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Oxygen isotope studies of achondrites

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Abstract-Oxygen isotope abundances provide a powerful tool for recognizing genetic relationships among meteorites. Among the differentiated achondrites, three isotopic groups are recognized: (1) SNC (Mars), (2) Earth and Moon, and (3) HED (howardites, eucrites, diogenites). The HED group also contains the mesosiderites, main-group pallasites, and silicates from IIIAB irons. The angrites may be marginally resolvable from the HED group. Within each of these groups, internal geologic processes give rise to isotopic variations along a slope- $\frac{1}{2}$ fractionation line, as is well known for terrestrial materials. Variations of Δ^{17} O from one planet to another are inherited from the inhomogeneities in the solar nebula, as illustrated by the isotopic compositions of chondrites and their constituents. Among the undifferentiated achondrites, five isotopic groups are found: (1) aubrites, (2) winonaites and IAB-IIICD irons, (3) brachinites, (4) acapulcoites and lodranites, and (5) ureilites. The isotopic compositions of aubrites coincide with the Earth and Moon, and also with the enstatite chondrites. These bodies apparently were derived from a common reservoir, the isotopic composition of which was established at the chondrule scale by nebular processes. Isotopic similarities between chondrites and achondrites are seen only for the following instances: (1) enstatite chondrites and aubrites, (2) H chondrites and IIE irons, and (3) L or LL chondrites and IVA irons. The isotopic data also support the following genetic associations: (1) winonaites and IAB-IIICD irons, (2) acapulcoites and lodranites, and (3) ureilites and dark inclusions of C3 chondrites. An attempt to reconcile the whole-planet isotopic compositions of Earth, Mars, and the eucrite parent body with mixing models of their chemical compositions failed. It is not possible to satisfy both the chemical and isotopic compositions of the terrestrial planets using known primitive Solar System components.

1. INTRODUCTION

The solar nebula was heterogeneous with respect to the distribution of the three stable isotopes of oxygen (Clayton et al., 1973). The heterogeneity manifests itself today in the form of deviations from the expected mass-dependent variations in isotopic abundances, and is seen on all scales studied, from micrometer-sized crystals to thousands-of-kilometersized planets (Clayton et al., 1976, 1977). Oxygen isotope variations in whole-rock samples of chondritic meteorites can be traced back to two principal sources: (1) interactions of gas and dust in the solar nebula and (2) interactions between minerals and fluids in the parent bodies (Clayton et al., 1991). This information is preserved because the chondrite parent bodies have not undergone large-scale melting, homogenization, and differentiation, and thus retain signatures of their early histories. Achondritic meteorites, on the other hand, have been melted, and in some cases homogenized and differentiated, so that their pre-accretional internal isotopic variations are usually not preserved. It is thus inherently more difficult to determine their precursors and to relate them to nebular processes. These observations clearly apply also to planetary bodies, such as Earth, Moon, and Mars.

Although melting and homogenization can make it difficult to interpret oxygen isotopic compositions in terms of the origins of parent bodies, these processes greatly enhance the power of oxygen isotopes to serve as planetary fingerprints, which distinguish one parent body from another. If a parent body undergoes extensive melting and isotopic homogenization, as has occurred, for example, on the Earth and Moon, then all subsequent isotopic variations on that body are subject to a simple mass-dependent fractionation law:

$$\delta^{17} O = 0.52 \delta^{18} O, \tag{1}$$

where the δ -values are referred to any sample of the parent body. If we use the terrestrial SMOW standard for reporting all data, then, on a three-isotope graph, all terrestrial rock samples fall on a line:

$$\delta^{17}O_E = 0.52\delta^{18}O_E$$
 (subscript E = Earth) (2)

and samples from a parent body X fall on another line:

$$\delta^{17} O_X = 0.52 \delta^{18} O_X + \Delta^{17} O_X, \tag{3}$$

where $\Delta^{17}O_x$ is a constant for body X. For example, for the SNC parent body (source of the shergottites, nakhlites, and Chassigny), $\Delta^{17}O \approx 0.3\%$, and for the HED parent body (source of the howardites, eucrites, and diogenites), $\Delta^{17}O \approx -0.3\%$ (Clayton and Mayeda, 1983). (Updated values of these constants are provided in this paper.) For any individual meteorite, $\Delta^{17}O_x$ can be determined as ($\delta^{17}O - 0.52\delta^{18}O$), and thus meteorites can be grouped according to their values of $\Delta^{17}O_x$, each group corresponding to a single oxygen reservoir, and possibly to a single parent body.

Analytical data for samples from a single planet, e.g., Earth, must show some scatter with respect to the simple law of Eqn. 1. At least four sources of variation are known for terrestrial materials: (a) analytical uncertainty, (b) variation of the proportionality constant, (c) departure from the

| CLASS | GROUP | METEORITE* | SAMPLE No.º | δ ¹⁸ 0† | δ ¹⁷ 0† | ∆ ¹⁷ 0†† | MEAN | (SD) | N‡ |
|-------|--------------|--------------|----------------|--------------------|--------------------|---------------------|------|-------|----|
| SNC | SHERGOTTITES | ALH 77005 | | 3.88 | 2.18 | 0.16 | | | 2 |
| | | EET 79001A | | 4.23 | 2.46 | 0.28 | | | 2 |
| | | EET 79001B | | 4.34 | 2.56 | 0.30 | | | 2 |
| | | LEW 88516 | | 4.14 | 2.39 | 0.26 | | | 1 |
| | | QUE 94201 | | 4.97 | 2.31 | 0.19 | | | 1 |
| | | Shergotty | USNM 321 | 4.61 | 2.76 | 0.36 | | | 4 |
| | | Zagami | BM 1966.54 | 4.55 | 2.75 | 0.38 | | | 2 |
| | | Y 793605 | | 4.01 | 2.30 | 0.21 | | | 2 |
| | | Mean | | | | | 0.27 | (.08) | |
| | NAKHLITES | Lafayette | Me 2116 | 4.73 | 2.79 | 0.33 | | | 1 |
| | | Lafayette px | | 4.57 | 2.57 | 0.19 | | | 2 |
| | | Lafayette ol | | 4.03 | 2.40 | 0.28 | | | 1 |
| | | Nakhla px | Cairo | 4.55 | 2.69 | 0.32 | | | 3 |
| | | Nakhla ol | | 4.10 | 2.49 | 0.36 | | | 3 |
| | | Mean | | | | | 0.30 | (.07) | |
| | CHASSIGNITES | Chassigny | Me 1546 | 3.91 | 2.33 | 0.30 | | | 3 |
| | | Mean | | | | | 0.30 | | |
| | OTHER MARS | ALH 84001 | | 4.53 | 2.58 | 0.22 | | | 2 |
| | | Mean | | | | | 0.22 | | |
| | | Class Mean | | | | | 0.28 | (.07) | |

Table 1. Oxygen Isotopic Compositions of Achondrites and Irons

linear approximation, and (d) non-mass-dependent isotope effects. Clayton and Mayeda (1983) assessed the uncertainties in their analytical procedures, and found a standard deviation of 0.07% in Δ^{17} O in replicate analyses of terrestrial samples. The same analytical procedures were used for data presented in this paper. The proportionality factor, 0.52, in Eqn. 1 is a mean value based on measurements of terrestrial rocks and minerals; the ratio of ¹⁷O/¹⁶O fractionation to ¹⁸O/ ¹⁶O fractionation may vary over a range of about 0.505 to 0.530 due to different mass dependence for different processes (Matsuhisa et al., 1978). Variations in the factor may occasionally produce Δ^{17} O variations up to 0.2% in individual samples, but values this large are rare. In addition, the linear Eqn. 1 is an approximation to a more accurate power-law relationship:

$$\left(1 + \frac{\delta^{17}}{1000}\right) = \left(1 + \frac{\delta^{18}}{1000}\right)^{0.52}.$$
 (4)

The linear approximation is in error by 0.05% for δ^{18} O = 20% and becomes worse for larger δ -values.

The effects of non-mass-dependent isotopic fractionation (Heidenreich and Thiemens, 1983) have not been observed in natural terrestrial materials, with one striking exception: atmospheric ozone (Schueler et al., 1990; Krankowsky et al., 1995). Although the details are still unknown, it appears likely that some variant of the molecular-symmetry-based mechanism discovered in the laboratory by Thiemens and Heidenreich (1983) also operates in the Earth's atmosphere. Isotopic anomalies in atmospheric ozone may be transferred to other atmospheric molecules such as CO₂ by exchange reactions (Wen and Thiemens, 1993), but no example has yet been found in which such effects have been preserved in terrestrial rocks and minerals. A possible example of preservation of this effect in martian rocks has been described by Karlsson et al. (1992), who found that water extracted from SNC meteorites has Δ^{17} O greater than that of the host rocks.

The net effect of the processes which can lead to departures for terrestrial rock and mineral compositions from Eqn. 1 is very small. A study of mantle-derived rocks and ancient sediments revealed no resolvable departures from this relationship (Robert et al., 1992).

The usefulness of Δ^{17} O in classifying achondrites and distinguishing among parent bodies depends on the magnitude of its variation from one parent to another, and thus on the likelihood of accidental coincidences. In meteorites, we do not know a priori whether or not the parent body has undergone internal isotopic homogenization, so that the measured range of Δ^{17} O values for a given meteorite class (or a given parent body) contains both the effects described above and the intrinsic internal variability if the body was not homogenized. For the data in Table 1, the mean of Δ^{17} O and its standard deviation are presented for each group. Values of 0.05 to 0.10 are typical of the standard deviations for isotopically homogenized parent bodies. Larger values indicate isotopic inhomogeneity in the parent body (e.g., ureilites and acapulcoites/lodranites).

Comparison between the Earth and Moon may be informative. All measurements of Δ^{17} O of lunar rocks and minerals are consistent with $\Delta^{17}O = 0$; i.e., lunar and terrestrial mateDownload English Version:

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