increase in the metallic liquid with temperature. The competition between oxygen, sulfur, and carbon for sites in the metallic liquid in our experiments is strongly resolved in favor of sulfur and carbon. Thus, if these two elements are present, high oxygen abundance in the mantle is apparently a doubtful criterion to establish that oxygen will be the major light element in the core.

The steepness of the solvus between liquid metals and silicates becomes measurably less dramatic at high temperatures. Consequently, we envision that there are regions of pressure–temperature–oxygen fugacity space in which the chemical distinction between silicates and metals is diminished and partition coefficients approach unity. However, our experiments do not predict in detail the siderophile element signatures of the Earth's upper mantle. Complete assimilation and dissolution of chondritic materials at high temperatures could, however, explain the pervasive, homogeneous distribution of mantle noble siderophiles that is attributed in some models to a late veneer. Finally, if our experiments are a useful guide, it does not appear that oxygen successfully competes with sulfur and carbon for sites in metallic liquids. This observation casts some doubt on the likelihood that oxygen is the dominant light element in the Earth's core.

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Orphan Strontium-87 in Abyssal Peridotites: Daddy Was a Granite

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The ^{87}Sr/^{86}Sr ratios in some bulk abyssal and alpine peridotites are too high to be binary mixtures of depleted mantle and seawater components. The apparent excess, or "orphan," ^{87}Sr appears to be separated from its radioactive parent. Such observations were widely held to be analytical artifacts. Study of several occurrences of orphan ^{87}Sr shows that the orphan component in abyssal peridotite is located in the alteration products of olivine and enstatite in the peridotite. The orphan ^{87}Sr is most likely introduced by infiltration of low-temperature (<200°C) seawater bearing suspended detrital particulates. These particulates include grains of detrital clay that are partly derived from continental (that is, granitic) sources and thus are highly radiogenic. Orphan ^{87}Sr and other radiogenic isotopes may provide a tracer for low-temperature seawater penetrating into the oceanic crust.

The available sources for Sr in oceanic and alpine peridotites are the mantle reservoir (^{87}Sr/^{86}Sr ≅ 0.7023) and seawater (^{87}Sr/^{86}Sr ≅ 0.7092). If these are the only sources, the Sr isotopic composition of abyssal peridotites must lie somewhere in between. The first measurements of Sr isotopes and Rb and Sr concentrations in peridotites (*1, 2*) immediately turned up the problem of excess ^{87}Sr in altered alpine peridotite. Although some rocks showed ^{87}Sr/^{86}Sr ratios consistent with a mantle origin, many had ^{87}Sr/^{86}Sr values much higher than those of seawater (*3*). Some of these samples could not have evolved in a single stage from the bulk Earth. Ultramafic nodules, by contrast, have ^{87}Sr/^{86}Sr ratios comparable to those in basalts. This led to the conclusion that alpine peridotites are not related to basalts, either by partial melting or by crystal accumulation (*4*). The most favored explanation for the high ^{87}Sr/^{86}Sr ratio of alpine peridotites was partial melting, resulting in an increase of the Rb/Sr ratio in the residual peridotite. Nonetheless, alpine peridotites were not considered parental to basalts (*2, 4, 5*).

This conclusion extended to abyssal peridotites as well (*6*). On the basis of isotopic evidence, as well as petrographic and trace-element data, abyssal peridotites strongly resembled the alpine peridotites and thus could not be considered parental to mid-ocean ridge basalts. At the time, the lowest ^{87}Sr/^{86}Sr ratio measured on an alpine or abyssal peridotite was 0.7046 (*1*), which was approximately the average of measurements for basalts. The highest alpine peridotite ^{87}Sr/^{86}Sr measurement was 0.7864 (*4*), and the highest abyssal peridotite measurement was 0.7227 (*6*). Ancient continental crust underlying the mid-ocean ridges was proposed to account for the high ^{87}Sr/^{86}Sr ratios (*7*). These results posed a minor but significant problem for plate tectonics, which required a genetic link between abyssal peridotites and mid-ocean ridge basalts.

The first indication that the high ^{87}Sr/^{86}Sr problem in abyssal and alpine peridotites was an alteration effect came when clinopyroxene from an abyssal peridotite turned out to have a significantly lower ^{87}Sr/^{86}Sr ratio than the bulk peridotite (*8*). Metamorphic minerals plagioclase and hydrogrossular had isotopic compositions that showed significant seawater interaction (0.7088 and 0.7089, respectively), but clinopyroxene from that rock had a much lower ^{87}Sr/^{86}Sr ratio of 0.7067 (*8*). Later reanalysis of leached clinopyroxene reduced this number somewhat to about 0.7052 (*9*).

Subsequently, consistently mantle-like Sr isotopic compositions were found by analysis of clinopyroxenes separated from alpine peridotites (*10*). This showed that alpine peridotites were in fact a plausible residual material for basalt, in terms of Sr isotopes. Since that time, a variety of other lines of evidence (*11–16*) have amply demonstrated the genetic link between abyssal peridotite and mid-ocean ridge basalt. The earlier measurements were ascribed to analytical error or to a metasomatic component. We refer to these ^{87}Sr/^{86}Sr ratios as containing orphan ^{87}Sr because there is a radiogenic component of unknown origin. We present new measurements that suggest an origin for the orphan ^{87}Sr reservoir.

Orphan ^{87}Sr has not been observed in unaltered peridotites or in leached mineral separates from alpine or abyssal peridotites (*8, 10, 16*). This strongly suggests that the process of alteration itself has a fundamental role in the formation of the orphan ^{87}Sr reservoir in abyssal peridotites. In order to

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Table 1. First magnetic fractions and whole rocks (*17*).

Cruise	Sample	^{87}Sr/^{86}Sr	2σ	K (ppm)	Rb (ppm)	Cs (ppb)	Sr (ppm)	K/Rb	Rb/Sr	^{87}Rb/^{86}Sr	Age* (Ma)
					First magnetic fractions						
RC2709	6–2	0.72634	4	202.2	1.116	–	3.6	181.2	0.310	0.875	1765
Vulc5	41–29	0.71115	3	154.7	0.377	2.91	5.9	408.5	0.064	0.180	761
PS86	6–37	0.70920	2	44.7	0.105	0.90	167.1	425.1	0.001	0.002	2935
PS86	6–37†	0.71198	2	22.7	0.108	12.5	1.7	210.1	0.062	0.1753	1958
IO11/76	56–58	0.71033	2	594.2	1.249	–	9.9	475.7	0.126	0.356	306
AII107	40–35	0.70919	2	–	–	–	24.9	–	–	–	–
					Whole rocks						
RC2709	6–2	0.70622	2	72.6	0.121	1.08	4.1	600.0	0.0292	0.0824	0.6
Vulc5	41–29	0.70945	7	110.2	0.171	0.53	20.7	643.3	0.0083	0.0234	2.3
PS86	6–37	0.70914	3	43.3	0.044	0.22	311.5	984.1	0.0001	0.0004	5.0
IO11/76	56–58	0.70910	2	128.8	0.178	1.50	6.8	723.6	0.0259	0.0733	9.7
AII107	40–35	0.70904	49	–	–	–	–	–	–	–	17.0

*For first magnetic fractions, these are two-point ages calculated from whole rock and first magnetic fraction data. For whole rocks, these are crustal ages, the maximum age from sample locations, ridge geometrics, and spreading velocities. †Result after leaching in cold 2 N HCl.

study this process, we separated a first magnetic fraction from several abyssal peridotites that consisted mostly of secondary phases (Table 1) (*17*). Four out of five of the first magnetic fractions show ^{87}Sr/^{86}Sr values higher than that of seawater. Sample PS86:6–37 had a demonstrable orphan ^{87}Sr/^{86}Sr component only after we leached it in cold 2 N HCl. Sample AII107:40–35 is the only one that shows no evidence of orphan ^{87}Sr. Although its first magnetic fraction has a higher ^{87}Sr/^{86}Sr ratio than the whole rock, this may be attributable to the presence of clinopyroxene in the whole rock.

The times required to achieve the measured ages given, starting at the isotopic composition of the whole rocks ("two-point" isochrons), are shown in Table 1. As these ages suggest, the data plot near a ~1500-million-year "isochron" (Fig. 1A). Most of these ages are far too high to have originated in the oceanic crust by single-stage evolution in a closed system. The age of alteration is approximately equal to the age of the crust from which the sample is taken. As is obvious from Table 1 and Fig. 1A, the calculated ages are much higher than the crustal ages, in some cases by orders of magnitude.

To test whether the high ^{87}Sr/^{86}Sr ratios could be the result of contamination or metasomatism in the mantle, we carried out leaching experiments on clinopyroxenes from all of the samples studied here. We determined ^{87}Sr/^{86}Sr and ^{143}Nd/^{144}Nd ratios in both residual clinopyroxene and leachate solutions (*16, 18*). The ^{87}Sr/^{86}Sr values for the leached clinopyroxenes ranged from 0.7028 to 0.7032, and the ^{143}Nd/^{144}Nd values ranged from 0.5131 to 0.5129. In all cases, the ^{87}Sr/^{86}Sr and ^{143}Nd/^{144}Nd values of the leached clinopyroxenes were within the field defined by mid-ocean ridge basalts. In no case was there evidence of contamination at a mantle level that would account for the high ^{87}Sr/^{86}Sr ratios documented here.

These results leave two possibilities. First, the orphan ^{87}Sr may have grown in the abyssal peridotite in a phase with a high Rb/Sr ratio, with the subsequent removal of the parent rubidium. Secondly, the orphan ^{87}Sr may have originated outside the sample, derived either directly from ancient continental lithosphere underlying the mid-ocean ridge or indirectly from detrital sediments. In this case, the carrier of the high ^{87}Sr/^{86}Sr values must be transported from an original reservoir with a high Rb/Sr ratio into abyssal peridotite.

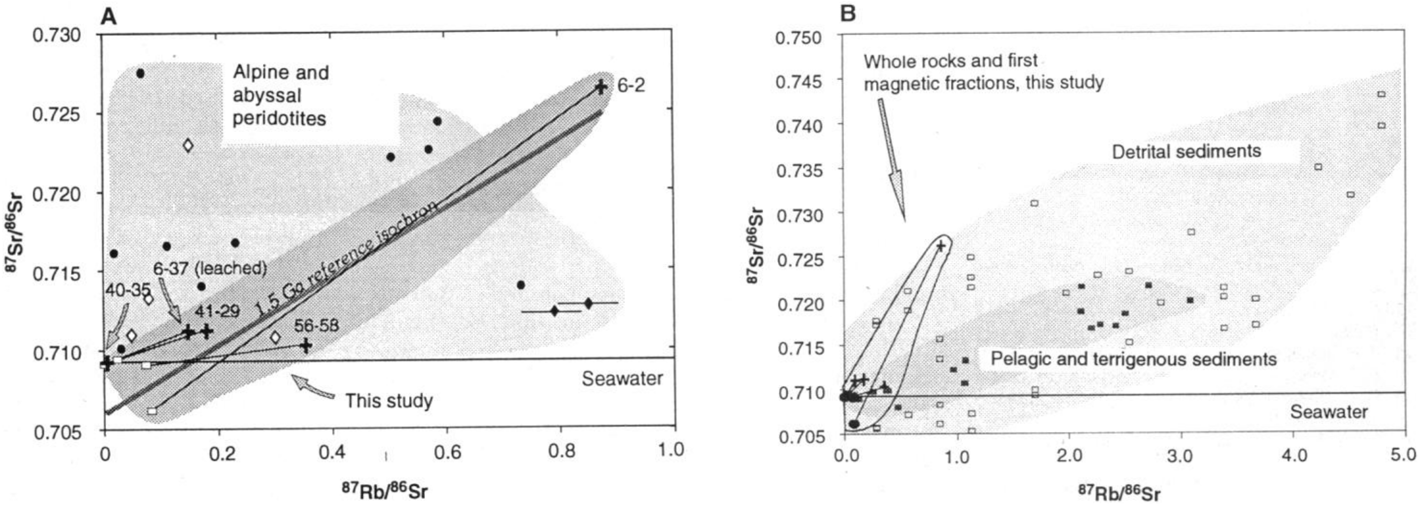

Fig. 1. Isotopic correlation of abyssal peridotites and first magnetic fractions. (**A**) Our data (dark gray region) from abyssal peridotite whole rocks (□) and their associated first magnetic fractions (+) are connected by fine lines. A 1500-million-year (1.5-Gy) reference isochron is shown as a heavy line. Whole-rock alpine peridotite data (light gray region) from (*1, 2*) (◆) (horizontal error bars) and (*4, 5*) (●). Additional whole-rock abyssal peridotite data from (*6*) (◇). (**B**) Peridotites and detrital sediments. Our data (enclosed by black curve) from whole rocks (●) and first magnetic fractions (+) are connected by fine lines. Data from pelagic and terrigenous sediments (*21, 22*) (■) (dark gray region) and detrital sediments (*20*) (□) (light gray region). The isotopic composition of modern seawater is shown as a solid, horizontal line in both (A) and (B).

For the high ^{87}Sr/^{86}Sr ratios to have grown in situ, the following conditions must have occurred: First, the peridotite must have had sufficient Rb added during alteration to raise the Rb/Sr ratio to a high value. Second, to form the high ^{87}Sr/^{86}Sr ratios, there must have been a significant period of closed-system isotopic growth. Third, another metasomatic event must have removed the Rb responsible for the high ^{87}Sr/^{86}Sr ratio. Thus far, no abyssal peridotite with a ^{87}Sr/^{86}Sr ratio higher than that of seawater has ever been found to have a sufficiently high Rb/Sr ratio or a sufficiently old crustal age for the calculated model age and the crustal age to be even close to agreement. This observation requires that Rb was removed from the peridotites subsequent to its addition and radioactive decay in every observed case.

Of the possible metamorphic minerals in an oceanic environment, celadonite has one of the highest Rb/Sr ratios, at about 1.5 to 6.5 (*19*). Sample IO11/76:56–58 would require an Rb/Sr ratio of ~2.9 to produce the observed ^{87}Sr/^{86}Sr value within 9.7 million years. Sample RC2709:6–2 in 0.6 million years would require a Rb/Sr ratio of 706. Such high Rb/Sr ratios in abyssal peridotite are not plausible.

A major possible source of orphan ^{87}Sr in the oceans is detrital clays in oceanic sediments. If derived from ancient continental sources, the detrital sediment suspended in seawater can have ^{87}Sr/^{86}Sr values as high as any measured in abyssal peridotites or higher (Fig. 1B). The alteration products of the breakdown of olivine have low concentrations of Sr [less than 1 part per million (ppm)]. Detrital sediment typically has a high Sr concentration [200 to 300 ppm and higher (*20–22*)]. Thus, small amounts of a fine-grained fraction of pelagic detrital sediment infiltrated into the peridotite would be sufficient to radically change its ^{87}Sr/^{86}Sr.

The first magnetic fraction data and the bulk rock data of (*2*) can to a great extent be explained as a mixture of a seawater component and a detrital sediment component with a wide range in ^{87}Sr/^{86}Sr. Sample RC2709:6–2 is somewhat problematic given that it lies outside the field of deep sea sediments. Sediments with a similar isotopic composition are known, though not from the deep sea. The sample is from an area where there have been few measurements of sediment Sr isotopic compositions. Analysis of sediments from that area should resolve the problem.

If the sediment component is introduced mechanically, the peridotite acts as a filter for sediment bearing near-bottom water circulating at low temperature. Peridotite weathering (alteration at <200°C) involves significant mass transfer and metasomatism at high water/rock ratios (up to 10^5). This results in an increase in porosity and permeability of altered peridotites with progressing low-temperature alteration (*16*, *23*). The transport of large quantities (>10^5 rock volumes) of seawater bearing fine-grained detrital clays through the rock would provide a plausible mechanism for the mechanical transport of sufficient quantities of detrital material into the rock to produce the observed Sr isotopic compositions. For typical particulate concentrations in the near-bottom nepheloid layer of (*24*), water/rock ratios of 10^2 to 10^5 would be sufficient to generate whole rock ^{87}Sr/^{86}Sr ratios of ~0.712, for example.

An alternate pathway for the introduction of detrital Sr into peridotite is seawater that has undergone isotopic exchange with detrital sediments. Such a formation water would acquire the isotopic characteristics of the detrital sediments with which it exchanged and could carry detrital Sr into an abyssal peridotite exposure. There is evidence that pore waters exchange isotopically with detrital sediments (*25*); however, pore waters with ^{87}Sr/^{86}Sr ratios high enough to produce the isotopic compositions documented here have not been observed.

Orphan ^{87}Sr is likely to be derived from detrital sediment. This is introduced either as a fine suspended load or by formation water equilibrated with sediment. Either way, it is probably carried by cold seawater during the low-temperature (<200°C) alteration of the peridotite. The weathering, erosion, and introduction of the clay into the peridotite were probably geologically recent events. The orphan ^{87}Sr can thus be said to have grown in the high Rb/Sr environment of a granitic rock over most of its history. With this pathway for the introduction of a detrital component into the crust, we may be able to trace the circulation of seawater by tracking a sedimentary component in altered igneous rocks. In this study, orphan ^{87}Sr serves as a geochemical tracer because Sr is low in abundance in peridotite. In other rocks, such as gabbros, which have high Sr concentrations, other sediment-derived isotopic signatures such as Os or Pb may be useful tracers of low-temperature seawater circulation within the crust.

The presence of a detrital orphan component of abyssal peridotites provides a previously unrecognized source of trace element and isotopic contamination in oceanic igneous rocks. Such a component may be more widespread than was previously thought. An orphan continentally derived component introduced at low temperature provides some explanation for unusual signatures seen in Sr and Nd isotopes in oceanic peridotite (*4*, *26*), Os isotopes in mid-ocean ridge basalts (*27*, *28*), and Pb isotopes in gabbros (*29*).

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