Reports

Summary of Arguments for a Hot Moon

Abstract. *Analysis of physical and chemical observations shows that the interior of the moon is now and always has been hot. It is close to the melting point at each level beneath a cool outer zone.*

A key datum in our understanding of the moon's history is the state of its interior at present. Many interpretations have been made, and some scientists even today consider the moon to be cool or cold inside. My purpose in this report is to summarize the physical and chemical observations bearing on this question. When the pertinent observations are collected, I conclude that they point unambiguously to the following model.

1) Very early in its history, the moon became melted and chemically differentiated.

2) The first observable melting produced a relatively light-colored, low-density material, which was exuded upward and formed a moonwide surface layer roughly 10 to 16 km thick. The lavas were probably melted by the energies of the moon's accretion and by subsequent compaction, possibly aided by short-lived radioactivity.

3) The outer layers of the moon quickly solidified and became thicker and more rigid as time went on.

4) Impact cratering continued at a declining rate all through these early stages.

5) By the time the giant impact craters, such as Mare Imbrium, Mare Humorum, Mare Humboldtianum, Mare Crisium, Mare Nectaris, and Mare Moscoviense, were formed, the moon was solid and quite rigid to a depth of at least 50 km, and probably to several hundred kilometers.

6) The impacts that produced the circular mare basins were made in solid rock, and the basins were formed dry. They did not release lava from below. There may have been a minor amount of impact melting, but, if so, this was distinct from the mare lavas.

7) A substantial period of time elapsed before the vast lava flows of the maria occurred.

8) The lavas came roughly 10^9 years after the formation of the moon. They came not in one vast surge but in hundreds or thousands of localized floods. They were driven upward by the volume increase due to melting.

9) Some smaller lava flows continued to appear throughout the balance of geologic time.

10) The lavas that formed the maria came from a considerable depth, probably close to 200 km but possibly deeper. This second melting presumably was caused by radioactive decay.

11) In general, the outer layers of the moon are cool and very highly viscous. The viscosity drops with increasing temperature and increasing depth.

12) In the deep regions, the present temperature is just below the melting point at the particular pressure involved. In other words, the moon is now hot and always has been hot.

13) The lavas of the maria were formed by a small partial melting of deep-seated ultrabasic materials.

Briefly these conclusions were reached as follows. The Surveyor 7 analysis of the highlands area material near Tycho (*1*) indicated that it was distinct from the mare material analyzed by Surveyor 5 (*2*). It was less basic.

Mixed in the dust and fines brought back by Apollo 11 (*3*) and Apollo 12 (*4*) was a minor amount of anorthositic material whose chemistry is clear evidence of a chemical differentiation. The anorthositic material is considered to have been derived from the uplands. We do not know that the uplands are composed largely of anorthosite. We do know that they are less basic than the maria and that their lower density is consistent with the approximate condition of isostatic equilibrium over most of the moon found by Muller and Sjogren (*5*). The uplands appear to have been formed by a partial melting and chemical differentiation of the body of the moon. This differentiation occurred before the formation of the earliest craters which are much older than the maria. Therefore the melted upland surface was produced almost in the beginning of the moon's history. This means that the early moon was hot enough for very extensive melting to occur and that the hot melt spread over the original surface. If the moon were originally entirely melted, the anorthositic layer, being less dense, floated on the deeper and more basic material. This latter interpretation seems doubtful. The moon may have accreted from hot or cold particles, but the body of the moon promptly became hot. The gravitational energies released during accretion heated the outer parts of the moon to a considerable depth. If the moon accreted during 20 million years, which is about the longest permissible time (*6*), the rate of accumulation of mass averaged considerably more than 10^4 times that from the earliest detectable cratering (*7*). The heat may not have penetrated to the moon's center originally because the center grew under low gravitational pull, but, as the moon grew in size, the impacts released much heat. In any event, calculations show that a moon with a cool center and a hot outer portion would increase in temperature until the inner parts became very hot, particularly if radioactivity were present (*8, 9*).

The present outer surface layer of low-density and less basic material is relatively thick. This may be derived from the relative heights of the uplands and the maria and from the general nature of the mascons (*10*). We do not yet have fully satisfactory solutions to the problems of the mascons, but density variations must play a prominent part. The fillings of the mascon basins, whatever their nature, are considerably denser than the surface material removed to allow the mascon to form (*11*).

As judged by the amount of isostatic adjustment due to the superimposed loads of the maria, the thickness of the anorthositic layer is about 10 to 16 km.

If this is so, its volume is about 0.017 to 0.027 that of the moon. In order to produce this light surface layer, a partial melting of about 9 percent is required (*12*). Consequently, a minimum volume of 0.19 to 0.30 of the moon had to be heated to the melting point if the light layer material was to be driven off. Geometrically this volume corresponds to the outer 120 to 200 km of the moon or to the inner 1000 to 1160 km, depending on where the melting occurred. In any event, a very large fraction of the moon became very hot in the period before the earliest detectable cratering.

Impact cratering may be traced to very early in the moon's history (*7, 13, 14*). It has been concluded that the oldest craters are deformed by sinking rims and rising bottoms (*15*), because impact melting, gravitational collapse, erosion, and mass wasting were found inadequate to account for the evolving shapes of craters. These long-term changes in form are characteristics of isostatic adjustments. The rate of adjustment declined throughout geologic time, and the viscosity of the outer layers has averaged 16 times higher since the formation of the maria than before (*7*). Weertman (*16*) has found that the viscosity of the moon appears to decrease with increasing depth.

Isostatic changes in crater form permit distinction of a sequence of events throughout geologic history. The rate of infalls declined from nearly 300 on a relative scale in the beginning, to about 10 at the time Oceanus Procellarum was formed, to about 0.8 now (*7, 14, 15, 17*).

Without exception, the giant impacts that produced the circular mare basins took place in solid rock (*13*, p. 305; *18*). The shape of the instantaneous impact crater is essentially independent of its size. The Mare Imbrium basin was about 50 km deep relative to the undisturbed ground level (*10, 19*). It quickly began to distort under isostatic stresses, but a rather long period of time passed before the lavas came. The entire shelf area within the Apennines-Carpathians-Alps border was rejuvenated. All older craters in this area were destroyed. This rebirth of the surface also has been nicely shown at Mare Orientale, where the shelf area has not been lava filled. While the Mare Imbrium crater and shelf were dry, the large craters listed in Table 1 were formed (*13*, p. 307; *18*).

Table 1. Craters formed within the Imbrium structure before the lavas arose.

Crater	Diameter (km)
Sinus Iridum	266
Plato	101
Archimedes	80
Cassini	57
Sinus Gay-Lussac	35
North of Aristillus	29
Wallace	23
Archimedes M	11

Similar listings may be made for each of the giant basins. They are clear evidence that a long period of time occurred between the impacts that formed the giant basins and the coming of the lavas. The upwelling of the lavas was not triggered by the impact. The giant impacts rejuvenated the surface within the basins and nearby. Any preexisting craters would have been destroyed. The listed craters are all lava-filled. There is no evidence on the lava surface near the listed craters of the explosive violence shown by postmare Aristillus. The lavas came considerably later and partially drowned the craters formed subsequent to and within the giant craters. Crater counts suggest a dry period of some 1 to 5 $\times$ 10^8 years. Since the lavas were not impact melts, since the maria came much later than the giant impacts, and since the impacts occurred in rock at least 50 km thick, the lavas of the maria came from deep within the body of the moon.

Crater counts on the postmare craters, combined with the Apollo 11 and Apollo 12 ages, tell us that the lavas came in the period of roughly 3.7×10^9 (*20*) to perhaps 3.3×10^9 years ago (*21*). However, smaller flows can be seen on the surface of the older flows, which indicates that the lavas did not all come at once but came as a multitude of separate flows. Some of these later flows show so few craters on them that they may be quite recent. An interesting example of three sequential flows is near Triesnecker (*22*) and a relatively young flow is in the Palus Putredinis region of Mare Imbrium.

When rock melts, it generally expands about 10 percent. The density of 3.0 g/cm^3 for Apollo 11 basalt magma at 1150°C (*23*) permits it to rise hydrostatically from source to surface regions. On cooling, it contracts first to gabbro (3.36 g/cm^3) (*24*). This expansion is ample to split superimposed layers and to force hot liquid magmas upward until they debouch onto the lowest portions of the moon's surface. The lavas rose through myriads of cracks and, when these cracks intersected the deepest parts of the surface, the lavas appeared. They are observed to lie in low regions—that is, in circular maria and crater bottoms; wherever they developed and solidified, their superimposed weight caused the surface to adjust isostatically (*13*, p. 232). The typical lava surface is depressed about 1 to 2 km or more below normal ground level.

Measurements of surface and subsurface temperatures show that a solar cycle is superimposed on a moonwide temperature of about −180°C, not far below the surface (*25*). Such a very low temperature implies a slow escape of heat from the interior. Calculations (*8, 9*) suggest a temperature that increases with depth and reaches about 1000°C at perhaps 200-km depth. Many similar calculations have been made, which have taken into account both radiation and conduction, and all reach the same general conclusions. Consequently, any recent lavas have come from below this depth, and it is probable that the maria themselves came from magmas generated at similar depths. They certainly did not come from shallow depths.

Conversely, all calculations indicate that, once the moon became melted or nearly melted deep beneath the surface, there was no way to get rid of the heat fast enough to allow this deep layer to cool significantly. The moon is now and has always been hot, and the temperature is close to the melting point of the deep-seated materials, as is indicated by the long sequence of basic and ultrabasic lava exudations. Ultrabasic rocks generally are considered to have the SiO_2 abundance of less than 45 percent by weight and also to have a relatively high abundance of MgO. Rose *et al.* (*26*) report nine Apollo 11 rocks with SiO_2 in the range of 38 to 42 percent and with MgO in the range of 6 to 8 percent. The indicated temperatures are in the range of 1000° to 1500°C.

This is a brief summation of the physical observations and arguments that have led me (*13, 18, 27*) to the picture of a presently hot moon. The chemical observations are more subtle.

In April 1970, Urey (*12*) attempted

to account for the presence of the Mare Tranquillitatis type of material and for the anorthositic layer on the moon by postulating a cold moon of chondritic composition which underwent a surface, or near-surface, melting. One method of producing such melting was suggested to be a superbright phase of the primitive sun. Age determinations (*20, 21*) now show that the uplands are roughly a billion years older than the maria. A hot flash from the sun probably did occur as it approached the main sequence, but it came so early that it could not be a cause of both periods of melting. If the earlier melting was sun-induced, it would be necessary to explain how such a heat pulse could penetrate roughly 100 km into the moon to permit the anorthositic layer to be derived from the more basic main body of the moon. It would be necessary to melt a large fraction of the moon to permit a chemical differentiation large enough to permit the separation of the anorthositic layer. In any event, Urey's hypothesis is not sustained. Alternatively, he has attempted to explain the lava as a surface melting within a very hot hydrogen atmosphere (*28*) or as impact melting (*29*). The arguments against the latter have been presented, and the effect of a hot hydrogen atmosphere would be to reduce the surface FeO to metallic Fe, contrary to observations.

Conversely, a melt of the composition of the maria cannot develop from the less basic material of the uplands. The uplands are older than the maria and covered the original surface. Therefore the lavas of the maria came from below.

With regard to the origin of the mare lavas, Ringwood (*8*) points out that it is not valid to assume a "chondritic mix" for the moon. It is now certain that the moon is strongly depleted in a wide range of volatile elements including the alkali metals, as compared with chondrites. It is also depleted in iron and probably has a different ratio, $(Ca + Al)/(Mg + Si)$, than do chondrites. This depletion must exist throughout the moon, since the lavas now on the surface came from deep within the moon, and depletions of this type are not caused by partial melting (*24*).

The ranges of pressure and temperature conditions within the moon are readily accessible under controlled laboratory conditions. Thus we have a fairly good idea of what the possible phase equilibria are and what chemical processes are involved in making lunar basalt.

Ringwood and Essene (*24*) showed that Apollo 11 basalt transforms to eclogite at an unusually low pressure (12 kb at 1100°C). The deeper regions of some maria and related intrusive feeder structures beneath the maria may be composed of eclogite. However, the high density of the eclogite modification precludes the Apollo 11 basalt from being a major component of the moon. Therefore, the Apollo 11 basalt must be a partial melt product of the main mass of the moon. Composition of near liquidus phases was determined by means of an electroprobe microanalyzer. The data found by Ringwood and Essene, together with the mean density and moments of inertia of the moon, provide strong constraints upon the mineralogy and composition of the source regions from which the basalts were derived. They concluded that the moon was largely composed of orthopyroxene and subcalcic clinopyroxene, together with some olivine. Source material of this composition would be capable of generating the major and trace element composition of Apollo 11 basalt by a minor amount of partial melting at depths of 200 to 400 km.

Ringwood (*30*) estimated the percentage of partial melting to be from 1 to 2 percent for Apollo 11 basalts and perhaps 3 to 5 percent for Apollo 12 basalts. With these data, it can be demonstrated that very large volumes of the moon had to become heated in order for the lunar basalts to be driven up to the surface.

We can calculate an approximate volume of the lavas of the maria. The density of the uplands surface layers seems to be about 2.8 g/cm^3 and the mare material about 3.3 g/cm^3. We have not measured the average depth. We do know that in southern Oceanus Procellarum and Mare Nubium (*31*), the lavas are shallow and approach 2 km in depth. In much of Oceanus Procellarum, the lavas are much deeper, as few crater rims project. In the central basins of the great circular maria, the lava depths are considerably greater. An average depth may be approximated as follows for the maria. Isostatic equilibrium is here assumed, and the Orbiter results of Muller and Sjogren (*5*) indicate that, over most of the maria, this balance has been achieved.

If we arbitrarily assume the average depth to be 3 km, we may calculate

$$\frac{3.3}{2.8} = 1.1785$$

from which we derive

$$0.1785 \times 3 \text{ km} = 0.5355 \quad \text{km}$$

which equals the depression of the maria below the mean sphere, on the assumption of an average lava thickness of 3 km. The actual average depression appears to be about 2 km deep, and, inasmuch as the maria cover about one-third of the front face, then 1⅓ km is depression of the maria and ⅔ km is the elevation of the rest of the surface. Therefore,

$$1.33 \div 0.5355 = 2.48$$

Therefore 3-km thickness times 2.48 equals 7.44 km equals the average thickness of the maria. This figure, rounded off to 7 km, seems reasonable when we consider that the circular mare basins are probably 20 to 50 km deep.

Now the area of the visible half of the moon's face is almost 1.9×10^7 km^2, and the area of the maria is about one-third of this, or 6.3×10^6 km^2. The volume of the maria is thus 4.4×10^7 km^3. The volume of the entire moon is 2.2×10^{10} km^3, and therefore the portion of the melt that extended up over the surface was equal to one part in 500 of the volume of the moon. Now the lava of the maria is the result of a partial melt of the material of the body of the moon. Moonwide average figures may be from 2 to 4 percent partial melting. Consequently, the volume of the moon which became hot enough to drive off the maria lavas was equal to 10 to 5 percent of the entire volume of the moon, if we assume that all melted material actually reached the surface. This assumption is highly improbable, and therefore an even higher percentage of the moon was heated essentially to the melting point of orthopyroxene.

From both chemical and physical reasoning, it has been concluded that the melting occurred at a depth in the moon of from 200 to perhaps 400 km. Only 69 percent of the moon is inside a radius of 1538 km, and only 45 percent is inside 1338 km. If we combine this fact with the probability that a significant part of the material melted never reached the surface, we must conclude that we now have evidence that at the second melting a minimum volume in the range of 20 to 40 percent of the portion of the moon below 200 to 400 km became hot enough to melt the mare basalt. This is so large a fraction of the moon that it is highly probable that everything below the deep outer layers became extremely hot very early in the moon's history and again when the maria were formed. Since the moon was hot when the basalts rose, it is still hot. No calculations I have seen suggest that a hot moon could cool significantly at depths of a few hundred kilometers within geologic time. Even convection could only help to cool the inner moon to the point where it could no longer convect.

The thesis that the lavas of the maria came from surface heating no longer should be considered valid. The deep moon must be hot, although not necessarily molten. The surface is cool, and we may infer a transition zone in the outer few hundred kilometers.

Ralph B. Baldwin
Oliver Machinery Company, Grand Rapids, Michigan 49502

References

1. A. L. Turkevich, E. J. Franzgrote, J. H. Patterson, *Science* **162**, 117 (1968).
2. ——, *ibid.* **165**, 277 (1969).
3. Lunar Sample Preliminary Examination Team, *ibid.*, p. 1211.
4. ——, *ibid.* **167**, 1325 (1970).

P. M. Muller and W. L. Sjogren, *ibid.* **161**, 680 (1968).

A. L. Albee *et al.*, *ibid.* **167**, 463 (1970).

7. R. B. Baldwin, *Icarus*, in press.
8. A. E. Ringwood, *Earth Planet. Sci. Lett.* **8**, 131 (1970).
9. G. J. F. MacDonald, *Science* **133**, 1045 (1961).

R. B. Baldwin, *ibid.* **162**, 1407 (1968).

J. A. O'Keefe, *ibid.*, p. 14; H. C. Urey, *ibid.*, p. 1408; D. U. Wise and M. T. Yates, *J. Geophys. Res.* **75**, 261 (1970).

12. H. C. Urey, address presented before the NATO Advanced Study Institute, Newcastle upon Tyne, England, April 1970.
13. R. B. Baldwin, *The Measure of the Moon* (Univ. of Chicago Press, Chicago, 1963).
14. W. K. Hartmann, *Icarus* **12**, 131 (1970).
15. R. B. Baldwin, *ibid.* **11**, 320 (1969).
16. J. Weertman, *Rev. Geophys. Space Phys.* **8**, 145 (1970).
17. R. B. Baldwin, *Icarus*, in press.
18. ——, *Bull. Va. Polytech. Inst. Part A* **56**, No. 7 (1962).
19. R. J. Pike, thesis, University of Michigan, Ann Arbor (1968).
20. Lunar Sample Analysis Planning Team, *Science* **167**, 449 (1970).
21. D. Papanastassiou and G. Wasserburg, *Earth Planet. Sci. Lett.* **8**, 269 (1970).
22. R. B. Baldwin, *Astron. J.* **69**, 377 (1964).
23. D. F. Weill, I. S. McCallum, Y. Bottinga, M. J. Drake, G. A. McKay, *Science* **167**, 635 (1970).
24. A. E. Ringwood and E. Essene, *Geochim. Cosmochim. Acta*, in press.
25. R. W. Shorthill, *Boeing Sci. Res. Lab. Rep. D1-82-0196* (1962).
26. H. J. Rose, Jr., *et al.*, *Science* **167**, 520 (1970).
27. R. B. Baldwin, *Pop. Astron.* **50**, No. 7, 1 (1942); *ibid.* **51**, No. 3, 1 (1943); *The Face of the Moon* (Univ. of Chicago Press, Chicago, 1949).
28. H. C. Urey and G. J. F. MacDonald, *Sci. J. Univ. Calif.* **5**, No. 5, 60 (1969).
29. E. J. Öpik, *Ir. Astron. J.* **8**, 38 (1967); H. C. Urey, *The Planets* (Yale Univ. Press, New Haven, 1952).
30. A. E. Ringwood, personal communications.
31. R. B. Baldwin, *Publ. Astron. Soc. Pac.* **82**, 857 (1970).

24 August 1970

Soft Parts of Cephalopods and Trilobites: Some Surprising Results of X-ray Examinations of Devonian Slates

Abstract. *X-ray studies of slates from Bundenbach and Wissenbach (Lower and Middle Devonian, West Germany) revealed a surprising amount of pyritized soft parts of cephalopods and trilobites. The tentacles of cephalopods, the appendages, the intestinal tract, and the structure of the interior of facet eyes of trilobites* (Phacops *species and* Asteropyge *species) were found in a well-preserved state.*

The discovery of soft parts of Paleozoic fossils is a very rare event (*1*). During an extended x-ray investigation of Devonian fossils from the famous localities of Bundenbach and Wissenbach (Lower and Middle Devonian, West Germany) many unprepared slates were found in which soft parts and extremely fine structures of the embedded fossils are preserved. These structures, consisting of a very fine-grained pyrite, are so delicate that they are normally destroyed by the usual mechanical preparation of the specimens. Examination with soft x-rays (25 to 40 kv) showed that this pyritized material gives good contrast because of the high absorption coefficient of pyrite. From these observations it was concluded that the preservation as pyrite depends largely upon the locality. It is not known what initiated the formation of pyrite in those parts of the buried animal which contained large amounts of albumin and other constituents with high sulfur content. The products of decomposition were adsorbed on the enclosing fine sediment and held in this

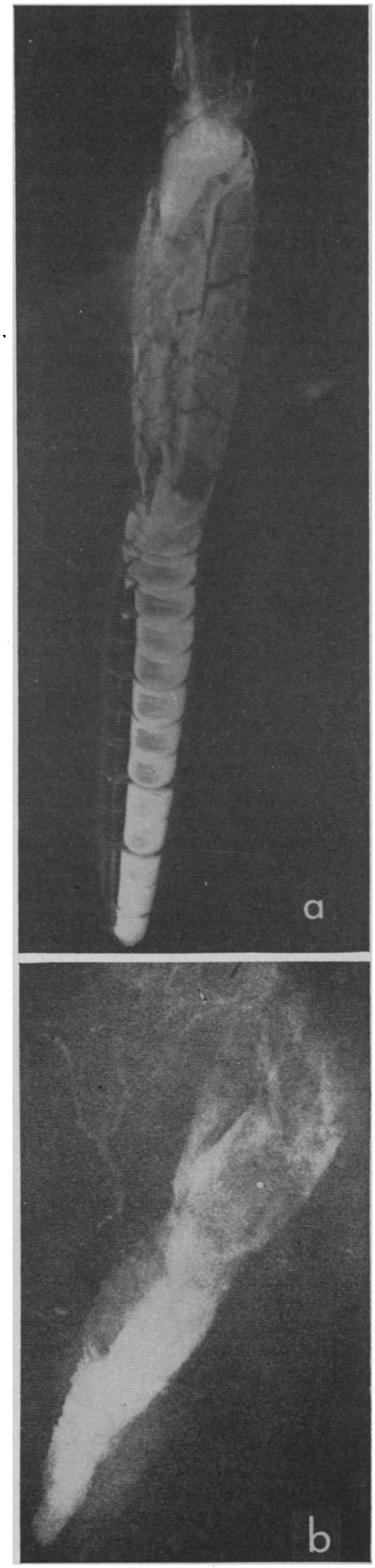

Fig. 1. (a) *Lobobactrites* sp., a cephalopod with tentacles. Right of the center, a very young cephalopod. (b) Very young embryonic stage of a cephalopod (*Orthoceras*).