Oxygen isotopic constraints on the composition of the moon

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Abstract—The mean oxygen isotopic composition of 5 Apollo 17 soils, one Apollo 17 breccia and one Apollo 12 soil is $\delta O^{18} = 5.63 \pm 0.05$ and $\delta O^{17} = 3.8 \pm 0.2‰$. These values are within several tenths of a part permil of the composition of a large fraction of the lunar interior. High-temperature condensate aggregates from Allende and other C2 and C3 chondrites are vastly enriched in O^{16} compared to this composition. The moon cannot be a mixture of ordinary chondrites and Allende inclusions, nor can it be derived from such a mixture by chemical fractionation processes. The moon's isotopic composition *is* consistent with a mixture of high- and low-temperature condensates but the refractory fraction would have to be free of the O^{16}-rich component so prevalent in the meteoritic aggregates, a fact which makes such models less attractive than they once seemed.

Introduction

THE HIGH CONCENTRATIONS of refractory elements and low abundances of volatiles in the lunar rocks have prompted some investigators to propose that part or all of the moon accreted from materials enriched, relative to the solar system abundances, in high-temperature condensates such as the Ca–Al-rich inclusions found in the Allende meteorite. See, for example, Gast *et al.* (1970), Anderson (1972, 1973), Krähenbühl *et al.* (1973), Wänke *et al.* (1973, 1974), Ganapathy and Anders (1974). These models have been criticized by Clayton *et al.* (1973b) and Grossman and Larimer (1974) because the oxygen isotopic compositions of lunar rocks differ enormously from those of Allende inclusions. The purpose of this paper is to compare the δO^{18} and δO^{17} of lunar, meteoritic, and terrestrial materials and to discuss the origin of the moon in the light of these data.

Analytical Methods

Analytical procedures for lunar samples were the same as those used in this laboratory for previous Apollo missions (Onuma *et al.*, 1970). O^{17}/O^{16} variations were measured as outlined by Clayton *et al.* (1973a). Oxygen isotopic compositions are reported in the δ-terminology, as permil (‰) deviations from the SMOW standard. Uncertainties in δO^{18} and δO^{17} are estimated to be ± 0.1 and ± 1 permil, respectively.

RESULTS

Table 1 shows the oxygen isotopic compositions of five Apollo 17 soils, one Apollo 17 breccia, and one Apollo 12 soil. The mean of these analyses is $\delta O^{18} = 5.63 \pm 0.05$ and $\delta O^{17} = 3.8 \pm 0.2$.

Table 1. Oxygen isotopic compositions of lunar samples.

Sample No.	Description	δO^{18} (SMOW)	δO^{17} (SMOW)
70019,10	Breccia	5.53	3.7
73221,8	Soil	5.64	3.4
73281,7	Soil	5.66	4.3
76501,14	Soil	5.68	4.1
79221,25	Soil	5.61	3.3
79261,17	Soil	5.64	3.7
12033,55	Soil	5.62	4.0

DISCUSSION

Oxygen isotopic composition of the moon. The oxygen isotopic compositions of the common rock-forming minerals in lunar igneous rocks vary little from one sampled locality to another (Onuma *et al.*, 1970; Taylor and Epstein, 1970, 1973; Clayton *et al.*, 1973b). The δO^{18} of pyroxene is always between 5.3 and 5.8‰. Olivines are between 4.9 and 5.1‰ and plagioclases between 5.6 and 6.4‰. These minerals come from igneous rocks which are crystallization products of partial melts of source regions in the lunar interior. The small range of δO^{18} implies that the source regions of the rocks from all the sites are very similar in oxygen isotopic composition. Because isotope fractionations are very small at basalt melting temperatures, the δO^{18} of a mineral in a surface igneous rock is within several tenths of a part permil of the δO^{18} of the same mineral in the source region. Assuming that the source regions are composed largely of a mixture of olivine, pyroxene, and plagioclase, a large fraction of the lunar interior has a δO^{18} of $\sim 5.5 \pm 0.2‰$. Furthermore, the presence of large quantities of melilite (Anderson, 1973) in the lunar mantle would not alter this conclusion since unpublished data on terrestrial basalts show that coexisting melilite and pyroxene have the same isotopic composition at these temperatures.

A chemical isotope effect is known to be associated with the surfaces of lunar soil particles (Epstein and Taylor, 1971). This causes bulk soil samples to be enriched in δO^{18} by 0.3 to 0.5‰ (Taylor and Epstein, 1973) and in δO^{17} by 0.15 to 0.25‰ relative to igneous rocks. When the compositions of the soils in Table 1 are corrected for this effect, their δO^{18} remains very close to that of the lunar interior inferred above. The δO^{17} values for the soils must also be within a few tenths of a part permil of the δO^{17} of the mantle because the isotopic compositions of the igneous rocks are related to those of the source regions through chemical effects.
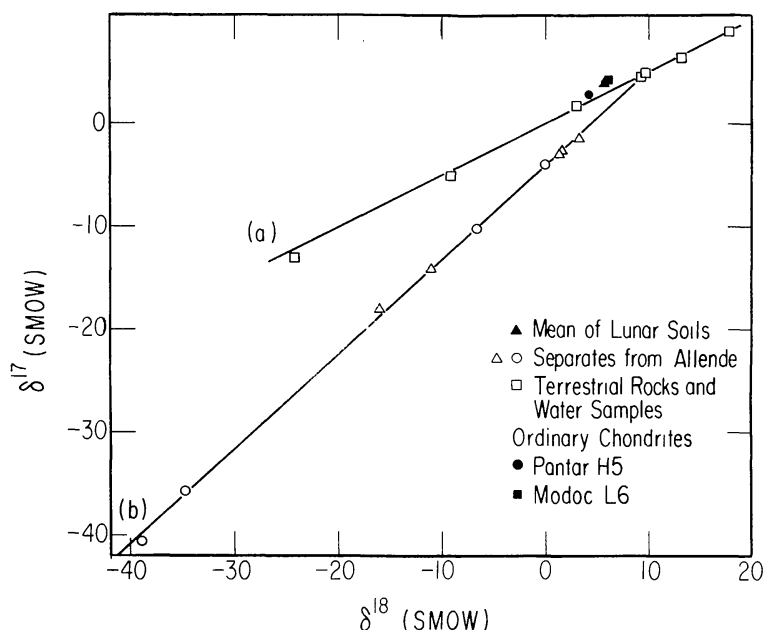
Fig. 1. Oxygen isotopic compositions of lunar samples, terrestrial rock and water samples, ordinary chondrites and separated inclusions from the Allende meteorite. Note that the lunar composition does not lie on any of the mixing lines which can be drawn between Allende inclusions and H-group chondrites.

Oxygen isotopic compositions of solar system materials. Figure 1 is a plot of δO^{17} versus δO^{18} for a variety of samples from different bodies in the solar system. The data for terrestrial rocks and water samples plot along line a which has a slope of $+\frac{1}{2}$. This is expected for all samples derived by chemical processes from the same batch of nucleosynthetic material. The mean composition of the lunar samples in Table 1 is seen to lie close to line a, but slightly above it. The refractory Ca–Al-rich aggregates in Allende, however, fall along line b which has a slope of $+1$. These objects are vastly lower in δO^{17} and δO^{18} than the moon. In particular, the lunar point does not fall on any of the family of mixing lines which can be drawn between the H-group ordinary chondrites and any of the Allende inclusions. The L-group chondrite data point is indistinguishable from the lunar point. The moon cannot be a mixture of Allende inclusions and ordinary chondrites, a model suggested by Wänke *et al.* (1973). Nor can the moon be derived by chemical fractionation processes from such a mixture since lines joining the lunar point to points on the mixing lines do not have slopes of $+\frac{1}{2}$.

There is a suggestion in our data that the lunar point is displaced slightly from the terrestrial chemical line a. This may indicate that the moon is depleted in the O^{16} component by about 2‰ relative to the earth. Although this is an extremely tenuous conclusion, it may have very important cosmochemical consequences if supported by future measurements.

Oxygen isotopic constraints on the origin of the moon. The fact that line b has a slope of unity indicates that nuclear processes, rather than chemical ones, dominated the establishment of the isotopic compositions of the Allende inclu-

sions. Thus, these compositions are *not* the natural consequence of chemical isotope effects accompanying the condensation of high-temperature minerals from the solar nebula. Clayton *et al.* (1973a) suggested that line *b* is a mixing line between "ordinary" solar system oxygen, having an isotopic composition at the intersection of lines *a* and *b*, and pure, or nearly pure, O^{16}. They proposed that the O^{16} entered the solar nebula bound in preexisting interstellar grains which escaped vaporization and isotopic homogenization during the early high-temperature stage of the nebula. These grains were then incorporated by the minerals of the Allende inclusions as they condensed from the cooling nebula. Alternatively, O^{18} and O^{17} may have been consumed in nuclear reactions during an early stage of the sun, with the resulting O^{16}-enriched gas ejected into the nebula where it was implanted into the surfaces of condensate grains. Whatever the origin of the O^{16}-rich component, it is clearly a contaminant in the sense that its presence in the inclusions masks the isotopic compositions which they would have inherited purely from the chemical isotope effects of condensation.

By projecting their compositions alone line *b* onto line *a*, one can correct the oxygen isotopic compositions of the inclusions to the same O^{16} content as the moon. The average inclusion would then plot at the intersection of the two lines, $\delta O^{18} = 9.6\%_0$.

If this inclusion is assumed to have condensed at 1400°K (Grossman, 1972), then the mean equilibration temperature of the materials which accreted to form the moon can be inferred from the difference in the δO^{18} between the intersection and the lunar point. Using the principles behind the cosmothermometer of Onuma *et al.* (1972), and a nebular δO^{18} of $\sim 14\%_0$ (Onuma *et al.*, 1974), this temperature is $\sim 880°K$.

The oxygen isotopic composition of the moon is thus consistent with a mixture of high-temperature condensates with materials having lower equilibration temperatures. *The high-temperature condensates incorporated by the moon, however, must have had vastly different oxygen isotopic compositions than the Allende inclusions.* In fact, every chondritic high-temperature condensate so far measured (Clayton *et al.*, 1973a) is vastly enriched in O^{16} compared to the moon. The lunar equilibration temperature derived above is considerably below the condensation temperatures of nickel–iron and magnesium silicates in the pressure range 10^{-3}–10^{-5} atm. These phases could have made up a significant fraction of the lower temperature component.

Another important difference between the lunar rocks and the Allende inclusions is their radically different refractory siderophile element contents. As predicted by condensation calculations (Grossman, 1973), the Allende inclusions are enriched in refractory platinum metals, for example, by factors of 12–24 relative to the C1 chondrites (Grossman, 1973; Wänke *et al.*, 1973). The lunar rocks are depleted in these elements by factors of 10^2–10^4 compared to C1 chondrites (Ganapathy *et al.*, 1970), suggesting that a very efficient, post-accretional removal process was active. The possibility that nickel–iron is a significant fraction of the lower temperature component suggests that this process could have been the gravity separation of a siderophile element-rich Fe or Fe–S melt.

The validity of any lunar model in which the moon accretes from substantial quantities of high-temperature condensates rests critically on the assumption that these materials can form somewhere in the nebula in a way which does not lead to the incorporation of the O^{16}-rich component. This component has been found in every chondritic high-temperature condensate measured, including the Ca–Al-rich inclusions in Allende (Clayton *et al.*, 1973a).

If the source of the O^{16}-rich component was interstellar grains, its relative absence from the lunar rocks implies that the grains were not present in the part of the nebula where protolunar material condensed. This may mean that a thorough mechanism was available for the separation of preexisting grains from that part of the nebula before condensation began. Alternatively, it may mean that all preexisting grains were evaporated and their unique isotopic compositions homogenized with the rest of the nebula. This would imply that the moon formed from materials which condensed in a region of the nebula which had a higher initial temperature than the condensation site of the Allende inclusions. This, in turn, suggests that the moon condensed closer to the center of the nebula than the meteoritic condensates.

If the O^{16}-rich component was produced in the sun and then implanted into the condensate grains, the relative absence of it from the lunar rocks implies that protolunar material had a smaller mean surface/volume ratio than the Allende condensates at that stage of nebular evolution.

In summary, it is possible that the moon accreted from a super-chondritic proportion of high-temperature condensate, but such models seem less attractive than previously because this component with the required isotopic composition has never been found in meteorites.

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