DENSITY WITHIN THE MOON AND IMPLICATIONS FOR LUNAR COMPOSITION*

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Abstract. Density models for the Moon, including the effects of temperature and pressure, can satisfy the mass and moment of inertia of the Moon and the presence of a low density crust indicated by the seismic refraction results only if the lunar mantle is chemically or mineralogically inhomogeneous. If C/MR^2 exceeds 0.400, the inferred density of the upper mantle must be greater than that of the lower mantle at similar conditions by at least 0.1 g cm^{-3} for any of the temperature profiles proposed for the lunar interior. The average mantle density lies between 3.4 and 3.5 g cm^{-3}, though the density of the upper mantle may be greater. The suggested density inversion is gravitationally unstable, but the implied deviatoric stresses in the mantle need be no larger than those associated with lunar gravity anomalies. Using $C/MR^3 = 0.400$ and the recent seismic evidence suggesting a thin, high density zone beneath the crust and a partially molten 'core', successful density models can be found for a range of temperature profiles. Temperature distributions as cool as several inferred from the lunar electrical conductivity profile would be excluded. The density and probable seismic velocity for the bulk of the mantle are consistent with a pyroxenite composition and a 100 MgO/(MgO + FeO) molecular ratio of less than 80.

1. Introduction

Density has traditionally been a very useful quantity by which to characterize the material making up the interior of a planet. In the Earth, realistic models for the distribution of density with depth came only after the velocity of seismic waves was determined throughout the Earth's interior. Classical density models were derived from simple relationships between seismic velocity and either density of the radial derivative of density. More recent models for the Earth have been generated as solutions to the inverse normal-mode problem.

In the case of the Moon, the seismic-wave velocity is not yet known throughout the lunar interior. Nor have any normal modes or surface waves yet been identified on the lunar seismograms. Thus one's approach to the question of the density distribution in the Moon, and associated implications for lunar structure, composition, and evolution, must be less direct than has been possible for the Earth. A common theme in the literature has been to begin with a preferred compositional model for the lunar interior and then to check whether the predicted density distribution matches the relevant constraints to within acceptable tolerances (e.g. Ringwood and Essene, 1970; Mason, 1972; Anderson, 1973). An alternative philosophy has been to consider a family of density models that match the appropriate contraints and to derive information on lunar structure or composition from characteristics common to the

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models (e.g. Gast and Giuli, 1972; Solomon and Toksöz, 1973). In this paper, we will follow the latter approach.

Specifically we shall examine some simple density models for the Moon that are consistent with the lunar mass and moment of inertia, with the latest information on the seismic velocity of the lunar crust and mantle, and with assorted estimates of temperature in the lunar interior. Though the mass and moment of inertia are, of course, only integral constraints on the lunar density distribution, they are nonetheless powerful ones when combined with the seismic refraction results. The seismic velocity structure is reasonably well known to a depth of 120 to 150 km in the Oceanus Procellarum region (Toksöz *et al.*, 1972a, 1974). While this structure may not be correct for other regions of the Moon, the material shallower than 120 to 150 km constitutes roughly 20 to 25% of the lunar mass.

Several of the conclusions reached in this paper re-iterate ones made in our earlier work (Solomon and Toksöz, 1973). In particular, if the lunar moment of inertia C/MR^2 equals or exceeds 0.400, then the thick, low-density crust implies a density inversion in the lunar mantle due to a change in chemistry or mineralogy with depth.

New material presented here includes the implications for density models of recent seismic evidence for a thin, high velocity layer beneath the lunar crust (Toksöz *et al.*, 1974) and for a zone of partial melting below 1000 km depth (Nakamura *et al.*, 1973). We also examine in detail the consequences of a dense, iron-rich central core, a possibility that would be enhanced by a lunar moment of inertia in the lower range of values recently proposed. Further, we use current bounds on density and seismic velocity in the interior of the Moon to limit compositional models for lunar mantle.

2. Some Constraints on Lunar Density

The primary constraints on spherically symmetric models for the distribution of density within the Moon are the lunar mass and moment of inertia and the measured density of lunar surface rocks. Additional, though not unequivocal, constraints are the likely discontinuities in density inferred from the seismic velocity structure and the expected behavior of lunar material as a function of temperature and pressure. These constraints are discussed in detail in this section.

2.1. Mass, moment of inertia

We adopt the values $GM = 4902.7 \pm 0.1$ km^3 s^{-2} for the product of the gravitational constant and the mass of the Moon, and $R = 1737.5 \pm 0.6$ km for the mean lunar radius. These quantities are taken from Kaula (1971), and indicate a mean lunar density of 3.344 ± 0.004 g cm^{-3}. Only the least significant figure in each of these quantities is likely to be modified by future data or refinements in analysis (cf. Sjogren and Wollenhaupt, 1973).

The moment of inertia of the Moon is not as well known as we might wish. We shall generally use the oft-quoted value of $C/MR^2 = 0.402 \pm 0.002$, first given by

Kaula (1969) from the gravity-field coefficients of Michael *et al.* (1970) and since reproduced by Michael and Blackshear (1972). Other recent solutions for the lunar gravity field (Liu and Laing, 1971; Sjogren, 1971), relying on essentially the same Lunar Orbiter tracking data as that used by Michael and Blackshear, have given lower values for C/MR^2, between 0.393 and 0.397. Thus while we shall concentrate much of our attention on the particulars of density models giving C/MR^2 equal to or greater than 0.400, we shall devote some space to the implications of the proposed lower values.

2.2. Crustal thickness and density

From the travel times and amplitudes of seismic waves from artificial impacts on the Moon's surface the distribution of P and S wave velocity in the upper 120 km has been determined for the Fra Mauro region of Oceanus Procellarum (Toksöz *et al.*, 1972a, 1974). The velocity profiles indicate a two-layer crust about 65 km thick. In the upper crust, to a depth of 25 km, the velocity increases rapidly with depth, indicating that fracturing and porosity play an important role. In the lower crust between 25 and 65 km depth, the velocity is nearly constant at about 7 km s^{-1} for compressional waves.

Comparison of the velocity profile with laboratory measurements of compressional velocity in lunar and terrestrial rocks at appropriate pressures allows some estimate of the most likely rock types for each layer (Toksöz *et al.*, 1972a, 1974). The velocities in the upper crust fall within the range of the velocites of mare and nonmare basalts and thus the upper crust in the Procellarum region is probably similar in composition to basaltic rocks found at the surface. The lower crust has a velocity similar to those measured in terrestrial gabbros and anorthosites and to values recently determined for lunar anorthosites (Todd *et al.*, 1973). The likelihood of abundant anorthositic gabbros in the lunar highlands (Wood *et al.*, 1970; Adler *et al.*, 1972a, b) and the spectral evidence for highland material underlying some mare craters (McCord *et al.*, 1972) make such a material a strong candidate for the primary constituent of this layer.

Using the composition of the lunar crust inferred from the seismic velocity information, we can estimate the density of crustal material. The measured densities of mare basalts, corrected for porosity, are 3.3 to 3.4 g cm^{-3} at standard temperature and pressure (Kanamori *et al.*, 1970; Stephens and Lilley, 1970). The density of highland rocks, which occupy about four-fifths of the Moon's surface area, must be somewhat less, as pointed out by O'Keefe (1968). The density of plagioclase-rich rocks is dominated by anorthite, with density $\varrho = 2.76$ g cm^{-3}. Measured and model values of anorthosites and nonmare basalts are, respectively, 2.8 to 2.9 g cm^{-3} and about 3.0 g cm^{-3} (Wood *et al.*, 1970, 1971). Further, this difference in density of 0.3 to 0.4 g cm^{-3} between mare basalts and highland rock is compatible with a 25-km thickness for the upper crust and the observed 3-km difference in mean elevation between highlands and maria (e.g. Kaula *et al.*, 1972). Assuming a four to one ratio of plagioclase-rich highland rocks to mare basalts and no significant contribution from porosity, the mean density of the upper crust of the Moon is then about 3.0 g cm^{-3} (see also Turkevich, 1971). We shall adopt this value in calculations below.

From the rapid increase in seismic velocity with depth in the upper crust, there must be a substantial porosity, particularly in the upper 10 km. According to A. M. Dainty (personal communication, 1972), the variation of velocity with depth is roughly consistent with a porosity given by $p = 0.31 \exp(-z/3.9)$, where z is in km. We shall include the effects of such a porosity distribution in some of the density models below.

The density of the lower crust is somewhat less certain than that of the upper crust, and we shall treat this quantity as a variable in the range 2.7 (anorthite) to 3.4 g cm^{-3} (mare basalt). The clear match between the seismic velocity of the lower crust and that measured in the anorthite-rich rocks from the Moon, however, suggests that a narrower range of uncertainty, perhaps a 2.8 to 3.0 g cm^{-3}, is preferable. The nearly constant velocity in the lower crust indicates that significant porosity is absent in that layer.

We should note that a total crustal thickness of about 65 km may not be representative of the entire Moon. The observed offset between the Moon's center of mass and center of figure can be explained if there is a greater thickness (100 km or more) of low density crust on the lunar farside than on the nearside (Kaula *et al.*, 1972; Wood, 1973), though of course the offset can be explained by more extensive lateral density variations at depth (e.g. Ransford and Sjogren, 1972). Wood (1973) has extended such arguments further and proposed a crustal model for the Moon that reproduces the differences between the principal moments of inertia and the mascon gravity anomalies as well as the departure of the center of mass from the center of figure. In his model, the crust is thinner at the lunar poles and beneath mascon maria than beneath non-mascon maria (e.g. Oceanus Procellarum).

2.3. Seismic velocity in the mantle

The seismic velocity in the uppermost mantle of the Moon is ambiguous at present (Toksöz *et al.*, 1973). There is evidence both that the mantle to a depth of 120 to 150 km has an average compressional-wave velocity of 7.6 to 7.7 km s^{-1} and that the very top of the mantle may be composed of material with velocity near 9 km s^{-1}. If the high-velocity material is present, then it appears from travel-time and amplitude considerations to be confined to a layer no more than about 20 km thick. A layer with velocity as high as 9 km s^{-1} must contain large amounts of such dense minerals as garnet (Toksöz *et al.*, 1972a; Anderson and Kovach, 1972). This layer, if present, may consist of a high-pressure assemblage of lower crustal material. The smaller value for compressional velocity, on the other hand, is in the lower range of values normally attributed to material of the Earth's mantle.

An important additional constraint on models for the deep interior of the Moon is the observation that high frequency shear waves, which can be seen from moonquakes deeper than 900 km (Latham *et al.*, 1973), are missing on records of waves which have propagated below about 1000 km depth (Nakamura *et al.*, 1973). A sharp drop in shear wave Q at that depth is indicated, and the preferred explanation is that the Moon has a partially molten core 700 km in radius. There may also be a drop in compressional wave velocity at 1000 km depth (Nakamura *et al.*, 1973), though the magnitude of the decrease is sufficiently small (less than 0.4 km s^{-1}) to preclude complete melting.

2.4. Temperature

Density is a function of both pressure and temperature. The latter variable is especially important in a body as small as the Moon, but while the subject of the lunar temperature profile has received a great deal of recent attention, the issue is far from settled.

We shall treat the temperature distribution as imprecisely known, and calculate density models for several temperature profiles approximately spanning the range of distributions currently in vogue. Four such profiles are shown in Figure 1. Profile A is from a thermal history model (Solomon and Toksöz, 1973) designed so that the surface heat flow matches the Apollo 15 value (Langseth *et al.*, 1972). Profiles B and

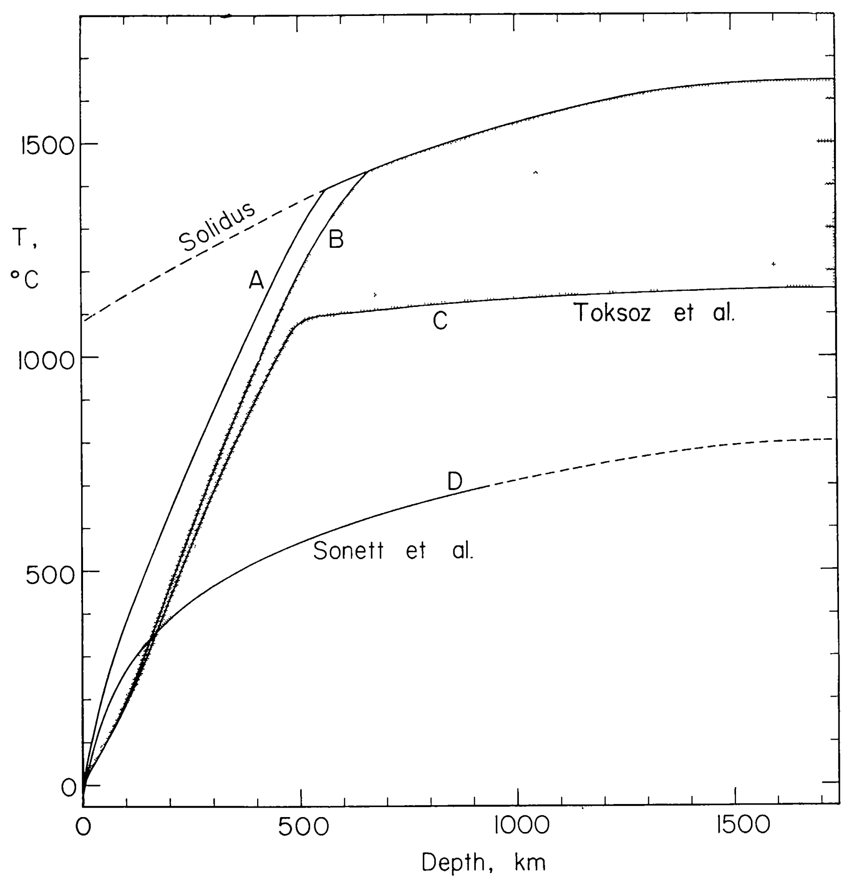

Fig. 1. Some estimates of present-day temperature in the Moon. Profile A is the final temperature from a thermal history model of Solomon and Toksöz (1973) matching the Apollo 15 heat-flow value (Langseth *et al.*, 1972). Profiles B and C are from Toksöz *et al.* (1972b); the range of present-day temperature profiles in their models is indicated by the stippled region. Profile D is from Sonett *et al.* (1971). The solidus is that of mare basalt (Ringwood and Essene, 1970).

C are two present-day temperature distributions from thermal models of Toksöz *et al.* (1972b); the total range of final temperature profiles in the models of Toksöz *et al.* is indicated by the shaded region (including only models with bulk uranium concentration equal to 23 ppb). Profiles B and C (Figures 5 and 12 of Toksöz *et al.*, respectively) have the same initial temperature curve and bulk radioactivity; they differ in that the thermal evolution model resulting in profile C is based on the assumed dominance of solid-state convection as an energy transport mechanism below the lithosphere. Profile D is the temperature profile inferred by Sonett *et al.* (1971) from the lunar electrical conductivity distribution and is shown roughly extrapolated to the Moon's center. Such a profile, a reasonably typical 'cool' Moon, has also been proposed by Tozer (1972), who argued that efficient energy transfer by solid-state creep would produce a present-day temperature distribution substantially below the solidus of mantle material.

3. Calculation of Density Models

We consider models of the Moon consisting of chemically homogeneous layers, which for convenience are numbered consecutively with depth. For each model, the equations of hydrostatic equilibrium and conservation of mass are integrated numerically using a fourth order Runge-Kutta method. The integration step-size in each layer is chosen so that the total mass and moment of inertia of the model Moon can be calculated to four-figure accuracy. The starting density in one or two layers is perturbed until a fit is obtained to the mass or to the mass and moment, respectively; generally four or five iterations are sufficient. Temperature is taken from an assumed profile. Densities at each discrete depth are corrected for compression and thermal expansion. No explicit account is made in the density calculation of partial melting where present in a temperature model.

In the lunar crust, compressibility and thermal expansion are assumed constant and are taken from reported measurements, corrected to low porosity, on mare basalts (Stephens and Lilley, 1970, 1971; Baldridge and Simmons, 1971). Values used are, respectively, 1.4 Mbar^{-1} and $2.2 \times 10^{-5}\,°\text{C}^{-1}$. In the lunar mantle, the pressure and temperature dependence of compressibility and thermal expansion are generally included. All parameters used are those of either forsterite (Skinner, 1962; Kumazawa and Anderson, 1969) or of bronzite (Frisillo and Barsch, 1972; Frisillo and Buljan, 1972); where not stated otherwise, the forsterite values will be employed.

4. Models with Homogeneous Mantles

We begin with models in which the lunar mantle is chemically and mineralogically homogeneous, at least to the extent that the entire Moon from a depth of 65 km to the center has a uniform density at standard temperature and pressure. (As used below, 'standard' temperature is $-20\,°\text{C}$, the near-surface temperature of the Moon according to Langseth *et al.*, 1972.) If a temperature distribution in the Moon and densities, respectively ϱ_1 and ϱ_2, at standard conditions for the upper and lower crust are assumed, then the density ϱ_3 (again at standard conditions) of the homogeneous mantle follows from the total mass of the Moon, and the moment of inertia may be readily calculated. Since we regard the density of the lower crust as somewhat uncertain, however, it is more proper to state that ϱ_3 and C/MR^2 are functions of ϱ_2.

This is illustrated in Figure 2 for three-layered models of the Moon in which several effects on density are progressively included. For model 1, density is assumed to be uniform in each layer, independent of pressure and temperature. For models 2a and 2b, the pressure and temperature dependence of density is included; ϱ_2 and ϱ_3 are then the densities that lower crustal and mantle material would have at $P=0$ and

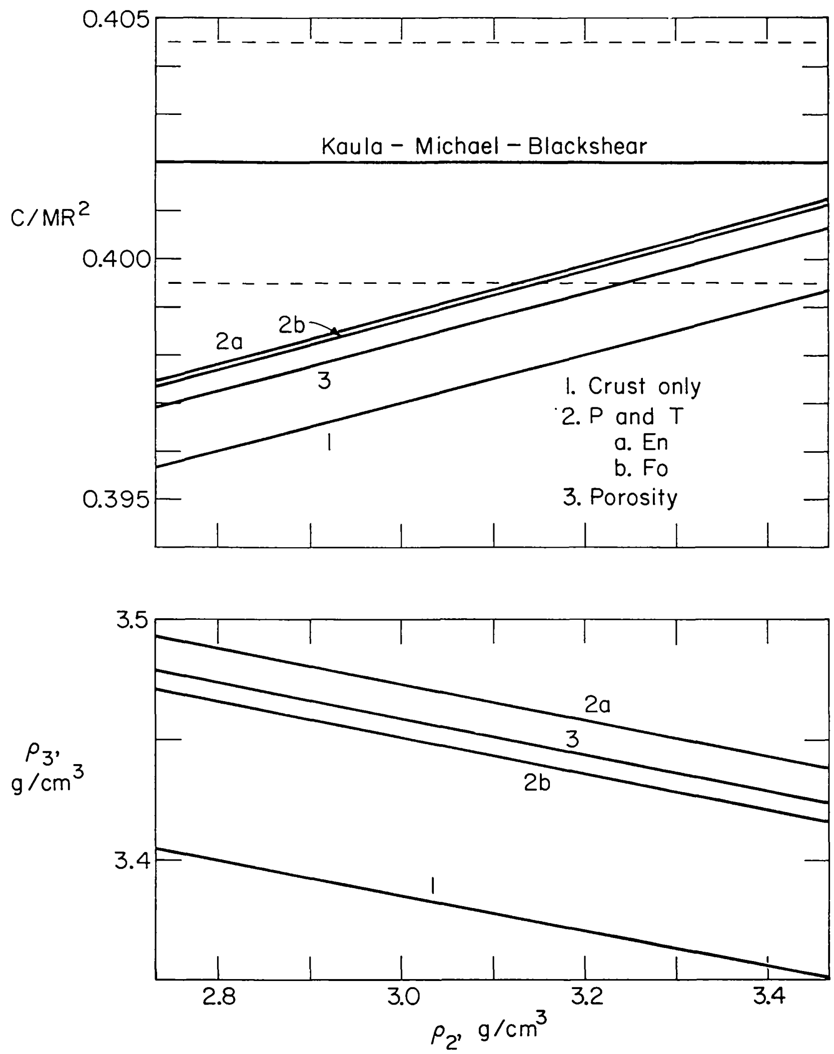

Fig. 2. The variation of the moment of inertia C/MR^2 and the density ϱ_3 of the mantle, assumed homogeneous, as a function of the density ϱ_2 of the lower crust for lunar density models of various degrees of sophistication. A value of 3.0 g cm^{-3} is assumed for ϱ_1. Model 1 has constant-density layers. Model 2 includes the effects of temperature and pressure; compressibility and thermal expansion in the mantle are taken equal to those of orthopyroxene (En_{80}) for 2a and olivine (Fo_{100}) for 2b. Model 3 is similar to 2b except that a relationship for the porosity of the upper crust has been included. The limits of uncertainty in C/MR^2 (Kaula, 1969; Michael and Blackshear, 1972) are indicated.

$T = -20\,°C$. The temperature profile chosen is one which maximizes C/MR^2 (see Figure 5); it is similar to profile B in Figure 1. Curve 2a is obtained if the lunar mantle has compressibility and thermal expansion equal to those of bronzite ($En_{80}Fs_{20}$); curve 2b is obtained if the mantle has compressibility and thermal expansion equal to those of forsterite. The temperature dependence of thermal expansion in bronzite is not well-known (Frisillo and Buljan, 1972), so is not included.

It is clear that incorporating the effects of temperature and pressure can raise both C/MR^2 and ϱ_3, and that calculations based on uniform-density layers (e.g. Gast and Giuli, 1972) can be misleading. For the temperature distribution used, C/MR^2 increased by 0.002 and ϱ_3 by 0.07 to 0.09 g cm^{-3} relative to model 1. The difference between C/MR^2 for models 2a and 2b is negligible, and indicates that for considerations of moment of inertia the precise composition of the lunar mantle, if ultrabasic, is not important. There is a small disparity in ϱ_3 (0.02 g cm^{-3}) between models 2a and 2b, a consequence primarily of the different thermal expansion for forsterite and bronzite and of the omission of any temperature dependence in thermal expansion when the bronzite parameters were used.

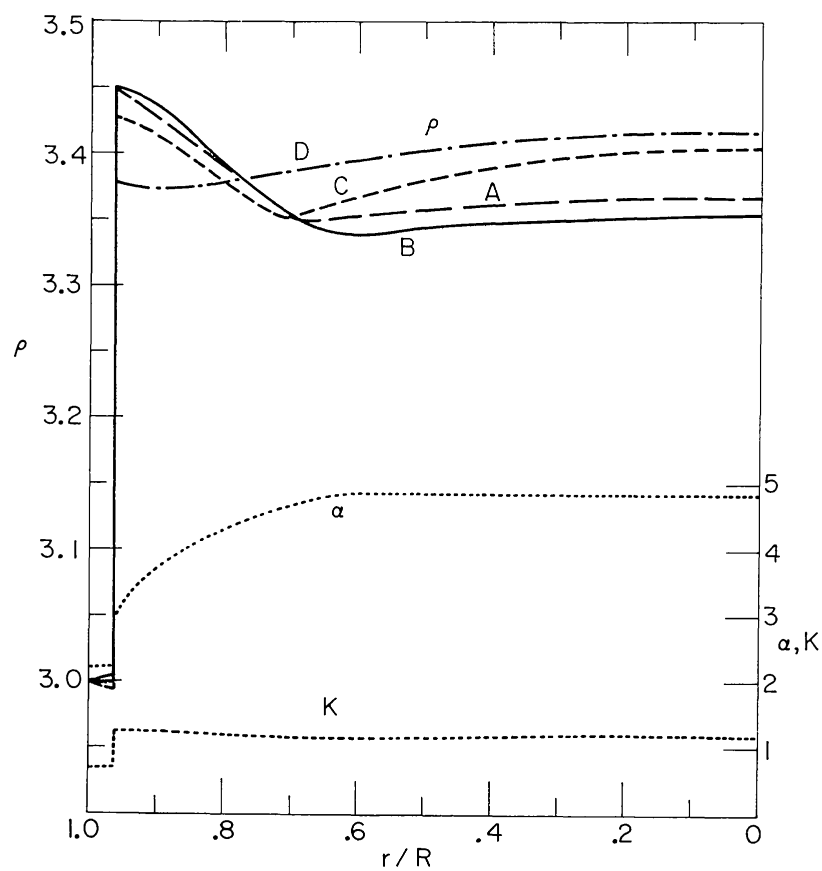

Fig. 3. The variation of density (in g cm^{-3}) with normalized radius in the Moon for several models with homogeneous mantles. The letters A through D correspond to the temperature profiles of Figure 1. A density of 3.0 g cm^{-3} is assumed for ϱ_1 and ϱ_2. The bulk modulus K and thermal expansion α are those of forsterite; the particular values used in conjunction with the density distribution appropriate to temperature profile B are shown (K is in Mbar, α in 10^{-5} °C^{-1}).

For model 3, an expression for porosity in the Moon's upper crust (see Section 2.2) is included. The mean density in the upper crust is lowered to 2.7 g cm^{-3}, and C/MR^2 and ϱ_3 are respectively lessened by 0.0005 and increased by 0.008 g cm^{-3} in models otherwise similar. It should be mentioned that if the seismic velocity profile in the upper crust is due only to narrow cracks and joints, then the porosity and its effects on density models may be negligible (Warren, 1972).

The relationship between density and temperature distributions in the Moon merits discussion in some detail. Density models corresponding to each of the four temperature profiles of Figure 1 are shown in Figure 3: for all models ϱ_1 and ϱ_2 are taken equal to 3.0 g cm^{-3}. For profile D, the effects on density of temperature and pressure nearly balance out in the lunar interior; the total range of density is only about 1% in the mantle. For the warmer temperature profiles, temperature dominates pressure in the lithosphere while the pressure effect is slightly greater than that of

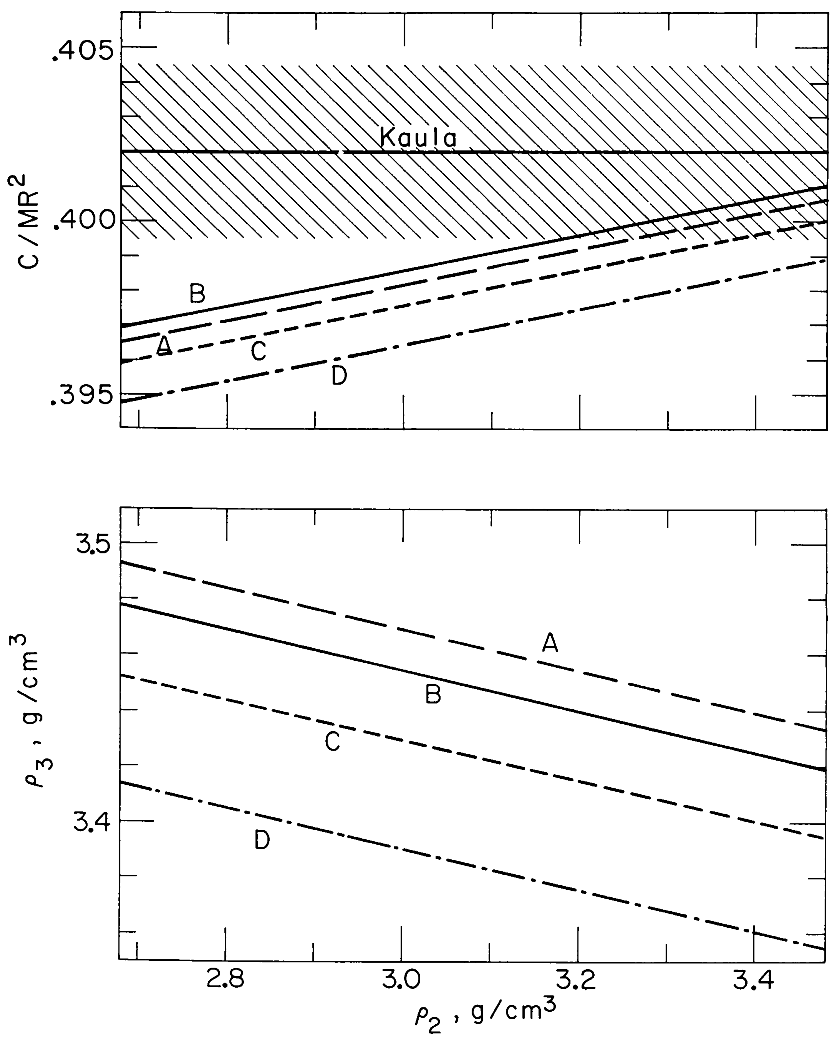

Fig. 4. The variation of the moment of inertia C/MR^2 and the density ϱ_3 of the mantle as a function of the density ϱ_2 of the lower crust for models of the lunar density profile with homogeneous mantles. The letters A through D correspond to the temperature profiles of Figure 1. A density of 3.0 g cm^{-3} is assumed for the upper crust. The shaded region outlines the limits of uncertainty in the observed moment of inertia of the Moon (Kaula, 1969; Michael and Blackshear, 1972).

temperature in the zone of melting and/or convection. The total variation of density due to pressure and temperature in the lunar mantle for these temperature profiles is 2 to 3%. The values of bulk modulus and thermal expansion used for profile B are also shown.

The moment of inertia for all of the models of Figure 3 is less than 0.400. Further, for models with uniform mantles to match the Kaula-Michael-Blackshear value of C/MR^2 of 0.402 ± 0.002, the crustal density must be unreasonably high. This is shown in Figure 4. For temperature profiles as cool as profile D (Sonett *et al.*, 1971), Figure 4 illustrates that the Moon's moment of inertia cannot be matched by a density model that has a uniform mantle and yet satisfies the Moon's total mass. For warmer temperature models, the moment of inertia can be fit to with the present uncertainty over limited ranges of density values for the mantle and lower crust. For profiles C, B and A, a value for C/MR^2 of 0.400 is obtained if ϱ_2 exceeds, respectively, 3.38, 3.18 and 3.27 g cm^{-3}. These densities are too high for a gabbroic anorthosite or anorthositic gabbro, and they would have to be raised if we included the effect of porosity in the upper crust. We might note that the density of the mantle, at standard conditions, is in the range 3.40 to 3.45 g cm^{-3} for the collection of models represented in Figure 4 that satisfy $C/MR^2 \geqslant 0.400$. Such a mantle density is thus similar to the model spinel pyroxenite ($\varrho = 3.42$ g cm^{-3}) proposed by Ringwood and Essene (1970) as the material of the lunar mantle.

That temperature profile A gives lower values for C/MR^2 but higher mantle densities than does profile B is due to the fact that higher temperatures in the former profile are confined to the uppermost 600 km. The effect of this temperature difference may be seen in Figure 3. For the same values of ϱ_1 and ϱ_2, the mantle above 600 km is less dense and the lower mantle more dense using profile A than if temperature curve B is adopted.

This observation is systematized somewhat in Figure 5. There we show the moment of inertia C/MR^2, the mantle density ϱ_3 and other related parameters for a family of thermal history models (Solomon and Toksöz, 1973). All models begin with the same initial temperature profile, include differentiation of radioactive heat sources and simulated convection upon melting, and preserve constant scaling of Th/U and K/U throughout the Moon. The models differ only in the average uranium concentration. Density models were generated as above assuming $\varrho_1 = 3.0$ and $\varrho_2 = 3.1$ g cm^{-3}. The most important feature of Figure 5 for purposes of this discussion is a maximum in C/MR^2 as a function of uranium content, or alternatively as a function of heat flow. Models with an average present-day uranium concentration near 30 ppb (parts per billion) give C/MR^2 almost large enough to match the Kaula-Michael-Blackshear value to within the stated uncertainty. Such a uranium content corresponds to a surface heat flow of 15 erg cm^{-2} s^{-1}, half the Apollo 15 value (Langseth *et al.*, 1972). For thermal models with higher or lower concentrations of radioactive heat sources, it becomes more difficult for a density model with a uniform mantle to tive $C/MR^2 \geqslant 0.400$. We should note that the values for C/MR^2 in Figure 5 might be raised slightly by adopting a solidus for the lunar mantle higher in tempera-

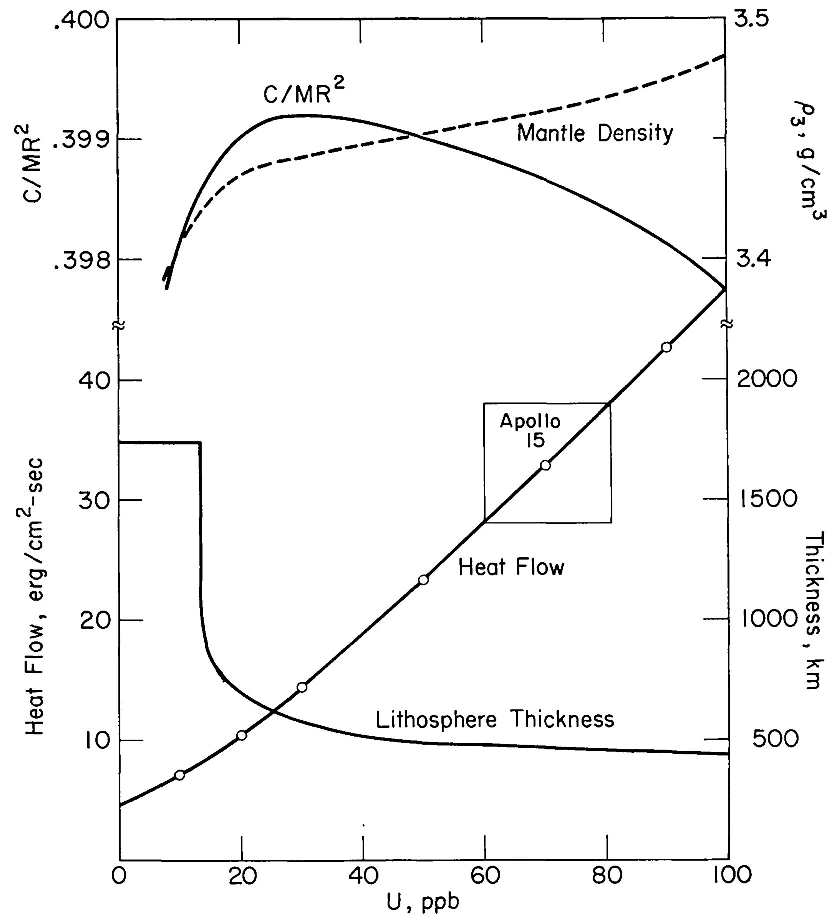

Fig. 5. The variation of C/MR^2 and other features of the Moon with average present-day uranium concentration for a family of lunar thermal history models (see text and Solomon and Toksöz, 1973, for details). The moment of inertia and mantle density ϱ_3 were calculated from the appropriate temperature profile and the lunar mass, assuming a homogeneous mantle and adopting $\varrho_1 = 3.0$ and $\varrho_2 = 3.1$ g cm^{-3}. The 'lithosphere thickness' is the depth to the point at which the temperature first approaches within 100°C of the basalt solidus. The box outlines the range of uranium concentrations that match the Apollo 15 heat flow of 33 ± 5 erg cm^{-2} s^{-1} (Langseth *et al.*, 1972). Circles denote the specific models which were investigated and from which the curves were drawn.

ture than that of mare basalt, but they would be lowered if more reasonable lower densities were adopted for the lunar crust.

For all proposed temperature profiles for the Moon, none of the density models with uniform mantles gives a value of C/MR^2 equal to or greater than 0.400 unless implausible densities are assumed for the crust or unless the crust is significantly thinner than 65 km. Thus, we consider below density models in which the mantle is inhomogeneous. Specifically, we treat two-layered mantle models that fit both the Moon's mass and moment of inertia.

5. Some Consequences of $C/MR^2 \geqslant 0.400$

As an approximation to an inhomogeneous lunar mantle, we assume the mantle is composed of two homogeneous layers. Given the densities of the crust and a temperature profile, the densities of the upper and lower portions of the mantle are single-valued functions of the depth to the boundary between the two layers. To satisfy the mean lunar density and $C/MR^2 \geqslant 0.400$, the upper mantle must be denser than the lower mantle at similar pressure and temperature conditions. The precise density contrast between upper and lower mantle is a strong function of other parameters used in constructing the density model.

This is illustrated in Figure 6, where we plot ϱ_3 and ϱ_4, respectively the densities at standard conditions of the upper and lower mantle, vs the normalized radius of the lower mantle. A value of C/MR^2 of 0.400 was used to calculate ϱ_3 and ϱ_4. Curves are shown for the several temperature distributions discussed earlier; densities of the upper mantle generally exceed 3.5 g cm^{-3}; lower mantle densities are consistently less than 3.45 g cm^{-3}. These compare with a mean lunar density, at standard

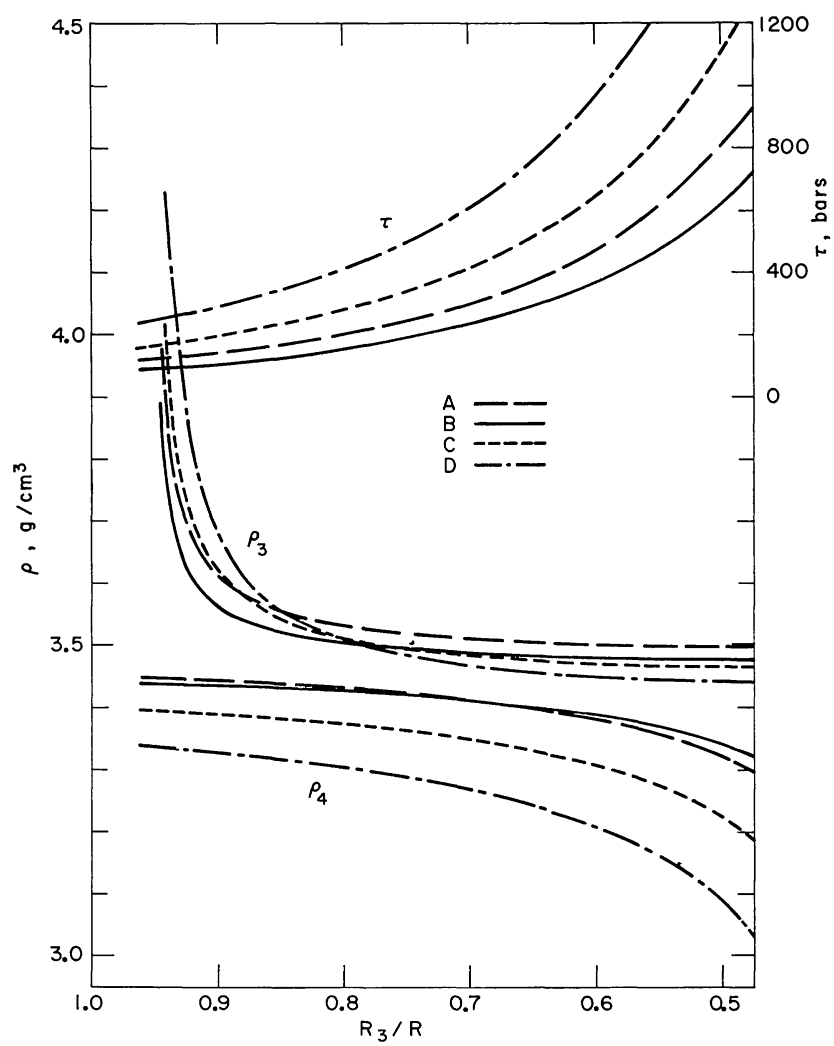

Fig. 6. The variation of densities ϱ_3 and ϱ_4 of the upper and lower mantle as a function of the normalized radius of the lower mantle. Density models match the mass and moment of inertia (0.400) of the Moon. A density of 3.0 is adopted for both the upper and lower crust. The letters A through D correspond to the temperature profiles of Figure 1. Also shown is the maximum shear stress τ implied by the density inversion in the lunar mantle.

conditions, of 3.35 to 3.42 for the temperature models considered. Minimum density contrast occurs at a lower mantle radius of 0.7 to 0.8 R (350 to 500 km depth). This minimum contrast depends on the temperature model and the assumed lower crust density but is in the range 0.08 to 0.20 g cm^{-3} for the curves shown in Figure 6.

A density reversal in the lunar mantle is, of course, gravitationally unstable. Thus for such a reversal to have persisted for several billion years requires that the Moon maintained considerable strength over that time period, at least to a depth somewhat in excess of the upper to lower mantle transition. This requirement, in view of the likelihood that the deep lunar interior may convect by solidstate creep at moderate temperatures (Runcorn, 1962; Kopal, 1962; Turcotte and Oxburgh, 1969; Tozer, 1972), strongly favors a thin upper mantle.

To put this in more quantitative form, imagine that a thin dike of lower mantle material penetrates the upper mantle to the base of the crust. Then because of the density contrast, there will be a difference in pressure between the base of the dike and the base of the normal upper mantle. The maximum shear stress τ at the base of the dike, equal to half the pressure difference, is plotted for each temperature-density profile in Figure 6 as a function of lower mantle radius. The smallest values of τ are 90 to 250 bar, depending on the temperature model; τ increases with the thickness of the upper mantle, though is still within 10% of its minimum value for $R/R_3 = 0.9$ (upper mantle thickness 110 km). The smallest values of these shear stresses, for temperature profiles similar to A or B, are in the range of the maximum shear stresses estimated to occur beneath mascons (Conel and Holstrum, 1968; O'Keefe, 1968; Wise and Yates, 1970).

Thus if C/MR^2 is equal to 0.400, a thin, high-density mantle would be permitted for certain lunar temperature profiles. The existence of such a layer receives considerable support from the seismic refraction evidence discussed earlier. For temperature distributions similar to profile D or for larger values of C/MR^2 (see Figure 12 of Solomon and Toksöz, 1973), the values of τ are at least twice the maximum shear stresses inferred from the Moon's gravitational field, so that it would be somewhat unreasonable to suppose that such stresses could have been sustained in the lunar lithosphere for several billion years. Solomon and Toksöz (1973) have discussed a number of possible explanations of the density reversal necessary to give $C/MR^2 = 0.402$, including transformations of an upper mantle rich in Al and Ca to a high-pressure mineral assemblage relatively recently in lunar history or a lower mantle that began and remained at low temperatures because of a primordial depletion in radioactive heat sources. While such hypotheses make the implied gravitational instability somewhat more palatable, all of them impose some rather special constraints on the composition or evolution of the Moon.

The suggestion of partial melting at 1000 km depth (Nakamura *et al.*, 1973) also bears on the question of a density inversion in the lunar mantle. The implications of a decrease in density at 1000 km are explored in Figure 7. In the figure, the density of the upper mantle and the density reversal required at 1000 km depth to match the lunar mass and the indicated moment of inertia are shown for the four lunar tempera-

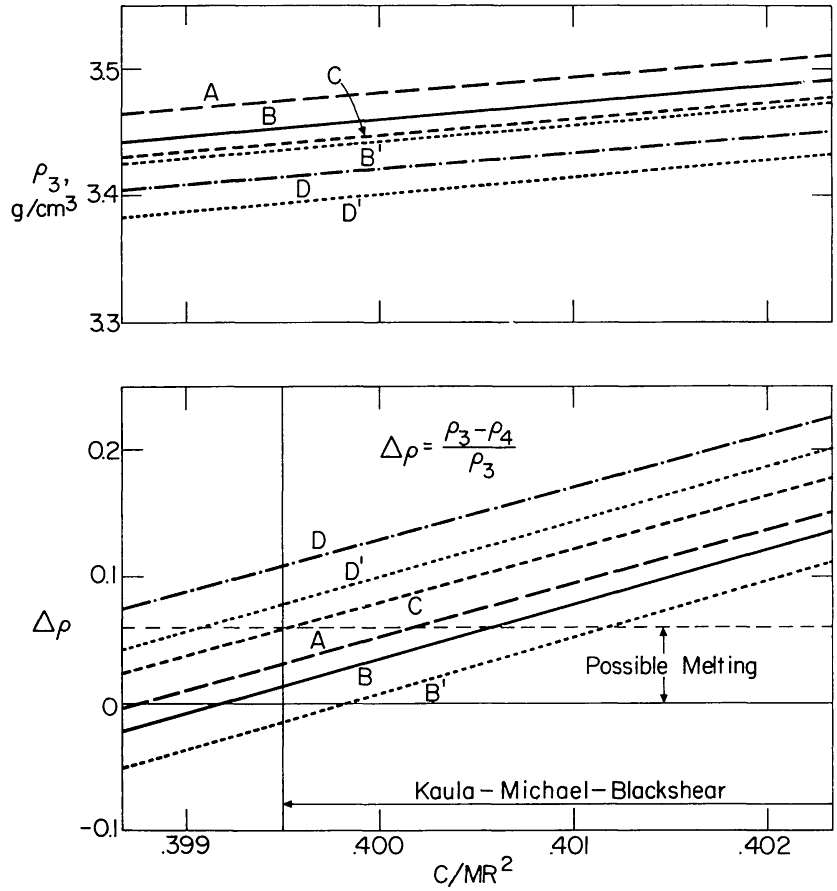

Fig. 7. The upper mantle density ϱ_3 and normalized density decrease $\Delta\varrho$ necessary to match the lunar mass and given moment of inertia if the density decrease is postulated to occur at 1000 km depth. The letters A through D correspond to the temperature profiles of Figure 1. The values $\varrho_1 = 3.0$ and $\varrho_2 = 3.1$ g cm^{-3} were used. The range of uncertainty in C/MR^2 (Kaula, 1969; Michael and Blackshear, 1972) and the range of $\Delta\varrho$ possibly attributable to melting below 1000 km are indicated. Curves with primed letters include a 3.7 g cm^{-3} layer, 20-km thick, immediately beneath the crust in the models. .

ture profiles of Figure 1. For two of the profiles we have repeated the calculations including a layer 20 km thick of density 3.7 cm^{-3} at the top of the mantle. As we might expect, a large change in density below 1000 km is necessary to give an appreciable change in C/MR^2. For this reason, some interesting conclusions can be drawn.

Even if fusion is complete below 1000 km depth, it would be unreasonable to attribute a density decrease of more than about 6 percent to the melting process (see a discussion of this point in Solomon and Toksöz, 1973). The small decrease in compressional velocity at 1000 km depth indicates that at most, partial melting is involved (Nakamura *et al.*, 1973). It is well known (e.g. Walsh, 1969; Solomon, 1972) that the relationship between melt concentration (i.e., density) and either velocity or Q is not single-valued. Thus we can only guess the maximum decrease in density at 1000 km permitted by the seismic data, but this maximum could easily be much smaller than 1%.

If a 6% drop in density at 1000 km depth is allowed, then models with homogeneous upper mantles and temperature profiles A, B or C can give $C/MR^2 \geqslant 0.400$. The same statement cannot be made for temperature profile D, even if the thin, high density layer is included. If the largest reasonable density decrease at 1000 km is closer to 1%, only temperature profiles A and B give C/MR^2 equal to or greater than 0.400, and then only if the high-density upper mantle layer is present.

6. Some Possibilities if C/MR^2 <0.400

Since values for the Moon's moment of inertia below the range given by Kaula (1969) and Michael and Blackshear (1972) have been proposed, it is worthwhile to consider what implications such lower values would have on the lunar density distribution. From Figure 4 we can see that if C/MR^2 equals 0.396 to 0.399, then density models with chemically uniform mantles can match the Moon's mass and moment of inertia for at least one of the proposed temperature profiles. Further, temperature profile D would return to some favor if C/MR^2 were less than 0.397. In general, a value for C/MR^2 between 0.396 and 0.400 would be rather dull, in that the moment of inertia would only weakly constrain the structure and temperature of the lunar mantle.

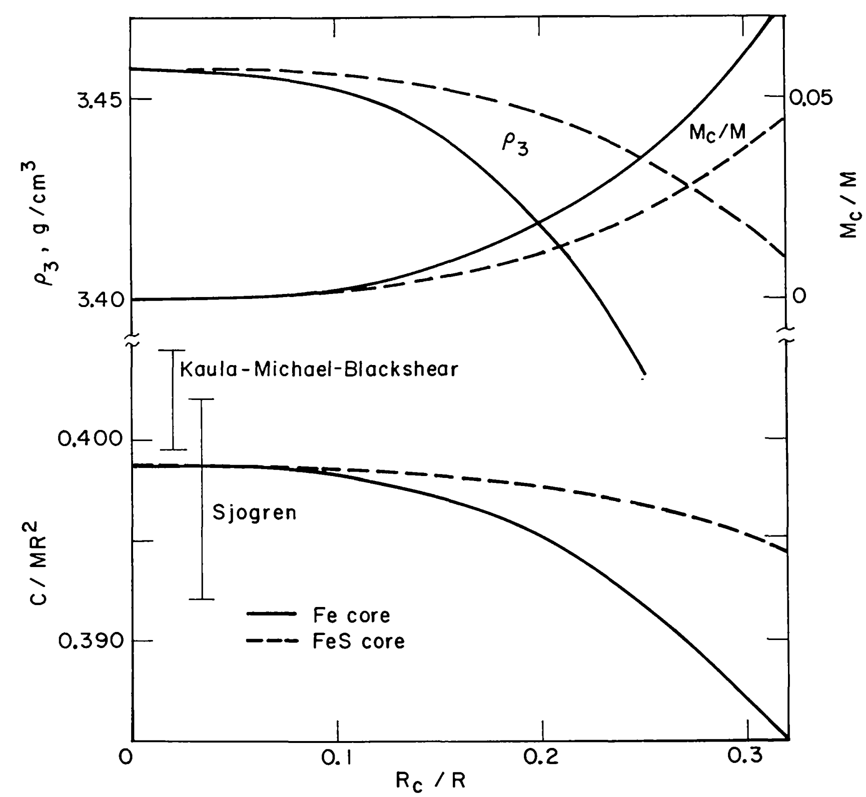

Fig. 8. The effect of an iron-rich core of radius R_c on the moment of inertia C/MR^2 and the mantle density ϱ_3 in two models of the Moon with homogeneous mantles. The two models differ only in the assumed core material. The mass M_c of the core as a fraction of the lunar mass M and two recent estimates (Kaula, 1969; Michael and Blackshear, 1972; and Sjogren, 1971) of C/MR^2 for the Moon are also shown.

On the other hand, a value for C/MR^2 of 0.395 or less, while improbable, would be more interesting. Either a relatively dense core or a crust significantly thicker on the average than near Oceanus Procellarum might be inferred. There is little point in quantitatively pursuing this discussion very far, but it is useful to consider briefly the effects on density models of a heavy central core.

In Figure 8 we show the moment of inertia and mantle density as functions of the radius of a lunar core. Curves are shown for cores of pure iron and pure troilite (FeS), which should bracket possible core densities. Temperature profile B was used, with the compressibility of α-Fe and the volume thermal expansion of iron (Clark, 1966) used for both core materials.

As pointed out by many previous authors, the mass and moment of inertia of the Moon, even if known to better accuracy than at present, are insensitive to the presence of a very small core ($R_c/R \leq 0.1$). If C/MR^2 exceeds 0.400, then of course a large core adds to the magnitude of a density reversal required in the lunar mantle. If C/MR^2 is 0.395 or less, as noted above, a dense central core is favored for all conceivable lunar temperature profiles. For profile B (see Figure 8), a value of C/MR^2 of 0.395 can be fit with a core radius of 0.20 R (pure iron) to 0.31 R (pure FeS). These cores would occupy, respectively, 1.8 and 4.0% of the lunar mass. Smaller cores would be obtained using either different temperature profiles or a larger value for the Moon's moment of inertia.

7. Velocity-Density Contraints on Mantle Composition

The relationships among density, seismic velocity and composition (Birch, 1961; and innumerable others) have, under certain circumstances, been very useful for elucidating some of the physical and chemical characteristics of the Earth's mantle. We can at least begin now the application of such techniques to the interior of the Moon.

Given in Figure 9 are the measured density and compressional velocity for some possible constituents of the lunar mantle and an approximate relationship between density and velocity for rocks from the Moon's surface (Mizutani *et al.*, 1972). Shown also in the figure are two regions in velocity-density space which encompass the models proposed for the lunar mantle. As noted in Section 2, the compressional velocity models for the upper 60 km of the Moon's mantle fall into two groups: those with a mantle velocity between 7.6 and 7.7 km s^{-1} and those with a high velocity (perhaps 8.6 to 9.2 km s^{-1}, M.N. Toksöz, personal communication, 1973) in a thin layer at the base of the crust and an uncertain decrease in velocity below that layer.

The density of mantle material may be similarly constrained. For the bulk of the lunar mantle, we consider as a possible density (at standard conditions) the value of ϱ_3 for any permissible density model represented in Figure 7; i.e. such that C/MR^2 is at least 0.400 with a density decrease at 1000 km depth no larger than 6%. We have also considered density models with temperature profiles not given in Figure 1 and

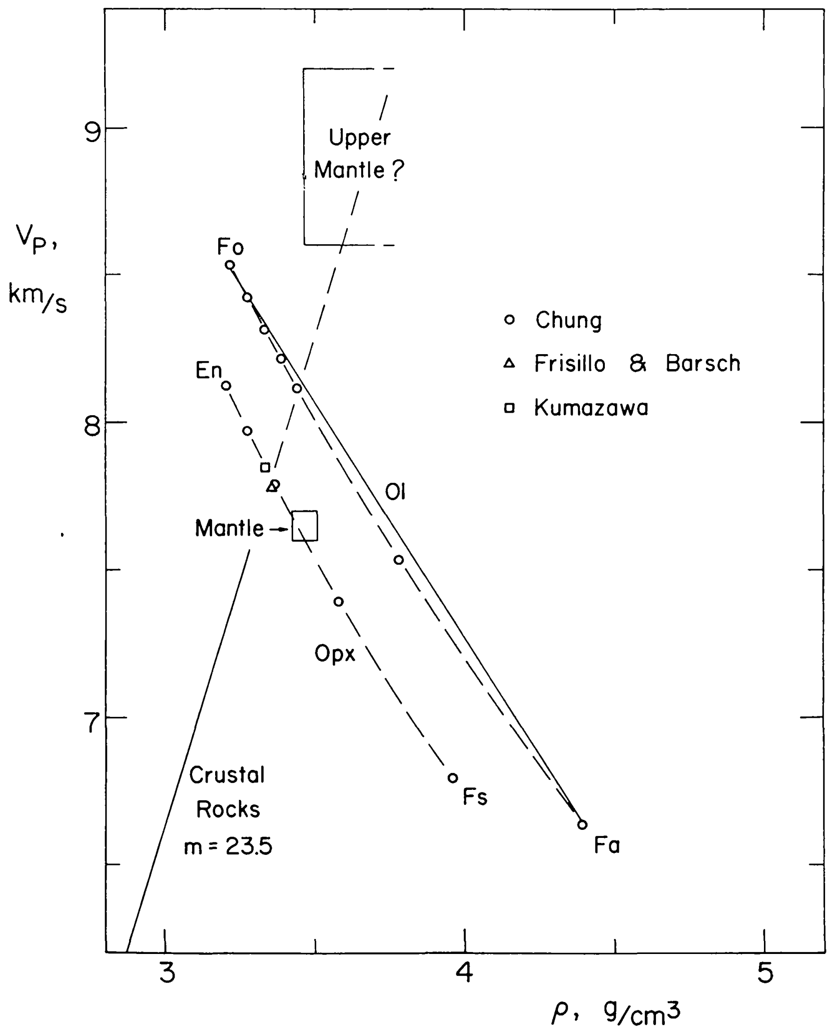

Fig. 9. Compressional-wave velocity and density in some pertinent rocks and minerals and in the Moon. The olivine (O1) data are from Chung (1970); the dashed line is a smooth curve through the data points (circles), the adjacent solid line was proposed by Liebermann (1970). The orthopyroxene (Opx) data are mostly Chung's (1971) results on synthetic polycrystalline samples; two points from single-crystal measurements (Kumazawa, 1969; Frisillo and Barsch, 1972) are also shown. The line for lunar crustal rocks of approximate mean atomic weight 23.5 is from Mizutani *et al.* (1972). The boxes delimiting the density and velocity of the bulk of the lunar mantle and of a possible high-velocity upper mantle layer are taken from arguments described in the text.

with material properties (thermal expansion and compressibility) equal to those of bronzite as well as forsterite. The range of ϱ_3 for all acceptable models is 3.42 to 3.50 g cm^{-3}.

Using the limits of 7.6 to 7.7 km s^{-1} for the bulk of the lunar mantle, the range of possible density and velocity values generates the surprisingly small box in Figure 9. Perhaps coincidentally, this box falls on the $V_p - \varrho$ line for orthopyroxenes (Chung, 1971; Kumazawa, 1969; Frisillo and Barsch, 1972) at a point corresponding to a 100 Mg/(Mg+Fe) atomic ratio of about 70. Thus the velocity and density of the mantle give at least encouragement and perhaps support to the theory of Ringwood and Essene (1970) and of Green *et al.* (1971) that the lunar mantle, the source region of mare basalts, is a pyroxenite or olivine pyroxenite with 100 Mg/(Mg+Fe) equal to about 75. It should be obvious that density and velocity alone do not uniquely determine the chemistry of a material, and other proposed compositional models for the lunar mantle (e.g. Anderson, 1973) may fall within the velocity and density constraints. Nonetheless, the velocity-density information can be very useful for eliminating possible compositional models. A dominantly olivine mantle, for instance, is ruled out by Figure 9, a conclusion which should surprise few.

A second box is included in Figure 9 for the possible high-velocity, high density layer in the uppermost mantle. The velocity constraints are cited above. The lower bound on density is determined by assuming the high-velocity material extends to great depth; i.e., it is the density ϱ_3 in models with uniform mantles which come very close to satisfying $C/MR^2 \geqslant 0.400$. The upper bound on density is indeterminate using the considerations of this paper. It is interesting, but probably not significant that the velocity-density region (i.e. box) for this high-velocity layer falls on the upwards extrapolation of the curve of Mizutani *et al.* (1972) for crustal rocks.

8. Conclusions

Density models for the Moon, including the effects of pressure and temperature, can be made to satisfy the mass of the Moon, the standard value for C/MR^2 of 0.402 ± 0.002 (Kaula, 1969; Michael and Blackshear, 1972), and the presence of a low-density crust some 65 km thick (Toksöz *et al.*, 1972a, 1974) only if the mantle of the Moon is chemically inhomogeneous. Specifically, the density of the upper mantle must exceed the density at similar pressure and temperature of the lower mantle by at least several percent. The need for an inhomogeneous mantle can be avoided if C/MR^2 is less than 0.400 or, alternatively, if the average lunar crust is considerably thinner than in the Fra Mauro region of Oceanus Procellarum. Because the mantle occupies a large fraction of the volume of the Moon, its density at standard conditions is constrained by the mean lunar density to lie in the range 3.4 to 3.5 g cm^{-3}. For the more likely models with inhomogeneous mantles, the upper mantle density probably exceeds 3.5 g cm^{-3}.

The density inversion suggested for the mantle of the Moon is gravitationally unstable. To match $C/MR^2 = 0.402$, the maximum shear stress at the base of a column or dike of lower mantle penetrating the upper mantle must be 2 to 5 times the maximum stresses associated with the Moon's gravity field. Such stresses could not generally have persisted throughout lunar history, most likely ruling out such a large value for the moment of inertia. To match $C/MR^2 = 0.400$, the maximum shear stress is comparable to the stresses inferred to be sustained beneath mascons for certain temperature profiles. A temperature distribution as cool as that of Sonett *et al.* (1971) would require too large a density reversal to have persisted for several billion years.

Recent seismic evidence suggests a thin, high density layer at the top of the mantle and a slight decrease in density due to partial melting at 1000 km depth. Density models with these features can give C/MR^2 equal to 0.400 or greater, but again only for some temperature profiles, not including the one proposed by Sonett *et al.* or others with similarly low central temperatures.

With the combined knowledge of the probable ranges in density and seismic velocity of the lunar mantle, some useful statements can be made about mantle composition. A predominantly olivine mantle can be ruled out, but a predominantly pyroxene mantle fits the present data well. Further, the Mg/Fe ratio indicated from the velocity and density information agrees with the value inferred from high-pressure melting experiments on mare basalts.

A more concrete treatment of the density distribution must await a more definitive determination of the lunar seismic velocity profile and, more importantly, a more accurate value for the Moon's moment of inertia. As long as workers active in the determination of the lunar gravitational field cannot agree on a value of C/MR^2 to within better than 1 or 2%, then firm handles on the nature of the Moon's upper mantle or tight contraints on the properties of a lunar core must elude us. The Lunar Orbiter experiments vastly improved our knowledge of the Moon's gravitational field, especially considering that the classical value for C/MR^2 indicated the frightening possibility that the Moon might be hollow. More accurate determination of the principal moments of inertia of the Moon, utilizing high-latitude and backside tracking data, should remain a high priority.

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