

Chapter 6

Isotope Geodynamics

The crust, mantle lithosphere and asthenosphere represent distinct chemical reservoirs. The evolution of these reservoirs through time can be traced by analysis of their radiogenic isotopic signatures. In this section we provide the essential concepts needed to utilize radiogenic isotopes as a constraint on the evolution of these reservoirs. Three decay schemes are used extensively in the study of geodynamics: the Rb-Sr scheme, the Nd-Sm scheme and the U-Pb scheme.

6.1 Geochemical reservoirs

Why are there distinctive geochemical reservoirs? The differentiated state of the present Earth is largely due to the formation and segregation of *partial melts* in the interior of the planet. Because magmas are less dense than the solid rocks in which they form, they naturally migrate towards regions of lower pressure. The key process in the evolution of distinct geochemical reservoirs formed due to the segregation of partial melts is the way in which elements are partitioned between the melt and the residual solid. A simple model for the variation in concentration of elements in magmas during partial melting based on batch melting in which the melt stays in equilibrium with the solid until the specified fusion proportion is:

$$\frac{C_l}{C_o} = \frac{1}{F + D(1 - F)} \quad (6.1)$$

where F is the fraction of fusion, C_o is the concentration of the element in the initial solid, C_l is the concentration of the element in the melt, D is the bulk distribution coefficient:

$$D = \sum_{i=1}^n w_i K_{D_i} \quad (6.2)$$

where

$$K_{D_i} = \frac{X_i^a}{X_{melt}^a}$$

and where n is the number of mineral phases, w is the weight fraction of mineral i , and X_i^a is the concentration of element a in phase i . The distribution coefficient of trace elements with respect to specific minerals is a function of ionic charge and radius. Different minerals have different affinities for individual trace elements and therefore different K_D values. The D values represent the effective *bulk distribution coefficient* for the whole mineral assemblage coexisting with the melt. Trace elements that have high D values are referred to as *compatible*, while those with $D < 1$ are referred to as *incompatible*. Note that what constitutes a compatible or incompatible element depends on the specific mineral assemblage. Thus, Sr is an incompatible element for mantle-melting involving olivine and pyroxene assemblages whereas it is a compatible element during crustal melting where feldspars are involved. The crust is the repository of trace elements which are incompatible with respect to the mineralogy of the Earth's mantle.

Equation 6.1 has a number of important special solutions (Figure 6.1). If $D = 1$ then $C_l/C_o = 1$ and there is no fractionation between source and magma. If $D \gg 1$ then $C_l/C_o < 1$ and the element is preferentially retained in the unmelted residue. If $D = 0$ then C_l/C_o tends to $1/F$ and the trace element is strongly partitioned into the melt phase and its enrichment is simply a function of the amount of melting. Figure 6.1 also illustrates the effect of extraction of partial melts on the relative concentration of trace elements in the unmelted solid residue. Note that highly incompatible elements are very quickly and efficiently extracted from the melting solid and the residues of such melting are highly depleted in these elements.

The solution to Eqn 6.1 for varying amounts of melting of two trace elements both of which have $D < 1$ but for which $D_{E_1} (= 0.01) < D_{E_2} (= 0.8)$ where both elements have $C_o = 5\text{ppm}$ is shown in Figure 6.1, Figure 6.1 illustrates the variation of the ratio $(E_1/E_2)_{liquid}$ as a function of F . Clearly, for small to moderate amounts of melting, there is a very significant fractionation of the two trace elements;

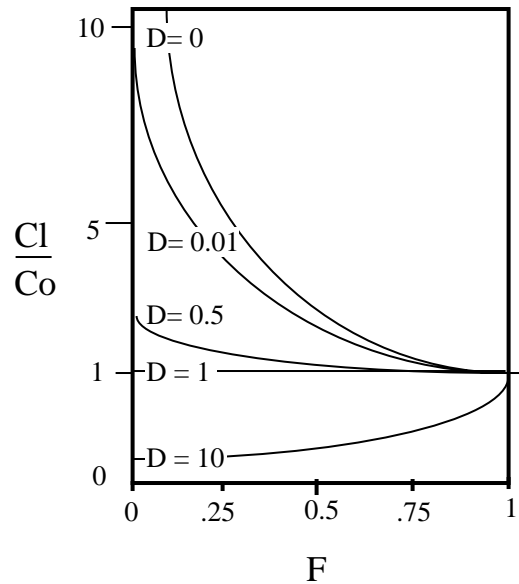


Figure 6.1: Solutions to Eqn 4.1 are shown as a function of the proportion of fusion the degree of enrichment or depletion relative to their source (C_o), of incompatible and compatible trace elements whose bulk distribution coefficients range from zero to ten.

and the ultimate destination of the melts when segregated (i.e., the outer regions of the Earth) will become preferentially enriched in E_1 compared to E_2 . By contrast, the residue of melting always has a lower ratio E_1/E_2 than either the original source or the derived melt.

It is important to note that Eqn 6.1 describes equilibrium batch partial melting which assumes that the liquid remains in contact and equilibrium with the residue up to the specified degree of fusion, at which stage it is totally extracted. This is a rather simplified model of the processes likely to operate in the earth and we will discuss some more physically realistic mechanisms later.

6.2 Radioactive isotopic decay

We can use the models of the type described above to infer something about the compositions of the source regions of magmatic rocks. However, as discussed in the previous section, partial melts of rocks

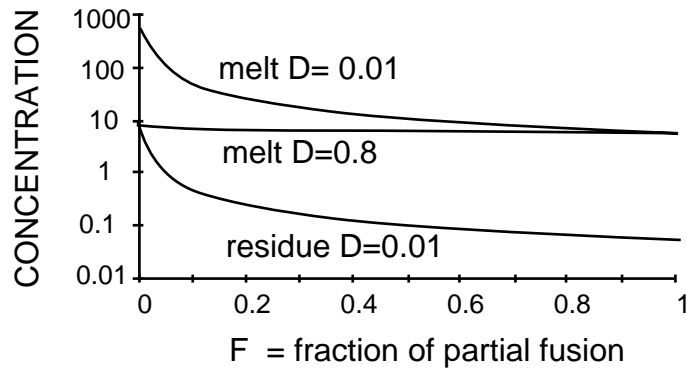


Figure 6.2: Concentration in the melt produced by progressive partial fusion of trace elements whose bulk distribution coefficients are 0.01 and 0.8. Initial concentration in the source was 5 ppm. Also shown is the concentration in the residual solid of the element whose $D = 0.01$.

only reflect the composition of those source rocks indirectly. By contrast, it is assumed that at the time of melting, the isotopic compositions of the liquid and its source rocks are the same and thus the initial isotopic compositions of magmas are the same as those of their source regions at the time of melting.

Naturally occurring radioactive isotopes decay spontaneously to yield stable daughter isotopes. The rate of decay is a physical constant unique to each specific decay scheme. We refer both to the decay constant, λ , and to the half-life, $t_{1/2} = (\ln 2)/\lambda$, of an isotope. The half life being the time taken for half the initial number of parent isotope nuclei to decay to daughter products.

The equation describing radioactive decay as a function of time is:

$$N = N_0 \exp(-\lambda t) \quad (6.3)$$

where N_0 is the initial number of radioactive parent atoms and N the number left after time t . The numbers of new daughter atoms, d^* , produced in time t , is given by:

$$d^* = N_0 (1 - \exp(-\lambda t)) \quad (6.4)$$

In most geological materials, there will already be some of the daughter element present at the start of the time interval of interest which must be added to by the products of radioactive decay. Thus the total radiogenic isotope concentration, d , of a rock or mineral is given

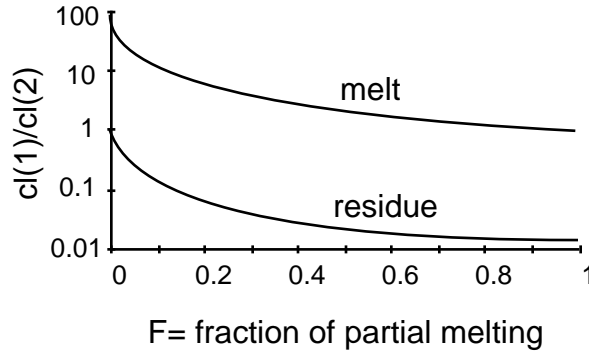


Figure 6.3: The concentration ratio of a highly incompatible element ($C_i(1)$) whose $D = 0.01$ and a moderately incompatible element ($C_i(2)$) whose $D = 0.8$ in the melt formed by progressive fusion.

by:

$$d = d_0 + N(\exp(-\lambda t) - 1) \quad (6.5)$$

and is clearly dependent on:

1. the amount of the parent isotope present at the start,
2. the amount of the daughter isotope present at the start, d_0 and,
3. the time interval during which this system was closed, t and at this stage the amount of the parent isotope left is N .

Isotopes in geological samples are analyzed by use a mass spectrometer as a ratio relative to a reference non-radiogenic isotope of the same element. In the Sr-system, the reference isotope is ^{86}Sr and in the Nd-system it is ^{144}Nd . The so-called isochron equation shows that after the passage of t years (in 10^9 units) the present day isotope ratio of a closed system depends on its initial ratio and the ratio which is purely a function of the concentration ratio of the parent and daughter elements. For the Rb-Sr scheme the isochron equation is

$$\left(\frac{^{87}\text{Sr}}{^{86}\text{Sr}}\right) = \left(\frac{^{87}\text{Sr}}{^{86}\text{Sr}}\right)_i + \left(\frac{^{87}\text{Rb}}{^{86}\text{Sr}}\right)(\exp(\lambda t) - 1) \quad (6.6)$$

(note that in this case ^{87}Sr is the radiogenic isotope and ^{86}Sr the reference and ^{87}Rb is the radioactive parent however we could equally

apply the same equation to the other decay schemes with appropriate decay constants).

6.3 Isotopic fractionation during melting

For elements of high atomic mass melting processes do not affect the ratios of two isotopes of the same element because the fractionation is the same for each. However, melting frequently does produce large fractionations between parent and daughter isotopes. The relationship between the isotopic ratio, α , the parent to daughter ratio, μ , which is a compositionally dependent term and the time of differentiation, t , in a closed system is given by

$$\frac{d\alpha}{dt} = \lambda\mu \quad (6.7)$$

where λ is the decay constant. The importance of this is considerable not only for the determination of specific rock ages, but also because it provides a connection between the chemistry of specific regions of the earth and their radiogenic isotopic compositions. As discussed in Section 4.1 the changes in concentration of trace elements in both extracted melts and in residues during fusion depends on the D value. Because parent and daughter elements will have different D values partial melting will produce melts and residues whose parent to daughter ratios are different from the source. For instance in general $D_{Rb} < D_{Sr}$ and $D_{Sm} > D_{Nd}$. Therefore melts have higher Rb/Sr and lower Sm/Nd ratios than their sources. Figure 6.3a illustrates the general process of the production of a new separate reservoir, M , by melting and melt extraction of a source, S , at time, TD , leaving a slightly modified source region, R . This specific example is true for a situation in which the parent isotope is more incompatible than the daughter (as is the case for Rb - Sr). If the opposite were true (as in the case of Sm-Nd), the extracted melt trajectory would be less steep than that of the residue. Figures 6.3b and c illustrate two other petrogenetic processes which are also relevant. Of particular importance is isotopic homogenisation which will occur by diffusion during geologically meaningful timescales at temperatures above the so-called blocking temperature. It is an important observation that even at temperatures above the blocking temperature, the rate of diffusional equilibration between adjacent solids or non-convective, non-mixing liquids is only of the order of a

meter in 10^7 years. For this reason the establishment of large separate geochemical reservoirs like the crust or asthenospheric upper mantle, requires both large scale melt extraction and convective heat and mass transfer.

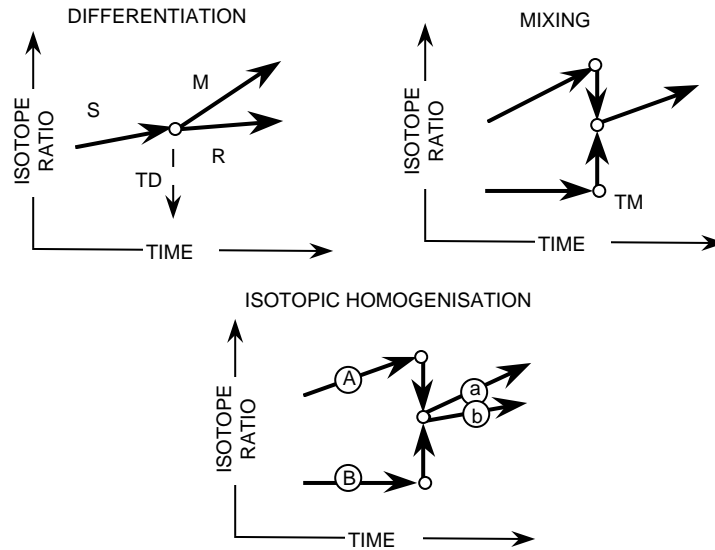


Figure 6.4: Generalized model for three broad types of temporal control on rock radiogenic isotopic composition. a) differentiation by partial melting, S = source, m = melt, TD = time of melting. b) mixing of two compositionally and isotopically distinct melts at time TM to form a homogeneous hybrid. c) Solid-state isotopic homogenisation of two adjacent rocks masses or minerals, A and B, due to metamorphism.

6.4 The chondritic earth

An important process in the chemical differentiation of the earth from an initial compositionally homogeneous globe has been the generation and segregation of partial melts to form several new separated reservoirs (crust, asthenosphere etc). A comprehensive geodynamic model of the earth requires knowledge of the age, life-span and source of these separate regions.

Because the earth has undergone continuous differentiation since its formation at about 4.55×10^9 years ago it is not a simple problem to determine what the bulk composition of the whole, undifferentiated planet was. We only have easy access to the outer crustal

portions of the earth and the entire crust only represents a small proportion of the earth's total mass. As we have already seen, this part of the earth has an unrepresentative enrichment of those elements whose distribution coefficients with respect to the mantle minerals are very low. For instance approximately 70% of the earth's entire supply of K, Rb and Cs all now reside in the crust.

By contrast, the Moon which is a considerably smaller planetary body, has undergone much less differentiation than has the earth. Its crust is much more like what we suppose its whole composition to be and the oldest dates obtained from Moon rocks are much closer to 4.55 Ga. than the oldest terrestrial dates (Amitsoq Gneiss from S.W. Greenland dated at 3.9 Ga.). This illustrates that some form of crustal re-cycling on earth has destroyed most of the original crust, activity absent from the Moon since the earliest Proterozoic. Meteorites are pieces of a fragmented planetary body (probably more than one) which probably occupied an orbit between earth and Mars. Like the Moon, this appears to have had only a short history of geological activity and was then broken-up to form the asteroid belt. Meteorites therefore represent a unique opportunity to directly examine and analyze samples from the interior of a terrestrial planet like earth. They fall into three groups:

- Metallic Iron-Nickel alloy meteorites; fragments of the core of an earth-like planet.
- Stony meteorites; these are mainly silicate-rich composed of olivine and pyroxenes and are derived from the equivalent of the earth's mantle.
- Chondritic meteorites; a unique class of meteorites which have had no high temperature history. They are carbon-rich and often contain significant concentrations of volatile compounds.

The stony meteorites provide mineral isochrons in both the Rb-Sr and Sm-Nd systems (O'Nions et al, 1979) that indicate the age of the terrestrial planets to be 4.5 -4.55 Ga. and indicate initial $^{87}\text{Rb} / ^{86}\text{Sr}$ and $^{143}\text{Nd} / ^{144}\text{Nd}$ ratios to be 0.69898 and 0.50682, respectively.

The chondritic meteorites have not been part of a planet that has differentiated core and mantle and their bulk composition is thought to be very similar to the inter-stellar dust from which the planets accreted. The first approximation of the whole-earth's composition

is therefore the composition of the chondritic class of meteorites and it is informative to present data on terrestrial samples normalized relative to this as illustrated by Figure 6.4.

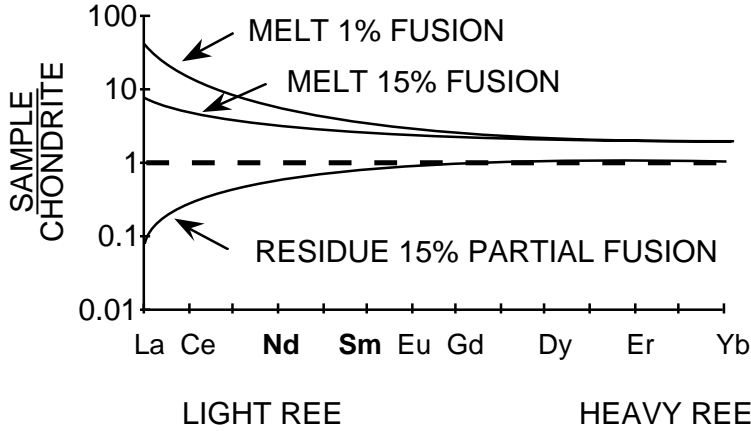


Figure 6.5: The dashed line represents the normalized rare earth content of a chondritic source and the normalized compositions of melts due to 1 and 15% partial fusion are shown. The residue in equilibrium with the 15% melt is the lower most curve.

Given that we are able to define the initial Nd and Sr isotopic composition and age of the Earth and we can use the composition of chondritic meteorites to define its Rb/Sr and Sm/Nd ratios, we are now in a position to predict the variation in the $^{87}\text{Rb} / ^{86}\text{Sr}$ and $^{143}\text{Nd} / ^{144}\text{Nd}$ ratios for this chondritic uniform reservoir (CHUR) as a function of time (McCulloch and Wasserburg, 1978). The "mantle" lines in the the two lower diagrams in Figure 6.4 are a representation of this reservoir.

It is very common that the Sr and Nd isotopic compositions of rocks are not quoted directly as ratios but are compared to the expected values of the chondritic reservoir. This is referred to as epsilon, ϵ , notation.

$$\epsilon_{Nd}(0) = \left(\frac{\left(\frac{^{143}\text{Nd}}{^{144}\text{Nd}} \right) (0)}{I_{chur}(0)} - 1 \right) \times 10^4 \quad (6.8)$$

where

$$I_{chur}(0) = I_{chur}(T) + \left(\frac{^{147}\text{Sm}}{^{144}\text{Nd}_{Chur}} \right) (e^{\lambda t} - 1)$$

$$\lambda = 6.54 \times 10^{12}$$

$$I_{chur}(0) = 0.512638 \quad \text{and} \quad \frac{{}^{147}\text{Sm}}{{}^{144}\text{Nd}}_{chur} = 0.1967$$

$\epsilon_{Nd}(0)$ (or $\epsilon_{Sr}(0)$) refer to the comparison between the present day value of the chondritic reservoir and the present day isotope value of the sample (ie the measured value). If this rock is not of zero age and if the age is known, then its initial isotopic composition at its age, T , can be calculated and compared to the chondritic reservoir at that time, $\epsilon_{Nd}(T)$ and $\epsilon_{Sr}(T)$. In the upper diagram of Figure 6.4, the axes could equally be presented as isotope ratio values or ϵ values. The "bulk earth" position in this diagram is the intersection point of $\epsilon_{Nd}(0) = 0$ and $\epsilon_{Sr}(0) = 0$. This figure shows the field modern day ocean island basalts (OIB) and mid ocean ridge basalt (MORB). These reflect the composition of the present day mantle and show that though some ocean island magmatism is sampling mantle whose composition is close to the predicted chondritic Earth, the modern mantle shows some very strong deviation from this model. This is particularly so for those portions of the mantle sampled by the voluminous magmatic production at the mid ocean ridges. On this same diagram from Figure 6.4, the field of continental crust is also shown. This has high ${}^{87}\text{Rb} / {}^{86}\text{Sr}$ (or positive ϵ_{Sr}) values and low ${}^{143}\text{Nd} / {}^{144}\text{Nd}$ (or negative ϵ_{Nd}) values. As the range of Sm/Nd ratios of continental rocks is relatively limited, Eqn 6.8 reduces to express ${}^{143}\text{Nd} / {}^{144}\text{Nd}$ as a function of time and this is illustrated as the age contours in the continental crustal quadrant in Figure 6.4. Note that this is not so for ${}^{87}\text{Rb} / {}^{86}\text{Sr}$ as there is a very large range of Rb/Sr values for common crustal rocks.

6.5 The depleted mantle

Figure 6.4 shows that for the partial fusion of a chondritic source like that assumed to represent the undifferentiated earth, the degree of total rare earth element enrichment in the melt diminishes with larger amounts of melting and that the residual source is depleted in these incompatible elements. More importantly, the melts have lower Sm/Nd ratios than the original pristine source and the residue has a higher Sm/Nd ratio.

To a simplified first approximation (with respect to the rare earth elements), Figure 6.4 illustrates the evolution of the two important

terrestrial geochemical reservoirs:

- a melt repository (= crust) and,
- a source region depleted in the crust forming elements by crustal extraction (= some part of the present mantle).

The lower two diagrams in Figure 6.4 show the present day range of Nd and Sr isotopic compositions of the continental crust with respect to a mantle reservoir. These represent simplifications of the true situation, because (Figure 6.4) the extraction of melts from the mantle bring about complimentary changes to the composition of that mantle. A mantle region that has suffered partial melting and melt extraction will have higher Sm/Nd and lower Rb/Sr and U/Pb ratios than the original chondritic unmelted mantle.

Mid-ocean ridge basalts are melts of the contemporary decompressed asthenospheric mantle beneath the oceanic ridges. Even though these are melts they have rare earth patterns which are like the solid residue in Figure 6.4 and are strongly LREE-depleted. This suggests they are the products of remelting of mantle that has already lost a melt component before. In Figure 6.4, the MORB field is at high ϵ_{Nd} and low ϵ_{Sr} confirming that their source mantle had achieved its depleted chemistry over a long period of geological time (perhaps 2 Ga). We then view the continents as being complimentary to the depleted mantle now being sampled by mid ocean ridge magmatism. The average age of the continental crust is given by:

$$\langle T \rangle = \int_0^T \tau_m(\tau) dt \quad (6.9)$$

where τ_m is the mass of increment of crustal addition aged τ .

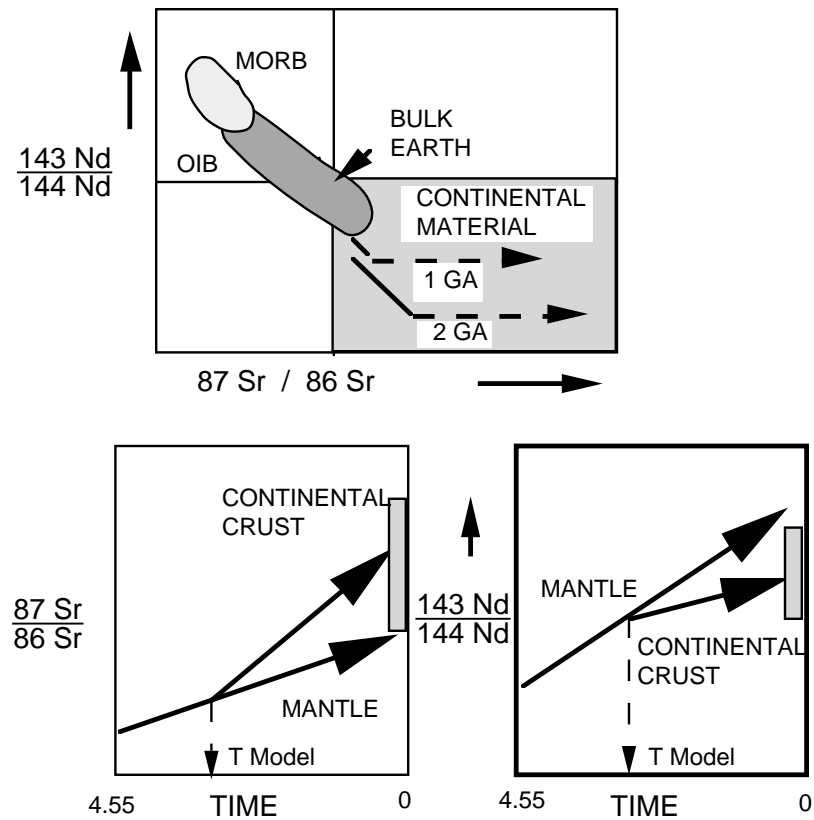


Figure 6.6: Upper: Present day Nd - Sr isotopic covariations for important terrestrial reservoirs. MORB is mid ocean ridge basalts, OIB is ocean island basalts. Lower: Sr- and Nd-time evolutionary diagrams. T_{model} is a model age. Data for the ocean island basalts suggests that in addition to the depleted domain there still exists a mantle reservoir that has not suffered relative Rb and Nd depletion. There is some suggestion that this is a lower mantle or lower upper mantle reservoir which still has a chondrite-like chemistry. This region is also a probable source of rare gas emissions including the now extinct ^{129}Xe suggestive of long term isolation without melt extraction. If this were true it would have important implications in constraining the extent and dimensions of mantle convective cells.