ON THE THERMAL HISTORY OF A MOON OF FISSION ORIGIN

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Abstract. Model calculations show that the thermal history of a Moon which originated by fission from the proto-Earth is the same as that for the Moon as it is currently understood. In particular, a fissioned Moon currently has a small percent of partial melt or at least near solidus temperatures below depths of 800 km in accord with the seismic data which show that the deep interior of the Moon has a very low Q. The models have moderate (20–50%) degrees of partial melting in the upper mantle (depths < 300 or 200 km) in the period between 3 to 4 $\times 10^9$ years ago and, therefore, can account for the mare filling epoch. Finally the heat flow of the models is 18 ergs cm^{-2} s^{-1} which is close to the average of 19 ergs cm^{-2} s^{-1} derived from the Apollo heat flow experiments. These findings add further support for the fission origin of the Moon.

1. Introduction

The purpose of this paper is to present the results obtained from a study of the thermal history of a Moon of fission origin. The general structure, petrology, and initial development of a fissioned Moon, and its U, Th, and K content are discussed in a series of previous papers (Binder, 1974, hereafter referred to as paper I; Binder, 1975a, hereafter referred to as paper II; Binder, 1975b, hereafter referred to as paper III; Binder, 1976a, hereafter referred to as paper IV; Binder, 1976b, hereafter referred to as paper V; and Binder, 1976c, hereafter referred to as paper VI). The models developed in this study are limited to those which are spherically symmetric, but have radially dependent variations in the concentrations of the heat producing isotopes of U, Th, and K and in the thermal conductivity of the mantle and crust. As such, the models depict the general thermal history of the Moon, but do not show the secondary effects on the thermal history due to either the regional variations in the concentration of U, Th, and K in the crust (paper III) or the changes in the thermal conductivity in the crust caused by the major basin forming impacts. However, the results obtained in this study show that the general thermal history and the present thermal conditions of a fission model are consistent with those known for the Moon.

2. Computational Methods

The computations of the thermal models were carried out for a spherically symmetric body using a finite difference solution to the heat conduction equation

$$\rho C_p \frac{\partial T}{\partial t} = \frac{1}{r^2}\frac{\partial}{\partial r}\left(r^2 K \frac{\partial T}{\partial r}\right) + H(r, t), \tag{1}$$

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where C_p is the specific heat, ρ is the density, T is the temperature, r is the radius, k is the thermal conductivity and $H(r, t)$ is the heat-source term. A numerical solution to Equation (1) was obtained for the various models studied using the method presented by Reynolds *et al.* (1972).

3. Model Parameters

In order to calculate a thermal history for the Moon, it is necessary to define a variety of parameters. A number of these are model dependent, such as the initial temperature distribution and the distribution of the heat producing isotopes in the model, and are derived from the discussions of the structure and development of a fissioned Moon given in papers I–VI. Additional physical parameters which are needed for the study, but which are not uniquely related to the fission model, have been taken from the literature.

A. Structure of the Moon

As described in papers I, II, IV and VI, if the source region of the mare basalt magmas, i.e. the upper mantle, is dominated by pyroxene (e.g., Ringwood and Essene, 1970; Ringwood and Green, 1975), then a Moon of fission origin should have (1) a dunite (Fo_{95}) lower mantle below a depth of 300 km, (2) a pyroxene-rich peridotite (60–85% pyroxene, $Mg' \sim 70$*) upper mantle between 60 and 300 km depths, (3) a KREEP rich layer ($\sim$ 5 km thick) at the interface of the upper mantle and crust (this layer existed only during the first few hundred million years of lunar history (papers I and IV)), and (4) a 60 km thick feldspathic crust.

However, as is shown in paper V, the upper mantle is most probably dominated by olivine and not pyroxene and extends to a depth of only 200 km. Based on this conclusion, it is shown in paper VI that the entire mantle of a fissioned Moon is olivine-rich (65–75% olivine) with the lower part (below 200 km) being more magnesium-rich ($Mg' =$ 90–95) than the upper part ($Mg' \sim 70$). Thus, the only difference between these models, in terms of their thermal histories, is that the upper mantle (mare basalt magma source region) extends to a depth of only 200 km for the second, probable model as compared to 300 km for the first model; this difference is found to be insignificant.

B. Amount and Distribution of U. Th, and K

If the Moon fissioned from the Earth, then there is a simple relationship between the total amounts of U, Th, and K in both bodies. Based on the analysis given in paper III, a fissioned Moon should presently have about 35 ppb of U, 125 ppb of Th and 100 ppm of K and the heat produced by the decay of the radioactive isotopes of these elements would result in a lunar heat flow of about 13 ergs cm^{-2} s^{-1}. These bulk concentrations of U, Th, and K are used for all the models computed in this study.

According to the discussions given in papers II and IV, the distribution coefficients

* $Mg' = 100\,Mg/(Mg + Fe)$.

of U, Th, and K for olivine and orthopyroxene are so small that the orthopyroxene peridotite lower mantle would be essentially devoid of these elements. However, as pointed out in paper I, it is expected that filter pressing would not have been 100% effective in the lunar lower mantle (or elsewhere in the Moon) due to the very low gravity there. As such, some of the original melt would have been trapped between the olivine and orthopyroxene crystals and thus the lower mantle should contain small amounts of U, Th, and K. Based on the assumption that the amount of trapped original melt is on the order of 10%, then the concentration of U, Th, and K in the lower mantle is 10% of that of the bulk Moon. The computed values are 3.5 ppb of U, 12.5 ppb of Th, and 10 ppm of K for the lower mantle.

It is assumed for some of the models computed in this study that the trapping of melt was effective only below the depth ($\sim$ 800 km) where the seismic data indicate that a small degree of partial melt exists at present in the Moon (Nakamura *et al.*, 1974). In these models about 1.5% of the U, Th, and K of the Moon is located in the lower mantle. In other cases, it is assumed that the trapping of melt was effective throughout the lower mantle and in these models, the lower mantle contains up to 7% of the Moon's U, Th, and K. Finally, one set of models was computed which contained no trapped U, Th, and K in the lower mantle. This was done in order to provide baseline models needed to evaluate the effects on the Moon's thermal history due to the production of small amounts of heat in the deep interior of the Moon.

The amount of U, Th, and K in the upper mantle of each model in which the upper mantle extends to a depth of 300 km was taken directly from paper IV. These models have 25 ppb of U, 90 ppb of Th, and 72 ppm of K in their upper mantles. For those models in which the upper mantle extends to a depth of only 200 km, the concentrations of these elements are increased to 40 ppb of U, 144 ppb of Th, and 112 ppm of K. Based on these values about 20% of the Moon's U, Th, and K is found in its upper mantle.

The next zone in the models is the KREEP layer at the interface of the crust and upper mantle. Assuming that this zone is 5 km thick, a value which is in accord with the estimates given in papers II and IV, this zone would contain 2.69 ppm of U, 9.53 ppm of Th, and 7630 ppm of K. However, as discussed in paper I and IV, the evidence indicates that the KREEP zone existed only during the first few hundred million years of lunar history and that the KREEP, U and Th components of this zone were mixed into the crust during its early modification. Since there is no a priori way of judging the mixing time scale, one set of models was computed assuming that the KREEP layer existed only for the first 10^8 years of lunar history and then was instantaneously and uniformly mixed into the crust. A second set of models were computed assuming that the instantaneous mixing of the KREEP zone and the crust occurred 5×10^8 years after the Moon began its thermal development. These two types of models, while perhaps not completely bounding all possible mixing models, allow the evaluation of the general effect of the presence of a very high concentration of U, Th, and K (some 60% of that of the total Moon) at a depth of about 60 km on the thermal histories of the models.

According to the models in paper IV, the crust originally contained about 20% of the lunar U, Th, and K, but since the KREEP was mixed into it, the crust now contains nearly 80% of the heat producing elements. As a result, the concentrations of U, Th, and K in the crust are 250 ppb, 890 ppb, and 710 ppm, respectively.

There are two final points to be made about the U, Th, and K concentrations and distributions in the models. First, the values quoted above are all the present day values. The model computations take into account the decay of the various isotopes. Second, since the primary differentiation of a fissioned Moon occurred during the first 10^8 years of lunar history (paper I, also see 3B below), it is not necessary to consider the upward migration of U, Th, and K in models as is done for non-fission models which start out as undifferentiated bodies (e.g., see Toksöz and Solomon, 1973).

C. Initial Temperature

As discussed in paper I, if the Moon formed by fission, it was most probably completely molten and at super-liquidus temperatures just after it separated from the Earth. The first order analysis of the initial cooling and solidification of a fission Moon presented in paper I indicates that the Moon would have reached the solidus everywhere in its interior after about 10^8 years and it was during this short period that the Moon differentiated and formed its upper and lower mantles and crust (papers I, II, IV and VI). Since the age of the Moon is close to 4.6×10^9 years (Tera and Wasserburg, 1974, 1976), the Moon was everywhere at the solidus about 4.5×10^9 years ago, a value which is in essential agreement with the results of Tera and Wasserburg (1974, 1976) who find that the crust and source region of the mare basalt magmas became closed systems between 4.4 and 4.6×10^9 years ago. Thus, the model computations have $t = 0$ at 4.5×10^9 years ago when the Moon was everywhere at the solidus.

Since the composition of the mantle of a fissioned Moon is essentially that of pyrolite (papers V and VI); the pressure dependent solidus temperature of pyrolite (Green and Ringwood, 1967) is used as the basis for determining the initial temperature distribution in most of the lunar mantle. Additional solidus–temperature information was obtained from the data on ultrabasic melts with high anorthite contents presented by Ringwood (1976). Such melts have compositions similar to that proposed for a fissioned Moon during the latter part of the differentiation sequence when the feldspathic crust and upper mantle formed (papers IV and VI). Further, based on the liquidus temperatures presented for the above mentioned feldspar-rich ultrabasic melts (Ringwood, 1976) and those for Ringwood and Essene's pyroxenite (1970), another ultrabasic material related to pyrolite, it is estimated that the pyrolite liquidus lies about 200 °C above the solidus at all pressures found in the Moon. This estimate is used as an additional boundary condition for the modelling (see Section 3D) and in determining the initial temperature distribution in the Moon.

Using these data, two limiting initial temperature distributions for the models are defined: the first profile is based on the differentiation sequence proposed in papers I, II, and IV. In this sequence refractory, magnesium-rich olivine sank from the surface to form the lower mantle. In this case, the initial temperature of the model's center is that of the initial crystals forming from a pyrolite melt, i.e., close to that of the pyrolite liquidus. At higher levels in the mantle, the initial temperature profile approaches that of the pyrolite solidus since the less refractory components of the melt were deposited in the upper parts of the mantle. Finally, in the upper 200 km of the Moon, the initial temperature profile changes from the solidus temperature of a pyrolite melt to that of the feldspar-rich melt. Based on these considerations, the model's initial central and surface temperatures at 4.5×10^9 years ago were 1850 and 1175 °C, respectively and this profile represents the expected upper limit for the initial temperature distribution in a fissioned Moon.

The second temperature profile, which is also the lower limit profile, is based on the differentiation model presented in paper VI. In this case, the initial temperature is essentially that of the pyrolite solidus throughout the Moon, except in the uppermost 200 km where, as for the profile discussed above, the temperature is essentially that of the solidus of the feldspar-rich residual melt. In this case, the initial central temperature of the model is 1725 °C and the surface temperature is, as before, 1175 °C.

D. Solidus–Liquidus Relationships and Partial Melting

In general, the initial temperature profiles defined in Section 3C are also the solidus temperatures used in the modelling. However, a few models, for which the effects on the cooling of the lower mantle due to the trapping of interstitial liquid is considered, have a modified solidus. In each of these cases, the initial temperature profile is well above that of the pyrolite solidus and so the solidus of the trapped interstitial melt is assumed to be 100 °C lower than the initial temperature profile.

In previous thermal history studies of the Moon (e.g., Toksöz and Solomon, 1973) it is assumed that melting occurred only at solidus temperatures, i.e., that the temperature difference between the liquidus and solidus is zero, and that as additional heating occurred the temperature remained constant and all the energy went into the heat of fusion of the rock until total melting occurred. At that point, the temperature was still held constant and any additional heat energy was transferred upwards in the model to simulate liquid convection in the totally molten zone. As such, this type of thermal modelling takes into account neither the increase in the temperature of the Moon above the solidus as partial melting occurred nor the changes in the thermal gradients in the upper mantle of the Moon which accompany partial melting. In order to more accurately model the melting process and its effects on the thermal history, we have considered the fact that, as is discussed in Section 3C, the liquidus lies approximately 200° above the solidus. Within this interval, the degree of partial melting varies non-linearly with increasing temperature since the initial melting (at least at pressures less than 10 kb where anorthite is a stable phase in a pyrolite-like material, see Figure 2 of Ringwood and Essene (1970) or Figure 2 of Green and Ringwood (1967)) occurs at a constant temperature at the anorthite–olivine–pyroxene peritectic. After the peritectic melting is complete, the slope of the melting vs temperature curve is controlled by the olivine–pyroxene cotectic reaction and by the heat of fusion of pyroxene as more pyroxene enters the melt and finally the slope is controlled by the heat of fusion of olivine for higher degrees of partial melting. However, for simplicity, these complications are ignored and it is assumed, as a rough approximation, that the temperature increases linearly with increased heating in the 200 °C interval between the solidus and liquidus. Also, it is assumed that the degree of partial melting is proportional to the increase in temperature between the solidus and liquidus. Finally, within this interval, the computations are carried out accounting for both the specific heat of the co-existing melt and crystals and the heat of fusion of the crystals as discussed in Section 3E.

As will be shown in Section 4, the liquidus is not reached in any of the wide range of fission models computed for this study; as such the transfer of heat upwards in the models by simulated convection in a totally molten zone was not necessary in these computations. However, in a real magma, convection certainly can occur when the degree of partial melting is high, but before total melting has been achieved. According to our models, the maximum degree of melting in the Moon is between 30 and 50% and at these relatively low values, it is doubtful that large scale convection could have occurred. As such, convection even in partially molten zones, was not considered in the computation of the models.

E. Specific Heat and Heat of Fusion

As discussed above, the computations include not only the specific heat of the rocks at subsolidus temperatures but also specific heat and the heat of fusion of the materials between the solidus and liquidus. In Equation (1), the specific heat appears in the left hand side of the equality, but, of course, there is no consideration of the heat of fusion in the heat conduction equation. However, the heat of fusion is, especially in the case under consideration where the melting occurred over a wide temperature range, physically equivalent to a very large specific heat of the material. As such, the specific heat and heat of fusion are combined to make an effective specific heat for the temperature interval between the solidus and liquidus.

From Toksöz and Solomon (1973), the specific heat of rocks is 1.2×10^7 ergs gm^{-1} °C^{-1} and the heat of fusion is 400×10^7 ergs gm^{-1}. Since the melting interval is taken to be 200° and since it is assumed in this paper that the degree of melting increases uniformly over this interval, 2×10^7 ergs gm^{-1} are required to melt the additional 0.5% of the original solid for each degree C the temperature increases between the solidus and the liquidus. These 2×10^7 ergs gm^{-1} °C^{-1} plus the 1.2×10^7 ergs gm^{-1} °C^{-1} specific heat used for the solid as well as the liquid phases, combined to give an effective specific heat of 3.2×10^7 ergs gm^{-1} °C^{-1} for the melting interval.

The value of 1.2×10^7 ergs gm^{-1} °C^{-1} for the specific heat is used in the computations for all subsolidus temperatures and the 3.2×10^7 ergs gm^{-1} °C^{-1} effective specific heat is used for all parts of the models where melting, and, of course, the resolidification of a partially melted zone occurred with the following exception. As discussed in 3D, a few models were computed for which it is assumed that 10% of the original melt was trapped as interstitial liquid in the lower part of the models and that the solidus of this fluid is

100 °C lower than the initial temperature of those parts of the models. Thus, in those parts of the models, it is assumed that 40 ergs gm^{-1} due to heat of fusion were yet to be released as the interstitial liquid solidified (all the models show only cooling, never heating, towards the models center, see Figures 1–4) and that since this solidification took place over a 100 °C interval, the effective specific heat of this regions was 1.24×10^7 ergs gm^{-1} $°C^{-1}$ until complete solidification occurred, after which the 1.2×10^7 ergs gm^{-1} $°C^{-1}$ value was used.

F. Thermal Conductivity

The thermal conductivity for dunite as defined by Schatz and Simmons (1972) is used throughout the mantles of all the models since olivine is the major constituent of the bulk of the mantle. According to Schatz and Simmons, the conductivity is sum of the lattice conductivity (K_L) and the radiative conductivity (K_r) where

$$K_L = \frac{4.184 \times 10^7}{30.6 + 0.21\,T(\mathrm{K})} (\mathrm{w\,cm^{-1}\,K^{-1}}), \tag{2}$$

$$K_r = 0 \quad \text{for} \quad T < 500\,\mathrm{K}, \tag{3}$$

and

$$K_r = 230(T - 500°) \quad \text{for} \quad T \geqslant 500\,\mathrm{K}\ (\mathrm{w\,cm^{-1}\,K^{-1}}). \tag{4}$$

Based on these relationships, the thermal conductivity of the lunar mantle varies between 3 and 4.5×10^5 ergs $cm^{-1}\,s^{-1}\,K^{-1}$, depending on the temperature.

Generally, the thermal conductivity used in model calculations for the lunar interior are also used for the crust. However, the lunar crust consists of feldspathic norites (with less than 10% mare basalts) and is well brecciated in its uppermost parts. Thus, the use of thermal conductivities applicable for ultramafic rocks of the interior is questionable at best. According to the measurements of Mizutani *et al.* (1972) and Mizutani and Osako (1974) the thermal conductivities of feldspathic norite 14311 and anorthorite 77017 are 0.5 to 1×10^5 ergs $cm^{-1}\,s^{-1}\,K^{-1}$ and 0.1 to 0.2×10^5 ergs $cm^{-1}\,s^{-1}\,K^{-1}$, respectively and are not very temperature sensitive. However, the above values are for rocks collected at the surface which have pores and cracks which lessen the thermal conductivity. When these effects are corrected for, Mizutani *et al.* and Mizutani and Osaka find a value of about 2×10^5 ergs $cm^{-1}\,s^{-1}\,K^{-1}$ for both 77017 and 14311. Since the pore spaces and cracks close quickly with increasing pressure and are essentially completely closed by a depth of 10 to 20 km, we have adopted a value of 2×10^5 ergs $cm^{-1}\,s^{-1}\,K^{-1}$ for the crusts of most of the models. However, because pores and cracks do reduce the conductivity in the first 10 to 20 km of the crust, one set of models was computed with $K = 10^5$ ergs $cm^{-1}\,s^{-1}\,K^{-1}$ for the outer 20 km.

G. Other Considerations

In addition to the parameters discussed above, there are several others and certain boundary conditions which need be mentioned. Among these is the boundary condition that the mean, constant surface temperature of the Moon is −20 °C (Langseth *et al.*, 1972).

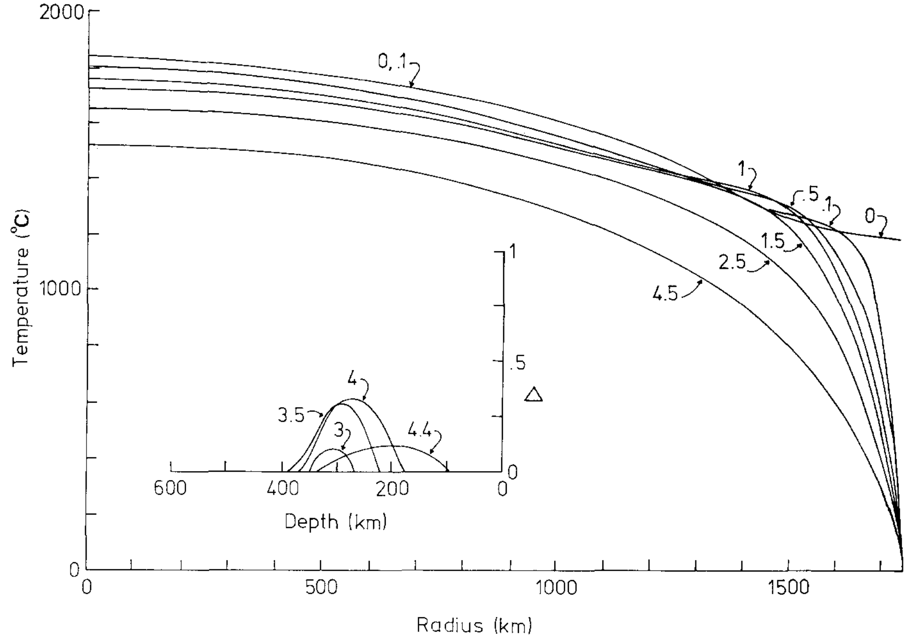

Fig. 1a. Thermal profiles for model Moon 1a as a function of time since the Moon was everywhere at the solidus 4.5×10^9 years ago. The time, in billions of years since 4.5×10^9 years ago, is indicated for each profile. The insert gives the degree of partial melting (Δ) as a function of time and depth as derived from those profiles which, in part, lie above the solidus defined by the $T = 0$ profile. The time, in billions of years before the present, is indicated for each curve in the insert. See text for further details.

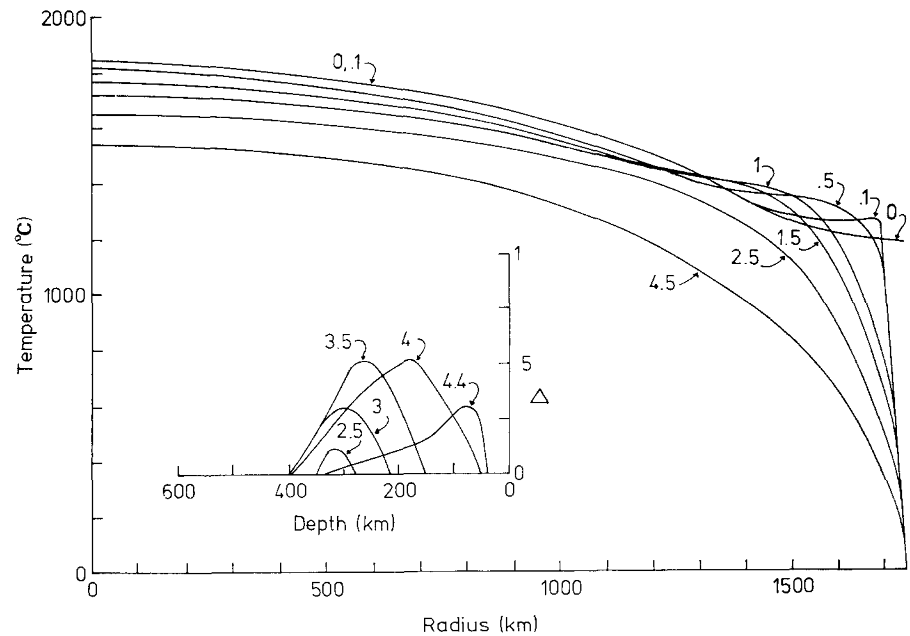

Fig. 1b. Thermal profiles for model Moon 1b, otherwise same as Figure 1a. See text for further details.

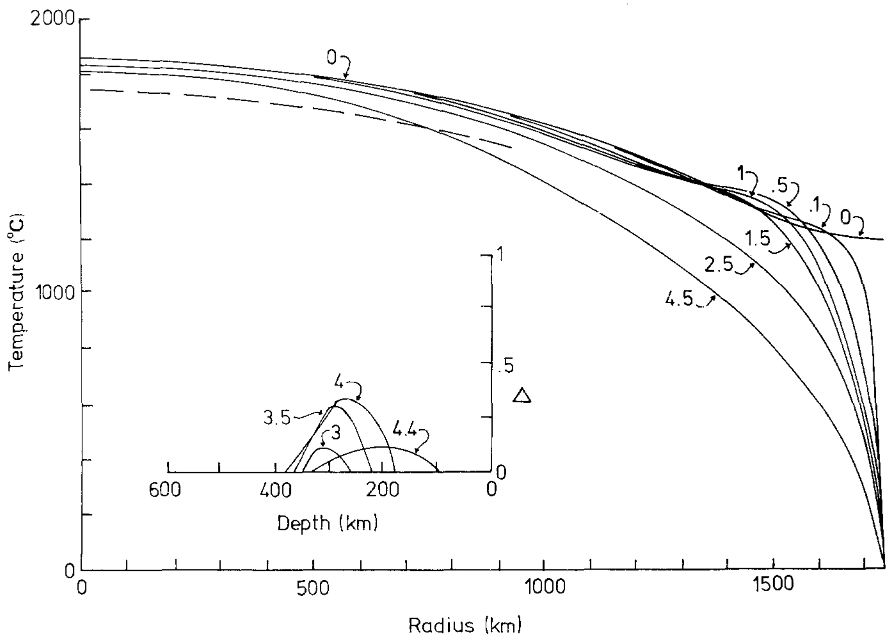

Fig. 2a. Thermal profiles for model Moon 2a, otherwise same as Figure 1a. See text for further details.

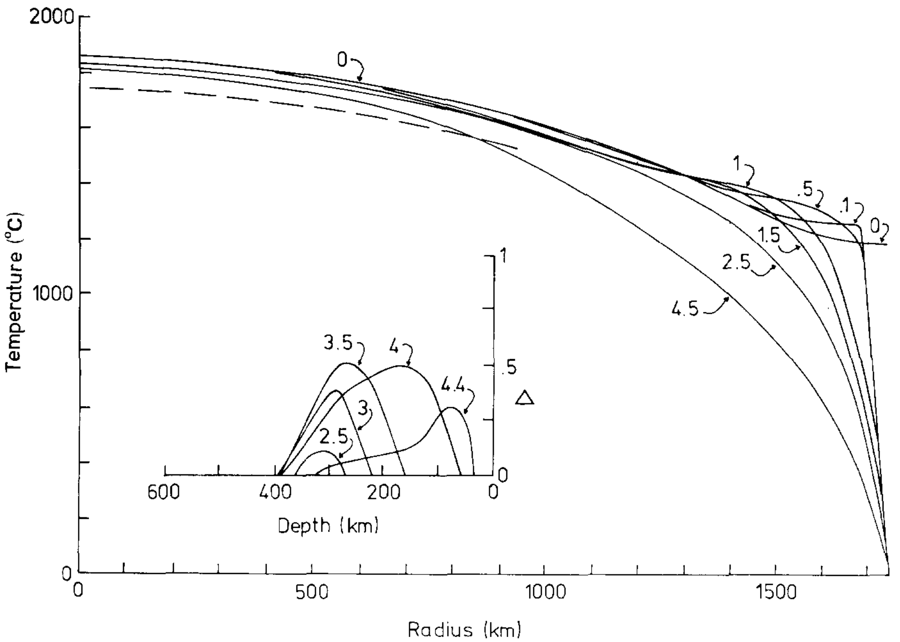

Fig. 2b. Thermal profiles for model Moon 2b, otherwise same as Figure 1a. See text for further details.

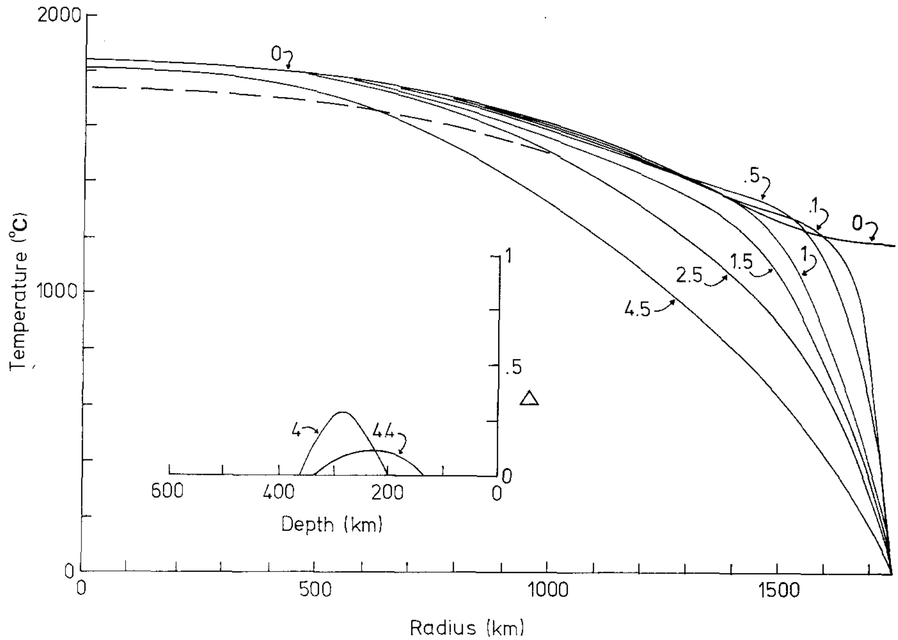

Fig. 3. Thermal profiles for model Moon 3b, otherwise same as Figure 1a. See text for further details.

No consideration was given in this investigation to the possibility that solid state convection played a role in the thermal history of the Moon. This is the case since models (e.g., Toksöz and Solomon, 1973) which include solid state convection have present day interior temperatures which are far too cool to be able to account for the partial melting or near solidus temperatures implied by the seismic data (e.g., Dainty *et al.*, 1975).

Finally, all models which can be considered to be representative of the real Moon must have degrees of partial melting of their upper mantles in the range of 20–50% (paper V) during the time interval from about 3.3 to at least 4×10^9 years ago in order to account for the mare basalt epoch. This partial melting has to occur in the depth range of 60–200 km for the most probable models (papers V and IV) and 60–300 km for less likely models (papers I, II, and IV). Also the current temperature of the model must be at near solidus temperatures or slightly above the solidus for depths greater than 800 km (Nakamura *et al.*, 1974; Dainty *et al.*, 1975) in order to account for the zone of low Q and/or partial melting in the deep lunar interior.

4. Thermal Models

The following are discussions of several pairs of thermal history models, each pair of which was computed to evaluate the effects of the variation of one or more parameters on the thermal evolution of a fissioned Moon. The members of each pair differ only in that in the first (model a) the KREEP layer is assumed to have mixed into the crust

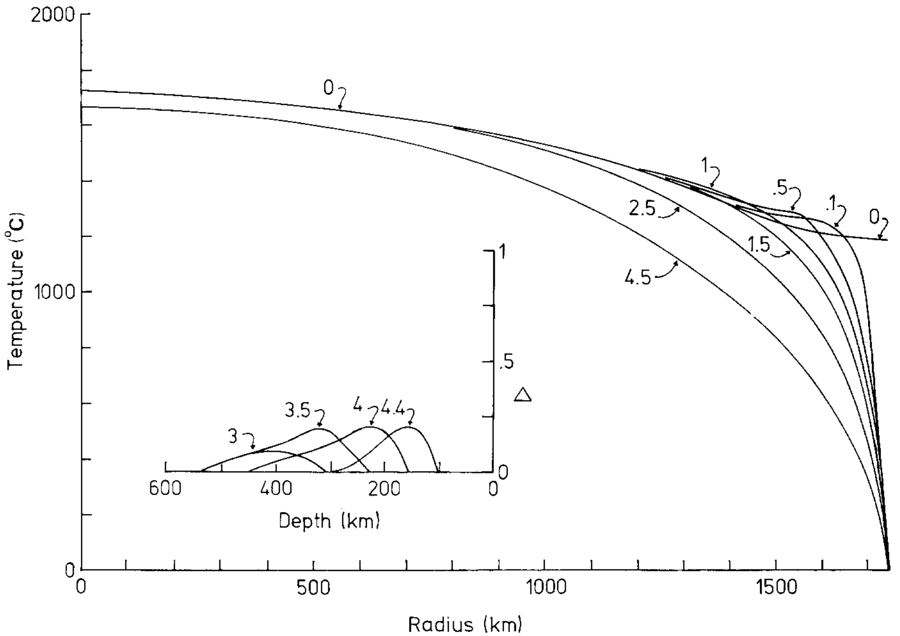

Fig. 4a. Thermal profiles for model Moon 4a, otherwise same as Figure 1a. See text for further details.

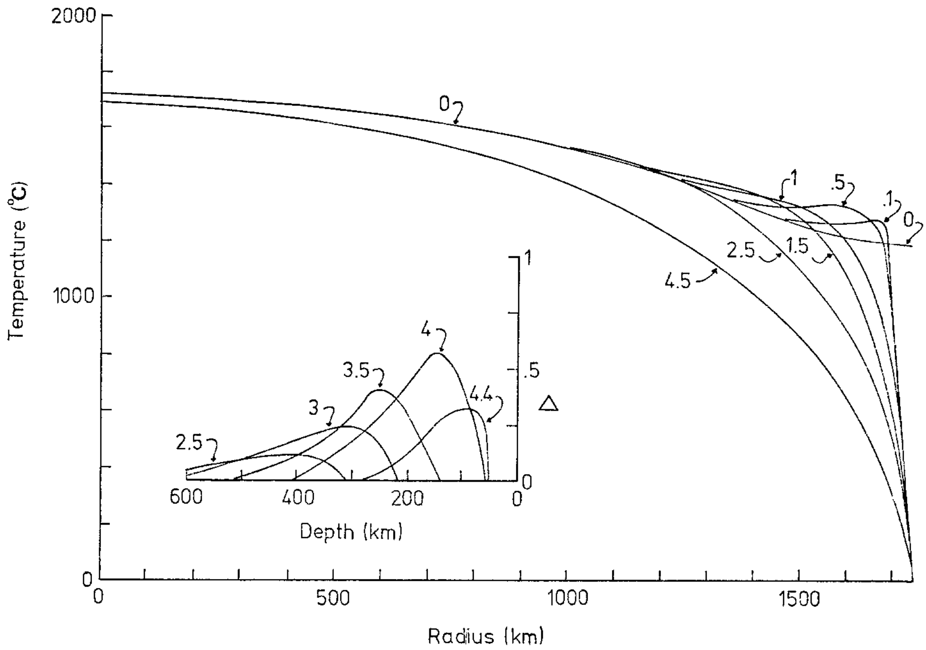

Fig. 4b. Thermal profiles for model Moon 4b, otherwise same as Figure 1a. See text for further details.

10^8 years after $t = 0$ and in the second (model b) this event occurred 5×10^8 years after $t = 0$ (see Section 3B). The parameters varied are (1) the thermal conductivity of the crust, (2) the amount and distribution of U, Th, and K in the upper and lower mantle, (3) the solidus temperatures in the center of the Moon, and (4) the thickness of the upper mantle. All other parameters are constant and are as described in Section 3.

A. MODELS 1a AND 1b

These models were computed using the baseline assumptions that the conductivity of the crust is 2×10^5 ergs cm^{-1} s^{-1} K^{-1}, that the lower mantle is devoid of U, Th, and K, and that the upper mantle extends to a depth of 300 km. As shown in Figures 1a and b and as clarified in the inserts which give the degree of partial melting (Δ) vs depth for times at 4.4, 4, 3.5 and 3×10^9 years ago, the interior of both models 1a and 1b are presently 350 °C below the solidus and hence are far too cold to be able to account for the low Q zone below depths of 800 km. Model 1a, in which the KREEP layer and crust were mixed together 4.4×10^9 years ago, is not capable of producing the full range of partial melting (20–50%) required to account for all the mare basalt magmas in the period between 3.3 and 4×10^9 years ago.

Model 1b, in which the KREEP materials were mixed into the crust at 4×10^9 years ago, easily has sufficiently high degrees of melting (~ 50%) to account for the full range of mare basalt magmas if the source region extends to a depth of 300 km and may be able to account for the magmas if it extends only to a depth of 200 km (paper V).

The difference between the melting behavior of the upper mantles of models 1a and 1b (and all other model pairs) is due to the blocking effect of the heat generated by the high concentration of U, Th, and K in the KREEP layer between 55 and 60 km depths. As long as this layer is intact, a large amount of heat is generated at this depth and this prohibits the development of a steep thermal gradient between top of the mantle (depth $\geqslant 60$ km) and the surface. As long as there is no steep gradient between the upper mantle and surface, the heat (1) generated in the upper mantle, (2) that which is conducted into the upper mantle from the lower mantle, and (3) in some cases, that which is conducted into the upper mantle from the KREEP layer can not really get out and the temperature of the upper mantle rapidly elevates. The sooner this blocking effect of the KREEP layer is removed, the quicker a steep gradient is established in the upper part of the Moon and the smaller is the amount of partial melting in the upper mantle. Thus, model 1b has higher degrees of partial melting in the upper mantle and the partial melting occurs over a longer period of time than for the 1a, an effect which is true for all a and b model pairs.

In summary, while the 1b model can account for the generation of the mare basalt magmas, its deep interior is too cool at present to account for the low Q zone below a depth of 800 km and model 1a fails on both accounts.

B. MODELS 2a AND 2b

This pair of models is similar to 1a and 1b except that at all depths below 800 km, interstitial melt (10%) is trapped in the matrix of crystals. As discussed in Sections 3B and 3D the amount of interstitial liquid carries 1.5% of the total U, Th, and K of the Moon and the solidus (dashed lines in Figures 2a and b) of this liquid is assumed to be 100 °C below the initial temperature.

As can be seen by comparing Figures 2a and b with 1a and b, respectively, the early thermal history and partial melting of the upper parts of models 2a and 2b are essentially the same as for the corresponding 1a and 1b models. The essential difference is that, due to the addition of a small amount of U, Th, and K to the deep interior, the present temperatures below 800 km are generally above the solidus and thus, these models can easily account for the small amount of melt or the low Q of the deep lunar interior.

As is the case for the 1a model, the 2a model does not account for the generation of all the mare basalt magmas. Model 2b does properly model the thermal history of the Moon, both from the stand point of the generation of the magmas as well as the modelling of the low Q zone in the deep lunar interior.

C. Model 3a, 3b and Related Models

Models 3a and 3b are similar to models 2a and 2b except that the conductivity of the crust was computed using the Schatz and Simmons (1972) relationships (see Equations (2), (3), and (4)). At the temperatures found in the crust, these relationships give K between 3 and 4×10^5 ergs cm^{-1} s^{-1} K^{-1} for most of lunar history. These values are 50 to 100% higher than the value of 2×10^5 ergs cm^{-1} s^{-1} K^{-1} found for crustal rocks as discussed in Section 3F. Though relatively small, this increase in K has a dramatic effect on the lunar thermal history and clearly shows that the general practice of using mantle values of K for the crust leads to unrealistic models.

Figure 3 gives the temperature curves for model 3b and can be compared directly with Figure 2b. It is clear from such a comparison that the increased thermal conductivity of the crust of model 3b compared to that for model 2b allows the upper part of the Moon to cool off very rapidly and that even the blocking effect of the heating in the KREEP layer (which in model 3b, like all b models, remains intact until 4×10^9 years ago) could not keep the upper mantle hot enough to account for the generation of the mare basalt magmas. The situation is even worse for the 3a model in which the KREEP layer is mixed into the crust 4.4×10^9 years ago and the KREEP heating blockage is essentially absent. In this case (not shown) the upper mantle never reaches super-solidus temperatures and thus, as expected, model 3a is even a poorer approximation to the lunar case than is model 3b.

However, it is interesting to note that, despite the strong cooling effect of the relatively high K of the crust on the lunar upper mantle, the relatively small amounts of U, Th, and K in the lower mantle do succeed in keeping the temperature high enough to be able to account for the partial melting and low Q zone below 800 km in model 3b.

Because of the strong effect on the models, due to relatively small changes in the thermal conductivity of the crust, an additional pair of models were computed in order to evaluate the effects on the models of the addition of a 20 km thick zone of brecciation with $K = 1 \times 10^5$ ergs cm^{-1} s^{-1} K^{-1} (see Section 3F) at the top of the crust. These models (not shown) show very large degrees of melting (up to 70%) to a depth of 400 km over the period from 2 to 4.5×10^9 years ago – regardless whether or not the KREEP layer was mixed early (case a) or later (case b) into the crust. These models are clearly too hot to correctly account for the more limited degrees of partial melting ($\leqslant 50\%$) which produced the mare basalt magmas, the shut off of mare volcanism at $\sim 3.3 \times 10^9$ years ago and the limited depth ($\lesssim 200$ km) of the source region of the mare basalt magmas – at least within the modelling constraints using in this study. However, it is clear that the insulating effects of the crust and especially the zone of brecciation in the outer crust can dominate the thermal evolution of the Moon, a point which is discussed further in Section 5.

D. MODELS 4a AND 4b

The final pair of models which we will describe are those in which the upper mantle extends only to a depth of 200 km and hence which correspond to the petrologic model of a fissioned Moon described in paper VI. In addition to the decrease in the maximum depth of the upper mantle and the corresponding increase in the U, Th, and K in that region (see 3B), models 4a and 4b contain 10% interstitial fluid throughout the lower mantle. Also, the initial temperature profile used for these models is that of the pyrolite solidus (see Section 3C) and as such, it is assumed that there is no temperature difference between the initial temperature and the solidus temperature of the interstitial melt. Further, the 40 ergs gm^{-1} of heat of fusion is ignored since previous modelling showed that the heating due to the small amount of U, Th, and K in the deep interior totally over-shadows the effects on the models of this small amount of heat of fusion. However, all these effects are minor in comparison to structural changes in the model and even this has only a small effect on the thermal histories of models in comparison to models 2a and 2b.

As is the case for model 2a, the present temperature of the deep interior of model 4b is just below its initial temperature at depths below 800 km and as such the low Q or partially molten zone below these depths is accounted for, but the upper mantle cools off too quickly to be able to account for the generation of the mare basalt magmas.

In contrast, model 4b, like model 2b, has both a currently hot interior and partial melting of up to 50% at depths of $\lesssim 200$ km occurs during most of the mare filling epoch. As such, the thermal history of model 4b also adequately approximates that of the real Moon.

5. Discussion

According to the above descriptions, models 2b and 4b successfully account for the thermal history of the Moon. While this is true in the general sense, these models do have at least two deficiencies. First, as discussed in Section 3B, the assumption that the KREEP layer was intact until 4×10^9 years ago is probably a limiting case. It is more likely that the intense early modification of the crust by cratering and the migration of the low melting KREEP material from the KREEP layer into the crust (papers I and IV) was a continuous process which started at 4.5×10^9 years ago and ended about 4×10^9 years ago. Hence, the thermal curves and partial melting curves for the real Moon are most probably intermediate between those of the a and b model pairs. Secondly, as shown in Figures 2b and 4b, the degrees of partial melting at $\sim 3.3 \times 10^9$ years ago at depths of 200 km are small and, hence, these models do not properly account for the generation of the Apollo 12 and 15 magmas which were probably produced by 35–45% partial melting at 3.3×10^9 years ago (paper V). Thus, from both standpoints, the 2b and 4b models are a little too cool unless the magmas did come from depths as great as 300 km, though this seems unlikely (paper V). However, as shown by the modelling, the temperatures and degrees of partial melting of the mare basalt magma source region are sensitive not only to the time scale of the redistribution of the KREEP materials, but also *very* sensitive to the thermal conductivity of the crust. As such, it is clear that the models can be fine-tuned, by varying slightly the redistribution of U, Th, and K in the crust and KREEP layer and the thermal conductivity of the crust, in order to exactly produce the required sequence of partial melting in the upper 200 km of the Moon.

Further, the extreme sensitivity of the thermal history of the outer 200 km of the Moon to the exact distribution of U, Th, and K and the thermal conductivity of the crust, indicates the important role that regional variations in these parameters must play in the generation of the mare basalt magmas. As discussed in paper III, it is most likely that, due to variations in the thickness of the original KREEP layer and due to the redistribution of KREEP by large basin forming impacts, the U, Th, and K content of the crust and remnants of the KREEP layer varies on the 100 to 1000 km scale by at least a factor of ± 3. Arkani-Hamed (1974) has shown that similar variations in the U, Th, and K content of the crust, due to impact basin formation, as well as the simultaneous formation of insulating ejecta blankets around the basin do promote partial melting in the upper mantle in the immediate vicinity of the basins. Conel and Morton (1975) have also investigated the insulating effects of basin ejecta blankets and find that they significantly alter the thermal history of the basin regions. It is, therefore, clear that the generation of the mare basalt magmas, i.e., their depth of origin, the time scale and the degrees of partial melting as modelled in this paper, are certainly influenced by the basin forming events. Since this is the case, it is not necessary to reproduce the exact mare basalt melting sequence with spherically symmetric models alone. As such, models 2b and 4b are considered to adequately reproduce the basic thermal history of the Moon.

6. Heat Flow

As discussed in paper III, if, as is generally assumed, that the Earth and the Moon are in thermal steady state (Langseth *et al.*, 1973), simple scaling indicates that a Moon of fission origin must have a present day heat flow of about 13 ergs cm^{-2} s^{-1}. However, the heat flow values derived from the various models computed for this study range from 17.8 to 18.6 ergs cm^{-2} s^{-1} and average 18.2 ergs cm^{-2} s^{-1}. The 5 ergs cm^{-2} s^{-1} increase in the heat flow of the models over that produced in a steady state condition from the decay of radioactive U, Th, and K simply indicates that the Moon is still losing part of the heat it had when it formed. This can be readily understood from inspection of the figures of this paper and by noting that, due to the initial differentiation of the Moon, almost all the U, Th, and K are located in upper mantle and crust. As can be deduced from the figures, the relatively large amounts of heat which were generated in the upper 200–300 km of the Moon blocked the flow of heat from the interior by keeping the thermal gradient at depths below 300 km small until about 2×10^9 years ago. Since that time, the gradual increase in the gradient has reached increasingly larger depths in the lower mantle and the heat of the deep interior is being released.

In addition to the various constraints and boundary conditions used in the development of the models, the Apollo 15 and 17 heat flow measurements (Langseth *et al.*, 1976) provide an additional test of the validity of the models. According to Langseth *et al.*, the heat flow at the Taurus–Littrow area and the Hadley–Apennine area is 16 ± 4 ergs cm^{-2} s^{-1} and 22 ± 5 ergs cm^{-2} s^{-1}, respectively. The mean of these two sets of measurements, i.e., 19 ergs cm^{-2} s^{-1}, is essentially the same as the 18 ergs cm^{-2} s^{-1} heat flow derived from the modelling presented in this paper. Thus, the observed heat flow and that predicted for a Moon of fission origin are in good agreement.

However, following the discussion given in paper III, the estimates of the bulk U, Th, and K content of a fissioned Moon are based on the assumption that at least the Earth is in thermal steady state (Langseth *et al.*, 1973). If the Earth is not in a steady state, but like the Moon is cooling, then the analysis presented in paper III gives an upper limit to the U, Th, and K content of the Moon and the 18 ergs cm^{-2} s^{-1} model heat flow is an upper limit for that of the average Moon. While a detailed discussion of the thermal history of the Earth, its current thermal state, and the implications therefrom for the U, Th, and K content of a fissioned Moon are well beyond the scope of this paper, if 25% of the current terrestrial heat flow were due to original heat, then by scaling, the current lunar heat flow would be about 15 ergs cm^{-2} s^{-1}, i.e., only slightly below the 18 ergs cm^{-2} s^{-1} derived assuming the Earth is in steady state.

7. Conclusions

Based on the models developed in this study, we find that the thermal history of a Moon of fission origin is consistent with that of the Moon as we now understand it. In particular, (1) the current solidus or near solidus temperature of the deep lunar interior, (2) the generation of the mare basalt magmas (the time scale, depth of generation, and degree of partial melting), and (3) the heat flow of the Moon are all accounted for by these models. As such, these results add further support to the hypothesis that the Moon originated by fission from the proto-Earth.

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