lieve to be due to a continuously high sedimentation rate.

Carbon-14 ages for the base of the upper carbonate unit are about 3,000 years B.P., and for the organic unit about 7,000 years B.P. The sedimentation rate for these two units within the "normal" sequence is thus about 10 cm per 1,000 years. Sedimentation rates for other cores, where the sequences are extended or interrupted, can be established if the base of one of the units can be identified. This has been done for the upper carbonate unit, and using this method, we have determined the sedimentation rates for the most recent sediments of the Black Sea (Fig. 4).

Radiocarbon analyses of the total dissolved carbonate present in the oxygen-containing surface waters of the Black Sea (0 to 100 m) show essentially recent ages, and an isotopic abundance δ for carbon-14 of -20 ± 10 has been calculated for prebomb times (*6*). It can thus be safely concluded that the organic matter produced in such waters reflects this zero age. We thus assume that the age of the organic matter at the sediment-water interface is essentially zero. We also assume that our cores sampled the entire carbonate unit (*7*).

Sedimentation rates greater than 30 cm per 1,000 years occur in the eastern part of the basin and along the Turkish coast. Turbidites are common in cores from the eastern part of the basin. Cores collected along the Turkish coast are similar to those shown in core c (Fig. 3), probably reflecting the high sedimentation rate due to the many rivers along this mountainous coast.

Sedimentation rates are relatively low in the western part of the Black Sea. Sediments from the Danube, Dniester, and Dnieper rivers are apparently trapped in the estuaries and on the broad continental shelf of that area and presently do not reach the deeper parts of the basin. Seismic reflection profiles in this area indicate that these rivers were an important source of sediment in the past and that the rivers had built a fan almost across the basin (*8*).

In the results reported here we have touched only on the general aspects of the recent sediments. As more detailed studies proceed, widespread events of shorter duration will certainly become evident. For example, several thin white layers are present in the organic-rich unit; the uppermost of these layers is composed of coccolithophorids; the lowermost band is aragonite of a so far unidentified origin. Clearly, the Black Sea sediments offer rare possibilities for the establishment of detailed correlations and the interpretation of changes in a recent depositional environment.

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Thermal Conductivity of Lunar and Terrestrial Igneous Rocks in Their Melting Range

Abstract. *The thermal conductivity of a synthetic lunar rock in its melting range is about half that of a terrestrial basalt. The low conductivity and increased efficiency of insulating crusts on lunar lavas will enable flows to cover great distances without being quenched by high radiant heat losses from the surface. For a given rate of heat production, the thermal gradient of the moon would be significantly steeper than that of the earth.*

Any attempt to interpret the origin and behavior of lunar magmatic rocks requires an evaluation of their thermal properties at high temperatures. Previous measurements of the thermal diffusivity and conductivity (*1*) of Apollo 11 specimens have been made at temperatures below 150°C. We have obtained data up to 1500°C for a synthetic sample of lunar composition and have made comparable measurements on a variety of terrestrial volcanic rocks.

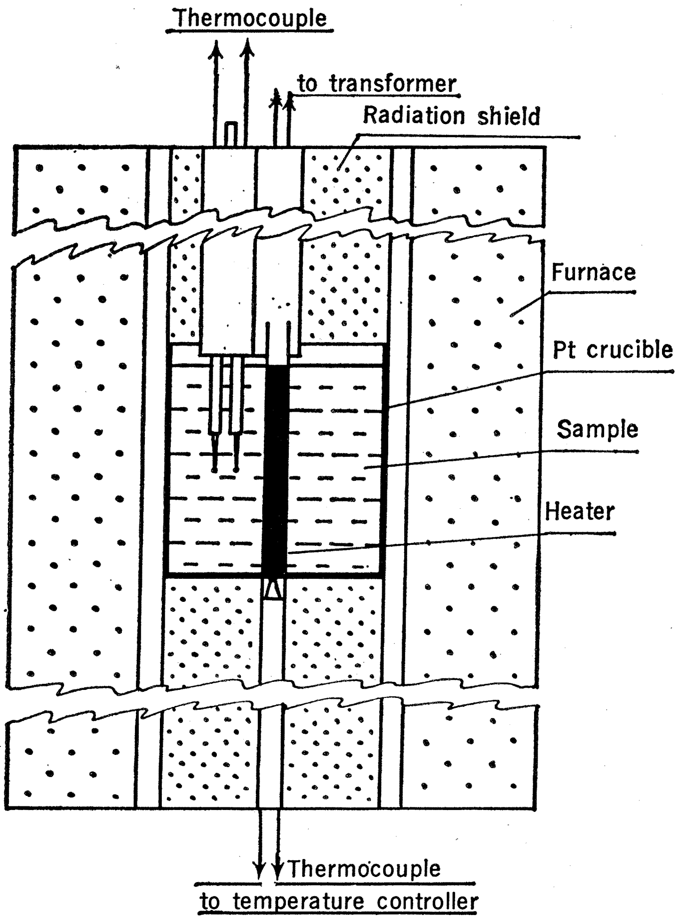
The sample of lunar composition used in our measurements was prepared according to the reported major-element analysis of Apollo 11 specimen 22. The sample was prepared by the method described in our earlier report on the viscosity of the same sample (*2*). By adding small amounts of sample during the melting it was possible to eliminate bubbles. Samples of basalt and andesite were found to have negligible porosity at all temperatures. The synthetic lunar sample contained no bubbles when quenched from 1000°C but developed about 3 percent interstitial pore space on crystallizing. The porosity of rhyolite was about 3 percent, and the measured thermal conductivity values were corrected to zero porosity (*3*).

We used a radial heat-flow technique (Fig. 1) in which conductivity k is calculated from the relation (*3*):

$$k = \frac{IV}{2\pi LJ(T_1 - T_2)} \ln \frac{R_2}{R_1}$$

where I and V are the electrical current and potential through a platinum resistance heater of length L (5 cm), J is the mechanical equivalent of heat, R_1 and R_2 are the radial distances of the inner and outer thermocouples, and T_1 and T_2 are the temperatures of the inner and outer thermocouples. The sample was placed in a cylindrical platinum crucible 5 cm in diameter and 5.5 cm high with the heater positioned at the axis of the crucible and with the two thermocouples at distances of about 0.7 and 1.1 cm from the heater. A temperature difference of less than 10°C was produced by 10 watts of power supplied to the heater.

Initial measurements were made at 1500°C, and measurements were repeated at successively lower temperatures, several hours apart, so as to allow the system to reach a steady state. A second series of measurements was made as the temperature was raised, and close agreement was obtained between heating and cooling cycles. A



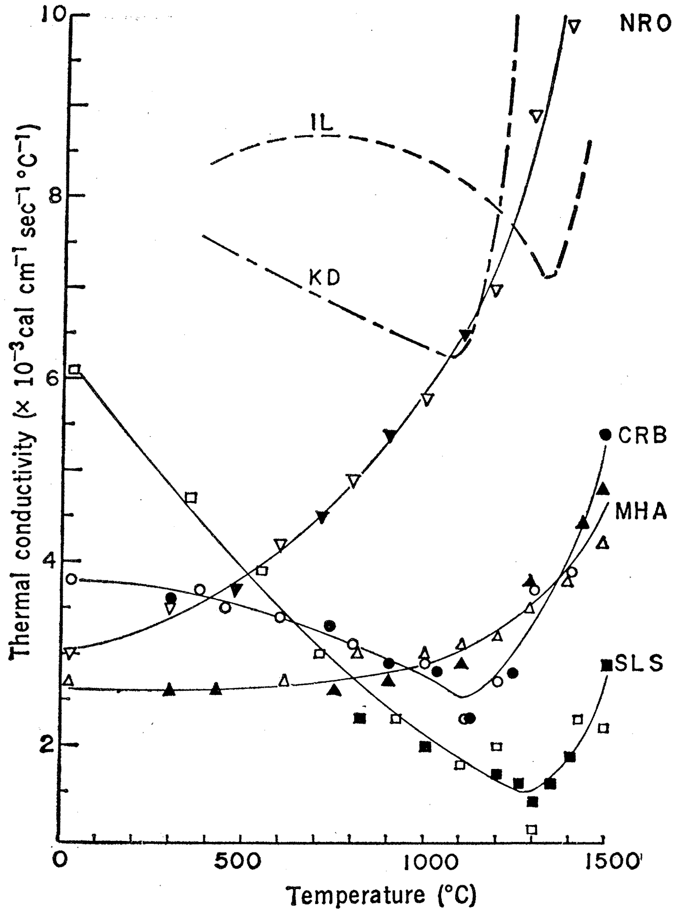


Fig. 1 (left). Schematic diagram of apparatus used in measuring the thermal conductivity of rocks at high temperatures. Fig. 2 (right). Measured thermal conductivity of rocks as a function of temperature. *SLS*, a synthetic lunar sample with the composition of Apollo 11 sample 22; *NRO*, a rhyolite obsidian from Newberry caldera; *CRB*, a Columbia River basalt; and *MHA*, an andesite from Mount Hood; *IL* and *KD* are samples of lherzolite and dunite studied by Kawada (*7*). (Solid symbols) Values measured at successively higher temperatures during the heating cycles; (open symbols) corresponding measurements made at stages of the cooling cycle.

plot of temperature drop between the thermocouples as a function of heating rate is linear and passes through the origin, an indication that convection could not have had an important effect (*4*). The accuracy of the system was established through a comparison of values obtained for a U.S. National Bureau of Standards artificial glass with those obtained by Kingery (*5*) for silicate glasses of similar composition.

Thermal conductivity data are shown in Fig. 2. At 1400°C all samples except those of ultramafic rocks were entirely molten. The sample of lunar composition began to crystallize between 1300° and 1380°C and was probably completely crystalline at 1100°C; the basalt crystallized in the interval between 1275° and about 1050°C. The low viscosity of these two rocks enabled us to approach equilibrium between the crystal and liquid phases in the 2- or 3-hour interval between measurements, but nucleation and growth of crystals in the andesite and rhyolite are so slow that an excessive amount of glass was present at temperatures below the liquidus. The andesite began to crystallize at 1225°C but still contained glass below 1000°C, and the rhyolite was essentially all glass throughout the entire range of the measurements. The effect of the glass would be to give the sample a lower conductivity than it would have under conditions of crystal-liquid equilibrium.

The outstanding feature of the data in Fig. 2 is the very low conductivity of molten volcanic rocks of basic composition, especially the sample of lunar composition. In its melting range (between 1100° and 1380°C) the lunar rock has a conductivity of about 1.5×10^{-3} cal cm^{-1} sec^{-1} °C^{-1}, nearly half the value for a molten terrestrial basalt. At lower temperatures there is a strong negative temperature dependence of conductivity for the lunar rock. At temperatures above the liquidus, radiative heat transfer increases markedly in all of the samples, but the transition from a negative to a positive temperature dependence becomes progressively less pronounced as the rocks become more siliceous. This latter relation is due, at least in part, to the large glass content of the siliceous rocks at low temperatures.

Lavas with the composition of the Apollo 11 samples must have a high fluidity (*2*), but the fact that their liquidus temperature is very high may cause them to be quenched and to cease to flow if the radiant heat losses from the surface were to result in cooling a large vertical portion of the flow. The remarkably low thermal conductivity we observed indicates that a hot crust formed on the surface would be an efficient insulator and would conserve internal heat, so that the fluid lava would continue to flow beneath the crust. For these reasons lunar lavas would be likely to flow beneath a thin crust, and individual flows could cover extensive areas of the moon's surface.

A further consequence of the low

thermal conductivity of the lunar rock is the implication it has for interpretations of the thermal regime of the moon's interior. If the Apollo 11 specimens are in any way representative of specimens at deeper levels or even if they contributed a significant fraction of the moon's original composition, it is clear that the thermal gradient resulting from a given rate of heat production on the moon could be substantially steeper than that on the earth. It seems improbable that the interior of the moon could be composed entirely of material as rich in iron and as opaque as the Apollo 11 samples (*6*), and the actual composition and thermal conductivity were probably intermediate between those of the lunar surface rocks and terrestrial ultramafic rocks, such as the dunite and lherzolite studied by Kawada (*7*). From the relations shown in Fig. 2 one can estimate the thermal conductivity of such an intermediate composition if the proportions of the component compositions are known. Extraction of the material with an iron-rich composition through partial melting and eruption at the surface must have produced a marked increase in the thermal conductivity in the levels from which the lunar lavas were derived.

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Mars: Is Nitrogen Present?

Abstract. *If the atmosphere is uniformly mixed, a mixing ratio of nitrogen to carbon dioxide of 5 percent is consistent with the observational data on the ultraviolet dayglow of Mars. If the magnitude of the eddy coefficient in the atmosphere is similar to that for the earth, this limit is reduced to less than 0.5 percent.*

The fractional abundance of nitrogen relative to carbon in the martian atmosphere is an important parameter in studies of the evolution of the planet. An extension of Rubey's (*1*) classic discussion of volatile gases in the atmosphere-ocean system of the earth suggests a mixing ratio of N_2 to CO_2 of a few percent if the atmospheres of Mars and the earth have a common origin and the amount of CO_2 in the martian polar cap is not large relative to that in the atmosphere. The difference in composition of the planetary atmospheres is due largely to the presence of oceans on the earth in which carbon is effectively precipitated as $CaCO_3$. If the N_2 mixing ratio in the martian atmosphere were much less than a few percent, a serious revision of current thinking on atmospheric evolution would be necessary.

The failure of the ultraviolet experiment on Mariners 6 and 7 (*2*) to detect emissions associated with nitrogen has already provoked such speculation. We discuss here the implications of the Mariner results by examining the detailed physical and chemical processes that lead to nitrogen emission in the martian airglow.

The strongest emissions from N_2 below 4000 Å are excited primarily by electron impact and have been extensively studied in the earth's atmosphere (*3, 4*). Electrons are produced by photo-ionization of atmospheric gases at high altitudes in the atmosphere. Their mean energy upon production is approximately 10 ev (*5*) and they are degraded in energy by collision processes, both elastic and inelastic. Some of these collision processes correspond to excitation of optical emissions from nitrogen, which is our primary concern. To compute the resulting emission spectrum one must first determine the energy distribution function for photoelectrons. The convolution of this quantity with an appropriate electron excitation cross section gives the excitation rate of optical emission per molecule.

We adopt models for the neutral atmosphere composed of CO_2 (*6*) with a small uniform admixture, *x* percent, of N_2 and a solar radiative flux appropriate to the Mariner 6 flyby. The rate of energy degradation for electrons in CO_2 has been estimated by Henry and McElroy (*7*) and by Strickland and Green (*8*); similar data for N_2 are given by Green and Barth (*9*) and by Dalgarno *et al.* (*5*). The distribution function derived for electrons with energy greater than 5 ev is shown in Fig. 1 for a height of 187 km in the martian atmosphere. For comparison we include a similarly computed distribution function for electrons at a height of 200 km in the earth's atmosphere. The interesting feature of these results is the sharp break in the martian distribution function near 7 ev, the threshold for excitation of electronic transitions in CO_2.

The threshold energy for excitation of the lowest electronically excited level of N_2, $A^3\Sigma_u^+$, is 6.5 ev. The $A^3\Sigma_u^+$ level is the upper level of the Vegard-Kaplan band system which appears as a weak emission feature in the dayglow of the earth. This feature is weak for the earth partly because the metastable $A^3\Sigma_u^+$ level is efficiently quenched by collisions with atomic (*10*) and molecular oxygen (*11*), constituents that are abundant in the earth's atmosphere but relatively rare in the martian upper atmosphere (*12*). Quenching by CO_2 is relatively inefficient (*11*), and most of the molecules in the $A^3\Sigma_u^+$ level produced above an altitude of about 100 km on Mars will radiate. If contributions due to cascade from higher levels are included, we predict an overhead intensity of 800*x* rayleighs for Vegard-Kaplan band emission in the martian atmosphere.

Unfortunately from the point of view of detection, this emission is distributed over many bands. The intensity in any given band is not expected to exceed 10*x* rayleighs. The stronger emissions, such as the (7-0) and (6-0) bands, occur near 1700 Å in a spectral region containing strong bands of CO produced by dissociation of CO_2 and by dissociative recombination of CO_2^+. We suggest that an intensity of 50 rayleighs, corresponding to an assumed N_2 mixing ratio of 5 percent at the emitting altitude, might have escaped detection in the preliminary study of Barth *et al.* (*2*) despite the high quality of their spectra.

The second most intense emission band of N_2, according to our calculations, is the first positive system associated with the transition $B^3\Pi_g \rightarrow A^3\Sigma_u^+$. We estimate that the overhead intensity due to electron impact would be