OXYGEN ISOTOPE SIMILARITIES AND DIFFERENCES BETWEEN THE EARTH AND MOON: CAN OXYGEN ISOTOPES DISTINGUISH METEORITES ON THE MOON

M. J. Spicuzza[1], J. M. D. Day[2], L. A. Taylor[2], and J. W. Valley[1], [1]Department of Geology & Geophysics, University of Wisconsin, 1215 W. Dayton St., Madison, WI 53706. spicuzza@geology.wisc.edu, [2]Planetary Geoscience Institute, Dept. of Earth & Planetary Sciences, Univ. of Tennessee, Knoxville TN 37996.

Introduction: Oxygen isotope systematics of planetary materials provide constraints on the evolution of the solar system. Recent advances in oxygen isotope analysis (laser-fluorination) provide higher precision data and an opportunity to further constrain the variation among solar system materials. These improved data have led to the need for a more precise method for reporting deviations from the terrestrial fractionation line (TFL) [1] and resulted in more tightly constrained determinations of the TFL [e.g. 2, 3]. Here we report oxygen isotope compositions of Apollo lunar mare basalts, as well as lunar and SNC meteorites. In addition, we suggest that the relative homogeneity of the oxygen isotope systematics of the Moon could be used to advantage in prospecting for meteorites (e.g. from the early Earth) which must have landed on the Moon over the last 4.5GA.

The prevailing theory for the formation of the Earth-Moon system invokes a cataclysmic impact between the proto-Earth and a Mars-sized planet [4,5]. These models indicate that the Mars-sized impactor contributed 70-90% to the mass of the Moon, with the remainder coming from the proto-Earth. Recent oxygen isotope studies of lunar materials conclude that the $\Delta^{17}O$ of the Moon is identical to Earth (within 0.01‰) [3, 6]. If these models are correct, then it restricts the oxygen isotope composition of the impactor to be nearly identical to that of the Earth. Mass balance constraints indicate that if the Mars-sized impactor accounts for only 50% of the mass of the Moon, it must have had a $\Delta^{17}O$ within 0.02 of that of the Earth. On the other hand, it has been suggested that the coincidence in $\Delta^{17}O$ between the Earth and Moon is more simply accounted for by homogenization of the two oxygen isotope reservoirs during the impact event [7].

Analytical procedures: Laser fluorination three-isotope oxygen analyses were performed at the University of Wisconsin using a 30W CO_2 laser with BrF_5 as the fluorinating agent. Oxygen purification was achieved utilizing an in-line Hg-diffusion pump (to remove any F_2) and standard cryogenic methods using 13x mole sieve. We analyzed a variety of terrestrial silicates with a range in $\delta^{18}O$ values from -0.5 to 22.9‰ in order to independently determine the slope of the TFL ($\lambda = 0.5259 \pm 0.0008$; 95% confidence level). λ is defined by the equation:

$$1000\ln [1+(\delta^{17}O/1000)] = \lambda * 1000\ln [1 + (\delta^{18}O/1000)] + 1000\ln (1 + k),$$

where k represents the deviation from the terrestrial fractionation line [1].

This determination provides the baseline for comparing the Earth with the Moon and other extraterrestrial materials, and eliminates the potential errors introduced by differences in analytical conditions between laboratories. Our determination agrees well with those reported from two different labs in [2]. As a proxy for whole rock, we analyzed 2-3mg pieces in an effort to minimize surface area (and reactivity) of the sample during the room-temperature pretreatment of the samples with BrF_5. The use of small chips (versus powders from larger and possibly more representative volumes) might be expected to introduce variability in $\delta^{18}O$ due to mode effects, especially in coarse grained rocks. However, our "duplicate" analyses of 2-3 mg chips differ by less than 0.11‰ in $\delta^{18}O$ except for one coarse-grained Apollo15 basalt (15555) and two Dhofar desert meteorites known to have been affected by terrestrial alteration.

Results: We report whole-rock oxygen isotope analyses of five Apollo 12, seven Apollo 15, and seven Apollo 17 mare basalts (Table 1). In addition, we report whole-rock oxygen isotope data for five paired LaPaz mare basalt meteorites, meteorite Dhofar 287, a lunar "soil" (Luna 24), whole-rock data for three SNC meteorites (Dhofar, Zagami, and MIL 03346,45) and orthopyroxene separate data from ALH84001.

The average $\delta^{18}O$ of seven Apollo 17 high-Ti lunar mare basalts (5.46 ± 0.11‰, 1σ) is significantly lower than the average for low-Ti basalts analyzed in this study. Five Apollo 12 low-Ti basalts yield an average $\delta^{18}O$ of 5.72 ± 0.056‰ (1σ) and seven Apollo 15 low-Ti basalts average 5.65 ± 0.11‰. In addition, the five paired LaPaz (low-Ti) mare basalt meteorites average 5.67 ± 0.05‰ (1σ).

Three SNC meteorites were analyzed as chips; Dhofar 019 ($\delta^{18}O = 5.85 \pm 0.27$‰, $\Delta^{17}O = 0.301$‰, n=2), Zagami ($\delta^{18}O = 4.71 \pm 0.01$‰, $\Delta^{17}O = 0.313$‰, n=3), and MIL 03346-45($\delta^{18}O = 4.69 \pm 0.09$‰, $\Delta^{17}O = 0.328$‰, n=2). Orthopyroxene separated from ALH84001 yielded $\delta^{18}O = 4.99 \pm 0.04$‰ ($\Delta^{17}O = 0.323$‰, n=3)

Discussion: The differences in average whole-rock $\delta^{18}O$ values for low- and high-Ti basalts are thought to reflect the different modal abundances of the constituent minerals and their relative fractionations. Following the model of Eiler [8], we calculated the equilibrium $\delta^{18}O$ value of olivine left in the residuum of a melt assuming the fractionation factor for the melts can be calculated by summing the fractionation factors of the normative mineralogy of the basalt. Fig 1 is a plot of the measured $\delta^{18}O$ – 5.1‰ versus the calculated value based on the normative mineralogy (5.1‰ represents the average value for MORB source mantle olivine [8]). With the exception of the lunar sample 15555, all samples plot within 0.16‰ of the average. We find this correlation convincing, especially when noting the reproducibility of the measurements is generally better than 0.10‰ and the potential for "mode effects" for such small sample volumes. We conclude that the source regions of high- and low-Ti basalts were in oxygen isotope equilibrium and that olivine in the lunar mantle source region has a value of ~5.15‰ ±0.15‰.

The average $\Delta^{17}O$ for lunar mare basalts in this study is +0.008‰ ± 0.012‰ (n=19). Interestingly, in the other recent high-precision study of lunar whole-rock [6], the average $\Delta^{17}O$ for lunar mare basalts was 0.006‰ ± 0.009‰ (n=14) and the combined dataset [3,6] yields an average

$\Delta^{17}O$ of 0.007‰ ± 0.011 (n=33) that is indistinguishable from the earth.

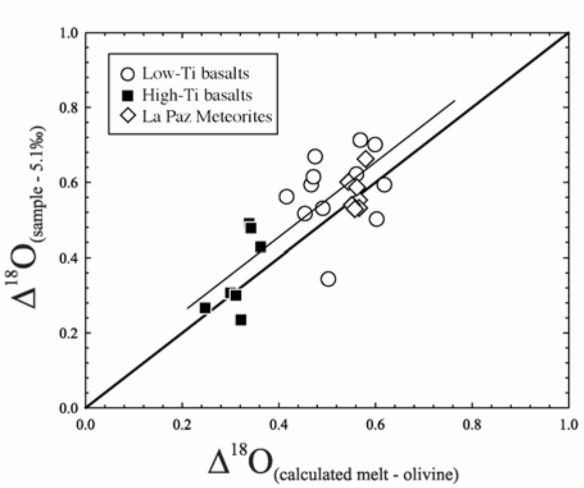

FIG. 1 Measured versus calculated whole-rock value normalized to terrestrial mantle olivine = 5.1‰ (mod. from [3]).

From the combined data set [3,6], the variation in whole-rock $\delta^{18}O$ for igneous rocks on the Moon is on the order of 1‰. The $\delta^{18}O$ of lunar volcanic glasses, KREEP, and mare basalts exhibit a range of 5.2‰ - 5.85‰. Highland rocks vary from 5.5‰ - 6.0‰, and breccias from 5.4‰ - 6.2‰. In contrast to the Moon, the $\delta^{18}O$ of rocks from the Earth's crust is widely varied with values ranging from < 0‰ to >30‰ overall, and <1%o to >15%o for magmas (figure 2), most importantly as a consequence of the interaction with the hydrosphere.

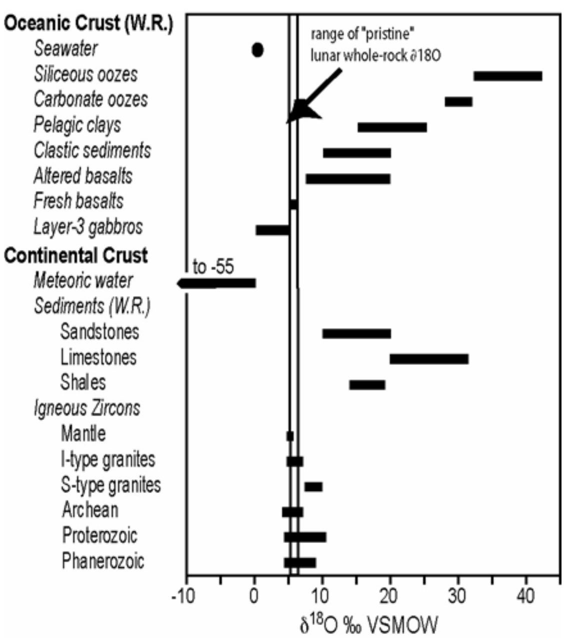

Figure 2 $\delta^{18}O$ of terrestrial rocks (mod. from [9])

Identifying meteorites on the Moon: The proposal of a lunar station raises the possibility of analysis of oxygen isotope ratios without return to Earth. This would be important if there is an effort to explore the moon for meteorites, Meteorites on the Moon are from many sources and may include terrestrial rocks older than 4 Ga. No rocks older than 4 Ga are known to be preserved on Earth and such samples are eagerly sought as they may contain information about the growth of continents, the early hydrosphere, and the emergence of life. Furthermore, ancient rocks on the Moon have been stored in near-pristine conditions if not affected by impacts (i.e., no hydrous alteration).

Samples with anomalous $\Delta^{17}O$ (i.e. $\Delta^{17}O \neq 0$) would be readily identified as meteorites. However, potential samples of the Earth could be distinguished based on distinctive $\delta^{18}O$. In fact, anomalous $\delta^{18}O$ values could earmark many types of meteoritic materials (including angrites, HED's, and SNC's) for sample return from the moon and further study (see fig. 3). One complicating factor would be to avoid samples formed by impact melting of the lunar regolith which may have been fractionated by space weathering and could have different $\delta^{18}O$ signatures.

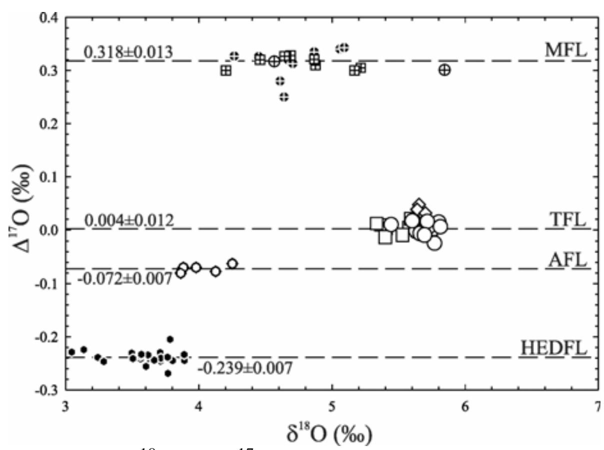

Figure 3. $\delta^{18}O$ vs $\Delta^{17}O$ (modified from [3]) MFL = Mars fractionation line, AFL = angrite fractionation line, HEDFL = HED fractionation line.

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