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Is the Moon Hot or Cold?

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Introduction

Baldwin (*1*) has summarized some of the evidence for the early and present internal thermal states of the moon. He argued that the interior of the moon, below about 200 to 300 kilometers, is presently "hot," that is, has interior temperatures close to or exceeding the lunar solidus curve. The evidence for a rapid differentiation of the moon about 4.6×10^9 years ago, the extensive igneous episode resulting in mare formation 3.7×10^9 to 2.8×10^9 years ago, the depletion of the moon in volatiles and its enrichment in calcium, aluminum, and the trace refractory elements all argue for a hot origin and high initial temperatures. A straightforward consequence of the lunar thermal inertia is that if the interior of the moon were ever hot it would remain hot to the present. The high surface concentrations of uranium, thorium, and potassium, and the Apollo 15 heat flow value of 33 ergs per square centimeter per second (*2*), indicate high present-day temperatures in the lunar interior (*3*).

In apparent contradiction to these conclusions, recent interpretations of the lunar conductivity profile, the nonhydrostatic shape of the moon, the existence of mascons, the remarkable aseismicity of the moon, and the absence of present-day volcanism have been interpreted as suggesting that the lunar interior is presently cold and, by implication, has always been cold.

Various attempts have been made to reconcile the conflicting evidence by postulating that the interior is cold and the surface manifestations of differentiation apply only to the outer reaches of the moon. By implication, the deep interior has always been cold and always deficient in uranium, thorium, and potassium. Remarkable constraints must then be placed on the accretional process on the moon with respect to bulk and trace element geochemical zonation (*4*).

We find that the basic observations do not demand a presently cold moon and are, in fact, consistent with a hot moon. We find that an iron-deficient, highly resistive, hot lunar interior, capped by a cool, rigid lunar lithosphere with a thickness of several hundred kilometers, can explain the relevant observations and is a reasonable model of the moon today.

Strength of the Moon

The nonequilibrium shape of the moon and the existence of lunar mascons suggest that the lunar interior possesses a long-term resistance to relaxation of the implied elastic stress differences. This, in turn, suggests that either a finite elastic strength or a high viscosity is a basic property of the lunar interior, but neither is consistent with its possessing hot temperatures (*5*). Several aspects of this crucial interpretation have not been generally appreciated.

It has been pointed out several times that the apparent departure of the moon from hydrostatic equilibrium is not so remarkable as might be supposed (*6*). Because of the smaller gravity on the moon and its lower mean density, it should be able to support much larger excess mass loads than the earth. In fact, the inferred mass excesses and the implied stress differences are less on the moon than on the earth.

Kaula (*7*) has shown that the moon is gravitationally smoother than the earth, provided that gravity anomalies are considered equivalent if the corresponding surface mass anomalies produce equal stresses. This is the "equal stress implication" (*8*) according to which gravity anomalies are scaled with the gravitational field of the appropriate planetary object. In terms of the second degree harmonics of the gravitational fields, Lorell *et al.* (*9*) have shown that the moon is several times smoother than the earth, and that Mars is considerably rougher than either. If these results are indicative of the bulk strengths of the planetary interiors, as implied by Urey and MacDonald (*5*), the qualitative conclusion is that the earth is stronger than the moon. If this measure of strength is, in turn, indicative of interior temperatures, the conclusion is that the moon is hotter than the earth, not colder.

Similarly, the lunar mascons, representative of the higher degree harmonics of the lunar gravitational field, are hardly dramatic features by terrestrial standards. Kaula (*7*) notes that there are at least 20 terrestrial "mascons" larger than the largest lunar mascon at Mare Imbrium, and that the largest of these is an order of magnitude larger than that at Mare Imbrium.

Moreover, it is not essential that the lunar gravitational anomalies, of whatever degree, be supported by the deep lunar interior. While the nonequilibrium shape of the moon implies stress differences of 15 to 20 bars if it is supported throughout the lunar interior, it implies stress differences of only 28 to 50 bars if it is entirely supported by an exterior shell 400 km thick overlying a deeper interior devoid of strength (*10*).

The shapes of the positive gravity anomalies associated with the lunar mascons imply mass excesses at depths

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of 25 to 125 km (*11*), in terms of a model of layered slabs of high-density maria fill. Depending on the density contrast, it is estimated that the thickness of the maria pile is 15 to 30 km. This model implies that the exterior of the moon is capable of supporting shear stresses in the 50-bar range at a depth of about 80 km. Urey and MacDonald (*5*) obtained similar stress differences in about the same depth range, 70 to 120 km, to support the higher degree harmonics of the lunar gravitational field.

More recent data on the lunar gravitational field strongly suggest that the lunar mascons are near-surface features (*12*). Hulme (*13*) has shown that the positive gravity anomalies associated with circular maria can be maintained even if the moon cannot support stress differences below depths of the order of 100 km. Baldwin (*14*) estimated that the crater shapes imply stress differences of 30 to 50 bars in the upper 50 km of the moon.

One alternative to the interpretation of Urey and MacDonald (*5*), then, is the existence of an exterior shell several hundred kilometers thick, capable of supporting long-term stress differences of 50 to 100 bars. This estimated strength is substantially less than measured rock strengths, which are of the order of several kilobars or more (*15*). It is also less than the strength inferred for the earth's crust and upper mantle, which ranges from 100 bars to over a kilobar (*16*). Estimates of the minimum strength required to maintain the nonhydrostatic shape of the earth range from 20 to 97 bars (*17, 18*).

The implied stress differences have probably not been maintained for billions of years by finite elastic strength; they have most likely been relaxing with time through slow anelastic deformation of the stressed exterior shell. To a first approximation, isostatic compensation can be modeled as a viscoelastic process. The time scale of the persistence of these features can then be translated into a minimum viscosity, variously estimated to be 10^{26} to 10^{27} poises (*5, 7*) for the case of a uniform moon. If the moon is modeled as a highly viscous layer over an interior of low viscosity (*19*), its nonequilibrium shape and the persistence of mascons imply viscosities of 2×10^{24} to 5×10^{25} cm^2/sec for a shell 200 km thick. These estimates are greater than the viscosity of the earth's upper mantle, 10^{21} cm^2/sec, but are less than or comparable to theoretical and experimental estimates of the viscosity of the terrestrial lithosphere (*20*). If the viscosity of a lunar lithosphere 200 km thick is 10^{25} poises, the relaxation times for the nonhydrostatic shape and the mascons are 2.5×10^9 and 1×10^9 years, respectively.

On the other hand, any estimates of relaxation times based on a linear relation between stress and strain rate (Newtonian viscosity) can be misleading; the stress–strain rate relation is probably highly nonlinear. Experimentally obtained creep laws for rocks have the form $d\varepsilon/dt \propto \sigma^5$, where $d\varepsilon/dt$ is the strain rate and σ is the stress (*21*). This leads to high strain and rapid stress release at high stress levels, and slow stress relaxation at low stress levels. If the stress differences were originally of the order of 1 kbar in the moon and have decayed to 50 bars in 4×10^9 years, one can calculate that the stress difference decayed to 500 bars in 4×10^5 years, to 100 bars in 2.5×10^8 years, and to 70 bars in 10^9 years. Another 2×10^9 years only results in a further decrease in stress of 16 bars. Stresses of a few tens of bars can therefore be regarded as essentially permanent (*22*).

The various estimates of implied stress differences in the moon are, of course, subject to some uncertainty. However, the observations discussed above are consistent with the existence of a rigid, cool lithosphere overlying an interior with little or no strength.

Lunar Electrical Conductivity Profile

Electrical conductivity profiles (*23, 24*) have seemingly provided the most direct and compelling evidence for a cold lunar interior. Apart from the uniqueness of the electrical conductivity profile itself, the nonuniqueness associated with inferring temperature from conductivity does not seem to be generally appreciated (*24*).

Laboratory measurements of electrical conductivity are generally interpreted with the relation $\sigma^E = \sigma_0{}^E \exp(-A/kT)$, where σ^E is the electrical conductivity, $\sigma_0{}^E$ is a constant that depends strongly on the bulk and impurity compositions and oxidation states (*25*), A is the activation energy, k is the Boltzmann constant, and T is the absolute temperature. For constant $\sigma_0{}^E$ the electrical conductivity is a strong function of temperature, low values of σ^E implying low temperatures.

Sonett *et al.* (*23*) have presented electrical conductivity profiles for the moon; they interpreted these profiles in terms of laboratory measurements of σ^E for lunar basalts and terrestrial samples of olivine and peridotite, and obtained temperatures less than 1000°C throughout the moon. On the basis of the claim that olivine and olivine peridotite "represent the least conducting geological material," they suggested that 1000°C is an upper bound on temperatures in the moon. More recent electrical conductivity measurements, however, have revealed still lower conductivities for olivine (Fig. 1).

Figure 1a is a summary of laboratory measurements of σ^E for several different materials. The hatched region represents the range of values obtained for the Apollo 11 and Apollo 12 lunar basalts (*26*). The unlabeled lines are for olivine and peridotite, from England *et al.* (*27*). These are the conductivity measurements on low-resistivity materials that have been used in determining the lunar temperatures. The curves labeled 1 to 5 in Fig. 1a are laboratory measurements for magnesium oxide, aluminum oxide, and three members of the olivine series with differing fayalite contents.

As shown in Fig. 1b, if the lunar conductivity profile is interpreted with any of these five materials, the corresponding lunar temperatures are considerably higher than those obtained by Sonett *et al.* (*23*) and Dyal and Parkins (*24*). Indeed, the temperature distributions so obtained are consistent with present-day temperatures obtained through thermal history calculations (*3*) and a presently hot lunar interior.

Of particular importance in the interpretation of the lunar conductivity profile are the amount and oxidation state of iron (*28*). The electrical conductivities of the end members of the olivine series differ by six orders of magnitude, forsterite being the more restive. The moon is clearly depleted in iron relative to terrestrial, solar, or chrondritic abundances. In terms of the olivine series, the likely contents of ferrous oxide in the lunar interior would still allow for a variation of three orders of magnitude in the resulting conductivity.

Equally important is the oxidation state of the iron. Material 4 of Fig. 1 is an olivine of 9.4 mole percent fayalite content, which is essentially free of the ferric ion. This material has a smaller electrical conductivity than other olivine samples possessing less total fayalite but more Fe^{3+}. The near absence of

Fe^{3+} in lunar materials suggests that they crystallized at extremely small partial pressures of oxygen. Thus, even if the bulk chemistry of the lunar interior indicated an FeO content equivalent to 10 mole percent fayalite, the lunar interior would still be highly resistive because of the lack of Fe^{3+}.

If so, the lunar conductivity profile is consistent with interior temperatures at or near the solidus curve (*28*). The inferred temperatures at depth are even higher if the FeO content of the lunar mantle is lower than that of the crust. The lunar conductivity profile suggests that this is the case. In the conductivity profile of Sonett *et al.* (*23*), the conductivity decreases by nearly three orders of magnitude between depths of 250 and 350 km. This is an indication of a compositional change, a phase change, or both, at this level. Sonett *et al.* (*23*) concluded that the material below the conductivity peak must be several orders of magnitude less conducting than dunite and that "such material is not known on earth." Actually, the conductivities of MgO and Mg_2SiO_4 at about 1150°C are close to those measured at a radius of 1400 km. The conductivity of more reasonable mineralogies such as calcium-aluminum garnet, diopside, spinel, and enstatite cannot be estimated at present. The precipitous decrease in electrical conductivity can most reasonably be interpreted as a decrease in FeO content from about 15 to 5 percent or from about 10 percent to essentially zero. This is consistent with the idea that the bulk of the moon is composed of the refractory materials that condensed before iron in the cooling solar nebula (*3, 29*).

Since the average FeO content of the lunar crust is approximately 10 percent and since little Fe^{3+} is in evidence from the lunar samples, it might be expected that material 4 gives a fair measure of the electrical properties of the outer part of the moon. Indeed, the temperatures inferred from the conductivity measurements on this material are remarkably consistent with those obtained from the thermal history calculation for the outer several hundred kilometers. The precipitous drop in lunar conductivity at a radius of 1500 km, however, gives a physically unacceptable local temperature minimum. This drop is probably indicative of a change in composition at a radius between 1400 and 1500 km. It is interesting that the condition for smooth temperature gradients can be met with any of the materials low in iron. A possible interpretation of this very rapid decrease in the lunar conductivity profile is that the FeO content of the lunar lower mantle is considerably less than that of the upper mantle or the crust, and that this decrease occurs in the depth interval from 200 to 300 km. If this is the case, the lunar interior must be highly resistive and hot to explain the observed conductivity profile. In the outer several hundred kilometers, any relative depletion of Fe^{3+} implies higher temperatures than those obtained in the same region by Sonett *et al.* (*23*).

Relative Absence of Present-Day Volcanic and Seismic Activity

The low level of lunar seismicity (*30*) has also been used as an argument for a cold interior (*31*), although the details of the argument have not been presented. Terrestrial experience indicates the contrary. Laboratory results show that at low temperatures rocks fail by brittle fracture. At high temperatures they yield by steady-state sliding or aseismic creep. Thomsen (*32*) has noted that estimates of the lunar pressure-temperature regime would place the outer several hundred kilometers of the moon well within the experimental stable-sliding regime. The absence of lunar seismicity can be attributed to lack of suitable stresses or to the nonbrittle behavior of the interior. The latter implies high temperatures.

The inverse correlation between seismic activity and proximity to the melting point is well known (*33*). Most terrestrial earthquakes occur in the cold, upper part of the crust. The absence of earthquakes at depths greater than 16 to 20 km in California, the absence of terrestrial earthquakes below 700 km, and the minimum in earthquake activity in the vicinity of the low-velocity zone can all be explained in terms of high-temperature phenomena.

On the other hand, the low level of lunar seismicity must mean that the stress differences necessary to cause fracture over dimensions of kilometers

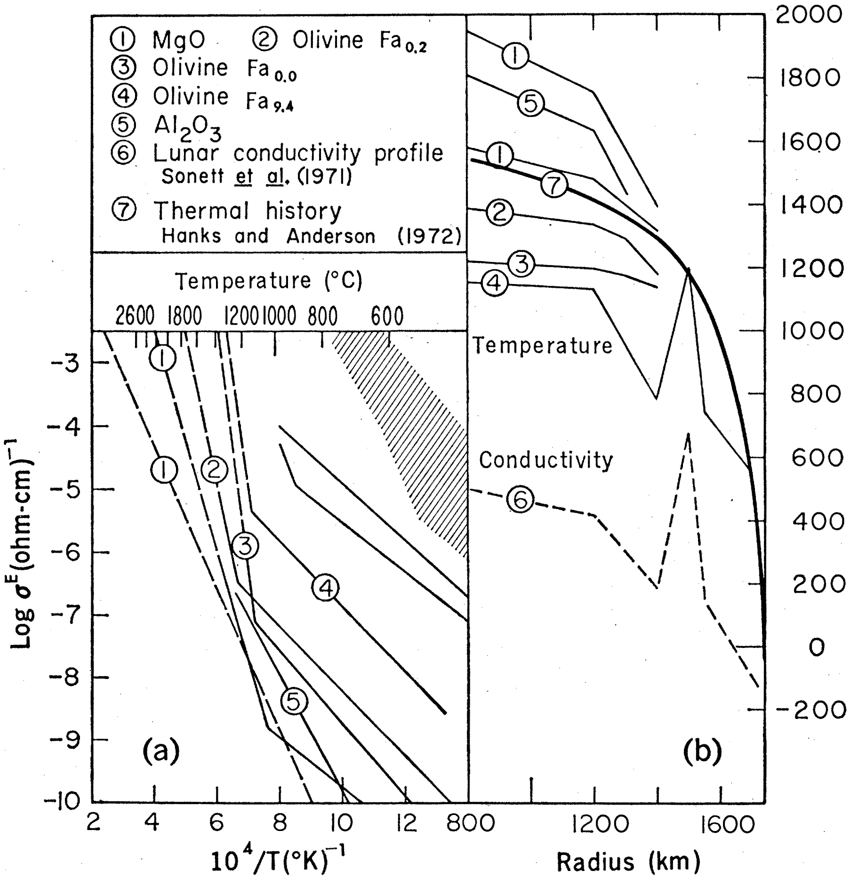

Fig. 1. (a) Electrical conductivities of MgO, Al_2O_3, forsterite (Mg_2SiO_4), and olivines of varying fayalite (Fe_2SiO_4) contents (*40*). (b) Electrical conductivity of the moon and inferred temperatures. The numbers on the curves refer to the compositions used in inferring the temperatures and correspond to the numbers on the electrical conductivity curves. Curve 7 is a thermal history model from Hanks and Anderson (*3*).

or more are not available in the moon. The existence of a rigid lithosphere that is capable of supporting stress differences of 50 to 100 bars does not conflict with the observation that many terrestrial earthquakes commonly have stress drops between several tenths of a bar and several hundred bars. The stress drop of an earthquake in no way constrains the preexisting stress except that the stress drop is a minimum estimate of the preexisting shear stresses.

It has also been suggested that the absence of lunar igneous activity after about 3.0×10^9 years ago implies that the interior has been cold, at at least not partially molten, since that time (*34*). Igneous activity certainly implies partial melting at depth in a planet, if impact melting is ignored, but the absence of igneous activity does not imply the absence of partial melting. We would not expect partial melting at depth to result in extrusive igneous activity if the melt content is small (less than 5 percent) or if stresses, either internal or external, are insufficient to break the overlying layer, the lithosphere. As noted above, the implied stress differences in the lunar lithosphere are far below the breaking strength of unfractured rock; from another point of view, the low level of lunar seismic activity suggests that suitable stress differences, including the effective dimension, are simply not available to rupture a lithosphere of several hundred kilometers.

Igneous activity on the earth is generally restricted to areas where the lithosphere is thin or where plates are colliding. It is absent, for example, on old shields which, on the basis of geophysical data, have thick lithospheres. Igneous activity is presumably shut off when the lithospheres gets too thick or the melt content of the interior gets too small, or both. Thermal history calculations show that the lithosphere thickens with time. Extrusive activity might also be triggered by rupturing of the lithosphere by large impacts or tidal stresses. These would both have been more effective in the early history of the moon.

On the earth, both seismicity and volcanism are closely related to the boundaries between moving plates. Shallow seismicity and basaltic magmatic activity occur in regions of plate separation; deep seismicity and andesitic volcanism occur when plates converge.

With respect to the moon the absence of seismicity and volcanism need only reflect the absence of relative plate motions, or plate tectonics as we know it on the earth. The absence of these phenomena in continental shield areas does not imply a cold terrestrial interior, but more likely reflects the great thickness of the lithosphere, 100 to 200 km. Plate mobility and penetration of the lithosphere by magma are probably consequences of the large thermal and mechanical inertia of the interior relative to the plate. The earth has a lithosphere that is negligible compared to the mass of the planet. The reverse is obviously true for the moon if the lithosphere is several hundred kilometers thick.

Thermal History Considerations

The most decisive evidence favoring a hot lunar interior—the evidence for a rapid differentiation of the moon, the extensive igneous episode resulting in mare formation, and the high surface concentrations of uranium, thorium, and potassium—is somewhat ambiguous: these phenomena need only have involved the outer 300 to 400 km of the moon, the deeper interior playing only a minor part in the development of surface phenomena. The distribution of the radioactive heat sources, however, greatly constrains the number of possibilities.

If the deep lunar interior has uranium concentrations greater than approximately 30 parts per billion it must be at least partially molten at the present time, even if it started cold. The surface concentrations of uranium are 10 to 100 times larger than this. If the single lunar heat flow value is at all representative of the average value, similar concentrations of uranium must persist to depths of 50 to 200 km. The average uranium concentration of the moon which is consistent with the surface concentration and the observed heat flow is 0.09 part per million (*3*).

The surface material of the moon and, by inference, the source region of the lunar basalts is enriched in the refractory elements relative to carbonaceous chondrites. If the lunar basalts and anorthosites are mixed in the proportions required to remove the europium anomaly (*35*), the average uranium content is 0.12 ppm, which is 12 times the average chondritic value. The moon is apparently depleted in materials more volatile than iron. If we assume that the moon is composed of only those materials more refractory than iron, we estimate that it should be enriched in such refractories as uranium, barium, strontium, and the rare earth elements by a factor of about 10 relative to carbonaceous chondrites, which are assumed to have their full complement of condensable material. This leads to an estimate of 0.1 ppm for the uranium abundance in the moon. Corresponding enrichments in thorium can be expected. The same reasoning leads to a depletion in potassium which, however, contributes only a minor amount of radioactive heating. Eucrites provide a close match to the lunar surface chemistry; they contain 0.1 ppm of uranium. These estimates are remarkably consistent with each other and suggest that uranium and thorium are strongly enriched in the moon relative to their chondritic or cosmic abundances. There is certainly no reason to believe that they might be depleted.

Other Considerations

Whether the deep interior of the moon is hot or cold has an important bearing on the overall composition of the moon and its origin. The chemistry of the lunar surface materials requires that their source region be enriched in calcium, aluminum, and the refractory trace elements such as uranium, thorium, barium, strontium, and the rare earth elements, and depleted in volatiles (*36*). The amount of this material, the inferred degree of partial melting, and the evidence for the great depth of the source region suggest that a substantial fraction of the moon is enriched in these refractory elements. The lunar seismic experiment has yielded data that is also consistent with the enrichment in calcium and aluminum of at least the outer 100 km (*37*). On the other hand, it has been suggested (*38*) that the whole moon cannot be enriched in calcium and aluminum because of the density associated with the garnet-rich assemblage that is stable at modest pressures. A ferromagnesian or chondritic interior has therefore been proposed (*4*), and it has been suggested that the exterior of the moon is composed of the initial high-temperature condensates. However, if temperatures are high the whole moon can be composed of high-temperature phases rich in calcium and aluminum, and there is no requirement for an ad hoc initial chemical layering of the above type. High temperatures move the stability field of plagioclase deep into the interior of the moon and this serves to decrease the mean density. A small core composed of the high-pressure phase assemblage can be tolerated without violating the mean density or moment of inertia. This removal of a constraint on the internal composition of the moon has been discussed by Anderson (*39*).

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Material	*T* (°C)	Confining pressure (kbar)	Ultimate strength (kbar)
Anorthosite	150	1.0	5.9
	500	5.1	9.4
Basalt	300	5.0	13.8
	800	5.1	2.6
Granite	150	1.0	3.3
	500	5.1	8.3

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41. Supported by NASA grant NGL 05-002-069. Contribution No. 2188, Division of Geological and Planetary Sciences, California Institute of Technology.

Louis Pasteur Sesquicentennial (1822–1972)

J. R. Porter

Born on 27 December 1822, Louis Pasteur spent his early life in Dôle and Arbois, France. The most unique feature of this period was his rigorous schooling, with his father—a tanner by trade—checking his lessons every evening. Between the ages of 13 and 18, Pasteur demonstrated great artistic ability, and he received a certificate in arts in 1840 at Besançon. His drawings were precise, and his attractive pastels and lithographs were so well done that art critics even today say he could have made a great reputation for himself in the arts (Fig. 1). Years later, from 1863 to 1867, he did serve at the École des Beaux Arts in Paris as professor of geology, physics, and chemistry in relation to the arts.

At the age of 21, Pasteur left his native province and went to Paris, where he came in contact at the Sorbonne with one of the most distinguished scientists of the time, Jean-Baptiste Dumas, who encouraged him to study chemistry at the École Normale Supérieure from 1843 to 1846. From this time until 1885, he made great and ingenious discoveries in at least nine different areas of science, and he influenced scientific thought and investigation in many fields. His major fields of research are carved in

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