

PRESENT KNOWLEDGE ABOUT THE THERMAL HISTORY
OF THE MOON

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Introduction

The discussion of the thermal regime in the moon may be approached in two ways. One may assume initial conditions and integrate the heat equation forward in time to determine whether certain things might have happened in the course of lunar history. Alternatively, one may assume present conditions and integrate the equations backward to determine something about the initial conditions. Clearly, the approach you take will depend on whether you think you know the initial conditions or the present conditions better. It may also depend on the particular hypothesis you wish to prove. It should be generally agreed that thermal evolution calculations are very tentative estimates of a situation we may never fully understand, and that it is against the ground rules to use any results to prove something else.

In this summary, we will show that the freedom available in postulating lunar composition is quite sufficient to produce any conclusion, i.e. early melting, late melting, or no melting at all. This result, which we have based on the most recent estimates of K in the earth, is the same result obtained repeatedly in the past by other authors. Readers desiring to look into this question in depth should consult Wasserburg et al (1964), MacDonald (1962), and MacDonald (1959).

The evolution of a solid planet heated from some initial temperature by long-lived radioactivities has been developed in some detail by Urey and by MacDonald in a number of papers. (A list of general references is given at the end). Very little in the way of new

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data or new insight has accrued in the past five years, so this paper must be of the nature of a review, save in the few instances where recent information on the possible composition of the moon may be discussed. Students of the subject are marking time for the present, in anticipation of the return of lunar samples. It should not be necessary to dwell on the importance of a direct determination of lunar composition; the circuitous discussion of lunar composition in this paper should be proof enough.

Current controversy revolves around two issues. There is, of course, the general problem of the time and mode of formation of the moon. Collateral to this is the nature and age of the lunar maria, which question has, perhaps generated more heat in the recent literature than any other. Opinion has become polarized over whether the radioactive heating has been sufficient to bring the moon close to melting, thus accounting for the maria plains of extrusive lava or fluidized ash-gas.

It has been frequently pointed out that the temperatures obtained by naively applying the forward time integration do not agree with evidence on the shape of the moon which seems to indicate that the moon has appreciable long term strength. Many students of the lunar surface, led by Kuiper and Levin, find the morphological evidence for volcanic maria so persuasive that they derive support for their ideas in this calculation. Others, particularly Urey and MacDonald, find the evidence for long-term strength sufficiently persuasive that they need to reduce the concentration of radioactive heat-producing elements by an appreciable factor. MacDonald (1962) discussed these questions, and

verified Urey's (1957) original assessment. It is the continuing interest in this subject, rather than any significant new findings, that brings us into the arena.

Composition of the Moon - major elements:

The mean density of the moon is 3.33, and, when decompressed, would be about 3.30. We may suppose that the moon is composed of materials similar to those found in either the earth, the meteorites, or the sun. It is not impossible that the moon will in certain ways be quite unlike any of these, and this discussion and many like it will be quickly forgotten. Today, however, we have no choice but to postulate similarities with objects whose composition we know more or less well. The mean density of completely oxidized chondritic meteorites is about 3.7, and the mean density of the earth's mantle, after effects of compression, including phase changes, are removed is about 3.35-3.4, although some uncertainties remain. The difference between the mantle and the meteorites lies primarily in the iron content. The missing iron in the mantle is presumed to have been reduced to the metal and separated to form the core (Ringwood, 1966). The near identity of the lunar with the mantle density leads to the supposition that the moon is identical to the mantle in respect to the major constituents: SiO_2 , MgO , $\text{FeO} + \text{Fe}_2\text{O}_3$, Al_2O_3 , and CaO . If true, this would suggest that the moon was formed from the same reservoir of material as the mantle, and is similarly depleted in iron. In any event, the moon cannot have the composition of the chondritic meteorites unless one makes a special assumption that low density minerals, such as are found in the carbonaceous chondrites, are present

in sufficient quantity to compensate the excess iron.

Ringwood (1966) has elaborated on these notions. He shows that the major element composition of the mantle can be obtained by chemical reduction of iron and segregation of a core, if one takes the Type 1 Carbonaceous Chondrites as representative of the initial solar composition of the material forming the earth, at least with respect to the relevant metals. Enough of the iron and silicon are reduced to metal such that:

- a) the mass and density of the core are correct
- b) the $\text{FeO}/(\text{FeO} + \text{MgO})$ molar ratio is that which is appropriate for the mantle - about 0.12.

The mantle composition obtained from this agrees quite well with the composition estimated on petrological grounds (Table 1).

Ringwood (1966) has proposed that the upper mantle is composed of a mixture of basalt and alpine peridotite. The former is considered to be the lowest melting fraction produced by heating of pyrolite, and the latter is identified as the refractory residuum. On petrological grounds, this model is quite reasonable, but neither phase equilibrium measurements or field evidence restricts the possible ratio of mixing as well as might be desired. Table 2 gives the compositions involved. Regardless of any uncertainty in the mixing ratio chosen for the pyrolite, it is seen that the degree of concordance between the pyrolite model mantle and the chondritic model of the mantle is not affected. There is therefore every reason to believe that the chondritic earth model is suitable, at least with respect to the major rock forming elements. Sodium and potassium do not fit this picture as well, being

Table 1

A summary of the reduction of iron in Type 1 Carbonaceous Chondrites and the resultant composition of the core. Based on Ringwood (1966) with modifications.

	<u>1</u>	<u>2</u>	<u>3</u>	<u>4</u>	<u>5</u>
SiO ₂	33.3	29.8	43.3	45.5	45.0
MgO	23.5	26.3	38.1	37.1	40.0
FeO + Fe ₂ O ₃	35.5	6.4	9.3	7.8	7.7
Al ₂ O ₃	2.4	2.7	3.9	4.4	3.8
CaO	2.3	2.6	3.7	3.2	2.3
NiO	1.9	-	-	-	-
Na ₂ O	1.1	1.2	1.8	.7	.5
K ₂ O	.55	.6	.9	.04	.03
Fe		25.9			
Ni		1.7			
Si		3.5			

All numbers are weight percent. 1. Composition of typical Carbonaceous Chondrites, calculated in the absence of C-, S-, and H₂O. 2. Reduction of all the Ni, enough Fe to produce the desired FeO/MgO ratio in the mantle, and enough Si to produce the desired ratio of (core mass)/(mantle mass). 3. The components of the mantle recalculated to 100%. 4. and 5. Mantle composition estimated by forming "pyrolite" (Ringwood, 1966) of a mixture of basalt and alpine peridotite. The two columns correspond to mixing ratios of 1:3 and 1:5. The basalt used is the low alkali tholeiitic basalt discussed by Engel et al., (1965). The potassium line was added by the present authors.

Table 2

Model Pyrolite compositions

	<u>1</u>	<u>2</u>	<u>3</u>	<u>4</u>
SiO ₂	44.0	49.0	45.5	45.0
MgO	47.0	7.3	37.1	40.0
FeO	7.4	6.9	7.8	7.7
Fe ₂ O ₃	.04	2.0		
Al ₂ O ₃	.25	17.3	4.4	3.8
CaO	.33	11.9	3.2	2.3
Na ₂ O	.01	2.8	.7	.5
K ₂ O	<.001	.16	.04	.03

All numbers are weight percent. 1. Composition of alpine peridotite after Ringwood (1966). 2. Composition of oceanic tholeiitic basalt after Engel et al. (1965). 3. and 4. Pyrolite obtained by taking basalt:peridotite ratios of 1:3 and 1:5, respectively.

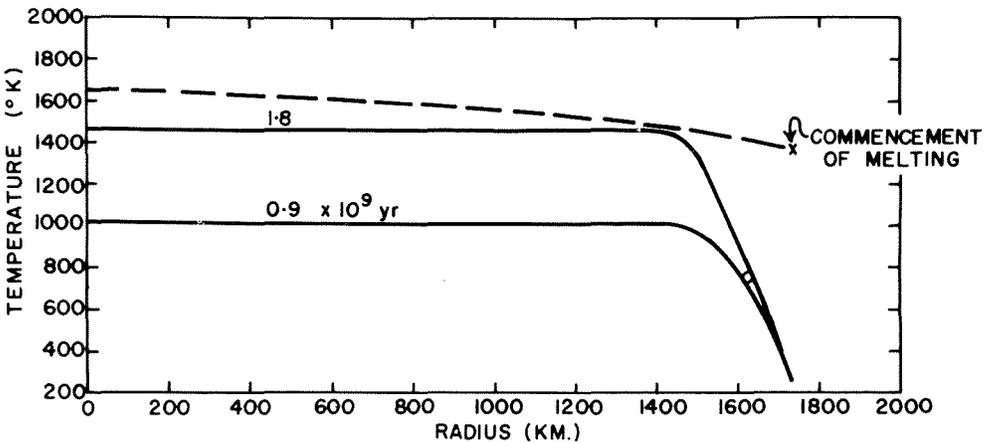


FIGURE 1

depleted by factors of 3 and 20, respectively. The work of Urey has shown the significance of relatively small discrepancies in meteoritic iron, but we see no way of estimating at this time the significance of the discrepancies between the two model earths. Problems of this kind can be better discussed when the composition of the earth is more precisely known.

At this time, prior to the arrival of the lunar samples, the least 'concocted' statement we can make about the moon is that:

1) The moon appears to have the same major element composition as the mantle, based on the "density coincidence" and the apparent connection between mantle and solar composition;

2) Both were obtained by fractionation of most of the Fe in the source material, which had solar composition (with respect to the elements Si, Fe, Mg, Ca, Al, at least).

This model suggests rather strongly that the earth and moon accreted as a joint system, with the missing lunar iron in the core. We do not discuss here the more general problems of lunar origin. It is sufficient to note that studies of the dynamical evolution of the earth-moon system do not seem to indicate simultaneous accretion of the earth and moon from a rotating mass of nebular material (MacDonald, 1964; Goldreich, 1966).

Composition of the moon - heat producing elements:

It is generally agreed that initial temperatures in the moon were substantially below the melting point. The evidence is of course circumstantial, but includes:

1) The fact that volatiles were not totally lost by the earth during accretion strongly implies that formation of the earth was in part a low temperature process. This in turn implies that the accretion was slow enough that most of the gravitational potential energy released could be reradiated on an equilibrium basis (MacDonald, 1959).

2) Due to its lesser mass, the amount of gravitational energy available in the moon is significantly less, and the corresponding heating less. Unless the moon accreted four to six orders of magnitude faster than the earth, its initial temperature was below the melting point. If it is estimated to have accreted in 10^6 years, the initial temperature would be about 330°K .

The assumption of an initially cold moon (say around 300°K) is a conservative one with respect to any melting which results from the calculation. Due to the low pressure gradient, both the melting point gradient and adiabatic gradient are low, so that we may normally neglect pressure effects in our discussion.

Long term heating is largely due to the decay of the following long-lived radioisotopes: MacDonald (1959)

<u>isotope</u>	<u>abundance</u>	<u>decay constant</u>	<u>total decay in 4.5 Ae</u>
U ²³⁸	99.27%	$1.5 \times 10^{-10} \text{yr}^{-1}$	2
U ²³⁵	.72	9.7	100
Th ²³²	100.	.5	1.2
K ⁴⁰	.012	5.3	13

While K is a mineral forming element in granites, these are found only in the outer terrestrial crust. In rocks related more directly to the mantle, and in chondritic meteorites, K is effectively a trace element, amounting to less than 1%. U and Th are trace elements, with abundances of several $\times 10^{-8}$. Due to their large ionic radii, these elements are preferentially excluded from the structure of most mantle minerals and are strongly enriched in the earth's crust. There are also indications that the upper mantle is depleted in alkalis with respect to the chondrites (Gast, 1960). The relatively simple behavior of the major components does not extend to the trace elements. The suggested cogeneration of the moon and the mantle may be of some help in estimating concentrations, but errors by factors of 3 or more are probably involved.

A lunar thermal history (in the sense of a calculation) is very simple, and can be illustrated by the calculation shown in Figure 1, based on a chondritic model for the heat-producing elements. At about 2.0×10^9 years, partial melting begins throughout all but a 250 km outer shell. The temperature rise in the center is given by the integrated heat production, since conductive cooling affects only the outer shell. Initial melting begins almost simultaneously throughout the insulated interior. Our discussion depends, therefore, on the assumed composition of the moon, in the elements U, Th, and K.

MacDonald (1962) and Urey (1957) have consistently pointed out the serious problem posed by the moon's departure from hydrostaticity. Partial melting is enough to guarantee that the affected zone would be unable to support stress differences of any magnitude for geological

time intervals. From laboratory and field evidence, it appears that regions substantially below the melting point can support stress differences of 50-100 bars for times of the order of 10^9 years. Caputo (1965) has computed the minimum stress difference to be expected in a moon with the observed gravitational moments and postulated outer shell thicknesses, assuming that the interior lacks long term strength. For a moon with various core radii he gets

<u>radius of core</u>	<u>minimum stress difference bars</u>
.25	18
.50	19
.75	28

Let us estimate that an outer shell of thickness 350 km (radius = .8) would be sufficient to support a stress of 50 bars. In his study of the chondritic model, MacDonald found that substantial reductions in radioactivity below the chondritic composition had to be postulated to prevent melting out to 85% or 90% of the lunar radius. The figure of the moon thus places an upper limit on the amount of internal heating which can be permitted, although exactly how much is not agreed upon.

One further point needs mentioning. If partial melting occurs early in lunar history, as at 2.0×10^9 years, subsequent heating and melting is implied in the remaining time. Fricker, Reynolds, and Summers (1966) have estimated how such a system might evolve, but were forced to substitute adjustable parameters for physical knowledge in the description of the redistribution of the melted fraction. Given the total integrated heat available in this early-melting (chondritic) moon, it is difficult to see how the moon could fail to become a strongly differentiated body. Yet the amount of volcanism implied by

granting that the maria are volcanic is not significant; they amount to no more than .03% of the lunar mass, which we may raise to .3% by arbitrarily assuming that most of the differentiated material is intruded at some depth. In addition, the Russian gamma ray data seem to indicate a potassium content which is everywhere below .2%. If the moon is differentiated, it is in a limited sense. The data are not incompatible with assuming the maria to be primitive basaltic differentiates of the ultrabasic lunar body, not unlike the terrestrial ocean floor basalts. On the earth, the pressure effect suppresses melting at most depths, while on the moon this effect is absent, suggesting that the moon could be 80-90% melted if it had the heat sources found in the chondrites. One would expect the total visible area of extrusives to be greater than it is, and one would expect a great deal more evidence of recent volcanism.

If one is willing to grant that the maria indicate a volcanic process, then these other problems are vitiated to some extent by adjusting the heat production to produce melting late in lunar history. This would imply a great deal less total heat production to account for. A longer time scale for heating also implies that the cool outer shell would be correspondingly thicker (as the square root of the time). An outer shell in excess of 400 km could support the stress differences present in the nonhydrostatic figure, and can be obtained by reducing the heat production by a factor of 2.

The potassium problem:

Before carrying these speculations further, it is appropriate to review the possible models of U, K, Th composition. It is in this area

that some progress seems to have been made in the past three years.

The coincidence of the present average terrestrial heat flow with the heat produced in a chondritic model earth (Table 3) was the basis for the inference that the earth had the chondritic radioactivity. Calculations of lunar heating (MacDonald, 1962) were therefore based on the "chondritic assumption", there being no difference whether the moon was assumed more closely related to the meteorites or the earth. As more data on the trace element compositions of meteorites and terrestrial materials become available, it became evident that a chondritic model was not appropriate to the earth as a description of trace element concentrations (Gast, 1960; Ringwood, 1966). Prominent among these problem elements are the alkalis. Wasserburg et al (1964) first discussed the thermal consequences of a non-chondritic earth. These authors tabulated the average composition of meteorites and groups of terrestrial rocks (Table 3), using data not available to MacDonald in 1962. Determinations by Heier and Rogers (1963) and Heier (1963) are the most significant in this compilation.

For a variety of terrestrial rocks, representing varying degrees of fractionation of the parent mantle material, the K/U ratio averaged 1.0×10^4 , compared with the chondritic value of 8×10^4 . Heating by this model would be more uniform in time, due to the relative reduction in the short-lived K^{40} . If we take Urey's (1964) latest compilation of abundances and compare with abundances estimated for a 3:1 pyrolite mantle, the alkalis appear depleted significantly with respect to solar or chondritic material (Table 4).

In this paper we have been discussing a mantle composed of

Table 3

Average K, U, Th composition of rocks and meteorites (after Wasserburg et al)

	<u>K%</u>	<u>U(ppm)</u>	<u>Th(ppm)</u>	<u>Th/U</u>	<u>K/U_h x10⁴</u>	<u>heat production(cgs)</u>
Chondrites	.0845	.011	.04	3.7	7.7	4.0x10 ⁻⁸
Achondrites: High Ca	.043	.081	.51	6.3	0.5	22.2
Granites	3.79	4.75	18.5	3.9	0.8	1050.
Basalts	.84	.6	2.7	4.5	1.4	154.
Eclogites: Low U	.036	.048	.18	3.8	0.8	10.3
High U	.26	.25	.45	1.9	1.1	44.5
Average terrestrial ratio	.01%	.01	.04	4.0	1.0	2.34
Oceanic tholeiite	.13	.065	.26	4.0	2.0	18.9
3:1 pyrolite	.032	.016	.065	4.0	2.0	4.7

Table 4

Atomic abundances of the alkalis (per 10^6 Si atoms)

	<u>solar</u>	<u>Type 1 CC</u>	<u>ordinary C</u>	<u>3:1 pyrolite</u>
Na	4.4×10^4	6.4×10^4	4.6×10^4	1.8×10^4
K	3160	3800, 3270	3460, 3990, 3420	1400
Rb	6.5		5.9	$<1.6^*$

* Based on Gast (1960), with K/Rb > 400 .

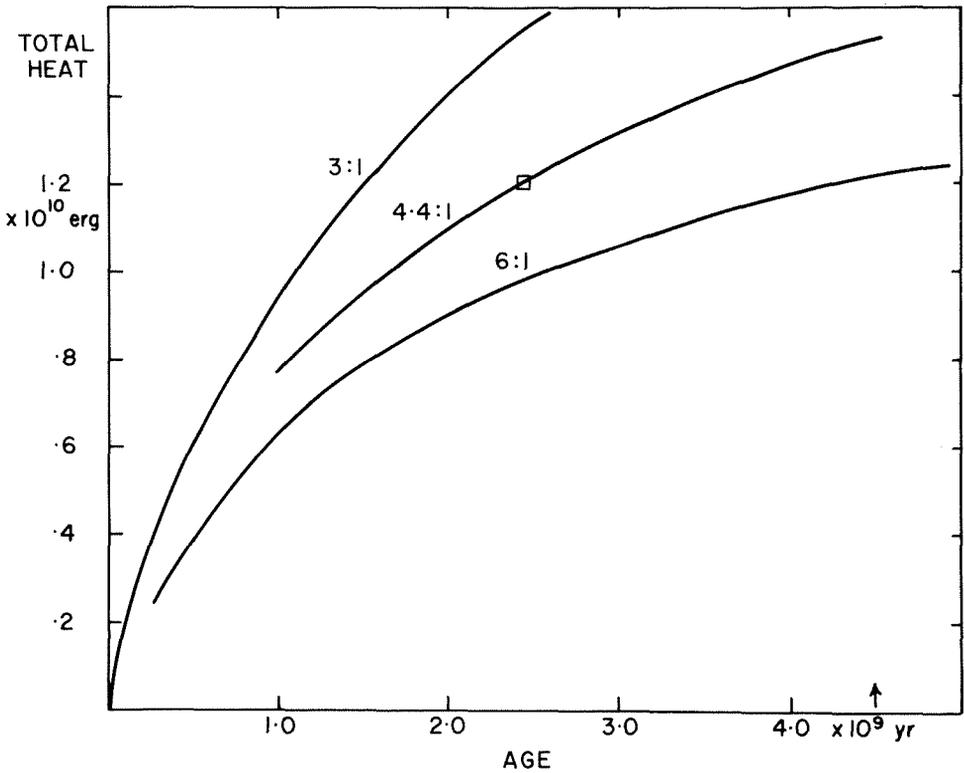


FIGURE 2

peridotite plus low potassium tholeiitic basalt. The potassium concentrations are significantly less than those discussed in recent papers on heat flow, and are connected with the recent discussion by Engel et al (1965) on oceanic basalts. The processes of partial melting and fractional crystallization which lead to differentiation of the upper mantle are reasonably well understood, but the determination of the mantle composition has lagged, due to problems in interpretation of observed igneous rock chemistry. Engel et al pointed out that the ocean-bottom tholeiitic basalts were probably the primary igneous differentiate of the mantle, since this class of basalt accounts for all but a small fraction of the total mass of oceanic basalt, and appears to be more primitive from a petrological point of view. The potassium in tholeiites runs from .1%-.2%, compared with values of .5% or more, which had been taken as typical of basalts. Since the K/U and Th/U ratios are reasonably well known, we can obtain a model composition for K, U, and Th in a pyrolitic mantle (Tables 1, 3, 4). The estimated heat production (Table 3) is close to that of the chondrites (the 2nd chondritic coincidence!), so this composition is again compatible with the observed terrestrial heat flow. The K/U ratio of 2×10^4 was an attempt to determine the preferred ratio for low-K basalts from the plots by Heier and Rogers. If the "terrestrial" ratio of 1×10^4 is taken, then the Th and U are doubled, and the heat production becomes effectively double the chondritic value. Clearly, what is needed now is measurements of K, U, and Th in low-K oceanic basalts, rather than the numerology which is required here.

Total heat production:

The total integrated heat production for pyrolite models is plotted in Figure 2. Assuming that a total of 1.2×10^{10} ergs is required to begin melting, we find that a 3:1 pyrolite mixture will reach melting in 2 Ae, and a 6:1 mixture will just reach melting in 4.5 Ae. Since pyrolite, at this stage, is merely a construct, designed to explain the composition of observed mantle differentiates, one has some freedom in selecting this ratio. Ringwood (1966) remarks that ultrabasic rocks are found with a range of ratios from 3:1 upward. One is now at liberty to select any given time for the onset of melting (say 2.4 Ae) and the mixing parameter can be suitably adjusted (in this case 4.4:1). Had the K/U ratio been taken as 1.0×10^4 , the heat production would be nearly doubled for all ratios. The model used here is more instructive, perhaps, because it demonstrates that a reasonable mantle analog for lunar composition is adjustable to produce any desired conclusion about lunar melting.

Preferred model:

We now indulge briefly in a statement of position, giving our current preferences in regard to the state of affairs on the moon.

1. While Gold has demonstrated that the maria could conceivably be produced by other than a volcanic process, we find his suggestions ingenious, rather than plausible. The properties of the maria have been discussed at great length, and we believe that they must be of internal origin, and must have been emplaced as fluids. Individual volcanic features, such as domes and "flows" are of a sufficiently small scale (comparable to terrestrial scales) that it seems indicated

that the maria formed by the superposition of a large number of individual extrusive events. The dome fields, some with summit craters, and the Alphonsus cinder cones are unmistakable features of internal origin, which are at present the best circumstantial evidence of a hot interior.

2. Lacking lunar samples at this time, we prefer to discuss the moon as a close relative of the mantle. Taking the radioactive abundances from the low-K tholeiites, we find that heat generation can be readily adjusted, within the uncertainties of the model, to provide for early or late lunar melting or none at all.

3. The nonhydrostaticity of the lunar figure argues that a solid shell 350 km or more in thickness be preserved if the interior reaches the lowest melting temperature. This is most easily arranged if melting first occurs late in lunar history.

4. We are not persuaded that the maria and other features of internal origin could be produced without the temperature in the interior reaching very near the melting temperature. One cannot eliminate the possibility that fluidized ash and breccia flows (nuee ardente) are the primary mechanism, rather than basalt as a liquid. The Russian gamma ray experiment appears to eliminate all terrestrial analogs except the low-K tholeiite and the entire class of ultrabasic rocks. If release of gas is an important phenomenon, it is hard to see how this can occur in a low temperature moon.

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Figure Captions

1. Temperature vs. radius in a chondritic moon initially at 0°C, at times 0.9 Ae and 1.8 Ae.
2. Integrated heat production per gram for pyrolites with various peridotite/basalt ratios, using composition for oceanic tholeiite in Table 3; K% after Engel et al (1965) and Th/U and K/U after Heier and Rogers (1963).