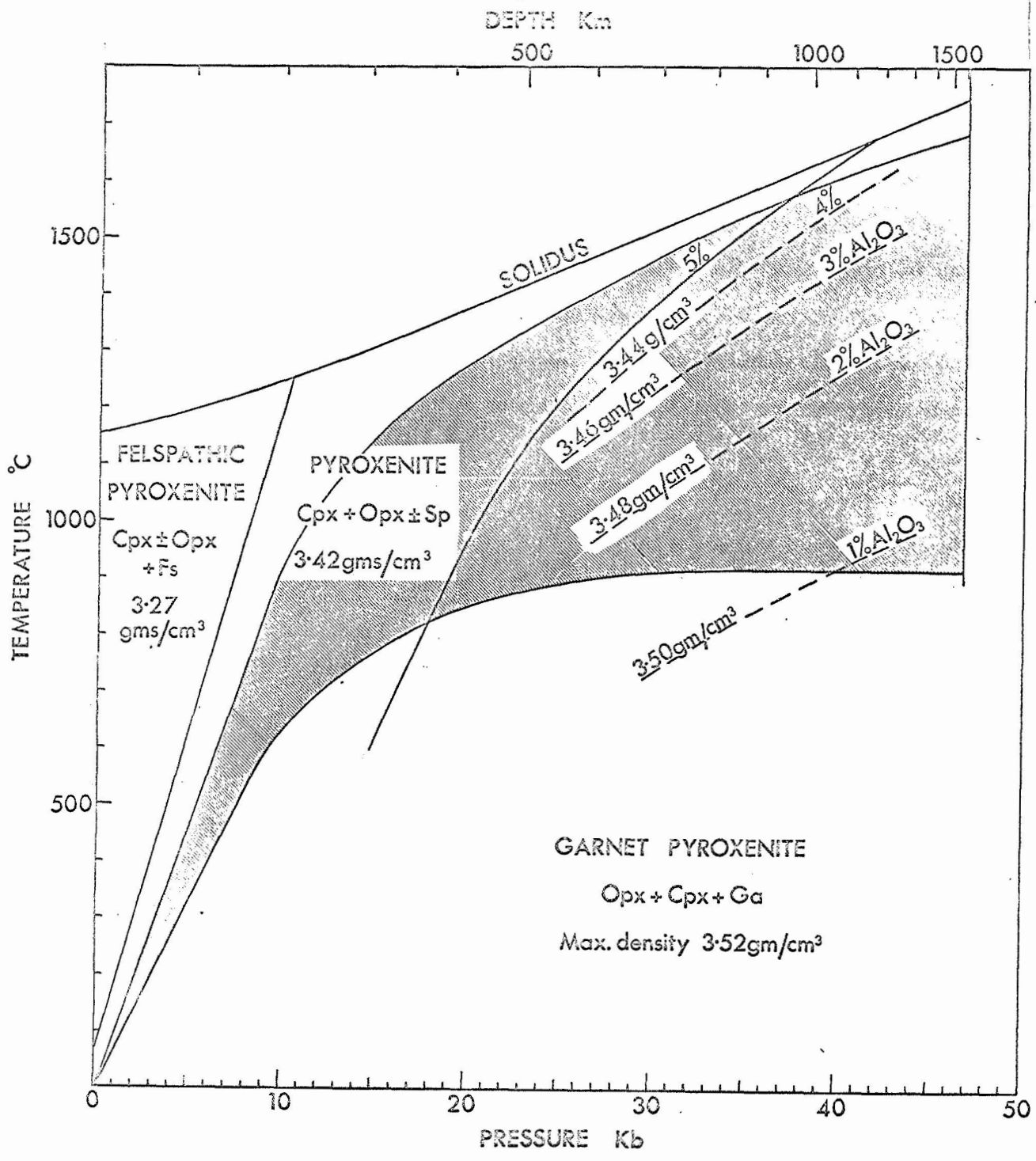
ability of the source material to explain the moon's density and moment of inertia.

The capacity of the above model of the lunar interior to provide explanations both of the major element chemistry of Apollo 11 basalts, and of the principal physical properties of the moon, was most satisfying. Was it also possible to explain the trace element distributions in Apollo 11 basalts in terms of the model?

Ringwood and Essene (1970b) argued that with a very small degree of partial melting, the incompatible elements would be strongly partitioned into the liquid phase and their high abundances in Apollo 11 basalts thereby explained. On the contrary, Gast et. al., (1970), Haskin et. al., (1970) and Philpotts and Schnetzler (1970) showed that if measured partition coefficients for the rare earths were used, it was not possible to explain their high concentrations in Apollo 11 basalt in terms of a single stage of partial melting from parent material possessing chondritic abundances. They argued that the parent material should possess abundances several times higher than chondrites. Also it was argued that the partial melting of a pyroxenite source region was unlikely to produce the observed europium anomaly. Finally, Smith et. al., (1970) claimed that the Ringwood and Essene partial melting model could not explain the high titanium content of the Apollo 11 rocks. (Titanium is enriched to a comparable degree to Ba, U, and the rare

earths, and is, therefore, behaving as an incompatible element).

These are weighty objections, which must indeed be met if the model is to survive. I suggest that the answer lies in the direction of a suggestion by Goles et. al., (1970). They point out that most of the incompatible elements in Apollo 11 rocks occur not dispersed throughout the major crystalline phases, as assumed in the equilibrium partition calculations used by Gast, Haskin and Philpotts, but in late stage accessory minerals often occurring at grain bounderies. It is considered probable, that in the source material of Apollo 11 basalts, a similar distribution of incompatible elements may occur. It seems entirely conceivable that with very small degrees of partial melting, conditions of surface or local equilibrium only, rather than bulk equilibrium, are achieved. Under the circumstances, the accessory minerals containing the incompatible elements (eg. zircon, whitlockite, geikielite) are completely or extensively incorporated into the earliest liquid to form, which accordingly is strongly enriched in these elements. Increasing degrees of partial melting serve only to dilute the earliest liquid with major components so that the abundances fall. (eg Figure 1.). This model appears capable of explaining the high enrichments of incompatible elements in Apollo 11 basalt and also the high titanium content, which requires that titanium was also present in an accessory mineral(geikielite) in the source regions. (Alternative

explanations of the titanium abundance are also possible--
(Ringwood and Essene, 1970b).

We come now to the celebrated europium "anomaly" (Figure 1).
It would be remarkable if such an "anomaly" did not exist. At the
redox state of Apollo 11 basalt, most europium occurs dominantly
as Eu^{2+} whereas under the more oxidizing conditions in terrestrial
systems, most europium occurs as Eu^{3+}. This is shown by the
experimental plagioclase-liquid partition coefficient for europium
of 1.0 for Apollo 11 basalt (Table 1) compared to values in the
range 0.06 to 0.4 in terrestrial rocks (Philpotts and Schnetzler
1970, Haskin et. al., 1970). The Eu^{2+} ion has crystal chemical
properties resembling Sr^{2+} and quite different from those of the
neighboring trivalent rare earths Sm^{3+} and Gd^{3+}. Accordingly,
the partition coefficient of Eu^{2+} is expected to be more similar
to that of Sr^{2+} (Table 1), than to those of Sm^{3+} and Gd^{3+}.

Ringwood and Essene (1970b), Essene et. al., (1970b) and Goles
et. al., (1970) suggested that most of the europium in the source
regions of lunar basalt occurs as a dispersed element replacing
Ca^{2+} in sub-calcic clinopyroxene and orthopyroxene. On the other
hand, they suggested that most of the trivalent rare earths occur in
accessory minerals, which accordingly possess a negative europium
anomaly. As discussed above, a small degree of partial melting

leads to preferential incorporation of the accessory minerals into the liquid which accordingly has a negative europium anomaly. The behaviour of europium in lunar magmas is thus considered to be analogous to that of strontium in terrestrial magmas which is depleted relatively to Ba^{2+} when alkali basalts form by small degrees of partial melting in the earth's mantle (Gast, 1968a).

The above explanation of the behaviour of europium is no more than a hypothesis which remains to be tested by experimental determinations of the Eu^{2+} and Sm^{3+} partition coefficients under the correct redox state. When these data become available, a key test of the Ringwood-Essene partial melting hypothesis will be possible. In the absence of these data, it is not possible to claim that the hypothesis is incapable of explaining the trace element abundances in Apollo 11 basalts.

(ii) Partial melting of plagioclase-bearing source region

Haskin et al., (1970), Gast and Hubbard (1970), and Philpotts and Schnetzler (1970) concluded that the trace element abundances in Apollo 11 basalt were more readily explained in terms of partial melting of a source material containing plagioclase, than by extensive fractional crystallization of felspar from a parent liquid. The presence of plagioclase was believed to be necessary in order to explain the europium anomaly. These authors found it necessary to postulate a source material enriched in rare earths, Ba, U, Th, and Sr over the chondritic abundance by a factor of 5 to 20. The

major phases in the source region were pyroxenes (Mg-rich)

plagioclase and olivine. The mineralogy were something like that

of a eucrite. This model thus implied a two stage magmatic history

for the moon,(i)Initial melting and differentiation leading to the

formation of an outer shell possessing an overall eucritic (basaltic

achondrite) composition. This probably occurred during the moon's

formation.(ii)Partial melting in this outer shell a billion years

later caused the formation of Apollo 11 basalts.

This model has two important attributes. It provides a ready

explanation of the europium anomaly, and it is capable of providing

an explanation of the nature of the lunar highlands (assuming them

to be anorthositic -- section 3).

The model also encounters some serious difficulties:

(i.) Apollo 11 basalt is not saturated with plagioclase

which does not appear until 30-50% of the rock has crystallized.

If Apollo 11 basalts had formed by partial melting of a

source rock in which plagioclase remained in the refractory

residiuum after the melting episode, then the magma is

necessarily saturated with plagioclase which must appear

on the liquidus. (Ringwood and Essene 1970b). Appealing

to a pressure effect does not help since pressure acts in

the opposite direction to that required (Section 2a).

(ii) The hypothesis fails to explain the depletion of

light rare earths in some lunar basalts (Gast and Hubbard,

(1970); Figure 1, Rock 10020). Likewise, it does not

readily explain the lack of fractionation between Ba

and Yb (Figure 1) in view of the substantial difference

in their plagioclase-liquid partition coefficients (Table 1).

(iii) It is difficult to understand how, even at

solidus temperatures, plagioclase can exist as a phase

at depths below 200 km. If the entire eucritic fraction

were segregated from a moon possessing the chondritic

abundances of non-volatile oxyphile elements, it would

form an outer shell less than 150 km thick. If a less

efficient degree of segregation is assumed—for example,

in the parental mineral assemblage proposed by Gast et.

al., 1970. (Olivine 40%, clinopyroxene 32%, plagioclase

22%, spinel 6%), plagioclase would not remain a stable

phase at the solidus at a depth greater than about 200 km.

(Moreover, the concentrations of incompatible elements

in this layer would then be too low to be consistant with

the Haskin-Gast partial melting models).

Models of this type encounter the problem of explaining the

the formation of Apollo 11 magma by a partial melting process in

the outer 200 km of the moon, i.e. in the lunar lithosphere, at a

period one billion years later than the formation of the moon.

The difficulties of accomplishing this objective were discussed

in section 2c where it was concluded that the Apollo 11 lavas must

have been derived from deeper than 200 km—i.e., from a region in

which plagioclase is no longer stable.

Gast et. al., (1970) and Haskin et. al., (1970) conclude from

measured rare earth partition coefficients that the source material

of Apollo 11 basalts, was strongly enriched in rare earths compared

to chondrites presumably owing to an earlier differentiation process.

They are a little vague about the degree of enrichment, but it can

not be inferred from their studies that to generate the Hi-Rb basalts,

it must have been between 5 and 20 fold. Taking the lowest estimate,

this would require total fractionation of rare earths from the entire

moon (assumed to possess chondritic abundances) into an outer shell

less than 150 km thick. If the source material contained more than

five times the chondritic abundances, the shell would be correspoond-

ingly thinner. The difficulties of producing Apollo 11 basalts

by partial melting in such a thin outer shell have already been

remarked.

Dickey (1970) has suggested a compromise between the two

hypotheses of lunar basalt origin mentioned above. He suggests,

in agreement with Gast and others, that early melting processes in

led to the formation of an outer shell of overall eucritic composition. His model which uses chondritic abundances, would not permit this shell to be more than 150 km thick. The eucritic shell further differentiates into a plagioclase-rich upper layer (anorthostic gabbro) and a pyroxene-rich lower layer, which consists mainly of cumulates. Apollo 11 basalts were subsequently formed by a small degree of partial melting of the pyroxenite lower layer.

This model accounts satisfactorily for the europium deficiency but encounters other problems. It is very difficult to explain the occurrence of widespread partial melting in the lunar lithosphere one billion years after the formation of the moon, when the outer 200 km had cooled extensively by conduction (Section 2c). Furthermore, the pyroxenite source region envisaged by Dickey is a cumulate in which the incompatible elements would be strongly depleted and fractionated among themselves compared to the parent magma. It is difficult to explain the high abundances of incompatible trace elements in Apollo 11 basalts and their limited relative fractionations (eg. Yb/Ba in Figure 2) compared to chondritic abundances by partial melting of such cumulates. Haskin et. al., (1970) have pointed out that the lack of strong fractionation of heavy from light rare earths in Apollo 11 basalts implies that high calcium clinopyroxene was not an abundant phase in the source region. (Section 2d, iv).

(iii) Discussion

In my opinion, the balance of arguments discussed above (2d (i) and (ii)) strongly favours the hypothesis that Apollo 11 basalts have formed by a small degree of partial melting of a pyroxene ± olivine source region at depths of 200-600 km and have ascended to the surface without extensive fractionation en route. This hypothesis provides a satisfactory explanation of their major element chemistry; it is capable of explaining most aspects of the trace element chemistry; it explains the density and moment of inertia of the moon, and is consistent with thermal history considerations. Furthermore, the process advocated is of a type which is known to occur widely on the earth—the alkali basalt suite, particularly the nephelinites, are believed to have formed by a very small degree of partial melting, comparable to that suggested for Apollo 11 basalts, in the earth's mantle, and display similar degrees of enrichment of incompatible elements such as Ba, U, and the light rare earths. Segregation of Apollo 11 basalt after only a very small degree of partial melting would be facilitated by its low viscosity which is only about a tenth of that of terrestrial basalt (Weill et. al., 1970). It is not yet demonstrated that the above partial melting hypothesis can explain the europium anomaly, nor has it yet been demonstrated that it is incapable of providing an explanation. This will provide a critical future test of the hypothesis.

On the other hand, the hypothesis of partial melting of an outer shell of previously differentiated eucritic material, whilst capable of explaining the europium anomaly, encounters a number of fundamental objections, two of which do not appear soluble even in principle.

In section 5, several boundary conditions for the origin of the moon will be derived on the assumption that the first hypothesis is correct. It is important to note that some of these boundary conditions could also be derived from the second hypothesis. Both have important attributes in common, requiring an early near-surface melting and differentiation process to produce the (presumed) anorthositic highlands, followed by a later extended phase of partial melting of internal origin to produce the maria. The mean lunar chemical compositions and oxidation states in both models are similar; the debate rests chiefly upon the depth at which the second phase of partial melting occurred and the thickness of the outer shell which was subjected to the primordial differentiation. One of the most definite conclusions about lunar history which can be drawn at the present stage is that the moon during the first billion years or so was subjected to extensive high-temperature crystal-liquid fractionation processes which must have affected a large proportion of its mass. The proposition that the moon was born cold, and has never been heated entensively to near the melting point except, perhaps in a thin,

near-surface zone (Urey, 1969) cannot possibly be sustained.

(iv)   Source Regions of Lunar and Terrestrial Basalts

Here we shall be concerned with some comparative aspects of the major element chemistry and mineralogy of the earth's mantle and the lunar interior. The coefficient of moment of inertia, I/MR^2, of the moon, $0.402 \pm .002$ (Kaula, 1970b), is very similar to that of a sphere of uniform density (0.400). To a first approximation, thermal expansion and compressibility cancel each other out in the moon. The above similarity thereby implies that providing ad hoc complex compensating structures are not introduced, phase transformations leading to mineral assemblages of higher density with increasing depth can play only a very minor role in the moon. The restriction that major phase changes in the lunar interior are not tolerable is an important boundary condition for the composition of this region. Ringwood and Essene (1970 a,b) demonstrated that for all practical purposes, this limits the mean Al_2O_3 and CaO contents of the lunar interior to less than 6 percent each, and that the moon, like the earth's mantle, is dominantly composed of iron-magnesium silicates.

There are, however, some important differences between the lunar interior and the earth's mantle. The $\dfrac{MgO}{MgO + FeO}$ ratio of the lunar

interior is probably between 0.74 and 0.80, compared to 0.88 in the
terrestrial interior (Ringwood and Essene, 1970b). Moreover, the
oxidation state of the lunar interior, being characterized by the
coexistence of metallic iron with ferromagnesian silicates, is much
lower than of the terrestrial mantle in which metallic iron is generally
absent and ferric iron occurs as a significant component of most min-
erals (Ringwood and Essene, 1970 a,b; Ringwood 1966a, 1970a).
Whereas olivine is generally believed to be the dominant mineral in
the earth's upper mantle, liquidus phases of Apollo 11 basalts
suggest that Mg-rich pyroxenes may predominate in the moon, although
the presence of olivine in the lunar source regions is not excluded.
(Ringwood and Essene, 1970b).

More subtle, but equally important differences exist. The Group
I Apollo 11 crystalline rocks contain on the average, less than 9%
of Al_2O_3 compared to the usual range of 12 to 18% Al_2O_3 in comparable
terrestrial basalts. This suggests that the (thermodynamic) activity
of Al_2O_3 in the lunar interior is less than in the earth's mantle.
Green and Ringwood (1967) have shown that aluminous pyroxenes play a
major role in basalt genesis in the earth. The low alumina content
of lunar basalts combined with the observations that the alumina
contents of clinopyroxenes on the liquidus of lunar basalts are much
poorer in Al_2O_3 than are the clinopyroxenes on the liquidus of terrestrial

basalts at similar pressures (Ringwood and Essene, 1970b; Green

and Ringwood, 1967) imply that the alumina contents of pyroxenes in

the source regions of lunar basalts are correspondingly smaller than

in the source regions of terrestrial basalts. This could be explained

by postulating a similar absolute amount of Al_2O_3 in the earth and

moon, but with the lunar interior being much richer in pyroxene than

the earth's mantle, so that the ratio of Al_2O_3 to pyroxene is lower

in the moon. Alternatively, if the less likely postulate is made

that olivine is an abundant phase in the lunar mantle, the amount

of pyroxene in the lunar mantle is correspondingly reduced without

altering the Al_2O_3 content of the pyroxene, so that the absolute

abundance of Al_2O_3 in the lunar interior is then smaller than in

the earth's mantle. Either way, we are obliged to assume that the

$\underline{\text{Pyroxene}}$ and/or $\dfrac{Al_2O_3}{MgO + FeO + SiO_2}$ ratios differ between lunar interior and
Olivine

terrestrial mantle. If we accept the earlier argument that the

CaO/Al_2O_3 ratios in earth, chondrites, achondrites, moon and sun

are the same an analogous conclusion follows—namely that the average

calcium content of pyroxenes in the source regions of lunar basalt

is smaller than in the source regions of terrestrial basalt.

Further evidence that the average $\dfrac{CaO}{MgO + FeO}$ ratio of pyroxenes

in the lunar mantle is less than in the earth's mantle comes from

rare-earth distributions.   In Apollo 11 basalts, the light *and heavy* rare earths

are greatly enriched but by about the same amounts, and the distribu-

tions are approximately parallel to the chondritic pattern. (Figure 1)

However, large differences exist between the partition coefficients

for heavy and light rare earths in the high-calcium clinopyroxenes

(approximately 16% CaO—Green and Hibberson, 1970) which are present

near the solidus in the earth's mantle (Gast, 1968; Haskin et. al.,

1970), and the heavy rare earths, eg. Yb, are not strongly discrimin-

ated against so that they do not behave as incompatible elements.

Under these circumstances, a small degree of partial melting pro-

duces a highly skewed rare earth pattern with strong relative enrich-

ment of the light rare earths, which is characteristic of terrestrial

basalts formed in this manner. (Gast, 1968).   The absence of skewing

in lunar basalts implies a much smaller amount, or absence, of

Ca-rich clinopyroxene in the lunar source region (Haskin et. al.,

1970).   On the other hand, the heavy rare earth partition coefficients

in Mg-rich pyroxenes are much smaller and permit these minerals to

be present in the source regions (Haskin et. al., 1970).
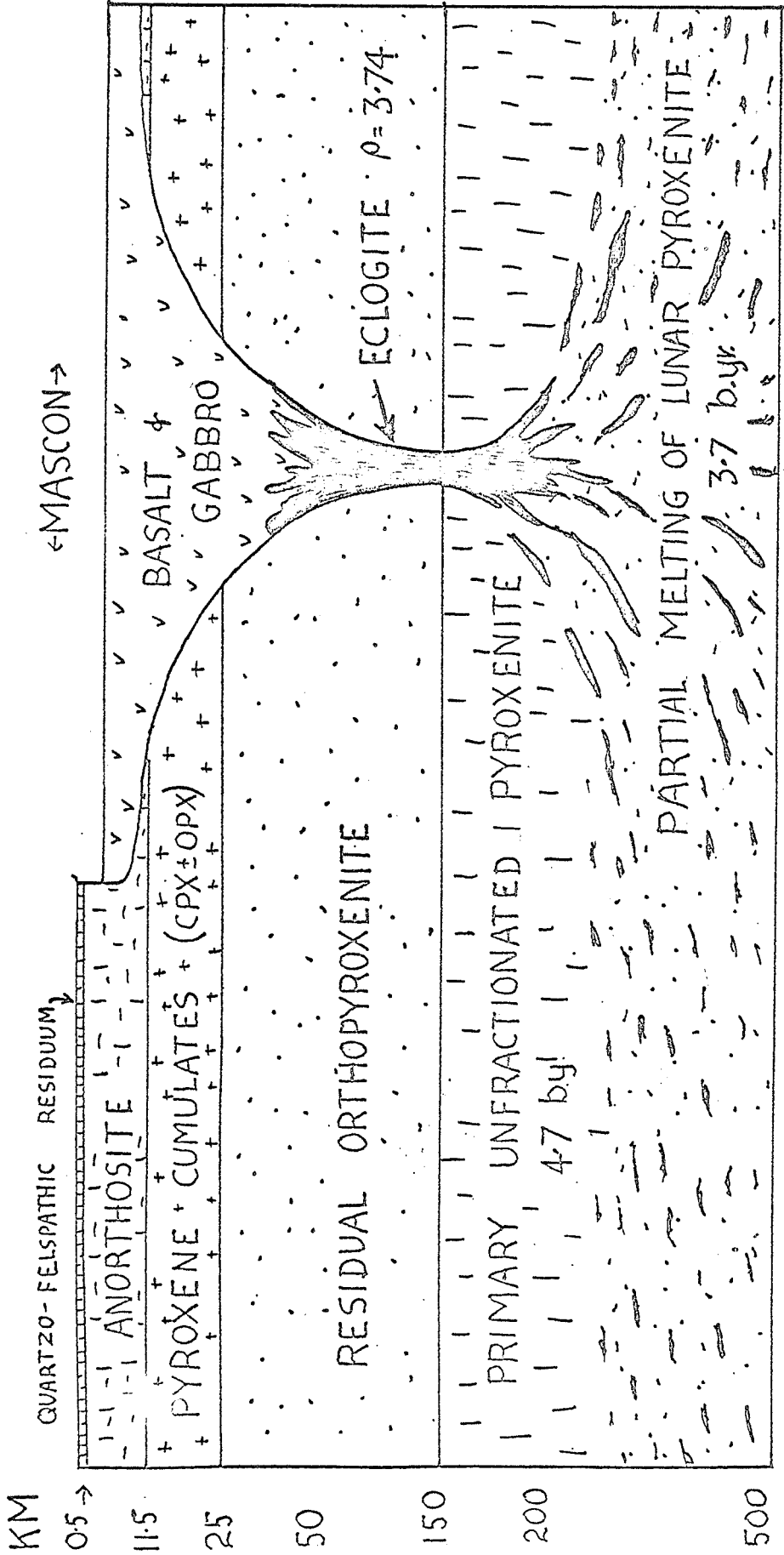
3. NATURE OF HIGHLANDS: THERMAL HISTORY

Several researchers (Anderson et. al., 1970; King et. al., 1970; VonEnglehardt et. al., 1970; Short et. al., 1970, Wood et. al., 1970a) suggested that the small proportion of plagioclase-rich rocks and glasses found in the fines were derived from the lunar highlands and that the latter were generally of "anorthositic" composition. The most comprehensive case for this hypothesis was stated by Wood et. al., (1970b) and need not be restated here. Although the evidence is circumstantial, the overall case is impressive and may be accepted as providing the basis for a tentative working hypothesis. The following discussion assumes that this hypothesis is correct.

Isotatic considerations (Kaula, 1970) imply that lunar highlands of presumed anorthositic composition are about 11.5 km thick. Assuming approximate chondritic abundances of oxyphile, non-volatile elements in the moon, an anorthositic crust of this thickness could be derived by efficient magmatic fractionation processes from an outer shell about 150 km thick containing about 20% of the mass of the moon. This would imply that the outer shell had been extensively melted, probably during, or soon after the moon's formation, to form a deeper zone about 125 km thick of residual orthopyroxene and olivine from which all the low-melting "basaltic" components had been removed to form an overlying shell of eucritic magma about 25 km thick. This layer of convecting eucritic magma then crystallized

Fig. 5. Proposed section (simplified and not to scale) of outer

   regions of moon with flow sheet illustrating two-stage magmatic

   history. The original layer (overlying the anorthosite) of

   quartzo-felspathic residual material containing high concen-

   tration of incompatible elements would have been largely

   destroyed and redistributed by subsequent impact craters.

   Formation of mascons by transformation of basalt and gabbro

   to eclogite in conical feeder structures beneath circular

   maria is discussed elsewhere (Ringwood and Essene, 1970b).

CONTINENT

SEA

KM
0·5 →

‹MASCON→

QUARTZO- FELSPATHIC   RESIDUUM

1·5

ANORTHOSITE

25

PYROXENE + CUMULATES + (CPX±OPX)

BASALT +
GABBRO

50

RESIDUAL   ORTHOPYROXENITE

150

ECLOGITE   ρ= 3·74

200

PRIMARY   UNFRACTIONATED   PYROXENITE

4·7 b.y.

500

PARTIAL   MELTING   OF   LUNAR   PYROXENITE

3·7 b.y.

and differentiated in a manner analogous to some mafic stratiform intrusives to form an upper layer rich in anorthosite and a lower layer dominantly of clinopyroxene, (from which orthopyroxene subsequently exsolved at lower temperature). The residual liquid ultimately resulting from this phase of crystallization differentiation was probably quartzo-felspathic in nature, and would have possessed very high abundances of incompatible elements. Small intrusions of such material may be widespread throughout the highlands. It is possible that rock 12013 is derived from one of these intrusions. A sketch of the proposed model is given in Figure 5.

The proposed igneous history combined with interpretations of lunar cratering (Baldwin, 1963) provide the following boundary conditions for the thermal history of the moon:

(1) Anorthositic highlands 11.5 km thick imply the occurrence of extensive melting in a shell about 150 km thick, probably some 4.7 billion years ago.

(2) Baldwin (1963) showed that older pre-mare craters were systematically shallower than younger pre-mare craters and concluded that the degree of isostatic adjustment during the cratering period decreased with time, implying increasing strength and decreasing average temperatures in the outer lithosphere during this interval.

(3)   Generation of Apollo 11 and Apollo 12 basalts by partial

melting at depths of 200 to 600 km from 1 to 1.4 billion

years after the formation of the moon requires heating

of this region, presumably by the long lived radioactive

elements, uranium and thorium.

(4)   Presumed rarity of extensive mare regions younger than

2 billion years old suggests subsequent cooling at depths

of 200 to 600 km. This is consistent with magnetic

observations which may indicate low electrical conductivity

and relatively low temperatures throughout the bulk of the

moon (Ness, 1970).

Ringwood (1966a) proposed a thermal history which is consistent

with these observations. (Figure 6). He suggested that gravitational

energy of accretion played an important role (See also, Baldwin, 1963,

p. 396). The mean gravitational energy disspated during formation

of the moon is approximately 400 cal/gm based on a free fall model.

The energy per gram increases as R^2, reaching about 690 cals/gm at

the end of accretion. Öpik (1961) has suggested that the moon may

have accreted in a period of about 100 years. Presumably the rate of

accretion also increased as the nucleus grew. Under these conditions

it is possible that a large proportion (eg 50%) of the energy of

Fig. 6. Notional temperature distributions within the moon at

different stages of its history in relation to solidus

and liquidus of possible pyroxenite source material (Table

2) for Apollo 11 basalt.

I. Soon after accretion; II. 0.5 - 1.5 b.y. after accretion;

III. Present day.
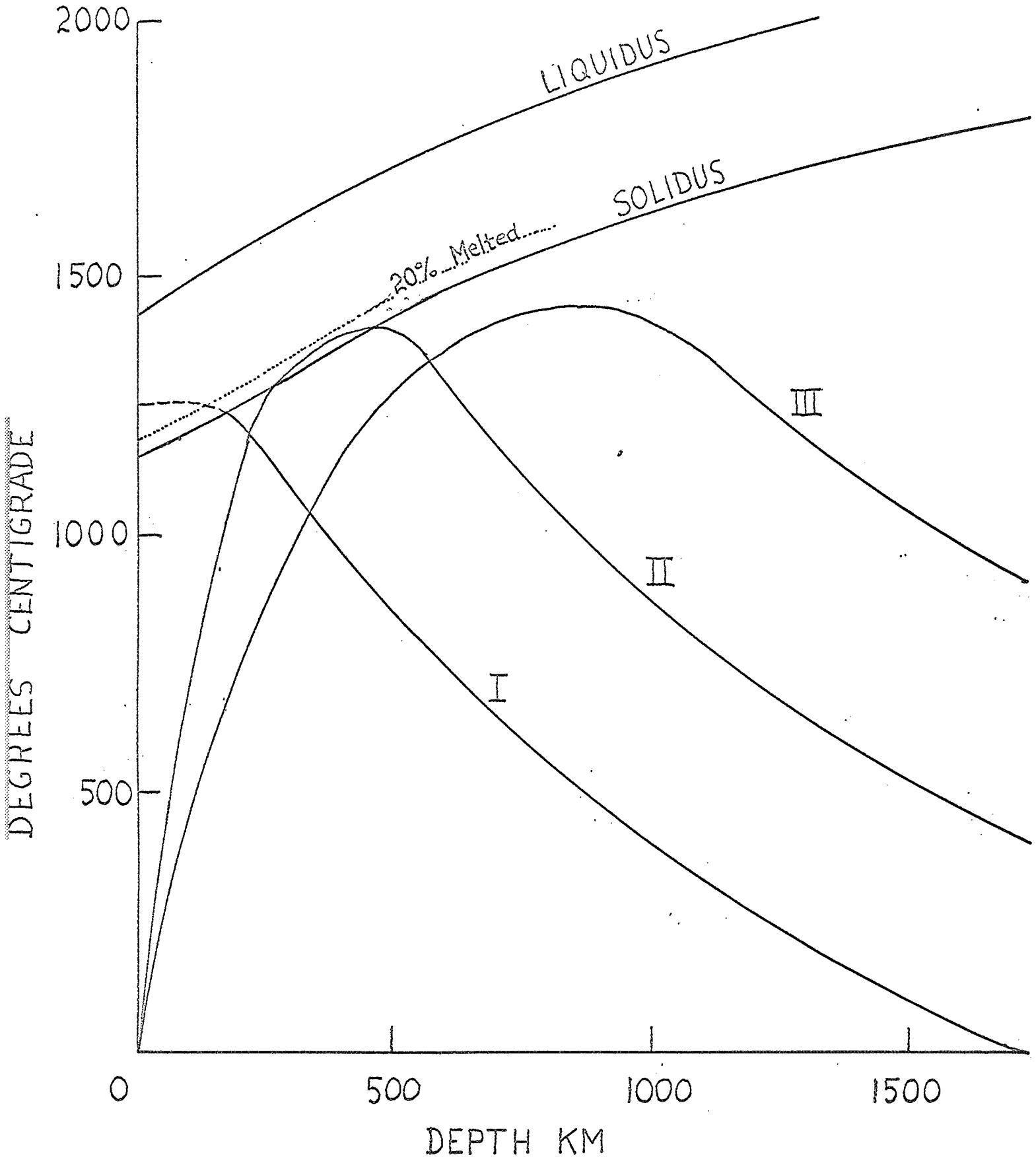
accretion was conserved thermally during the final stages of accretion. This could have led to extensive partial melting of the outermost shell of the moon during accretion. A suggested course of thermal evolution (after Ringwood 1966a) is given in Figure 6. After accretion, the temperature of the moon would increase from the centre outwards as in curve I, crossing the solidus in the outermost shell and permitting differentiation of the highlands. Solidification would occur a timescale of a few million years or less, and the outermost shell would cool by conduction on a timescale of 10^7 to 10^8 years. Increasing thickness of lithosphere during this interval accounts for Baldwin's observations on differential isostatic recovery of crater floors. Superimposed on the conductive cooling in the outer regions of the moon, heating by uranium and thorium occurred at depths below 200 km. It is postulated that this was sufficient to carry the thermal maximum through the solidus as in Figure 6, II, producing widespread volcanism and mare formation over an interval at least 1.0 to 1.5 billion years after accretion. (It seems possible that magmatic activity occurred episodically on the moon from 4.5 to about 3 billion years ago with the loci of the of the source regions of the magmas becoming deeper with time.)

Over a longer time scale, the sharp maximum decayed by conduction. Furthermore, differentiation of the interior and transfer of

radioactive heat sources to near-surface regions during the period

of mare formation drastically reduced the rate of heat generation

in the deep interior, leading to the temperature distribution

shown in curve III.

Wood et. al., (1970b) have suggested that the short lived

radionuclide Al^{26} was responsible for the melting event which

caused the formation of the anorthositic highlands. However, heating

by this source would not have been confined to the near-surface

regions but would have been effective throughout the body of the

moon leading to partial melting of the entire moon, and upward segre-

gation of the melted fraction. This model thus leads to complete

chemical differentiation, as indeed is postulated by Wood et. al.,

resulting in concentration of the heat sources into near-surface

regions, so that deep-seated cooling to depths of several hundred

kilometers occurs over a period of 10^9 years. It then becomes very

difficult to understand how the maria were formed by partial melting

processes .... in the lunar interior after this time interval.

Furthermore, Wood et. al.'s model implies that the source regions

of Apollo 11 basalts were strongly fractionated about 4.7 billion

years ago. The high absolute abundances and sub-chondritic patterns

of incompatible elements occurring in Apollo 11 basalts are not

consistent with this implication.

4. DEPLETION OF LUNAR BASALTS IN VOLATILE METALS

Analyses of Apollo 11 rocks (e.g. Ganapathy et. al., 1970) show that they are depleted by factors of 3 - 100 compared to terrestrial basalts in a wide range of metals which are well known to be relatively volatile under high temperature, reducing conditions. The depleted metals include Na, K, Rb, Cs, Zn, Cd, Hg, Bi, Tl, In, Ge, Pb, Sb and As. These depletions, when correctly normalized, are one of the most spectacular aspects of the comparative chemistry of Apollo 11 and terrestrial basalts.

Two explanations of this abundance pattern have been suggested. The strong depletion in volatile metals may be a primary feature of the lunar basalt source region in which case the abundance pattern assumes major genetic significance. (e.g. Ringwood, 1966a; 1970a; Ringwood and Essene 1970 a,b; Ganapathy et. al., 1970). Alternatively, it has been suggested that the volatile elements were "boiled off" during extrusion of the Apollo 11 basalts, and that the original composition of the lunar magmas was not significantly different from that of terrestrial basalt. (e.g. O'Hara et. al., 1970 a,b).

The strontium isotope ratios show definitively that the depletion of

rubidium relatively to strontium in Apollo 11 basalts occurred at

a very early stage of lunar history, and long before extrusion of

Apollo 11 lavas some 3.7 billion years ago (Ganapathy et. al., 1970;

Gast et. al., 1970; Hurley and Pinson, 1970; Ringwood and Essene,

1970b). It was pointed out by Hurley and Pinson (1970) that the

source region of Type B Apollo 11 rocks was characterized by Rb/Sr

ratios about one fifth of that of the earth's upper mantle and that

"these ratios came into existence at or close to 4600 million years

ago and not at the time of volcanism" (my italics).

A similar conclusion holds in the case of the enormous depletion

of lead which is observed in lunar basalts relatively to uranium, as

compared with terrestrial basalts. Tatsumoto (1970) showed that the

initial U^{238}/Pb^{204} ratio established in the moon 4.6 billion years

ago was much higher than the corresponding ratio in the earth's

mantle. He concluded that the gross depletion of lead relatively to

uranium in the moon is a primary feature established during the moon's

formation and that the Pb/U ratio in the moon has always been much

smaller than in the earth's mantle. Similar conclusions were

reached by Gopalan et. al., (1970) and by Compston et. al., (1970).

Although a very small amount of lead-loss might have occurred during

subsequent extrusion of Apollo 11 basalts, this was trivial in

amount compared to the primary fractionation and does not affect the above conclusion (Tatsumoto, 1970; Compston et. al., 1970).

The Apollo 11 basalts are characterized by remarkably constant sodium abundances, despite a wide range of cooling histories as indicated by textural variations from basaltic through doleritic to gabbroic. This is difficult to explain if solium was being lost by volatilization during cooling, (Goles et. al., 1970). Plagioclase in Apollo 11 basalts displays normal zoning with an increase of sodium and potassium towards the rims. Most of the potassium and rubidium in these rocks occurs in late stage interstitial material. It is clear that the concentrations of K and Rb in the residual magma increased by very large amounts during the course of crystallization of Apollo 11 magma. This behaviour corresponds to expectations for closed-system fractionation. It would not be readily explained if a five-fold depletion of Na, K and Rb had occurred by volatilization during crystallization of the magma.

It is concluded that the depletions of Pb, Rb, Na and K in Apollo 11 basalts compared to terrestrial basalts are characteristic of the respective source regions and are unrelated to the extrusion of Apollo 11 magma. It is reasonable to infer that the corresponding depletions of the other relatively volatile metals (Cs, Zn, Cd, Hg, Bi, Tl, In, Ge, Sb, As),

have a similar origin and significance.

5. SOME BOUNDARY CONDITIONS FOR THEORIES OF LUNAR ORIGIN

The preceeding sections have been devoted primarily to a critical discussion of some aspects of the chemistry and petrogenesis of lunar basalts - particularly those aspects relating to the conditions of formation of Apollo 11 basalts and the chemical and mineralogical nature of their source regions. The discussion has also been concerned with comparisons between lunar and terrestrial basalts and between their respective source regions. An attempt is made below to summarize the principal similarities and differences between lunar and terrestrial basalts, and, by inference, between the lunar interior and the earth's mantle, with the objective of defining some boundary conditions for theories of lunar origin.

(a) Overall Similarities between Earth and Moon

(i) The moon and the earth's mantle are both dominantly composed of ferromagnesian silicates with subordinate CaO and Al_2O_3.

(ii) The relative abundances of most of the non-volatile incompatible class of trace elements in the source regions of lunar and terrestrial basalts appear to have been similar and closely related to the chondritic abundances.

(iii) The absolute abundances of most of the non-volatile

oxyphile elements in lunar basalts fall within the

range of concentrations of these elements displayed

by terrestrial basalts (LSPET, 1969).

(iv) Elements which are comparatively volatile under high

temperature reducing conditions e.g. K, Rb, Pb, Tl, Bi,

In, are relatively depleted in the earth (by factors

of 5-10) compared to the probable abundances of

these elements in the primordial solar nebula (Gast,

1960, Ringwood 1966 a,b). Likewise, this group of

volatile elements is also relatively depleted in the moon.

The above similarities, particularly (iv) might be taken to

indicate that some of the fundamental chemical fractionation processes

which occurred when the earth and moon formed from the solar nebula

were similar, and tenatively, might point in the direction of a gene-

tic relationship between earth and moon.

(v) This suggestion is supported by the history of tidal

evolution of the earth-moon system which implies that

the moon was once only 2.8 earth radii distant from

earth (Gerstenkorn, 1955). This is almost identical

with Roche's limit, and it does not appear likely that

the similarity between these distances is a mere coin-

cidence as is implied by the capture hypothesis

(Öpik, 1961). Additional reasons for rejecting the

capture hypothesis are given in the next section.

If capture is rejected, the tidal history indicates

that the moon was probably born very close to the

earth and, therefore, a close genetic relationship

might be inferred. (e.g. Section 7).

(b)   Chemical Differences between Earth and Moon

Superimposed on the general resemblances which exist between moon

and earth (above) there are some very important specific differences:

(i)   The moon is strongly depleted in iron relative to the

earth. If a core is present, it cannot amount to

more than a few percent of the mass.

(ii)   Apollo 11 basalts are strongly depleted in many

siderophile elements relative to terrestrial basalts.

(e.g. Ganapathy et. al., 1970).

(iii)   The moon is much more strongly depleted in volatile

metals (e.g. Na, K, Rb, Cs, Pb, In, Tl, Zn, Hg etc.),

than the earth compared to primordial abundances.

(iv)   The $\dfrac{FeO}{FeO + MgO}$ molecular ratio in the source regions

of Apollo 11 basalt is probably between 0.20 and 0.26

compared to a probable value for this ratio of 0.12

in the source regions of terrestrial basalts (Green

and Ringwood 1967, Ringwood 1970b).

(v)   The mineralogy of the lunar mantle is probably

pyroxene-dominated as compared to a preponderance

of olivine in the terrestrial mantle.   The ratios of

Al_2O_3 and CaO to total pyroxenes are smaller in the

lunar mantle than in the terrestrial mantle.   Thus,

important differences exist in the relative abundances

of the major components: SiO_2, MgO, FeO, Al_2O_3 and

CaO between the lunar interior and the earth's mantle.

(vi)   Studies of the chemistry of rare earths in lunar and

terrestrial basalts (Haskin et. al., 1970; Gast and

Hubbard, 1970) indicate that calcium-rich clinopyroxene

is a less abundant phase in the lunar mantle than in

the terrestrial mantle.

(vii)   Terrestrial basalts and their source regions are much

more oxidized (oxygen fugacity of 10^{-8} to 10^{-9} atm.

at 1200° C) than lunar basalts and their source

regions ($fo_2 = 10^{-13.5}$ atm. at 1200° C).

(viii)   As a consequence of the low oxygen fugacity in the

moon, H_2O and CO_2 are unstable relative to H_2 and CO

at magmatic temperatures.   Water and carbon dioxide

are, therefore, very rare in lunar rocks compared to

terrestrial rocks.

6   THEORIES OF LUNAR ORIGIN

(a)   Capture Hypothesis

Gerstenkorn's (1955) study of lunar tidal evolution placed the
moon very close to the earth as recently as 2 billion years ago.
Because of the conflict between this date and the age of the earth,
he appealed to a capture mechanism for the moon on a retrograde
orbit slightly earlier than the time of closest approach. MacDonald
(1964) and Öpik (1969) pointed out that earth and moon would be
strongly heated and deformed during the time of close approach and
that the surface features of the moon must have been formed subse-
quently to this time. The ages of the Apollo 11 basalts and the
older ages inferred for the lunar highlands are in direct conflict
with the assumed timescale of lunar tidal evolution and consequently
the justification from this source for the capture hypothesis
disappears.

Urey (1962) argued for capture of the moon about 4.5 billion
years ago on different grounds. He maintained that the moon is a
"primary object" composed of the solar abundances of elements
(excluding gases). His case rested on an assumed agreement in

$\frac{Fe}{Mg + Si}$ ratios between the moon and the sun's atmosphere, in contrast
to the ratios in the terrestrial planets which were believed to be
much higher than in the sun. Recent remeasurements of the f-values
of iron (Garz et al, 1969) have removed the latter discrepancy and

and with it, the basis of Urey's hypothesis. The solar $\frac{Fe}{Mg + Si}$ ratio is now in agreement with the chondritic value within error limits, and furthermore, as demonstrated by Ringwood (1966a) and Ringwood and Clark (1970), satisfactory models of earth, Venus and Mars can be constructed from these abundances. The strong depletion of volatile elements in the moon also contradicts its identification as a "primary object". The evidence for the existence of lunar-sized primary objects originally rested upon the assumption (once also held by the author) that the diamonds found in some meteorites were produced under equilibrium conditions by static high pressures (Urey, 1956, 1966). This assumption is now known to be wrong since the diamonds were produced by shock waves (Lipschutz and Anders, 1961).

Capture of the moon by the earth is an event of low intrinsic probability. The two bodies must possess generally similar orbits and thus have been born in the same region of the solar system, and presumably, from the same parental material. The deficiency of iron in the moon, compared to the earth, Venus and Mars is thus inexplicable on the capture hypothesis, which furthermore does not offer any explanation for the remaining characteristic features of lunar composition discussed in section 5.

Singer (1968, 1969) and Alfven and Arrhenius (1969) have presented modifications of the capture hypothesis aimed at relieving one of several dynamical problems it faces (MacDonald 1966, Wise,

1969); namely the dissipation of energy and heating of earth and moon during the assumed close approach. Singer is obliged to abandon Gerstenkorn's time-scale for lunar tidal evolution which was the primary reason for turning to a capture-type hypothesis in the first place. Alfven and Arrhenius, on the other hand, make a number of highly speculative assumptions about the possibility of spin-orbit coupling. These doubtful assumptions reduce still further very low intrinsic probability (acknowledged by all) of the postulated capture mechanisms.

Modifications of this sort are not aimed at coping with the primary deficiencies of the capture hypothesis - inability to explain the low iron content of the moon and the other differences in composition between earth and moon. A hypothesis which merely sweeps these problems under the table is hardly worthy of serious consideration.

(b) Binary Planet Hypothesis

This hypothesis maintains that moon and earth formed independently in the same neighbourhood of the solar nebula from a common mixture of metal particles and silicate particles which were present in the nebula. It is assumed that during accretion, iron particles were somehow preferentially concentrated in the earth and depleted in the moon. Physical processes which have been invoked to explain this metal-silicate fractionation have been ad hoc and vaguely defined - e.g. Latimer (1950), Orowan (1969). A recent suggestion by Harris

and Tozer (1967) invoked ferromagnetic properties of metal grains to explain preferential aggregation and accretion in the earth. However Banerjee (1967) demonstrated that the mechanism was inadequate by a factor of 10^4 to cause the desired effects.

Irrespective of the iron fractionation problem, the binary planet hypothesis implies that the earth's mantle and moon have been formed from the same well-mixed silicate component. Accordingly basalts formed by partial melting in the earth's mantle and on the moon should display a generally similar compositional range. The major chemical differences (Section 5) between lunar and terrestrial basalts and between their respective source regions, particularly in major elements and volatile metals, are not readily explained on the basis of this hypothesis.

Ganapathy et al (1970) have recently proposed that the moon collected a much smaller proportion than the earth of volatile-rich material which they assume to have condensed from the solar nebula at a late stage. They suggest that this was due to the smaller capture cross-section of the moon compared to earth. This mechanism, however, would also be applicable during earlier stages of accretion of the moon from volatile-poor material, so that the ratio of volatile-rich to volatile-poor material would not be greatly affected. Moreover, the differences in major element composition between earth and moon are not explained.

(c) Fission Hypothesis

According to this hypothesis, the material now in the moon was
derived from the earth's upper mantle after segregation of the core.
Accordingly it would be anticipated that the chemical composition
of the lunar interior would be similar to that of the earth's upper
mantle and that the compositions of lunar and terrestrial basalts
formed by partial melting from their respective source regions
would also be generally similar. The large compositional differences
between lunar and terrestrial basalts and between their respective
source regions, particularly of volatile elements and major elements,
are not explained by this hypothesis, at least, in its earlier forms
which assumed that material from the earth's outer mantle was ejected
in a condensed state (i.e., not gaseous). Recent developments of
the fission hypothesis by O'Keefe (1969) and Wise (1969) suggest
that fission may be accompanied by formation and then loss of a
massive hot terrestrial atmosphere and that the moon may be the
residue of this atmosphere. These modifications have many aspects
in common with the hypothesis of Ringwood (1960, 1966a, 1970a) which
is discussed in the next section.

(d) Precipitation Hypothesis

The history of tidal evolution of the earth-moon system implies
that the moon was once only 2.8 earth-radii from the earth, which
is almost identical with Roche's limit. Öpik (1961, 1967) maintained

that this is unlikely to be a coincidence (as is implied by the capture hypothesis) and that it is more likely that the material now in the moon originally came from within Roche's limit where it existed previously in the form of a "sediment-ring" of earth-orbiting planetesimals – somewhat analogous to the rings of Saturn only relatively more massive. Öpik suggested that the moon was formed by the coagulation of the sediment ring after it had expanded beyond Roche's limit.
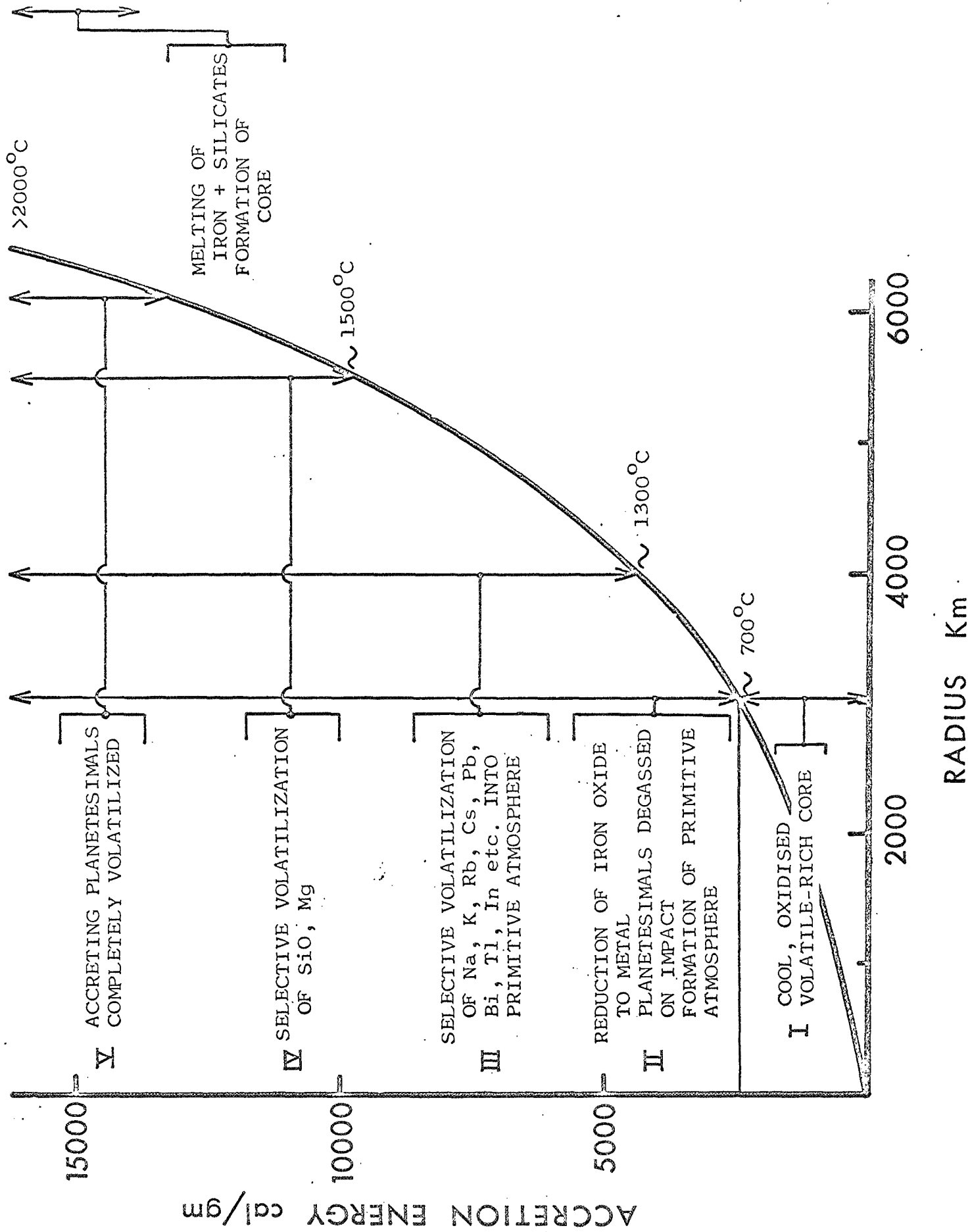
With the demise of the capture hypothesis, Öpik's interpretation looks all the more plausible. Öpik (1961, 1967) also suggested that the observed ellipticities of craters on the lunar highlands (see also Ronca and Salisbury, 1966) were caused by their formation on the moon when it was close to the earth and tidally distorted. From the degree of distortion, he inferred that the craters had been formed when the moon was passing outwards through a ring of earth-orbiting planetesimals at a distance of about 6 earth radii. It has frequently been suggested that the formation of highland craters represented the terminal stage of the moon's accretion. Since these craters post-date the melting and differentiation of the outer shell of the moon (section 3) it may be inferred that the moon was born much closer to the earth than 6 earth radii. Formation of most of the moon by coagulation of a sediment-ring could have occurred in a period of less than 100 years (Öpik, 1961). This

short timescale appears necessary if an appreciable part of the moon's gravitational potential energy is to be retained thermally to cause the early heating and melting as suggested in section 3.

The considerations advanced above strongly suggest a genetic relationship between the earth and the material from which the moon was formed. This is also suggested by the chemical resemblances between earth and moon noted in section 5. An acceptable hypothesis of lunar origin must be capable of explaining this inferred genetic relationship and must at the same time provide an explanation of the fractionation of iron and the other important chemical differences between earth and moon which were summarized in section 5.

The "precipitation" hypothesis developed by Ringwood (1966a, 1970a) may provide a possible framework for interpreting these relationships. A summary of this hypothesis is given below. A more detailed account is given in the preceding references. According to this hypothesis, the earth accreted in the solar nebula from planetesimals of primordial composition resembling the Type 1 carbonaceous chondrites. These contain completely oxidized iron together with large amounts of volatiles and carbonaceous compounds, and have retained the primordial abundances of most elements except for extremely volatile substances. It was assumed furthermore that accretion was completed just before the sun entered its T-Tauri

Fig. 7. Relationship between energy of accretion and radius of the

growing earth. The principal stages of accretion are also

shown in relation to the energy of accretion and approximate

surface temperatures.

The figure (rotated, read with RADIUS Km on the horizontal axis and ACCRETION ENERGY cal/gm on the vertical axis) shows a curve of accretion energy versus radius with the following annotations:

- >2000°C
- MELTING OF IRON + SILICATES FORMATION OF CORE
- ~1500°C
- ~1300°C
- ~700°C

Stages marked along the curve:

- V — ACCRETING PLANETESIMALS COMPLETELY VOLATILIZED
- IV — SELECTIVE VOLATILIZATION OF SiO, Mg
- III — SELECTIVE VOLATILIZATION OF Na, K, Rb, Cs, Pb, Bi, Tl, In etc. INTO PRIMITIVE ATMOSPHERE
- II — REDUCTION OF IRON OXIDE TO METAL PLANETESIMALS DEGASSED ON IMPACT FORMATION OF PRIMITIVE ATMOSPHERE
- I — COOL, OXIDISED VOLATILE-RICH CORE

Vertical axis labels: 5000, 10000, 15000 (ACCRETION ENERGY cal/gm)

Horizontal axis labels: 2000, 4000, 6000 (RADIUS Km)

phase characterized by rapid mass-loss and the generation of a solar wind some 10^6 to 10^7 times more intense than the present solar wind.

Formation of the earth under these boundary conditions is strongly influenced by the gravitational potential energy dissipated during accretion which in turn controls the chemical equilibria in the accreting material. The accretion energy per gram is plotted against radius of the growing earth in Fig. 7. It increases approximately as the square of the radius, reaching 15,000 cals/gm during the final stages.

During the early stages of accretion, the energy evolved is small and accretion is relatively slow. The temperature is accordingly low and is buffered by the latent heat of evaporation of volatiles, e.g. H_2O, in the planetesimals. During this stage (Fig. 7, I) a cool, oxidised, volatile-rich nucleus of primordial material, perhaps about 10% of the mass of the earth, is formed.

As the mass of the nucleus increases, the energy of infall of planetesimals becomes sufficient to cause strong heating on impact, leading to reduction of oxidized iron by carbon and formation of a metal phase. This is accompanied by degassing and the formation of a primitive atmosphere, mainly of CO and H_2 (Stage II, Fig. 7). With further growth (Stage III) both the temperature and intensity of reduction increase and metals which are comparatively volatile under high temperature reducing conditions (e.g. Na, K, Rb, Pb, Zn, Hg,

In, Tl) are volatilized into the primitive atmosphere. During

Stage IV, the surface temperatures exceed 1500° C and the

relevant equilibria (Ringwood 1966a) show that silicate minerals

are selectively reduced and evaporated into the primitive atmosphere,

whilst metallic iron continues to accrete upon the earth. Finally,

during Stage V, after segregation of the earth's core, which causes

a further evolution of 400-600 cals/gm of gravitational energy for

the whole earth, silicates from the outer mantle are directly evap-

orated into the primitive atmosphere. The mass of the primitive

atmosphere is about one quarter of that of the earth and it is

composed mainly of CO and H_2 with about 10 percent of volatilized

silicates.

It is assumed that this primitive atmosphere was dissipated

immediately after accretion or during the later stages of accretion

by a combination of factors (1) intense solar radiation as the sun

passed through a T-Tauri phase, (ii) mixing of the rapidly spinning

high-molecular-weight terrestrial atmosphere with the low-molecular-

weight solar nebula in which it is immersed, (iii) magnetohydro-

dynamic coupling resulting in the transfer of angular momentum from

the condensed earth to the primitive atmosphere, and, more specula-

tively, (iv) rotational instability of the atmosphere caused by

formation of the core (modified fission hypothesis - (Ringwood, 1960;

O'Keefe, 1969; Wise, 1969). The relative importances of these

processes are not known, but it seems likely that the intense solar wind from the T-Tauri phase of the sun played a major role.

As the result of a combination of these processes, the massive primitive atmosphere was dissipated. On cooling, the silicate components were precipitated to form an assemblage of earth-orbiting planetesimals resembling Öpik's sediment ring. A further fractionation according to volatility occurred during the precipitation stage, since the less volatile components were precipitated first at relatively high temperatures and close to the earth, whereas the more volatile components were precipitated at lower temperatures and further from the earth. The silicates precipitating at relatively high temperatures would probably have grown into relatively large planetesimals (10^2 - 10^7 cm diam.) which would tend to be left behind by the escaping terrestrial atmosphere. However the more volatile components precipitating at relatively low temperatures were more likely to have formed fine, micron-sized particles or smoke, which would be carried away with the escaping atmosphere by viscous drag, and hence lost from the earth-moon system. The moon then accreted from the sediment-ring of earth-orbiting planetesimals.

Ringwood (1970a) showed in greater detail that this "precipitation" hypothesis accounted for the fractionation of iron and silicates between earth and moon in the context of a close genetic relationship between earth and moon. Chemical fractionations within

the cooling primitive atmosphere were also shown to provide a basis for interpreting the strong relative depletions of volatile elements in the moon, the fractionation of some major oxyphile elements between moon and earth, the relative depletions of siderophile trace elements in the moon and the different oxidation state of the moon as compared to the earth's mantle.

Goldreich (1966) has shown that stable orbits of planetesimals within about 10 earth radii must lie in the equatorial plane because of the perturbing influence of the earth's rotational bulge. If the moon formed by coagulation of a sediment ring, it should have formed in the same plane, whereas tidal evolution studies show that the lunar orbit possessed a substantial inclination to the earth's equatorial plane when it was much closer to the earth. This has been cited as an objection to Öpik's sediment-ring hypothesis by Urey and MacDonald (1970). However Wise (1969) has pointed out that the difficulty can be avoided if the earth's rotational axis was tilted at about 10 degrees to the plane of the ecliptic <u>before</u> the moon was formed. The initial tilt of the earth's axis may have been a consequence of non-symmetric escape of the primitive atmosphere.

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