

SMALL SCALE OXYGEN ISOTOPE VARIATIONS IN ULTRADEEP (>300 KM) AND TRANSITION ZONE XENOLITHS

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The oxygen isotopic composition of 6 mantle xenoliths for which an origin in the transition zone can be deduced from their mineralogy (Haggerty, 1994) have been investigated. Our measurements, using the laser fluorination technique, demonstrate that ultramafic mantle rocks from the transition zone have an isotopic composition indistinguishable from that of the upper mantle (Mattey et al., 1994). Some of the xenoliths show, however, fractionation features which are unexpected. Individual mineral grains, selected to be clear and free of any visible signs of alteration, show evidence of isotopic disequilibrium on the sub-millimeter scale.

Garnets and olivines tend to have a fairly restricted $\delta^{18}\text{O}$ range, while clino- and orthopyroxenes on the other hand display much wider variations. The smaller variability of $\delta^{18}\text{O}_{\text{olivine}}$ compared to that of the coexisting pyroxene contrasts with the observation made in upper mantle xenoliths, where generally the variability of $\delta^{18}\text{O}_{\text{pyroxene}}$ is less than that of coexisting olivine (e.g. Kyser et al 1981, 1982, Rosenbaum et al. 1994). Examination of the ^{18}O fractionation between coexisting minerals on the mm scale revealed a range of $\Delta_{\text{cpx-ol}}$ from -3.11 to 0.2 o/oo within one specimen. $\Delta_{\text{opx-ol}}$ fractionations ranging from -0.49 to 0.03 o/oo are observed. Within one specimen fractionations consistent with high temperature equilibrium are found 20 mm away from fractionations which clearly represent disequilibrium.

Some of the observations can probably be attributed to secondary hydrothermal alteration. There are others, however, which are, with the evidence at hand, hard to reconcile with this concept. The most notable is the cpx-garnet fractionation. Based on data for the upper mantle, an average fractionation of 0.24 o/oo is expected (Mattey et al., 1994), which is consistent with theoretical computations (Kieffer, 1982, Zheng, 1993). In the samples studied the cpx-garnet fractionation is, however, consistently negative. Since a negative fractionation is observed in cases where fresh clinopyroxene resides in the center of fresh garnet grains other mechanisms need to be considered.

The exsolution of pyroxene from the majorite involves major structural changes. Not only must cation diffusion occur, but also changes in the oxygen-cation bonding, as well as a change in the density of the packing of oxygen. In the exsolution two phases are formed which differ by about 14% in the density of their O packing. Kinetic isotope effects could occur in the processes affecting oxygen atoms during the exsolution process.

The survival of isotopic disequilibria established in the mantle places limits on the time interval between the commencement of isotopic change in the mantle and the removal of the xenolith from it. E.g. the negative pyroxene-olivine fractionation in upper mantle xenoliths can be understood if isotopic re-equilibration occurs in a fluid dominated system in which pyroxene reacts by an order of magnitude faster than olivine. A condition for the preservation of the state of isotopic disequilibrium is, that the exchanging rock be removed quickly from the mantle after the interaction with the fluid started (Rosenbaum et al. 1994).

For a particular clinopyroxene grain (1mm radius) embedded in a garnet (5mm radius), a diffusion model was constructed. The observed isotopic composition of the clinopyroxene is 5.2 o/oo, while that of the garnet is 5.7 o/oo. For temperatures between 1200 and 1300 °C an equilibrium fractionation of 1.0003 between clinopyroxene and garnet is assumed, and a diffusion coefficient

of 10^{-14} cm²/sec for clinopyroxene (Connolly and Muehlenbachs 1988, Fortier and Giletti 1989, Farver 1989, and Elpick and Graham, 1990) and 10^{-17} cm²/sec for garnet (Fortier and Giletti, 1989).

Since the original state of disequilibrium is not known various initial isotopic composition differences were assumed. If initially the clinopyroxene was depleted by 2 o/oo with respect to the garnet (a rather unlikely case) within 3×10^6 years the presently observed state of disequilibrium would have been attained. If the initial disequilibrium was about 1 o/oo, the observed state would have been achieved within 400 000 years.

A series of computed isotopic composition profiles demonstrates how the presently observed isotope distribution would evolve during prolonged exposure to a temperature between 1200 and 1300°C. Within less than 1 million years isotopic equilibrium would have been established across the contact and within about 2 million years the isotopic contrast between the two minerals would have disappeared. Within 30 my, isotopic equilibrium would be expected to have been established.

The analysis suggest that the observed isotope distribution would have a life on the order of less than 10^6 years before it would have become significantly changed. It hence must have been produced either during or shortly before the eruption of the kimberlite and the rapid cooling of the xenoliths contained in it.

The observations made for transition zone mantle xenoliths demonstrate similarities as well as differences in the $\delta^{18}\text{O}$ record compared to that of upper mantle. The observed isotopic disequilibrium may have multiple causes including fluid-mantle interactions (Lowry et al. 1994) as well as kinetic effects in the exsolution of clinopyroxene from majorite. The extent of preservation of isotopic disequilibria places limits on the timing of the processes inducing them and the removal of xenoliths from the mantle.

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