ORIGIN, EVOLUTION, AND PRESENT THERMAL STATE OF THE MOON*

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ABSTRACT

The relative absence of lunar volcanism in the last 3 b.y. and the Apollo 15 heat flow measurement suggest that present-day temperatures in the Moon are approximately steady-state to depths of ~ 100 km. An exponential distribution of heat sources with depth may then be scaled by equating the surface heat flow to the integrated heat production of this exterior shell. Presumed present-day interior temperatures, as well as the present-day surface heat flow ~ 30 ergs/cm^2-sec, may be obtained with an initial temperature roughly corresponding to the Apollo 11 basalt solidus, the exponential scaling of heat sources, and a parameter $Q/U_oK = 1.6^{\circ}K/ppm(U_o)\text{-cm}$ [U_o = surface concentration of U in ppm, K = average interior thermal conductivity (ergs/cm-$^{\circ}$K-sec) and Q = present-day surface heat flow (ergs/cm^2-sec)]; the nonuniqueness is constrained by the observations of U_o and Q and inferences concerning K. The "best" models require strong concentration of heat sources in the upper 100-200 km, within the $\sqrt{K\tau}$ depth of ~ 300 km for which the buried heat sources may be felt at the Moon's surface. The concentrations of U for an originally homogeneous Moon are estimated to be ~ 9 x 10^{-8} gm/gm, close to that measured for eucrites and inferred for primitive inclusions of the Allende meteorite.

The estimated homogeneous concentrations of U, the chemistry of the lunar surface material and inferences to modest depth, and the short accretion time of the Moon necessary to provide large-scale differentiation at 4.6 AE suggest that the Moon had its origin in the rapid accretion of compounds first condensing from the protoplanetary nebula. Accretion of the Earth and Moon may well have kept pace with

condensation in the originally hot nebula. In the later stages of accretion-condensation, the Moon competed unsuccessfully with the more favorably disposed Earth.

The present thermal state of the Moon may well involve at least some partial melting through all of the lunar interior deeper than 200 km. This would eliminate the large density changes which would otherwise occur for Ca-Al rich compositions at depth. Such a present-day thermal configuration is neither inconsistent with temperatures inferred from electrical conductivity studies nor with the nonhydrostatic shape of the Moon. In the first place, the lunar interior is probably more deficient in total Fe than had previously been suspected, and in the second place the Moon is closer to being in hydrostatic equilibrium than is the Earth. The lack of present-day volcanism and the remarkable aseismicity of the Moon need only reflect the absence of plate motions, as they are known on the Earth, rather than a "cold" interior. For otherwise "hot" interiors, plate motions seem less likely on the Moon than on the Earth for simple geometrical reasons.

1. Introduction

The blossoming of definitive thermal histories for the Earth, the Moon and the other terrestrial planets has been plagued by several well-known difficulties. The more serious of these involve arbitrary and/or inadequate treatment of the initial temperature, convective heat transport, and the variation of thermal conductivity and the radioactive heat sources with depth. These uncertainties, however, may be at least partially offset by the application of geochemical and geophysical constraints to the origin and evolution of the planetary body under consideration. This approach has been used with some success to restrict a number of the possibilities for the origin and thermal evolution of the Earth (LUBIMOVA, 1958; MACDONALD, 1959; BIRCH, 1965; RINGWOOD, 1966; ANDERSON and PHINNEY, 1967; HANKS and ANDERSON, 1969; MURTHY and HALL, 1970; TUREKIAN and CLARK, 1969; ANDERSON _et_ _al_., 1971). In fact, it is fair to say that our present understanding of the Earth's early history is more firmly based on the qualitative understanding of these constraints than on the formalities of the actual temperature calculations.

We may reasonably expect that future development of new geophysical and geochemical constraints will prove even more fruitful in narrowing the possibilities for the Earth's early history. Even so, two fundamental limitations suggest that many aspects of the Earth's origin and initial evolution are forever lost to investigation at the present time. First, the Earth's geological record does not exist in recognizable form for times greater than $\sim$ 4 AE. Thus we are afforded not even observations of the immediate consequences of the Earth's early history. Second, there is accumulating evidence that the time scale of the global tectonic processes is such that convective overturn of the oceanic crust and upper mantle may have occurred many times since the Earth's origin. At best the continental masses have incomplete records of only several of these events, let alone a complete record of all of them. While this recognition is sufficient to condemn the general class of conventional thermal history models for the Earth, we are yet without the necessary apparatus to solve the formidable problem of nonsteady upper mantle convection. While current investigations of mantle convection will undoubtedly afford us some understanding of the past several hundred million years of the Earth's evolution, it is not overly pessimistic to assume that such calculations can never be reliably traced to the Earth's early history.

More recently, as a result of the Apollo missions and subsequent analysis of the lunar samples, a number of geophysical and geochemical constraints have been placed on the origin and thermal evolution of the Earth's moon. PAPANASTASSIOU and WASSERBURG (1971) provide a summary of the salient constraints (as of March 1971) and, without the benefit of formal thermal history calculations, inferred an origin and initial evolution of the Moon. TOKSÖZ _et al_. (1972) have generated conventional thermal history models on the basis of these and more recently obtained constraints, notably the present-day surface heat flow determination of 33 ergs/cm^2-sec at the Apollo 15 site (LANGSETH _et al_., 1971). Thermal history calculations provided some useful predictions of the evolution of the Moon prior to the Apollo landings. ANDERSON and PHINNEY (1967), on the basis of such calculations, suggested an early and extensive period of differentiation, a basaltic surface composition and a high surface concentration of uranium.

In fact, the Moon is a rather attractive subject for conventional thermal history calculations. Geologic manifestations of the Moon's early history are recorded in the rocks of age up to and including 4.6 AE. More importantly, the evidence for large scale melting of the outer reaches of the Moon at 4-3 AE but relative absence thereafter suggests that thermal history models based on conduction heat transport may even be appropriate, at least for the outer reaches of the Moon. (SODERBLOM and LEBOFSKY (1972) on the basis of a small impact erosion model, have suggested that this igneous episode, responsible for the major mare formations, extended until 2.8 AE and that smaller localized volcanic features are considerably younger.) If conduction heat transport is the main mechanism of energy transfer, the small size of the Moon allows heat generation of a significant fraction of the planetary volume to be felt at the Moon's surface in a readily calculable way. Even these casual observations provide us with a rationale for scaling the distribution of heat sources to depths of $\sim$ 100 kilometers in the Moon. We shall develop this idea in more detail in the next section.

The approach is admittedly circuitous. In the next (second) section, we scale several exponential heat source distributions and in the third section investigate them in terms of the resulting present-day interior temperatures and surface heat flows, using several values for K. The approach to this point excludes, in general, the effects of the early thermal history of the Moon, and we pick up this thread in the fourth section. With an estimate of the average uranium concentration of an originally homogeneous Moon and the remarkably short accretion time necessitated by the extensive differentiation at 4.6 AE, we propose in the fifth section an origin for the Moon in terms of a rapid accretion of particles first condensing from the protoplanetary nebula. Much of the argument here involves the chemistry of the lunar surface material and inferences to modest depth, the similarities between the lunar surface material and the basaltic achondrites, and the estimate of what compounds would first condense from an originally hot planetary nebula. In the sixth section, we return to the present thermal state of the Moon, specify what we "presume" to be present-day interior temperatures, and show that neither the nonhydrostatic shape of the Moon, present-day temperatures inferred from conductivity studies, the lack of present-day volcanism, nor the aseismicity of the Moon are incompatible with them.

2. The Scaling of the Lunar Concentrations of U, Th, and K

The evidence for extensive volcanic and igneous activity in the outer reaches of the Moon up to ~ 3 AE and its relative absence thereafter (PAPANASTASSIOU and WASSERBURG, 1971) is our primary constraint on the Moon's thermal history. Plainly, it suggests that mass transport may not be an important feature of the evolution of the Moon's outer several hundred kilometers in the past 3 b.y., thereby implying more than the normal credence for temperature calculations based solely on conduction heat transport. Moreover, it suggests that the temperature profile in this region is near steady-state. Certainly it is not getting hotter, and if the Apollo 15 heat flow measurement (LANGSETH *et al.*, 1971) is grossly representative of the lunar average, temperatures of 700-800°K are indicated at depths of ~ 100 km.

We begin with the steady-state approximation for temperatures in the Moon to depths of ~ 100 km. Then the heat lost through the Moon's surface must be approximately balanced by the heat generation in the region for which the steady-state temperature approximation is valid. The high surface concentrations of U, coupled with the evidence for extensive differentiation of the Moon at 4.6 AE (PAPANASTASSIOU and WASSERBURG, 1971) and again at 3.7-2.8 AE suggest that the concentration of heat sources decreases with depth. Following LACHENBRUCH's (1970) result, we assume that the concentration of heat sources decreases exponentially with depth. Then

$$Q = \frac{H_o}{R^2} \int_{r_o}^{R} r^2 e^{-\lambda(R-r)} dr \tag{1}$$

Here H_o is the surface concentration of heat sources, R is the Moon's radius, r_o is the inner radius of the shell for which the exponential distribution of heat sources is valid, and λ is the (spatial) decay constant, to be determined upon specifying r_o.

Several lines of evidence suggest that $R-r_o$ should be in the vicinity of 200-400 km or greater. PAPANASTASSIOU and WASSERBURG (1971) concluded that a major fractionation of the Moon occurred at 4.6 AE, resulting in the formation of the high K, Rb, U, and Th crust and the rare earth element abundance patterns. Presumably such a fractionation involved at least the outer several hundred kilometers of the Moon. RINGWOOD and ESSENE (1970) estimated that Apollo 11 basalts may have been formed by partial melting at depths of 200-400 km. A choice of $R-r_o$ of 200-300 km would also be indicated if the accretion models of HANKS and ANDERSON (1969) are invoked to provide melting upon accretion of the Moon. Uncompressed initial temperatures peak at .85 of the planetary radius for this accretion model. The accretion models for the Moon developed by MIZUTANI _et_ _al_. (1972) yield similar but slightly lower choices for $R-r_o$.

It does little good, however, to put r_o at depths greater than $\sim\sqrt{\kappa\tau}$, where κ is the thermal diffusivity and τ is the time scale for which the steady-state assumption must be approximately valid; the effects of heat sources at greater depths have not been significantly transmitted to the surface. Then,

$$R - r_o \leq 300 \text{ km} \tag{2}$$

for κ, thermal diffusivity, = .01 cm^2/sec and τ = 3 b.y. This choice for τ, while excluding the earlier igneous activity, may nevertheless be an overestimate for the time for which the steady-state approximation is valid. Below, we evaluate (1) for three choices of $R-r_o$: 100, 200, and 300 km.

H_o can be expressed in terms of the surface concentrations of uranium, U_o. Using the known relative abundances and unit heat production of the isotopes U^{238}, U^{235}, Th^{232}, and K^{40}, we may write

$$H_o = \left[.73 + .20 \frac{Th}{U} + .25 \times 10^{-4} \frac{K}{U}\right] 1.35 \rho U_o \tag{3a}$$

Here U_o is in ppm, and H_o is in ergs/cm^3-sec. With $\frac{Th}{U} = 4$, $\frac{K}{U} = 2 \times 10^3$ (for example, PAPANASTASSIOU and WASSERBURG, 1971) and ρ = 3.3 gm/cm^3, (3a) reduces to

$$H_o = 7.3\, U_o \tag{3b}$$

HAYS (1971) provides a convenient summary and useful discussion of estimates for the lunar concentrations of U^{238}, U^{235}, Th^{232}, and K^{40}.

It remains to specify U_o. Measurements on lunar samples have yielded values ranging from 0.2 ppm up to several ppm. HAYS (1972) summarizes these results. We will evaluate (1) for three choices of U_o: U_o = .25, .50, and 1.0 ppm. Subsequent temperature calculations will allow us to understand the effect of higher values of U_o.

Table 1 summarizes the evaluation of (1) and the determination of λ for the three choices of r_o and three choices of U_o discussed above. The general result of Table 1 is that, for a given value of Q, increasing U_o and $R-r_o$ lead to increasing λ's; that is, the concentration of heat sources decays more rapidly with increasing depth. The starred elements of the matrix indicate that no positive λ exists for that combination of U_o and r_o; that is, the concentration of heat sources must increase with depth. We exclude these possibilities from the outset.

We will investigate the consequences of the heat source distribution models summarized in Table 1 in the temperature calculations presented in the following section. We emphasize that these models of heat source distribution with depth serve only as a point of departure. While we feel that the line of reasoning followed is attractive, the resulting heat source models are certainly not unique and need not be correct. The best these heat source models can be is acceptable, and to be so they plainly must satisfy all other available constraints on the Moon's thermal history.

3. Thermal History Models for the Moon

For all of the thermal history models presented in this study, we take the planetary radius R = 1740 km, the surface temperature T(R) = 250^oK, and specific heat at constant pressure = 1.3×10^7 ergs/gm-K^o. We shall assume that the average present-day surface heat flow, Q, is given by the value obtained at the Apollo 15 site, 33 ergs/cm^2-sec. Further reference to Q shall always mean the present-day value, having units of ergs/cm^2-sec. Interior densities are scaled to match the mass and moment of the Moon, but are in the range 3.3-3.4 gm/cm^3, increasing with depth. The principal heat source variable is U_o, heat source concentrations at depth being determined by the scaling described in the previous section; $\frac{Th}{U} = 4.0$ and $\frac{K}{U} = 2 \times 10^3$. Generally, we will assume that the exponential scaling is valid throughout the lunar interior, despite the fact its "validity" is physically restricted by (2). An assessment of its validity, of course, will be very much a part of the discussion. Following arguments summarized by HAYS (1972), we begin with models for which the thermal conductivity K is uniform throughout the Moon with a value of 4.2×10^5 ergs/cm-^{o}K-sec. Subsequently, we will investigate the effects of both higher and lower interior thermal conductivities, and low thermal conductivity surface layers.

The calculations presented herein have been effected on a program originally written by R. A. Phinney and employed for similar purposes by PHINNEY and ANDERSON (1965,1967), ANDERSON and PHINNEY (1967), and HANKS and ANDERSON (1969). The resulting temperature calculations are for a radially inhomogeneous sphere. For the calculations presented here, the Moon has been subdivided into 50 layers, and temperatures are recalculated every 5×10^6 years. There is no formal accounting of melting phenomena in the temperature calculations presented below.

In the thermal history models presented in this section (Figs. 1-5) as well as Fig. 7, we take the initial temperature to be a profile with constant radial derivative with T (R,0) = 1353^{o}K. In the outer half of the Moon, this corresponds to the Apollo 11 basalt solidus (RINGWOOD and ESSENE, 1970), both in the absolute temperatures as well as the gradient. In the deep lunar interior, this initial temperature profile has a considerably larger gradient and overestimates the extrapolated basalt solidus by several hundred degrees at the center of the Moon. What we have in mind here is to overestimate heat flow from the deep interior to obtain a maximum deviation from "solidus" temperatures, for a deep interior generally deficient in heat sources. We shall return to this situation in a subsequent discussion of the present thermal state of the Moon.

We can rationalize this initial temperature distribution for the outer reaches of the Moon in terms of the evidence for large-scale differentiation at 4.6 AE, but the same assumption may considerably overestimate the initial temperatures in the deep interior, if the accretion models of TER HAAR (1948), HANKS and ANDERSON (1969) and/or MIZUTANI _et_ _al_. (1972) are appropriate to the Moon's formation. We do not consider this problem too critical, since thermal history calculations for the deep planetary interior are by and large a rearrangement of several assumptions, the most important of which is the initial temperature. We do, however, wish to avoid models that lead to large-scale melting in the lunar interior at the present time. To do so effectively, we will need to consider the initial temperature and lunar accretion process in more detail. We will reserve further discussion of the initial temperatures for the next section.

3.1 *The Effects of the Heat Source Distribution*

The Moon of Fig. 1 is a model for which the heat source concentrations have been set equal to zero. Q for this Moon is 5.7. We present this model to illustrate that the initial temperatures, in and of themselves, provide only a small fraction of Q, requiring that most of the heat flow through the surface be generated by near-surface heat sources. This model, of course, has cooled below the initial temperatures everywhere.

Figs. 2a, 2b and 2c are for the cases U_o = .25, .50, and 1.0 ppm with associated λ's of $.20 \times 10^{-2}$, $.90 \times 10^{-1}$, and $.20 \times 10^{-1}$ km^{-1}, respectively. These λ-values bracket the six values given in Table 1. For Figs. 2b and 2c, the effect of assuming the exponential scaling throughout the lunar interior is small, since the heat source concentrations at depths greater than 300 km is negligible. For Fig. 2a, however, it is plain that the deep burial of significant concentrations of heat sources (because of the small λ) drives temperatures to an unacceptably high level. ANDERSON and PHINNEY (1967), PAPANASTASSIOU and WASSERBURG (1971), HAYES(1972), and TOKSÖZ et al. (1972) have all noted that the deep lunar interior connot be subsolidus and possess U concentrations $\gtrsim$ 25-30 ppb. The interior temperatures of Fig. 2a are a variation of the same theme. Q for this model is also unacceptably high, 53. Setting the heat source concentrations of Fig. 2a equal to zero for $(R-r_o) \gtrsim 300$ km (Fig. 3) alleviates the high temperature problem somewhat, but not enough to avoid large-scale present-day melting in the range $.4 \lesssim \frac{r}{R} \lesssim .9$.

For Figs. 2b and 2c, the Q is 35 and 30, which agree well with the observed value of 33. This is not surprising because the heat sources were scaled to do just that. The steady-state approximation is vindicated in these cases because the heat sources lie sufficiently near the surface. For Figs. 2a and 3, however, the heat source concentration at depth is too great to prevent substantial temperature rises in the range $.7 \lesssim \frac{r}{R} \lesssim .9$.

The present-day temperatures of Fig. 2b exceed the "solidus" in the range $.5 \lesssim \frac{r}{R} \lesssim .8$, but not by more than 30°K. The present-day temperatures of Fig. 2c are sub-"solidus" throughout the Moon. Model 2b, however, should have experienced at least extensive ($.7 \lesssim \frac{r}{R} \lesssim .9$) partial melting prior to 3.0 AE. The extensive igneous activity at ~ 3.5-3.0 AE may have been the result.

3.2 *The Effects of Thermal Conductivity*

Fig. 4b is for the Moon of Fig. 2b with $K = 2.1 \times 10^{5}$ ergs/cm-$^{\circ}$K-sec, half of the value used to obtain Fig. 2b. It should provide for large-scale melting for $.7 \lesssim \frac{r}{R} \lesssim .9$ at the present, a situation difficult to reconcile with the relative lack of evidence for large-scale igneous activity since ~ 3 AE, unless a cool exterior shell thicker today than in the past presents a barrier to extrusion. The difficulty *is that we have to tolerate 50-100% variations in K, yet such* variations have a dramatic effect on the present-day lunar temperatures (~ 500°K between Fig. 2b and Fig. 4b at $\frac{r}{R} \approx .8-.9$).

Quite simply, the reduced thermal conductivity of Fig. 4b initially retards (relative to Fig. 2b) the flow of heat from the shell of high concentration of heat sources. In like measure, near-surface temperatures and temperature gradients are driven higher, thereby increasing the surface heat flow, in turn limiting the further rise in temperature. The variation is strongly buffered; the surface heat flow of Fig. 4b is not less than 78% of that for Fig. 2b (for times greater than 0.5 b.y.) and at present is 91% of that for Fig. 4b, an insignificant difference. The difference in near-surface temperatures, however, is dramatic because the heat sources are placed so near the surface and their concentrations are so high.

While the previous discussion affords us some understanding of reasonable variations of K for these lunar models, it does not provide a way out of the specific difficulty of the large, present-day temperatures of Fig. 4b. The standard remedy here is to examine a few more models. Our experience with Fig. 4b suggests that the Moon of Fig. 2c might tolerate a K of 2.1 x 10^5 ergs/cm-$^{\circ}$K-sec while the Moon of Fig. 2a might benefit from a K of 8.4 x 10^5 ergs/cm-$^{\circ}$K-sec. Such models are presented in Fig. 4c and Fig. 4a, respectively.

Fig. 4c is a successful Moon in the sense of Q (28) and present-day temperatures which are generally sub-"solidus." In fact, Fig. 4c is insignificantly different from Fig. 2b with respect to its present-day temperatures, and the difference in Q is only 20%, a tolerable variation. Fig. 4a is not so successful. Present-day temperatures are still too large, as is Q (59).

3.3 *A Scaling for Lunar Temperatures*

Assuming that the actual present-day interior temperatures are close to those given in Fig. 2b or Fig. 4c, we can evidently obtain them under the condition that the quantity

$$\frac{Q}{H_0 K} \propto \frac{Q}{U_0 K} = 1.6 \frac{{}^{\circ}K}{ppm(U_0)\text{-}cm} \tag{4a}$$

is approximately constant. The factor $\frac{Q}{U_o}$ fixes the exponential scaling of heat sources (Table 1), and our experience with Figs. 2b and 4c indicates that a decrease of $\frac{Q}{U_o}$ by a factor of 2 is approximately offset by a decrease in K by a factor of 2, with respect to the present-day interior temperatures. For Q = 33 ergs/cm^2-sec, our experience thus far is that U_o should be in the range of 0.5-1.0 ppm. Higher U_o's for the same Q provide for an increasingly cooler Moon, unless K is reduced in like measure. U_o's smaller than 0.5 ppm for the same Q provide for λ's too small, that is, heat sources buried too deeply, for present-day interior temperatures to be subsolidus, even with higher choices of K. Thus we wish also to restrict

$$\frac{Q}{U_o} \lesssim 100 \, \frac{\text{ergs/cm}^2\text{-sec}}{\text{ppm } (U_o)} \tag{4b}$$

For larger values of $\frac{Q}{U_o}$, the steady-state approximation on which (1) is based becomes increasingly bad.

Again emphasizing the proviso that the present-day interior temperatures of the Moon are approximately those given by Fig. 2b or 4c and that the scaling of heat sources is approximately valid, we have in (4) surprisingly strong restrictions on the quantities Q, U_o, and K. The constraints on Q and U_o, of course, are to be consistent with the observational results. Our preference for $0.5 \lesssim U_o \lesssim 1.0$ ppm for Q = 33 is well-centered with respect to the actual determinations. We may tolerate higher average values of U_o by reducing the interior K in the same proportion, or alternatively accepting slightly cooler present-day interior temperatures. Lower values of U_o would be expected to be associated with lower values of Q. It is difficult to estimate how representative the single Q value is of the average surface heat flow for the Moon. If the actual value is somewhat lower than Q = 33 ergs/cm^2-sec, we may tolerate lower values for U_o; if the actual value is somewhat higher we must restrict the minimum U_o allowable by (4b).

It is probably more difficult to estimate likely variations in the choice of K, mainly because its average value is the most inaccessible (of the parameters U_o, Q, and K) to direct measurement. Fig. 5 illustrates the effect of a low thermal conductivity surface layer. These models are the same as Fig. 2b, except that the surface layer (~ 35 km) is given a value of 3.15×10^5 ergs/cm-^{o}K-sec (Fig. 5a) and 2.1×10^5 ergs/cm-^{o}K-sec (Fig. 5b). The models provide for present-day temperatures in excess of the solidus temperatures in the range $.5 \lesssim \frac{r}{R} \gtrsim .8$ (up to 50^{o}K for Fig. 5a; up to 200^{o}K for Fig. 5b), not a critical difference.

Our intuition here is that the effect of a low conductivity surface layer is not crucial, provided that its thickness is sufficiently less than reciprocal λ. The qualifier 'sufficiently' depends on the relative difference in the K for this layer compared to the average interior K.

4. The Early Thermal History of the Moon

In the previous sections, we have generated a scaling for the depth distribution of heat source concentrations to understand present-day near-surface temperatures and surface heat flow. The deep interior temperatures are essentially unconstrained since we have generally side-stepped problems of the Moon's initial temperatures and, in a larger sense, the Moon's early thermal history. The steady-state approximation explicitly excludes these factors, the practical assumption being that their present-day effects are small. In this section, we consider these problems in more detail.

We begin with a Moon accreting uniformly with respect to composition according to the $Ct^2 \sin \gamma t$ accretion rate described by HANKS and

ANDERSON (1969). The constants C and γ are fixed by R and the total time for accretion t_{acc} respectively. The release of gravitational energy at the planetary surface (growing in time for $0 \leq t \leq t_{acc}$) results in temperature rises in part buffered by the re-radiation of energy to space. Temperatures are calculated according to eq. (1) of HANKS and ANDERSON (1969).

This accretion model is quite arbitrary, satisfying only the crudest physical constraints on the accretion process. If for other reasons, however, one can fix minimum temperatures that should result from the accretion process, one can determine a maximum value for t_{acc}. HANKS and ANDERSON (1969) thus estimated a maximum accretion time for an originally homogeneous Earth to effect a constraint on core formation. TOKSÖZ *et al*. (1972) have employed the same model to estimate a maximum t_{acc} in order that the Moon undergo large-scale differentiation upon accretion. MIZUTANI *et al*. (1972) have developed a similar accretion model from a different point of view. The principal variables are the particle density and relative velocities in the primordial gas-dust cloud. In any case, the accretion model is constrained by conditions which are presumed attributable to it. The basic premise is that the accretion process is the dominant source of whatever energy release is necessary to effect inferred conditions very early in the planet's history.

The Moon of Fig. 6 has the initial temperatures arising from the $Ct^2 \sin \gamma t$ accretion model with $t_{acc} = 2000$ years. The initial temperature of infalling material has been taken to be 278°K. "Solidus" temperatures are exceeded in the range $.8 \lesssim \frac{r}{R} \lesssim 1.0$, and the initial temperatures are set to the "solidus" temperatures in this region.

If the accretion process is invoked to explain the evidence for large-scale differentiation at 4.6 AE, which seems to be a reasonable if not the only explanation, the accretion process was remarkably swift. We will return to this point in the next section.

We assign the following depth distribution of U, taking again $\frac{Th}{U} = 4.0$, $\frac{K}{U} = 2 \times 10^3$. For $\frac{r}{R} \lesssim 0.8$, the U concentrations are for an originally homogeneous Moon. These are determined from

$$\bar{U} = \frac{U_0 \int_0^R r^2 e^{-\lambda(R-r)} dr}{R^3/3} \tag{5}$$

Again, we have assumed that the exponential scaling is valid throughout the lunar interior. The result is approximately

$$\bar{U} \simeq \frac{3U_0}{\lambda R} \quad , \quad \lambda R \gg 1 \tag{6}$$

The two successful heat source distributions of Table 1 fulfill this restriction, and

$$\bar{U} = \begin{Bmatrix} 0.086 \\ 0.091 \end{Bmatrix} \text{ppm}; \quad \bar{U}_0 = \begin{Bmatrix} 1.0 \\ 0.5 \end{Bmatrix} \text{ppm}, \quad \lambda = \begin{Bmatrix} 0.20 \\ 0.09 \end{Bmatrix} 10^{-1} \text{km}^{-1} \tag{7}$$

In Fig. 6, then, we assign $U = 9.1 \times 10^{-8}$ gm/gm for $\frac{r}{R} \lesssim .8$. For shallower depths, we take the total abundances for $.8 \lesssim \frac{r}{R} \lesssim 1.0$ and scale them exponentially with $\lambda = .90 \times 10^{-2}$. The resulting U_o = .26 ppm. The present-day concentration of U as a function of depth is sketched in Fig. 6.

Again, the deep burial of significant heat sources (91 ppb) drives interior temperatures to unacceptably high present-day temperatures. It is interesting that Q is nevertheless quite acceptable, 32. This situation underscores the nonuniqueness as well as inherent uncertainties associated with such thermal history calculations: a very acceptable Q may be obtained with very unacceptable interior temperatures. The interior temperatures of Fig. 6 also represent (with suitable scaling) the situation arising from the heating of a homogeneous Moon. Each successive temperature rise can be scaled by $\frac{\bar{U}}{9.1}$, the cumulative result being approximately the case for any other $\bar{U}$.

The most interesting feature of Fig. 6, however, is that at 3.6 AE, the entire Moon is at temperatures very close or exceeding the "solidus" except for $\frac{r}{R} \gtrsim .9$. This result is by and large independent of the near-surface scaling of heat sources. This situation could certainly be a logical prelude to the large-scale igneous activity at 3.7-2.8 AE. The formalities of Fig. 6, however, suggest that this event may be of considerably greater importance than the differentiation at 4.6 AE, in that the later event might involve most of the Moon. It would then be questionable whether the older surface rocks would survive this event.

To follow up on the possibilities that the major event in the Moon's thermal evolution occurred between 3.7-2.8 AE, we reconsider the Moon of Fig. 2b with an "origin time" of 3.25 AE (Fig. 7). That is, we assume that the exponential distribution of heat sources is set up at this time. In fact, present-day temperatures and Q for the Moon of Fig. 7 are insignificantly different from those of Fig. 2b. The diminished earlier concentrations of U (due to the smaller origin time) are approximately compensated by introducing the "solidus" temperatures at times nearer to the present.

The apparent lesson of Fig. 7 is that we have little control on the thermal evolution of the Moon prior to 3 AE. The presumed present-day temperature distribution and surface heat flow care little if the exponential distribution of heat sources and "solidus" initial temperatures are set up at ~ 3 AE or 4.6 AE. This result, together with Fig. 6, merely admits the possibility that the mare-forming magmatic episode may have been the more important event in terms of the Moon's differentiation. How pervasive this event was relative to the initial differentiation remains to be decided on further geological and geochemical observations.

Summarizing our numerical results thus far, we have seen that the evidence for lunar melting and differentiation at 4.6 AE requires a remarkably small accretion time for the Moon, if the energy of the accretion process is required to effect melting temperatures. The major event in the Moon's thermal evolution, however, may have been the magmatic event at 3.7-2.8 AE, at which point the exponential scaling of heat sources with depth may have been set up. We can understand presumed temperatures at the present day in the lunar interior, as well as the surface heat flow determination of 33 ergs/cm^2-sec, in terms of the exponential scaling of heat sources with depth and the quantity $\frac{Q}{U_o K}$. The present-day lunar temperatures and Q, however, are not particularly sensitive to the Moon's early history. Presumably, but not necessarily, the exponential scaling of heat sources was set up by a pervasive magmatic event, although if such an event reached only to depths of 300 km, it would still involve 43% of the Moon's volume. Finally the homogeneous concentrations of uranium U, as obtained from the integration of the presumed scaling, are similar to those obtained for the meteoritic class of eucrites. We develop these results, in conjunction with further geochemical observations, in the following section.

5. An Origin of the Moon

The average uranium content of the Moon (~ .09 ppm, as obtained from the successful thermal history models presented in the previous sections) is an order of magnitude larger than the average for carbonaceous chondrites. The U concentrations in achondrites, however, range up to 0.20 ppm; the meteoritic class of eucrites averages 0.10 ppm U, very close to the value estimated previously. TOKSÖZ *et al.* (1972) rejected eucritic source material for the Moon on the basis that the U content was too high to allow for acceptable deep interior lunar temperatures. We consider this conclusion premature. The Moon may well have eucritic concentrations of U, Th, and K and still have subsolidus present-day temperatures, if indeed this is a requirement, and present-day surface heat flows ~ 30 ergs/cm^2-sec. The proviso, of course, is near-surface concentration of these heat sources as suggested in the previous sections.

The similarities between the basaltic achondrites and the lunar surface material (and inferences to modest depth) with respect to mineralogy, petrographic fabric, and gross aspects of chemical composition are well-known. The significance of this, however, is not yet certain since certainly the lunar surface material and evidently the basaltic achondrites (DUKE and SILVER, 1967) represent the recrystallized melt bled from a more primitive, presumably more refractory parent material. What, if any, mineralogical, petrographical, and compositional characteristics these source materials may share is uncertain, although they must contain higher rare earth and uranium abundances than carbonaceous chondrites.

DUKE and SILVER (1967) suggested that the calcium-rich achondrites may have been derived from the Moon. Apart from their very high titanium content, the lunar rocks are quite comparable in chemical and mineralogical composition to some of the eucrites and quite different from ordinary and carbonaceous chondrites. The lunar basalt and eucrites both have pyroxene and calcic plagioclase as the major minerals. Both contain trace amounts of troilite and metallic iron, indicating crystallization at low oxygen fugacity.

The radiation ages of the Ca-rich achondrites range from 0.1 to 62×10^6 year. The distribution is similar to that of ordinary chondrites which implies that both groups came from bodies with very similar orbital elements. The shorter ages suggest that the starting orbit was Earth crossing. The observed radiation-age distribution is rather flat to $\sim 3 \times 10^7$ year; no ages above 7×10^7 year are found (HEYMANN et al., 1969). Monte Carlo calculations for a lunar origin give a much steeper distribution, but a lunar origin cannot be ruled out on this basis if only a few minor impacts occurred on the Moon in the last 10^7 year. HEYMANN et al. (1969) suggest an asteroidal origin for at least the unbrecciated eucrites.

In addition, TAYLOR and EPSTEIN (1970), on the basis of oxygen isotope data, suggest that the basaltic achondrites originated from a completely different oxygen isotope reservoir, possibly a different part of the solar system, than did the Earth, Moon and the chondritic meteorites. Moreover, the absence of extreme shock effects in eucrites and howardites argues against ejection from a body as massive as the Moon. We therefore consider a lunar origin for these achondrites as unlikely.

The Moon and parent material of the basaltic achondrites may well share a similar but independent origin. For both, we envision an origin resulting from the early accretion of the compounds presumed to condense first from a hot protoplanetary nebula. The depletion of volatiles in the Moon is now well-documented as its enrichment in U, Th, Ti, the rare earth elements, as well as Ca and Al. ANDERSON and KOVACH (1972), on the basis of lunar travel-time data, have concluded that the Ca-Al enrichment must persist to depths of at least 120 km. These results, together with the relative absence of iron in the Moon, are suggestive of a Moon which accreted from high temperature condensates. The approximate order in which the more abundant compounds condense or form by reaction with previous condensates is estimated to be as follows: perovskite ($CaTiO_3$), gehlenite ($Ca_2Al_2SiO_7$), spinel ($MgAl_2O_4$), andalusite (Al_2SiO_5) and anorthite ($CaAl_2Si_2O_8$). All of these condense before iron, $MgSiO_3$, Mg_2SiO_4 and Na and K bearing compounds and, of course, FeS and H_2O (LARIMER, 1967; LARIMER and ANDERS, 1967; CLARK <u>et al</u>., 1972). Since the abundance patterns described above are also appropriate to the basaltic achondrites, and perhaps more so to their parent material, we imagine a similar origin for them as well. The eucritic and the inferred lunar value for U are close to the value estimated by HOYLE and FOWLER (1964) for solar system nucleosynthetic material. If this value is appropriate to the local nebula from which the Earth-Moon system accumulated, U also appears to be a high temperature condensate.

We therefore envision the Moon and the achondritic source material as the accumulation of early condensates of a cooling nebula. Presumably, but not necessarily, the Moon accreted in an orbit highly inclined to the median nebular plane in the vicinity of the present Earth-Moon system. It competed unsuccessfully with the Earth in the later stages of the accumulation of both, perhaps because the Earth accreted in an orbit more nearly coplanar with the median plane of the protoplanetary nebula, thereby occupying a position closer to its center of gravity. The achondritic source material accreted in a still less successful manner, originally in orbits far removed from the Earth-Moon systems. ANDERSON and KOVACH (1972) have also argued for this origin of the Moon.

The estimated time for accretion, t_{acc}, of the Moon provides one constraint by which we can check the applicability of the above accumulation model for the Moon. Unless we are to invoke another means of compound separation, it is plain that the accretion process must keep pace with the condensation process. In particular, t_{acc} for the Moon should not be much different from the condensation time, t_{con}, of the reservoir of material from which the Moon ultimately accumulates. If only a small fraction of the initial reservoir (the high temperature condensates) accumulates to form the Moon, we may expect that t_{acc} for the Moon should be somewhat less than t_{con} for the original reservoir. CLARK _et_ _al_. (1972) estimated t_{con} for the planetary nebula to be 10^4 years, the estimate subject to considerable uncertainty. They considered this estimate to be an upper bound. A likely lower bound is the freefall time, which they estimated to be less than 100 years.

The minimum estimate for t_{acc} is 2000 years. How much smaller t_{acc} may be is to be decided on the basis of the extent of differentiation that the Moon suffered upon accretion. MIZUTANI et al. (1972) estimated t_{acc} to be less than 1000 years. Both of these estimates are roughly consistent with the idea that accumulation of the Moon kept pace with condensation of the planetary nebula in its early stages. If the lower limit of 100 years is the more appropriate estimate for t_{acc}, we can still imagine the lunar accumulation keeping pace with the nebula condensation, but we must tolerate large-scale melting and differentiation of the Moon upon accretion.

The reasoning here is somewhat circular, since we more correctly should use the condensation temperatures $\sim$ 1400-1700°K for the initial temperature of the infalling material. These temperatures alone would place much of the lunar interior near the RINGWOOD and ESSENE (1970) solidus. On the other hand, the assumption of the condensation temperatures implies that accretion did keep pace with condensation, in particular that $t_{acc} \sim t_{con}$. Thus we are led to accretion times for the Moon that are of the order of 10^4 years or less.

Thus a rapid accretion of the high temperature condensates appears to be a reasonable explanation of the Moon's origin. In the later stages of accretion-condensation, the Moon competes unsuccessfully with the more favorably disposed Earth and is therefore deficient in the later condensates iron, Mg_2SiO_4, and volatile fractions. The Moon should in no way represent Type I carbonaceous chondrites. We turn to inclusions in the Allende meteorite (CLARKE et al., 1970) as a possible clue to the primitive lunar material.

The mineralogy of irregular aggregates and some chondrules in this Type III carbonaceous chondrite meteorite is dominated by gehlenite ($Ca_2Al_2SiO_7$), titanium-rich aluminous pyroxene, anorthite and spinel. Other minerals include perovskite ($CaTiO_3$), grossularite ($Ca_3Al_2Si_3O_{12}$), ferroaugite and hercynite ($FeAl_2O_4$). Similar inclusions in the Leoville chondrites (KEIL et al., 1969) consist mainly of spinel, anorthite, perovskite and gehlenite, and reaction zones indicate that they were emplaced in the matrix at temperatures in excess of 1000°C. The Allende inclusions lack Fe, FeS, H_2O and are very low in K and FeO. The FeO is mainly in the ferroaugite and the hercynite. The rare earth element (REE) abundance patterns of a Ca-rich Allende inclusion (GAST et al., 1970) are similar to those in carbonaceous chondrites and eucrites but are enriched relative to both. The lunar interior must have REE abundances greater than carbonaceous chondrites (HASKIN et al., 1970). U and Th concentrations for an originally homogeneous Moon, as inferred from the REE abundances, must be comparable or slightly greater than eucritic values. This requirement is roughly satisfied by our estimate for $\bar{U}$.

The Allende inclusions contain about 40% gehlenite, 30% aluminous pyroxene, 10% anorthite and 20% spinel. Their density is about 3.2 g/cm^3. We propose that the Moon and protoplanetary nuclei have similar compositions. Partial melting of such a body would yield 10% anorthosite immediately and satisfy the models of WOOD et al.(1970) and ANDERSON and KOVACH(1972). The remaining material has the potential of yielding 12% of basalt with the Apollo 11 FeO and TiO_2 contents. Less differentiation is required if the average lunar basalt contains less FeO and TiO_2 than the Apollo 11 average. The remaining material would have a density of about 3.24 g/cm^3 in the low-pressure assemblage and a maximum density of 3.52 g/cm^3 if the gehlenite reacts completely with the aluminous pyroxene to form garnet. These values can be compared with the minimum and maximum densities of 3.27 and 3.52 g/cm^3 for the Ringwood-Essene model lunar pyroxenite from which they were able to construct a satisfactory lunar density model. We would not expect complete reaction to garnet until very high pressures. It remains to be seen, of course, whether the proposed composition would actually yield a suitable basalt upon partial melting. The melting is not necessarily an equilibrium process.

6. The Present Thermal State of the Moon

A wide variety of results, including element abundance patterns, major mare formation subsequent to the Moon's origin, and the volume of the Moon inferred to be necessary to provide the mare basalts, all suggest that the Moon is an extensively differentiated body. Unless the interior of the Moon has always been depleted in the radioactive heat sources, it seems inescapable that temperatures were near or above the solidus throughout much of the lunar interior early in its history. This is true even for an initially cold Moon possessing chondritic or terrestrial abundances of U, Th, and K (ANDERSON and PHINNEY, 1967). An extensive differentiation of the Moon, either at 4.6 AE or later at ~ 3.5-3.0 AE, sufficient to deplete the interior of its heat sources would require at least solidus temperatures at that time; this condition would not be much changed at the present, provided that convective heat transport is ineffective at temperatures slightly subsolidus.

On the basis of the calculations presented earlier, we estimate the present thermal state of the lunar interior to be as depicted in Fig. 8. Here we have plotted the present-day temperatures of Figs. 2b, 4c, and 7 to illustrate the previously discussed agreement. We have also drawn the previously discussed initial temperature "solidus" and the lower boundary of the Apollo 11 basalt-eclogite transition zone (RINGWOOD and ESSENE, 1970). The deep interior temperatures are as uncertain as the deep interior solidus temperatures; what we have done, however, is provide a maximum deviation from the solidus temperatures in the deep interior by providing a mechanism in the gradient of the initial temperature "solidus" profile for a large heat flow from the lunar interior. We could have effected a similar result with the use of higher interior conductivities (the inferred deficiency in Fe suggests that the higher radiative conductivities may be more appropriate for the Moon then previously suspected) and a more realistic solidus gradient.

The Apollo 11 basalt liquidus is perhaps 100^{o} higher than the solidus, and the energy absorbed by the latent heat of fusion would correspond to an additional temperature rise of $\frac{H_L}{C_p} \simeq 300^{o}$ for H_L, the latent heat of fusion = 400×10^{7} ergs/gm and C_p, the specific heat at constant pressure = 1.3×10^{7} ergs/K^{o}-gm. Thus complete melting requires energy production equivalent to some $400^{o}K$ above the solidus temperatures. For these models complete melting therefore appears to be unlikely anywhere in the Moon at the present or in the recent history of the Moon. On the other hand, we see no reason why temperatures for $\frac{r}{R} \lesssim .8$ need be significantly lower than solidus temperatures. That is, most of the lunar interior may involve at least some partial melting. We also see no difficulty for basalt (rather than eclogite) stability in the range $.82 \lesssim \frac{r}{R} \gtrsim .95$.

Such a temperature distribution is not inconsistent with the nonhydrostatic shape of the Moon nor the existence of mascons. Although the departures of the Moon from hydrostatic equilibrium seem impressive, the Moon is much closer to being in hydrostatic equilibrium than the Earth, when the lunar data are scaled appropriately for its self-gravitation. A comparison of C_{22} and S_{22}, when scaled to a standard object, reveals that the Moon is 2 to 6 times "smoother" than the Earth (LORELL *et al.*, 1972). The nonhydrostatic shape of the Moon, as well as the mascons, may well be supported by a cool exterior shell of several hundred km thickness. Using the intersection of present-day temperatures (Fig. 8) with solidus temperatures as a thickness measure, this exterior shell may have a thickness of ~ 200 km or greater; an exterior shell of 200 km represents 31% of the lunar volume.

We also feel that there is no conflict between the present-day temperatures of Fig. 8, and present-day temperatures inferred from electrical conductivity measurements (DYAL and PARKIN, 1971; SONETT *et al.*, 1971). These authors used relatively high iron content material in converting their values of electrical conductivity to temperature. Electrical conductivity is very sensitive to iron content. For example, according to KOBAYASHI and MARUYAMA (1971), a 10% addition of Fe_2SiO_4 to olivine increases the conductivity by 25 times. On the basis of the present model, the iron content of the lunar interior is very low, reflecting the low iron content of the initial condensates and the Allende inclusions and the subsequent further reduction due to removal of basalt. Adopting a decrease of 10% in the equivalent FeO content of the lunar interior we obtain temperatures of 1070°K and 2000°K where DYAL and PARKIN (1971) inferred 810°K and 1240°K. Thus, a slight decrease in the iron content of the lunar interior raises the inferred temperatures considerably. We feel there is no longer any conflict between the electrical conductivity profiles on the one hand and the thermal history calculations, heat flow value and evidence for early melting and differentiation on the other. All data and calculations are consistent with a hot lunar interior.

We also see no inconsistency between solidus temperatures and partial melting at depth and the relative absence of present-day volcanism and the remarkable aseismicity of the Moon (LATHAM *et al.*, 1971). Both the Earth's seismicity and volcanism may be viewed as consequences of planetary processes that give rise, on the Earth, to what is now know as global tectonics. In particular, most of the Earth's shallow seismicity may be associated with displacement discontinuities at plate boundaries, these "boundaries" being quite diffuse in continental margins. Deep and intermediate seismicity appears to be causally related to the lithosphere's descent into the asthenosphere. Likewise, the major centers of igneous and metamorphic activity also are confined, in the main, to plate margins. The notable exceptions are the centers of igneous and volcanic activity that MORGAN (1971) associates with mantle hot spots, or plumes, which in his view drive plate motions.

The implication here is that if plate motions were to cease, so would the bulk of the Earth's seismicity and volcanic/igneous/metamorphic activity. While it is generally agreed that plate motions on the Earth are thermally driven by a "hot" interior of large thermal inertia (MCKENZIE, 1969), such a "hot" interior appears to be only the necessary condition. An auxiliary condition must be plate mobility, that is, susceptibility to the driving forces from beneath. Our intuition is that plate mobility necessitates small plate mass and thermal inertia relative to the interior.

A lunar lithosphere of 200 km thickness constitutes 31% of the volume of the Moon. Its mass and thermal intertia are quite large relative to those of the lunar interior. For the Earth a lithosphere of 100 km thickness represents less than 5% of the planetary volume, and a lithospheric thickness of 50 km, perhaps more appropriate to oceanic regions, is less than 3% of the Earth's volume. The relative mass and thermal inertiae of these lithospheres are still smaller, since density in the Earth increases more rapidly with depth. We therefore suspect that plate motions are much less easily effected at the present time on the Moon than on the Earth for purely volumetric reasons.

We may also estimate the (tensional)stresses arising in the lithosphere (the membrane stresses) arising from a uniform pressure p exerted on the interior surface of the lithosphere of thickness h. The membrane stress σ is

$$\sigma \propto \frac{R}{h} p \tag{8}$$

For the same interior pressure p, which we might view as arising from the force system that ultimately drives plate motions, σ will be ~ 10 times less in the Moon than in the Earth, again for simple geometrical reasons. If such stresses are invoked, from time to time, to initiate lithospheric rupture prior to an episode of sea-floor spreading, the Earth's lithosphere is considerably more susceptible to rupture than is the Moon's lithosphere, if they possess comparable strengths in tension.

We do not mean to imply that the above discussion in any way represents the complete scaling of planetary convection. The discussion moreover has been confined to static considerations; at best it is applicable only to the initiation of plate motion, rather than to the maintenance of plate motions. It is only suggestive that, for otherwise "hot" interiors, plate motions and therefore seismicity, volcanism, and shallow igneous intrusion, are less likely on the Moon than on the Earth for simple geometrical reasons. The absence of these phenomena need not imply a cold lunar interior, just as the absence of these phenomena in continental shield areas may say more about the thickness of the local lithosphere than about sublithosphere conditions.

7. Origin, Evolution, and Present Thermal State of the Moon

In this section, we rearrange and summarize the several findings of this study in the form of a discussion of the origin, evolution, and present thermal state of the Moon. In terms of our present knowledge, it appears to be an attractive possibility, but even so it rests upon an edifice constructed, in part, from assumptions of unconfirmed validity and the basic imprecision of conventional thermal history calculations.

Inferences of the bulk chemistry of the Moon, rare earth element abundance patterns, the estimate for the homogeneous concentration of U, and the inferred accretion time suggest that the Moon had its origin in the rapid accretion of compounds first condensing from the hot protoplanetary nebula. Presumably, but not necessarily, the Moon accreted in an orbit highly inclined to the median nebular plane in the vicinity of the present Earth-Moon system. The Moon competed unsuccessfully with the Earth in the later stages of accretion-condensation, with a consequent lunar deficiency in the lower temperature condensates. The Moon accreted in times of the order or less than 10^4 years; immediately subsequent to its formation, the Moon was at least "hot."

It seems inescapable that temperatures through much of the lunar interior were near or above the solidus temperatures early in the history of the Moon. Interior solidus temperatures were probably cause and effect of a pervasive differentiation of the Moon, sufficient to deplete drastically interior concentrations of U, Th, and K. Interior thermal conditions would not be much changed at the present time, provided that convective heat transport is ineffective at temperatures slightly subsolidus. Present-day interior temperatures and surface heat flow are not particularly sensitive to the Moon's early thermal history and, in particular, cannot distinguish between the origin at 4.6 AE or an "origin" at 3.25 AE. Thus, the calculations admit the possibility that the present compositional and thermal state of the Moon is more directly related to the magmatic event at ~ 3.5-3.0 AE, rather than large-scale differentiation at 4.6 AE. Hopefully, however, this ambiguity may be resolved by further geochemical and geophysical observations.

At the present time, much of the lunar interior is at or near solidus temperatures, most likely involving at least some partial melting. In the range $.8 \leq \frac{r}{R} \lesssim .95$, basalt, rather than eclogite, may be the stable phase; if the phase equilibrium has not been established, basalt stability may persist to shallower depths. Neither the nonhydrostatic shape of the Moon nor present-day temperatures inferred from conductivity data militate against such a present thermal state. The absence of present-day volcanism and the Moon's aseismicity are as easily attributable to the absence of plate motions on the Moon as to a "cold" interior. For otherwise "hot" interiors, plate motions seem less likely on the Moon than on the Earth for simple geometrical reasons.

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TABLE I

Scaling of exponential distribution of U as a function of $R-r_o$ and U_o

			100	200	300	Shell thickness $R-r_o$, km
U_o[1]	Q/U_o[2]	Q/H_o[3]	.16	.31	.43	Shell volume ratio
.25	132	964	*	*	.200	$\times 10^{-2}$ km^{-1}
.50	66	482	*	.750	.900	$\times 10^{-2}$ km^{-1}
1.00	33	241	.160	.195	.200	$\times 10^{-1}$ km^{-1}

[1] U_o in ppm, for $Q = 33$ ergs/cm^2-sec.

[2] Q in ergs/cm^2-sec, U_o in ppm.

[3] Q in ergs/cm^2-sec, H_o in ergs/cm^3-sec.

*No positive value of λ exists.

Figure Captions

Fig. 1. Temperatures in the Moon. The number associated with each curve indicates billions of years after the Moon's origin at 4.6 AE (and similarly for Figs. 2-7). The initial temperature distribution (heavy solid curve labeled 0) is the "solidus" discussed in the text (and similarly for Figs. 2-5,7). The thermal conductivity (K) = 4.2 x 10^5 ergs/cm-^{o}K-sec, constant throughout the Moon (and similarly for Figs. 2,3,6,7). The value for Q equals the present-day surface heat flow in ergs/cm^2-sec for this model (and similarly for Figs. 2-7). In this model, the heat source concentrations have been set equal to zero. 0.1 units of radius fraction ($\frac{r}{R}$) corresponds to 174 km.

Fig. 2. Temperatures in the Moon for three heat source distributions corresponding to the cases U_o = 0.25 ppm, λ = 0.20 x 10^{-2} km^{-1} (a); U_o = 0.5 ppm, λ = 0.90 x 10^{-2} km^{-1} (b); and U_o = 1.0 ppm, λ = 0.20 x 10^{-1} km^{-1} (c), as discussed in the text. Present-day concentrations of U are indicated by the dashed line and the scale on the lower right sides of the figures (and similarly for Figs. 3-7). Other parameters as in Fig. 1.

Fig. 3. Temperatures in the Moon. The heat source distribution is the same as that for Fig. 2a, with the concentration set equal to zero for depths greater than 300 km. Other parameters as in Fig. 2a.

Fig. 4. Temperatures in the Moon for different average thermal conductivities. (a) K = 8.4, (b) K = 2.1, (c) K = 2.1 x 10^5 ergs/cm-^{o}K-sec. The heat source distribution for a, b, and

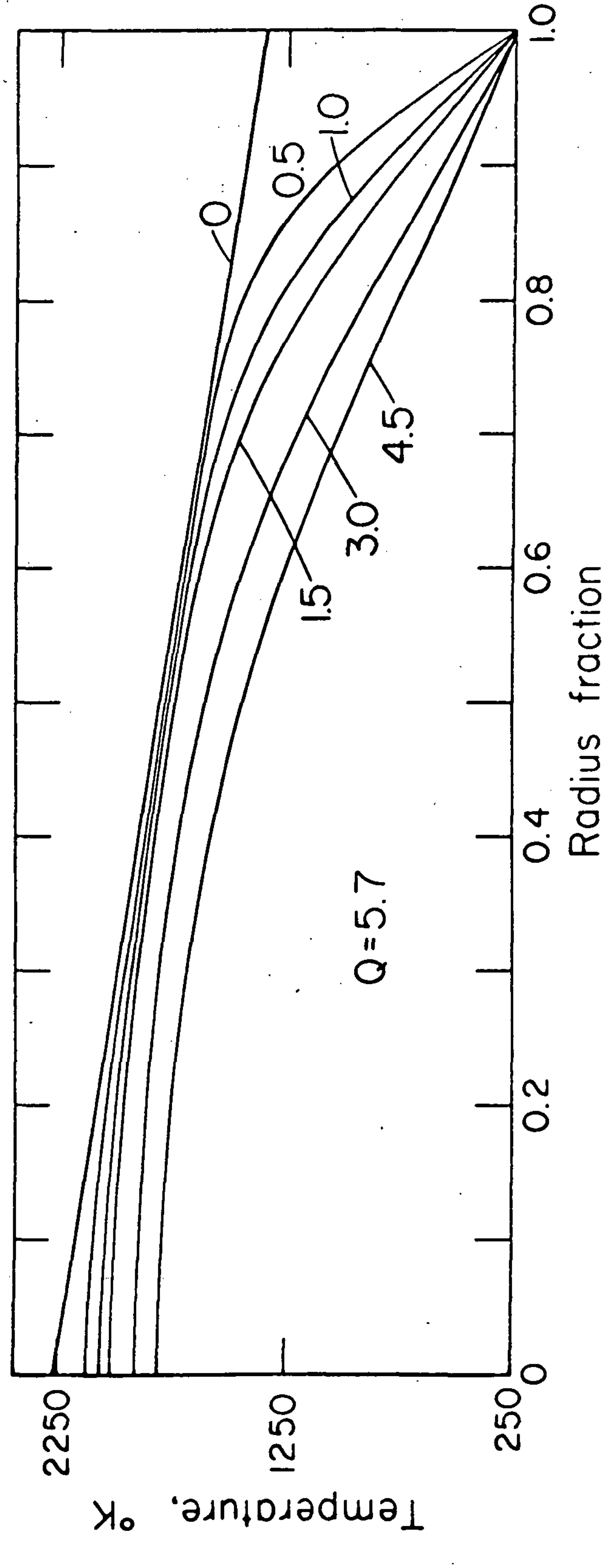
c are those given for Figs. 2a, 2b, and 2c, respectively.

Fig. 5. Temperatures in the Moon. The surface layer (~ 35 km thick) has a value of K = 3.15 x 10^5 ergs/cm-^{o}K-sec (a) and 2.1 x 10^5 ergs/cm-^{o}K-sec (b). Otherwise, these models have parameters identical to those of Fig. 2b.

Fig. 6. Temperatures in the Moon. Initial temperatures and heat source distribution as described in Section 4. Present-day U concentration as indicated. The lighter dashed curve is the initial temperature distribution for Figs. 1-5.

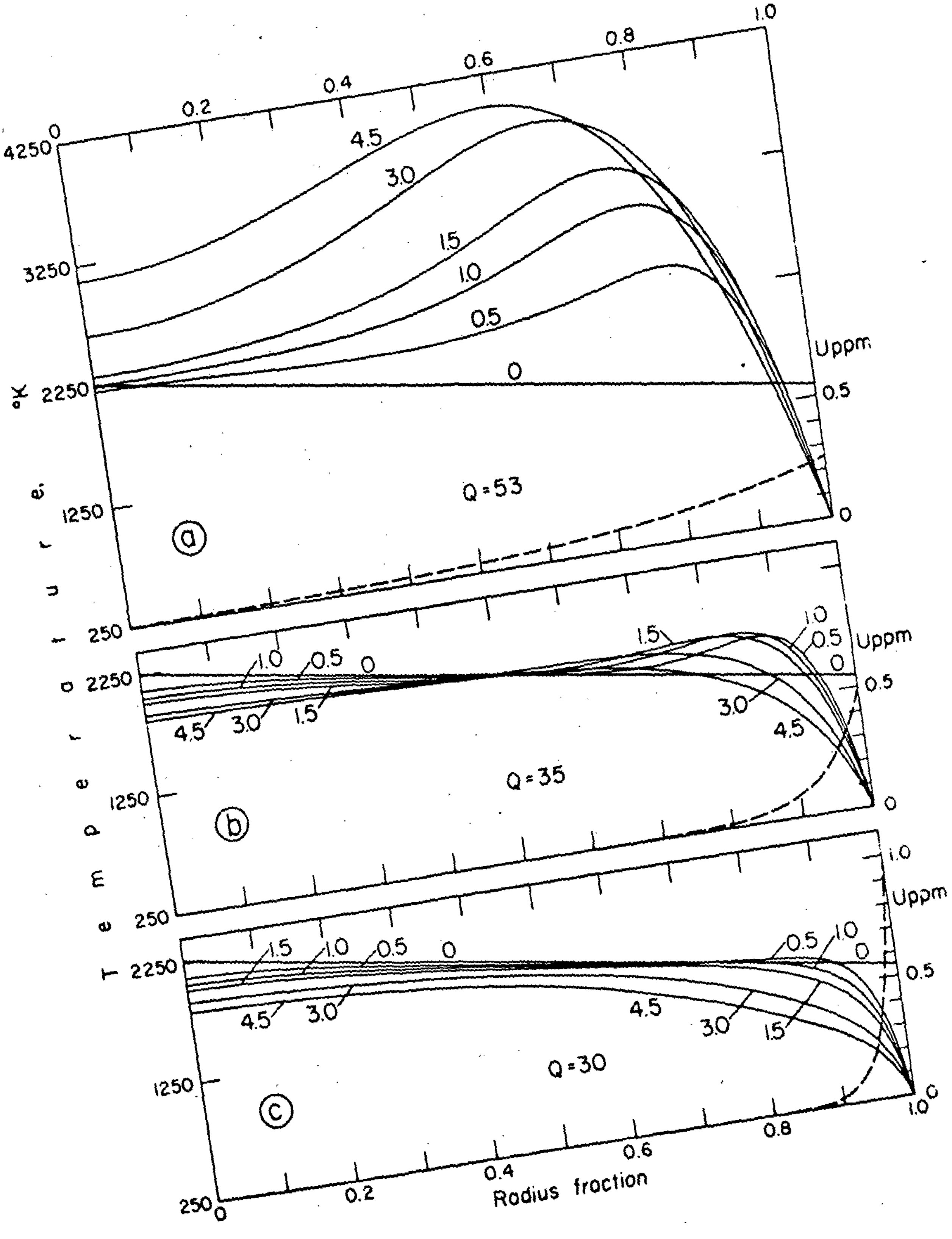
Fig. 7. Temperatures in the Moon. The "origin" for the Moon is taken to be at 3.25 AE (1.35 b.y. after the actual origin at 4.6 AE). Other parameters as in Fig. 2b.

Fig. 8. Present-day temperatures in the Moon as obtained from Figs. 2b, 4c, and 7. The heavy solid line labeled 0 is the initial temperature distribution of Figs. 1-5, as well as the initial temperature of Fig. 7. The second heavy solid line in the range $.82 \leqslant \frac{r}{R} \leqslant 1.0$ is the lower boundary of the basalt-eclogite phase transition (RINGWOOD and ESSENE, 1970).



Temperature, °K
2250
1250
250
0
0.2
0.4
0.6
0.8
1.0
Radius fraction
Q=5.7
0
0.5
1.0
1.5
3.0
4.5

Fig. 1



Temperature °K
4250
3250
2250
1250
250
0
0.2
0.4
0.6
0.8
1.0
4.5
3.0
1.5
1.0
0.5
0
Uppm
0.5
0
Q = 53
ⓐ
1.5
1.0
0.5
0
Uppm
0.5
4.5
3.0
1.5
1.0
0.5
0
3.0
4.5
Q = 35
ⓑ
1.0
Uppm
1.5
1.0
0.5
0
0.5
1.0
0
0.5
4.5
3.0
4.5
3.0
1.5
Q = 30
ⓒ
Radius fraction

F

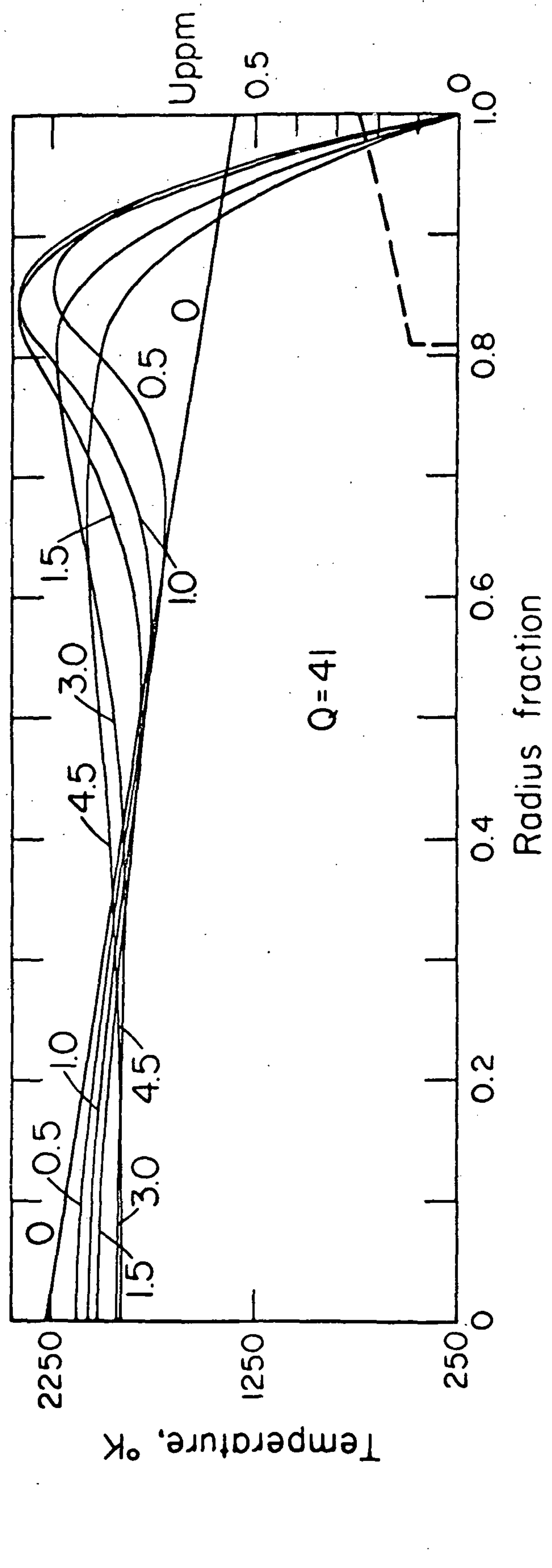
 Temperature, °K
2250
1250
250
0
0.2
0.4
0.6
0.8
1.0
Radius fraction
Uppm
0.5
0
Q=41
0
0.5
1.0
1.5
3.0
4.5
0
0.5
1.0
1.5
3.0
4.5

Fig. 3

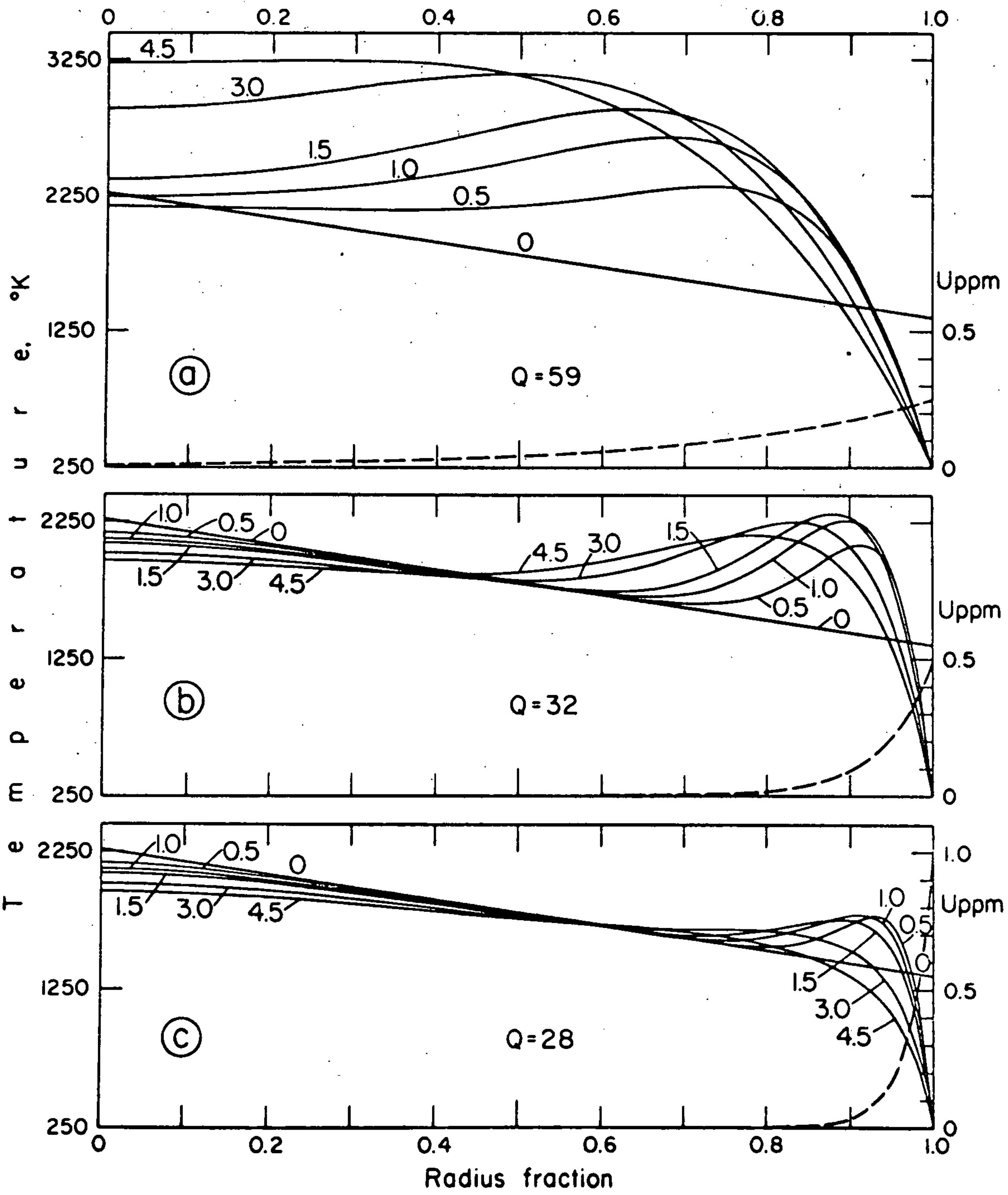
0
0.2
0.4
0.6
0.8
1.0
4.5
3.0
1.5
1.0
0.5
0
3250
2250
1250
250
Uppm
0.5
0
ⓐ
Q = 59
1.0
0.5
0
1.5
3.0
4.5
2250
1250
250
Uppm
0.5
0
ⓑ
Q = 32
1.0
0.5
0
1.5
3.0
4.5
2250
1250
250
1.0
Uppm
0.5
0
ⓒ
Q = 28
Temperature, °K
Radius fraction

Fig. 4

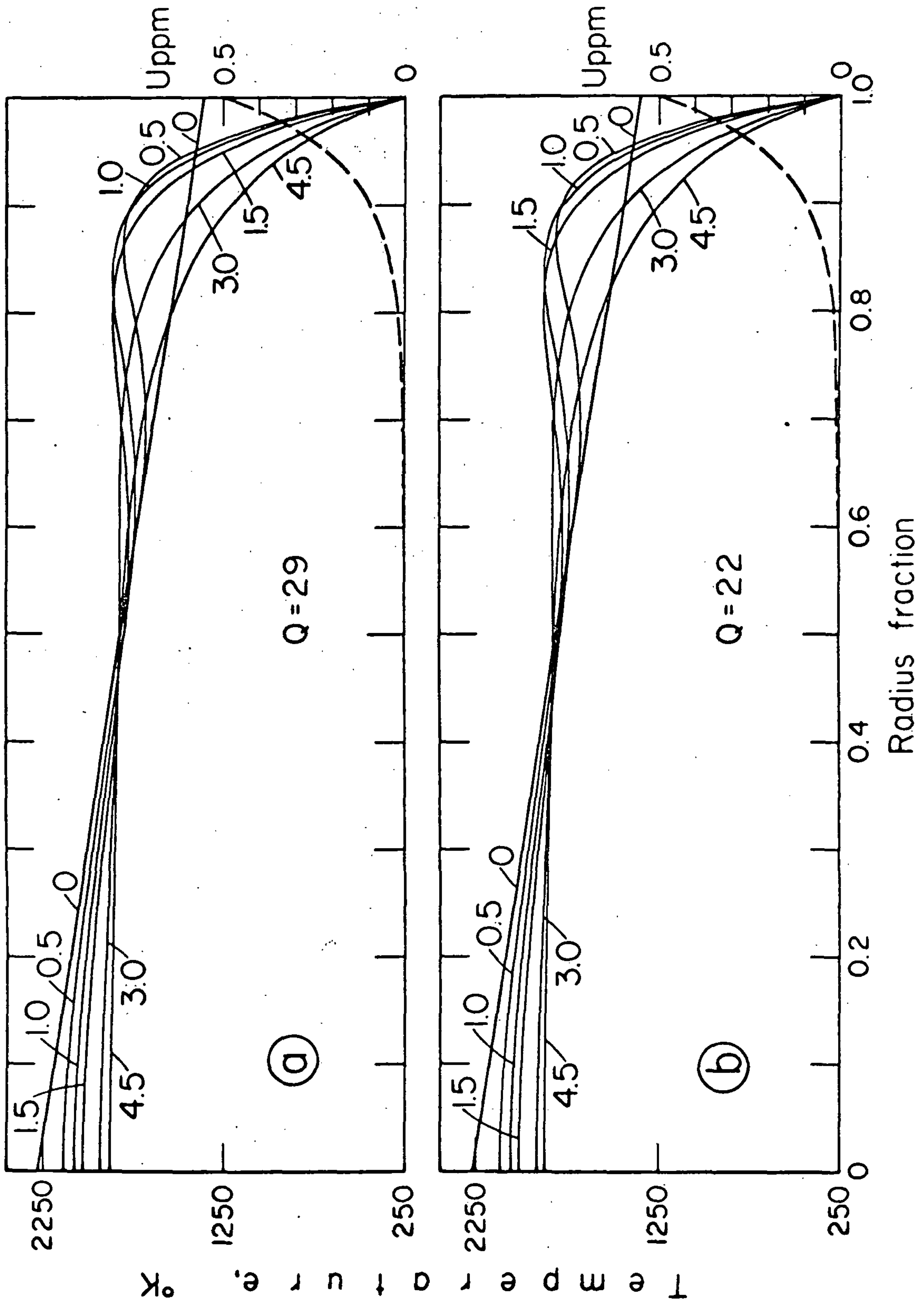
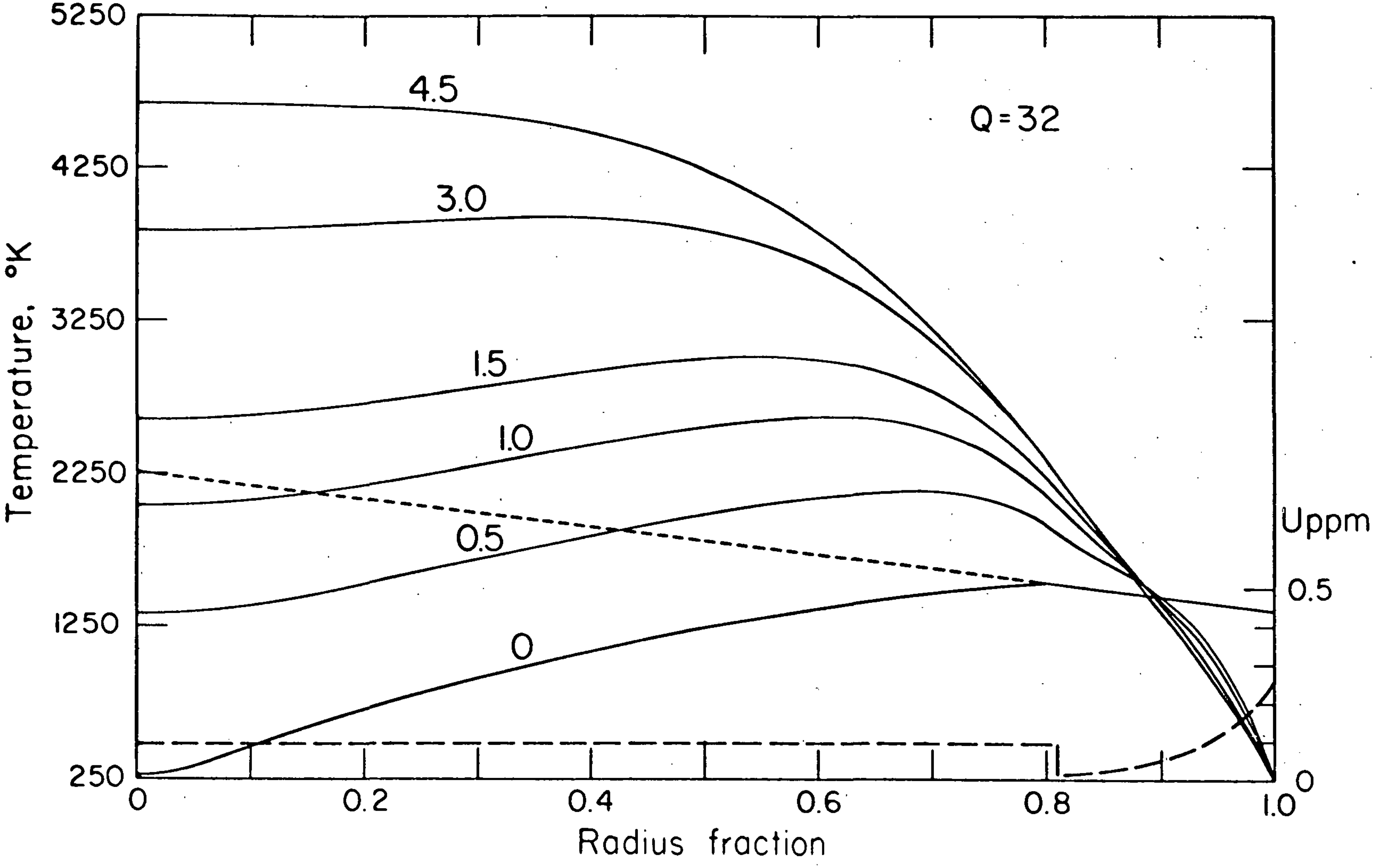

Temperature, °K
2250
1250
250
Radius fraction
0
0.2
0.4
0.6
0.8
1.0
Uppm
0.5
0
Q=29
Q=22
a
b
1.5
1.0
0.5
0
3.0
4.5

Fig. 5



Q=32
4.5
3.0
1.5
1.0
0.5
0
Temperature, °K
5250
4250
3250
2250
1250
250
Radius fraction
0
0.2
0.4
0.6
0.8
1.0
Uppm
0.5
0

Fig. 6

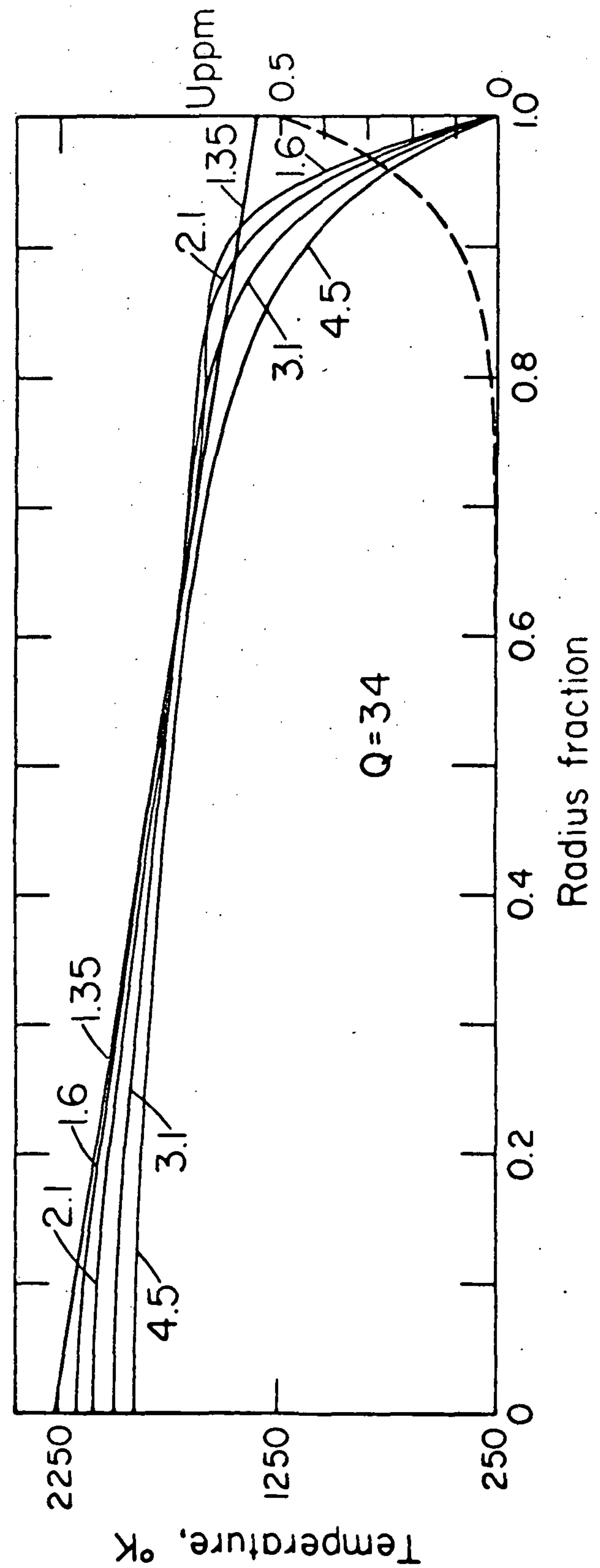

Temperature, °K
2250
1250
250
0
0.2
0.4
0.6
0.8
1.0
Radius fraction
Q=34
Uppm
0.5
0
1.35
1.6
2.1
3.1
4.5

Fig. 7

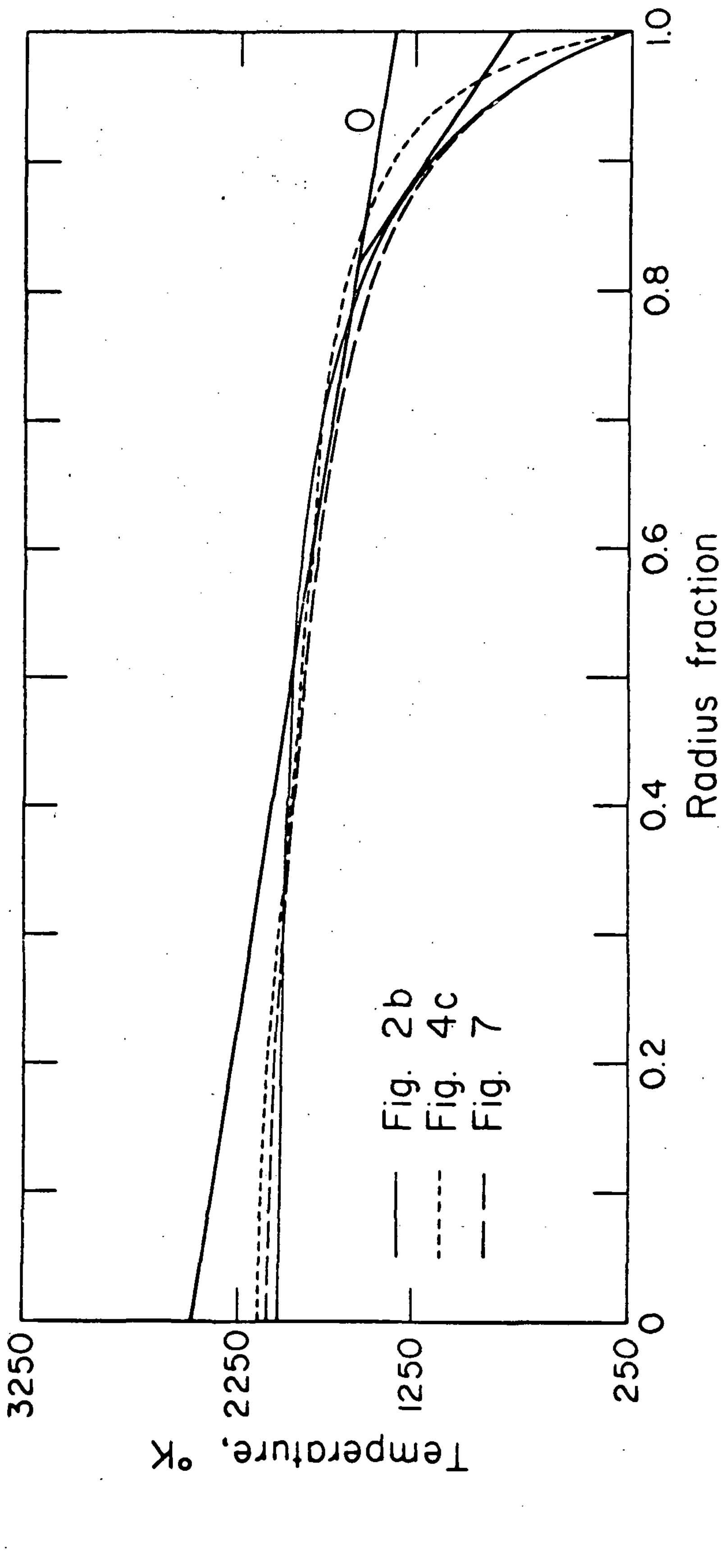

3250
2250
1250
250
Temperature, °K
0
0.2
0.4
0.6
0.8
1.0
Radius fraction
Fig. 2b
Fig. 4c
Fig. 7
0

Fig. 8