ARTICLE

Helium-3 from the Mantle: Primordial Signal or Cosmic Dust?

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Helium-3 in hotspot magmas has been used as unambiguous evidence for the existence of a primordial, undegassed reservoir deep in the Earth's mantle. However, a large amount of helium-3 is delivered to the Earth's surface by interplanetary dust particles (IDPs). Recycling of deep-sea sediments containing these particles to the mantle, and eventual incorporation in magma, can explain the high helium-3/helium-4 ratios of hotspot magmas. Basalts with high helium-3/helium-4 ratios may represent degassing of helium introduced by ancient (probably 1.5 to 2.0 billion years old) pelagic sediments rather than degassing of primordial lower mantle material brought to the surface in plumes. Influx of IDPs can also explain the neon and siderophile compositions of mantle samples.

The lighter isotope of helium (^{3}He) has been considered to be a primordial element that was trapped inside the Earth during accretion (*1*). Helium-3–rich magmas are often referred to as evidence for the presence of a primitive, undegassed mantle reservoir, which is invariably assumed to be the lower mantle. However, the large impacts and high temperatures associated with accretion likely cause much of the Earth's original inventory of volatiles to be lost so it has been puzzling that primordial ^{3}He and ^{20}Ne are still degassing from the Earth's interior (*2*). It has been fashionable recently to place the source of ^{3}He ever deeper in the Earth and further back in time. It has been proposed that volatiles were trapped in the original cold nucleus of the Earth or the iron core (*3*). One common model is that the reservoir with high ^{3}He/^{4}He ratios (primordial mantle) is at the base of the lower mantle and is brought to the surface by plumes. However, ^{3}He is also brought to the Earth by a variety of extraterrestrial (ET) carriers including interplanetary dust particles, solar wind, meteorites, and cometary debris and, as recently pointed out (*4*), might therefore enter the Earth's mantle at subduction zones along with other sediments. In this article I explore some implications of this process for the origin of terrestrial ^{3}He and other noble gases and the sources of various types of basalts.

The He concentrations in ET objects are at least two to four orders of magnitude higher than in terrestrial materials (Table 1). Most of the He in the latter is radiogenic ^{4}He, a decay product of uranium and thorium. The ^{3}He/^{4}He ratio varies from $>10^{-4}$ for ET objects to 2×10^{-8} for continental rocks. The atmospheric ratio (R_a) is 1.4×10^{-6}. Magmas from the mantle generally have ^{3}He/^{4}He ratios between 4×10^{-6} and 10^{-5}; the higher values are often found at hotspots, such as Hawaii, Iceland, and Yellowstone, and early in the eruption sequence (*5*). The basalts at midocean ridges, and most mantle peridotites, have a fairly constant ratio of 1.1×10^{-5}, or about 8 R_a; oceanic island basalts (OIBs) range from about 6 to 35 R_a (*5*). The midocean ridge basalt (MORB) reservoir is often referred to as the depleted (*6*), outgassed reservoir, and it is usually assumed to occupy the upper mantle. Paradoxically, MORB has higher ^{3}He contents than magmas attributed to the undegassed reservoir and higher ^{3}He/^{4}He ratios than some hotspot basalts (*5*).

Table 1. Helium contents of terrestrial and ET materials. Exponents for ^{3}He contents of each grouping are given on first row.

Material	^{3}He (cm^{3} STP g^{-1})	^{3}He/^{4}He ($\times 10^{-4}$)
	Extraterrestrial	
IDPs (*52*)	0.13×10^{-4}	2.4 ± 0.3
	0.35	14.5
	0.58	3.1
	0.28	2.3
	0.21	2.8
Lunar fines (*52*)	0.55	3.7
Cosmic (*14*)	0.4	4.0
Gas-rich meteorite (*52*)	0.036	2.4
Solar wind (*52*)		4.3
	Deep-sea sediments	
Bulk (*74*)	0.26×10^{-10}	0.661
Nonmagnetic (*74*)	0.21	0.52
Magnetic (*74*)	36.1	2.18
Magnetic (*18*)	220	3.1
Bulk (LSR)* (*14*)	0.5	1.0
Bulk (HSR)* (*14*)	0.001	0.005
	Igneous rocks	
MORB (*60*)	1.2×10^{-10}	0.11
Loihi (*53*)	0.09	0.45
Kilauea (*53*)	0.08	0.20
Reykjanes (*75*)	0.082	0.17
Samoan xenoliths (*38*)	0.12	0.31

*LSR, low sedimentation rate; HSR, high sedimentation rate.

Current Input of ET Helium to Earth

The flux of meteoritic and cometary material impacting the Earth is estimated to be 50 to 100 Gg year^{-1} (1 Gg = 10^{9} grams). This rate has been fairly steady over the past ~70 million years (*7*) except for brief periods of much higher flux. These rates have been established from Ir and Os abundances in deep-sea cores and from cratering statistics (*7–10*).

The incoming ET flux includes large objects, many of which fragment, burn up, and outgas in the atmosphere, and smaller objects, which can deliver encapsulated gases to the surface of the Earth. The He in very small particles (those <50 μm across are known as interplanetary dust particles or IDPs), along with other rare gases, is generally encased in magnetite (*11, 12*), a He-retentive material (*12*), or in magnetite-rich silicates. Experimental studies have shown that IDPs can retain half of their He upon heating to 650°C and still retain some at 950°C (*13*). As a result, IDP noble gases cannot be treated as ordinary volatiles, which are typically released at much lower temperatures during diagenesis and metamorphism.

IDPs deliver most of their He to the sedimentary columns. The ^{3}He content of deep-sea sediment cores ranges up to 6×10^{-11} cm^{3} g^{-1} (*14*). The ^{3}He/^{4}He ratio varies from 10^{-8} to 1.7×10^{-4}. Some of the highest ^{3}He ($>3 \times 10^{-11}$ cm^{3} g^{-1}) and ^{3}He/^{4}He values ($>10^{-4}$) have been found deep in sedimentary cores (and further back in time). This evidence further suggests that He can be retained in small particles against the influences of diffusion, seawater corrosion, and burial. In some areas, cosmic dust may make up 0.01 to 5 ppm of pelagic sediments (*14*). The ^{3}He concentration of deep-sea sediments is high ($>5 \times 10^{-11}$ cm^{3} g^{-1}) in regions of low sedimentation (<5 mm per thousand years) (*15*). The total amount of ^{3}He in oceanic sediments may be comparable to, or greater than, the amount in the atmosphere (*16*).

The average (ET) ^{3}He flux in small particles (<10 μm) to the deep sea has

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been 2×10^{-15} cm^3 cm^{-2} $year^{-1}$ over the past 40 million years [variation from 1×10^{-15} to 5×10^{-15} cm^3 cm^{-2} $year^{-1}$] (*17*). Using an ET flux of 50 Gg $year^{-1}$, based on siderophiles (8) and an IDP ^3He concentration of 4×10^{-5} cm^3 g^{-1} (*18*), the total ^3He delivered to the Earth is 2×10^6 cm^3 $year^{-1}$. Siderophiles, of course, survive atmospheric entry but some of the volatiles are lost by fragmentation and melting during atmospheric entry and weathering. The flux of ^3He to the seafloor in the small grains and fragments that are sampled in deep-sea cores is about 8×10^3 cm^3 $year^{-1}$ (*17*); thus the survivability rate of imported ^3He is 0.4%.

Large particles may also contribute ^3He (*19*) and will not be adequately sampled in deep-sea cores. Most of the ET mass brought into the Earth is contained in the large objects. Smaller particles, however, receive higher bulk radiation doses in space and thus likely have higher ^3He concentrations.

^{3}He Losses from the Mantle

New oceanic crust is generated at the rate of about 15 km^3 $year^{-1}$ and contains 6×10^{-11} cm^3 g^{-1} of ^3He (*1, 5*). Thus, about 3×10^6 cm^3 $year^{-1}$ of ^3He escapes from the so-called depleted or outgassed mantle reservoir. An unknown amount is lost by vesiculation before sampling. An unknown amount remains trapped in the oceanic crust and is recycled back into the mantle by subduction. The ^3He flux from non-midocean ridge volcanism is an order of magnitude less (*20*), and most of this is from island arcs. The hotspot ^3He flux is therefore much less than 10% of the mid-ocean ridge flux, or $<3 \times 10^5$ cm^3 $year^{-1}$. [Hotspots account for less than 10% of the terrestrial heat flow and magma production and with high $^3He/^4He$ ratios hotspot magmas contain about 0.4×10^{-11} to 4.0×10^{-11} cm^3 g^{-1} of ^3He (*5*).] The amount of ^3He currently outgassing from hotspots, therefore, is about one-tenth of the total ET ^3He influx to the atmosphere and about one or two orders of magnitude higher than the current rate that ^3He is incorporated in seafloor sediments from IDPs and, thus, potentially cycleable into the mantle at subduction zones.

One difficulty in the above comparison is that both mantle outgassing and ET influx rates vary greatly with time (*21*). Even if all the ^3He and other volatiles being outgassed from the Earth have been recycled (or cycled in the case of ^3He), there is no requirement that present input rates equal present output rates. It is the time-integrated fluxes that are important but these are not available. Recycled material may age for $>10^9$ years before returning to the surface (*22, 23*). Thus, it is the ancient IDP flux, rather than the modern flux, that is most relevant.

Subduction Zone Processes

The question arises, can the IDP in sediments survive subduction and how deep do they go into the mantle? Several studies have already shown that (i) pelagic sediments are efficiently recycled into the mantle (*24*) and their incompatible elements may show up as components in hotspot magmas 1 to 2 billion years later (*22*), (ii) some volatile and incompatible elements definitely survive the subduction process and make their way into the mantle (*25*), (iii) only a fraction of the subducted volatiles are returned promptly to the surface, (iv) noble gases are transported by hydrous fluids and trapped in metamorphic and mantle minerals (*26, 27*), and (v) noble gases are not as incompatible in mantle phases as are many other elements (*28*).

The fate of IDPs on the seafloor is just part of the general question of pelagic sediment recycling and its role in modifying the trace element and isotopic chemistry of the mantle. In order to affect mantle chemistry, the material in question must survive on the seafloor for periods of up to 10^8 years and avoid being scraped off at continent or trench margins. The highest ^3He concentrations in seafloor sediments are found in magnetic separates (*29*). Magnetite (Fe_3O_4) is stable on the seafloor and in the oceanic crust and should readily survive subduction in the relatively cool slab environment; additional Fe_3O_4 is manufactured, at high temperature, by reactions between H_2O and olivine (*30*). Even ^{10}Be and B, which are initially adsorbed on fine clays, are readily subducted into the mantle (*31*).

About half the sediments on the seafloor are eventually added to continental margins. These are mainly terrigenous sediments. Deep ocean pelagic sediments are preferentially subducted into the mantle (*24, 32*). Most of the volatile-rich material in the subducted sediments and altered oceanic crust is removed mechanically or in hydrous fluids before the slab reaches 200 km in depth. Only a fraction of these volatiles are accounted for in island arc volcanism; most of the remainder probably remains in the shallow mantle or is transported in fluids migrating back up the slab. The ET He appears to be more tightly bound than water or other subducted volatile elements (*12, 18*). If an appreciable amount of an element of concern survives subduction and is not incorporated in island arc volcanism, it is potentially available for future OIB magmatism, even if it does not reside in its original phase. The noble gases are surprisingly compatible in a variety of upper mantle mineral phases (*26, 28*) and, thus, may have a long residence time in the upper mantle. In contrast, they have been thought to have resided, since accretion, in the lower mantle and to be exhaled rapidly once brought into the shallow mantle by plumes.

In order for IDP to dominate the ^3He budget of the mantle source of high $^3He/^4He$ ratio hotspots, they must survive the initial subduction process and avoid the rapid extraction to the surface in island arc basalts (IABs), which is the fate of the most mobile elements (for example, B, Ba, Be, and Cs). On the other hand, the noble gases in subducted IDPs must not be so immobile that they cannot be released by melting at hotspots. Noble gases are moderately mobile in high-temperature aqueous fluids but are readily incorporated into mantle minerals such as amphibole, clinopyroxene, and olivine (*26–28*). Conditions for releasing, transporting, and retrapping of IDPs He may be different from those operating on U and Th, the sources of ^4He (*29*). It is probably easier to transport rare gases into the mantle in a relatively cold slab, buffered by dehydration reactions, than into the primordial mantle during the original high-temperature accretion and differentiation of the Earth. Nevertheless, at some point the IDPs react or melt and the rare gases will be mobilized. Hydrous fluids are capable of both transporting and depositing rare gases (*33*), without outgassing to the atmosphere, even under crustal conditions (*27*).

Some of the volatile elements that survive shallow subduction are, within a short period of time, included in magmas that erupt at island arc volcanoes. Most enter the mantle wedge between the subducting and overriding lithospheres. Thus, a large part of the shallow mantle may be affected by recycling, depending on the efficiency of sediment removal and fluid transfer, the role of the immediate magmatism, and residence times in the mantle. IABs are particularly rich in H_2O, CO_2, alkali metals, and other incompatible and large-ion elements. Small ions such as Be, B, C, N, and P can also be expected to return quickly to the surface (*34*).

According to some estimates, subduction zones deliver six times more water into the mantle than is returned promptly to the surface by island arc volcanism (*25*). This ratio may be similar for other terrestrial volatiles. The residence time of water in the mantle is about 1 to 2 billion years (*30*). Some fraction of the subducted IDP ^3He might be present in IABs but, in modern sediments, it is diluted by terrestrial He; high $^3He/^4He$ ratios are neither expected nor observed in IABs from mantle contaminated by modern sediment (*35*).

Different elements have characteristic and distinct recycling patterns during and after subduction (*31, 36, 37*). Basalts from convergent margin settings are strongly enriched in ^{10}Be, B, Cs, Rb, and Ba compared with those in other tectonic settings (*31, 37*). Other subducted elements such as Pb and Sr have much longer mantle residence times. They are more compatible and less mobile than the elements with extreme ionic radii. Helium does not have a particularly large radius and is relatively compatible in mantle minerals, suggesting that it can be trapped in lattice vacancy defects in crystals. Helium is not enriched in IABs (*35*).

Other elements are released from the slab at greater depth, by dehydration of higher temperature minerals, or by melting, and may also be involved in metasomatism at shallower depths. Mantle xenoliths from Samoa and Loihi are rich in ^{3}He (*38, 39*), presumably the result of transfer by metasomatic fluids (*33, 37*) rather than because of survival of primordial mantle.

Beryllium-10 also falls out of the sky. It is formed by cosmic-ray spallation of oxygen and nitrogen atoms in the atmosphere, transported in rain or snow to the Earth's surface, and concentrated in clay-rich pelagic sediments. Its incorporation into arc magmas directly indicates that oceanic sediments near the top of the sedimentary pile subduct to depths of order 100 km in the mantle and return to the surface in arc magmas in less than 10 million years. Boron is removed from seawater and adsorbed on sediments that enter subduction zones. It is one of the most mobile of elements (*31, 33*). Much of it is squeezed out and returned to the ocean before deep subduction, but, nevertheless, some also makes its way to the source regions of IABs where it becomes an important component in arc volcanism.

The lesson from B and ^{10}Be in IABs is that adsorbed and loosely bound elements that are readily mobilized by low-temperature fluids can make their way into the upper mantle in sediments. IDPs, which contain deeply implanted noble gases in refractory and retentive phases, should therefore be able to deliver their cargo, at least into the upper mantle.

Deep Recycling of Sediments into the Mantle

On average, arc igneous rocks appear to contain 2 to 5% of a recycled sedimentary component (*25, 31, 40*). The rate of sediment subduction is about the same as the rate of arc creation, so more than 95% of deeply subducted sediment remains in the mantle for some long period of time. According to Pb isotopes this residence time may be of order 1 to 2 billion years (*23*).

Weaver (*22*) has shown that isotopic and trace element data for OIBs can be satisfied by mixing ancient [1.5 to 2 billion years ago (Ga)] subducted oceanic crust plus some entrained sediment into the upper mantle OIB source. A few percent sediment is thought to be incorporated into the reservoirs that yield hotspot basalts. Somewhat more sediment can be involved in IABs; these generally erupt at locations directly over the sites of sediment injection. Sediments carry trace element and atmospheric noble gas signatures into the mantle that appear to be lacking in the MORB reservoir, which therefore may be relatively isolated and protected. Mantle affected by sedimentary contamination carries both an ET and an atmospheric noble gas signature, the latter dominating for the heavy noble gases. In addition, ^{4}He and ^{40}Ar are produced in situ in the sediments by radioactive decay, thereby decreasing the ^{3}He/^{4}He and ^{36}Ar/^{40}Ar ratios of subducted material unless the radiogenic products are not retained in the sedimentary column. On the other hand, if the mantle that provides high ^{3}He/^{4}He ratio basalts is primordial, as generally assumed, it should have primordial values for all isotopes, which it does not.

Based on Weaver's results (*22*), I assume that 10% or less of a sedimentary component is present in OIBs. As shown in Fig. 1, sediments containing 100 to 1000 parts per million of ET particles can raise the ^{3}He/^{4}He ratio of depleted basalts (I used MORBs) up to the level observed in OIBs. This is a conservative calculation. If sediment is mixed into a depleted peridotite upper mantle (*22*) before melting and melt extraction, the effect will be magnified inversely to the degree of melting (*41*). The ^{3}He/^{4}He ratio of refractory mantle rocks is about the same as in MORBs, and the total He content is typically an order of magnitude less. Therefore, He from IDP-rich sediments will dominate melts from a peridotite reservoir. The calculation also shows (Fig. 1) that hotspot basalts with low ^{3}He/^{4}He ratios can be satisfied by mixtures of MORB with a few percent terrigenous sediments. Thus, contamination of the upper mantle by pelagic and terrigenous sediments seems capable of explaining the geochemistry of high ^{3}He/^{4}He and low ^{3}He/^{4}He ratio basalts, respectively.

The obvious problem is that present-day sediments do not seem to have a large enough IDP component, unless this component is separated or concentrated after subduction. The IDP component in modern sediments is diluted by the current high terrestrial sedimentation rate. The concentration of IDPs in deep-sea sediments was probably much higher in the Precambrian than it is today (*4, 42*). Low organic sedimentation rates prevailed before the evolution of deep-sea microorganisms. Old ocean floor, antipodal to a supercontinent, can be expected to be particularly starved of sediments. Recycling of sediment-starved ancient oceanic crust is an attractive mechanism for explaining the isotopic and trace, element geochemistry of basalts on oceanic islands.

Recycling oceanic sediments also carries seawater and atmospheric O and Sr isotopic signatures into the mantle (*42*). Seawater

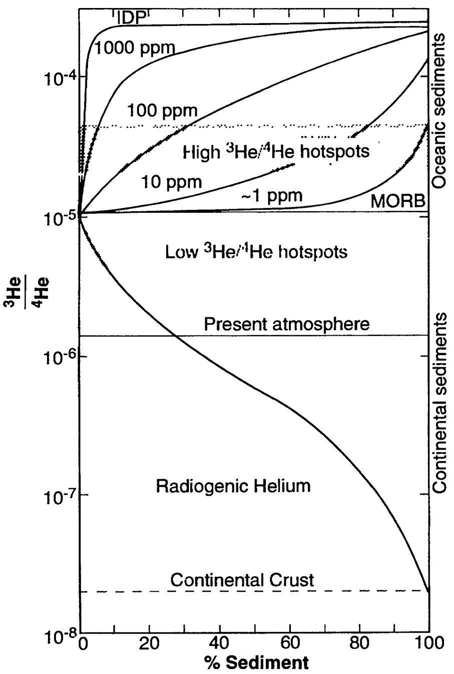

Fig. 1. ^{3}He/^{4}He ratios for various mixtures of MORB-type He and sediments. The upper mixing curves are for MORB and pelagic sediments with different fractions of IDPs ranging from ~1 to 1000 ppm and pure IDP. Deep-sea magnetic separates form the upper end of the range shown to ^{3}He/^{4}He $>10^{-2}$ (*14, 15, 17*). The lower curve is mixing between terrigenous sediments and MORB-like He. At least several percent of a contribution from sediments, in some form, is expected, based on modeling of trace elements and isotopes in oceanic island basalts. This degree of contamination of MORB also accounts for low ^{3}He/^{4}He ratios of hotspots. In order for the hotspots with high ^{3}He/^{4}He ratios to be satisfied by similar amounts of pelagic sediments, these sediments must have high ^{3}He/^{4}He ratios. If pelagic sediments contain 100 to 1000 ppm of IDP, they will have suitable ^{3}He/^{4}He ratios to explain high ^{3}He/^{4}He ratio hotspots by contamination of MORB. Smaller amounts of contamination are required if it is the source, rather than the magma, that is contaminated. The endmember compositions for ^{3}He are: IDP (10^{-5}), pelagic sediments (3×10^{15}), MORB (10^{10}), and terrigenous sediment (3×10^{12}), all in cubic centimeters per gram. The corresponding ^{3}He/^{4}He ratios are 2.5×10^{-4}, 10^{-8}, 10^{-5}, and 2×10^{-8} (see text). It is assumed that all of the IDPs are undegassed and that ^{3}He/^{4}He ratios do not decrease with time. The ranges for hotspots are from (*5, 45, 49*).

chemistry is controlled by continental runoff and hydrothermal venting. The continental contribution to seawater chemistry and deep-sea sedimentation rates varies with time, depending on continental elevation, tectonics, and erosion rates (*43*). The $^{87}Sr/^{86}Sr$ ratio in seawater was low near 800 Ma and before 2 Ga. Low seawater $^{87}Sr/^{86}Sr$ ratios, implying low continental erosion rates, are inferred for 500 million years ago to 1.5 Ga and for 2 to 2.6 Ga (*44*). The terrigenous component of sediments at these times should have been low. If organic sedimentation is also low (or the IDP flux high) then the $^3He/^4He$ ratio of subducted pelagic sediments will be high. Recycling of pelagic sediments during these times will introduce components with high $^3He/^4He$ ratios and low $^{87}Sr/^{86}Sr$ ratios into the mantle. Such a component has been inferred as a source for OIBs with high $^3He/^4He$ ratios, depleted Sr and Nd isotopic values, and moderately radiogenic lead isotopes (*45*). The time periods from 3.0 to 2.1 Ga and from 1.7 to 1.2 Ga may also have been periods of supercontinents, wide oceans, low sedimentation, absence of glaciations, tectonic stability, and low erosion (*44*). Seawater had a depleted Sr isotopic character consistent with low rates of continental erosion. Subduction of deep ocean sediments during these periods would have cycled material with a high $^3He/^4He$ ratio and a moderate $^{87}Sr/^{86}Sr$ ratio into the mantle. This material may remain in the shallow mantle because of its high buoyancy, low viscosity, and low melting temperatures and may be isolated for 1 or 2 billion years (*46*).

A Cosmic Reservoir in the Mantle?

If IDPs are an important contributor to the high $^3He/^4He$ ratio mantle reservoir, then this reservoir should also exhibit other cosmic characteristics including high contents of siderophile elements such as Re, Ir, and Os and near-chondritic isotopic $^{187}Os/^{186}Os$ ratios. Hotspot basalts have high siderophile contents. Hawaiian and Icelandic basalts exhibit high $^3He/^4He$ ratios and high Re and Os contents compared with midocean ridge basalts (*47, 48*). These OIB basalts also have nearly chondritic $^{187}Os/^{186}Os$ ratios (*48*). These properties of OIBs are consistent with the presence of an ET component in the OIB source region. IDPs in the mantle will dominate siderophile, as well as noble gas, geochemistry.

These characteristics could also be taken as support for genesis of basalts from plumes tapping a primordial, undegassed reservoir, perhaps the mantle layer next to the molten iron core. The high siderophile content of hotspot magmas seems to rule out such an origin, however, and favors shallow recycling of subducted IDPs.

This interpretation is further supported by other rare gas data. The $^{20}Ne/^{22}Ne$ ratio of the mantle is distinctly higher than that of the atmosphere. This is a well-known paradox for views of mantle and atmospheric evolution (*49*). For continuous degassing, the atmosphere and primitive mantle should have the same ratio because neither isotope is affected by decay to any significant degree. The $^{20}Ne/^{22}Ne$ ratio of the mantle however is more nearly solar than is the atmospheric ratio (*49–51*), which is inconsistent with mantle degassing as an explanation for the present atmosphere. The high $^{20}Ne/^{22}Ne$ ratios of basalts are consistent with a mantle having IDP (*18*), rather than atmospheric, $^{20}Ne/^{22}Ne$ ratios, and slightly favor an IDP over a solar ratio (*51, 52*). If so, the present Ne in the mantle may be a late addition and have little to do with the origin and composition of the present atmosphere.

Honda *et al.* (*53*) showed that Hawaiian basalts defined a linear trend on a $^{20}Ne/^{22}Ne$ versus $^{21}Ne/^{22}Ne$ plot (Fig. 2). One end-member of this trend is atmospheric neon [possibly from recycled sediment or post-eruptive contamination (*38*)], but the trend deviates from MORB, solar, or planetary values. They proposed, therefore, that hotspot rare gases were a mixture of at least three components, including a solar component. The trend, however, is almost exactly along a mixing line between atmospheric and IDP values. The $^3He/^4He$, $^{20}Ne/^{22}Ne$, and $^{21}Ne/^{22}Ne$ ratios of Loihi and Kilauea basalts all fall between the atmospheric and IDP (*38, 52–54, 58*) values. Thus, an IDP contribution to hotspot magmas explains the high $^3He/^4He$ ratios, the presumed solar Ne component, and the high and unfractionated trace siderophile patterns.

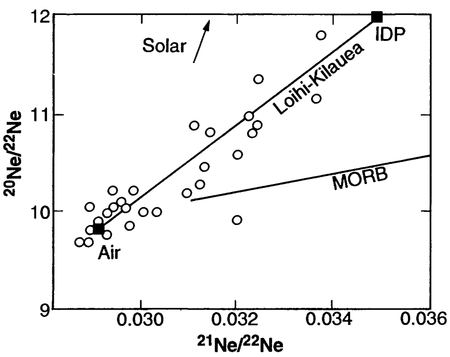

Fig. 2. Three-isotope plots for neon. Atmospheric (Air) and interplanetary dust (IDP) values are shown as squares. Points show high $^3He/^4He$ ratio basalts from Loihi and Kilauea. These points fall on a mixing line between Air and IDP, suggesting that Ne, as well as He, can be explained by introducing an IDP component into the mantle by subduction. The data are from (*52, 53, 73, 77*). Errors in the isotopic ratios are typically 10%.

The high ^4He and ^{21}Ne contents in MORBs and high $^{21}Ne/^{22}Ne$ and low $^3He/^4He$ ratios are consistent with radioactive growth of ^{21}Ne and ^4He over about 4.5 Ga in a depleted mantle reservoir (*55*). ^{21}Ne is produced by the irradiation of oxygen during U and Th decay and thus increases with time. The MORB reservoir appears, therefore, to have been isolated somehow, for about the age of the Earth, from the OIB reservoir and from the recycling of sediments that I suggest refreshes and re-enriches that reservoir (*56*). The OIB reservoir, therefore, is plausibly shallower than the MORB reservoir and may have been outgassed more recently or more continuously. With IDP subduction, however, it is also continuously regassed.

Hotspots with high $^3He/^4He$ ratios such as Loihi, Kilauea, and Samoa tend to have low $^{21}Ne/^{22}Ne$ ratios for a given $^{20}Ne/^{22}Ne$ ratio (*38, 53*). Rocks from other hotspot locations tend to fan out from the atmospheric value and fall between the MORB and Loihi-Kilauea trends (*38*). This pattern suggests mixing between MORB-like mantle (high $^{21}Ne/^{22}Ne$ and low $^3He/^4He$ ratios) and Loihi material. The Loihi-Kilauea neon trend suggests that subducted IDP-rich sediment, or fluids therefrom, may be mixing with, or metasomatizing, depleted peridotite (low U, so low ^4He production) rather than MORB or the MORB reservoir.

Although the part of the mantle melting to form MORBs has apparently been protected from recent recycling of sediments, as argued by Staudacher and Allègre (*57*), it may have been contaminated by IDP He and Ne early in Earth history. This is a possible explanation of the high noble gas contents of the degassed MORB reservoir relative to the primordial, undegassed OIB reservoir. The high noble gas content of MORBs may just reflect a higher incidence of IDPs in the early history of the Earth (*4*) and the low $^3He/^4He$ ratio may be a result of its age (^4He addition).

The observation that IDP Ne anchors one end of the mixing array (Fig. 2) involving basalts with high $^3He/^4He$ ratios suggests that the IDP component is a recent addition to the mantle or it has evolved in a low U environment, for example, a depleted mantle peridotite. The MORB reservoir with low $^3He/^4He$ ratios by contrast is either older or evolved in an environment less depleted in U. That high $^3He/^4He$ ratio basalts occur preferentially at hotspots is consistent with noble gas storage in a refractory part of the mantle or in refractory phases. Peridotite xenoliths from Loihi (*57*) and Samoa (*38, 39*) have high $^3He/^4He$ ratios, possibly the result of enrichment by metasomatic fluids. High temperatures or extensive melting may be required to release the trapped gases.

Mass Balance and the Late Veneer

It is difficult to do a complete accounting of the He budget of the Earth because He is constantly escaping from the atmosphere. There are several widely varying estimates of the current ^{3}He flux from the interior (*49, 59, 60*). Estimates of He outgassing based on crustal formation rates, oceanic He concentrations, and heat flow differ by orders of magnitude.

The heavier rare gases accumulate in the atmosphere. The amount of Ne in the atmosphere (air, oceans, sediments, and crust) is about 7.2×10^{19} cm^3, or 10^5 higher than could have been supplied by the present rate of impactors over the age of the Earth (*60*). ET neon also differs isotopically from atmospheric Ne. Sleep *et al.* (*61*) used lunar Ir and Ni abundances to calculate the total amount of material impacting the moon since solidification of the lunar crust. Scaling to Earth yields 5×10^{21} g as a lower limit for the total mass incident on Earth subsequent to 4.5×10^9 years ago. Crater counts give 10^{24} to 7×10^{26} g as the integrated flux of large bodies since 4.5 Ga (*10, 61, 62*). These integrated amounts imply that ET influx rates were 10^6 times the present rates at some time subsequent to crustal stabilization. The rate presumably decreased rapidly.

Much of the Earth's volatile and highly siderophile elements may have been brought in by this late veneer and subsequent subduction (*63, 64*). Chou *et al.* (*65*), using siderophile abundances in peridotites, estimated that the upper mantle contains 0.74% of a meteoritic component. For undegassed particles, this is adequate to raise the ^{3}He concentration of normal depleted mantle by three orders of magnitude and the ^{3}He/^{4}He ratio by a factor of 40. Estimates of the mass of the late veneer range from 10^{24} to $>10^{25}$ g (*63–65*). Much of this material was probably deposited on the seafloor and subducted. Some may be contributing to present-day ^{3}He degassing. About 4×10^{24} g of carbonaceous chondritic material suffices to explain the terrestrial abundance of such volatiles as C, N, H, Tl, Pb, and Cl (*63*). This amount would provide about 4×10^{19} cm^3 of ^{3}He and 6×10^{21} cm^3 of ^{20}Ne, on the basis of the IDP noble gas concentrations (*18*). These amounts are two to three orders of magnitude higher than implied by present rates of ^{3}He outgassing, integrated over 4.5×10^9 years, and the present ^{20}Ne inventory in the atmosphere. Chondritic material, and large objects, however, may have smaller concentrations of noble gases than IDPs, which are more uniformly exposed to solar radiation. On the other hand, IDPs may have received higher doses in the past.

If the atmospheric survivability rate for undegassed ET particles is 0.4% (as discussed above), then 16×10^{16} cm^3 of ^{3}He is buried, giving 3.6×10^7 cm^3 year^{-1} as the average ^{3}He flux from the mantle if spread over 4.5 Ga. Present observed and inferred rates of mantle ^{3}He flux are less than this lower bound. In sum, these calculations suggest that late accreting material is more than adequate to explain the ^{3}He flux from the mantle.

Discussion

I propose that subduction of IDP-rich sediments is the mechanism for introducing material with high ^{3}He/^{4}He ratios and presumed solar neon into the mantle. Magmas that sample this material yield basalts and gases with high ^{3}He/^{4}He ratios that have previously been attributed to deep mantle plumes. The sediment carrying high ^{3}He/^{4}He ratios may never get into the lower mantle. If the noble gases are stored in depleted, low-U peridotite, the ^{3}He/^{4}He ratio will decrease only gradually with time. Harzburgite and dunite xenoliths from the shallow mantle under Loihi and Samoa are enriched in noble gases and have high ^{3}He/^{4}He ratios (*38, 50, 58*); basalts from these hotspots also have high ^{3}He/^{4}He ratios.

If this hypothesis is correct the observation of ^{3}He in oceanic basalts cannot automatically be used as evidence of a primordial source or an ancient undegassed mantle. Even if there is still trapped primordial ^{3}He deep in the Earth's mantle, which is just now escaping to the atmosphere, the attribution of all ^{3}He to this source in degassing scenarios of the Earth may be dubious. Thus, high ^{3}He/^{4}He material may be a relatively recent addition to the Earth. ^{3}He anomalies, just as Ir anomalies (*65*), may be evidence for ET material.

Hot regions of the upper mantle, including the most active hotspots, occur where recent subduction cooling has not taken place (*66*). Current subduction cools the mantle and, in most places, introduces sediments with relatively low ^{3}He/^{4}He ratios into the mantle. Regions of the mantle unaffected by subduction will be hotter and will carry a more ancient subduction signature, particularly if capped by thick or unbroken lithosphere. Hotspots with high ^{3}He/^{4}He ratios may reflect the initial flushing out of the shallow mantle in such regions. This process would explain the association of high ^{3}He/^{4}He ratio magmas with major hotspots and the early products of Hawaiian volcanoes (*47*). However, the present model also predicts that basalts with high ^{3}He/^{4}He ratios will erupt in other places that tap the shallow mantle away from mature spreading ridges.

Although the concept of a primordial mantle has received much interest, there is little direct experimental, theoretical, or observational support for this idea. There is evidence for ancient depleted and enriched reservoirs (*67, 68*), recycling (*22, 23, 69*), and various enriched and depleted components in the upper mantle. Planetary assembly is thought to involve energetic accretion of hot particles. This process would seem to require wholesale melting, differentiation, and outgassing. It even appears that preplanetary material has been extensively heated and degassed (*70*). The Earth exhibits abundant evidence for a high temperature origin, early degassing, and efficient differentiation, including the presence of a magma ocean (*67*).

One model of mantle dynamics is that subducted crust and perhaps sediments are carried down to the base of the mantle, the presumed source of the plumes that are conjectured to provide hotspot magmas (*69*). The motivation for this long distance transport is the notion that most of the upper mantle is outgassed and depleted and that the noble gases can only be in the lower mantle. However, most of the incompatible and volatile-rich dehydration products of slab heating may remain in the shallow mantle. IABs and OIBs have similar trace element and isotopic characteristics. This similarity suggests that the recycled component is enriching the shallow mantle (EM). The devolatized oceanic crust could sink just to the base of the upper mantle (*66–68*).

If widely mixed into the shallow mantle, the recycled component would be available when the lithosphere thins or weakens or when this mantle is flushed out by ascending melts. The uppermost mantle is inhomogeneous because it contains various kinds of cycled and recycled material, of various ages and histories, and residual melts as well (*71*). This notion is generally accepted for the mantle wedge above subducting slabs and the continental lithosphere. These are the presumed mantle source regions for IABs and perhaps continental flood basalts, respectively. This shallow level enrichment may be a general characteristic of the asthenosphere (*68*), which imprints its trace element and isotopic characteristics onto hotspot or plume basalts that rise through it. This shallow cycling of sediments and dehydration products and deeper cycling of the refractory dehydration residues can be called bi-cycling. The shallow layer should be less homogeneous than deeper layers and material in it may have a shorter residence time. I predict that magmas and gases with high ^{3}He/^{4}He ratios may be found in extensional settings unrelated to hotspots or plumes. The Lau backarc basin may be such a place. It has basalts with high ^{3}He/^{4}He ratios (*72*).

The Samoa hotspot on the other side of an island arc may be related to lithospheric tearing caused by warping at the Tonga-Fiji arc, although it is usually considered to be the result of a deep mantle plume. High ^{3}He/^{4}He ratios are not expected in regions of recent subduction of modern sediments or where spreading has cleared out the shallow mantle (*35, 46*). The initial stages of crustal extension and continental flood basalt magmatism may also yield material with high ^{3}He/^{4}He and high ^{20}Ne/^{22}Ne ratios from the shallow mantle.

The subduction of ancient IDP-rich oceanic crust may have introduced a component into the mantle that has a high ^{3}He/^{4}He ratio and a moderate ^{87}Sr/^{86}Sr ratio. Such a signature has been recently identified in some hotspot magmas (*45*). The ET component also has a high ^{20}Ne/^{22}Ne ratio which I argue is also a relatively recent addition to the Earth. The atmospheric values of heavy noble gases in OIB follow naturally from the IDP-sediment mix. The presence of atmospheric, seawater, or sedimentary ratios of heavy noble gases in the primitive reservoir with high ^{3}He/^{4}He ratios has previously been another paradox associated with noble gas studies.

An atmospheric signature is invariably present in rare gas analyses of mantle samples. It is not clear when this signature was introduced. Explanations range all the way from an atmosphere-like plume or lower mantle component to contamination associated with eruption. The subduction of pelagic sediments unavoidably carries an atmosphere and seawater signal into the mantle, and this phenomenon has been much discussed in connection with many trace element and nonvolatile isotopic systems. The addition of He-rich and Ne-rich IDPs to the sedimentary inventory increases the power of the recycling hypothesis and explains why some rare gases in hotspot magmas appear to be atmospheric and others appear to be solar. This notion obviates the need for a separate undegassed reservoir to reconcile the seemingly disparate isotopic signals but does not rule out the possibility of recent atmospheric contamination that has nothing to do with the mantle.

Another puzzling aspect of basalts with high ^{3}He/^{4}He ratios, regardless of whether or not they are samples of primordial mantle, is their high and variable ^{40}Ar/^{36}Ar ratios (*50*). Pelagic sediments, and fluids from dehydrated oceanic crust, have high but variable K contents and K/U ratios compared with other mantle materials (*22*). If these are also the ultimate carriers of IDP noble gases, which metasomatize or contaminate shallow mantle depleted peridotites, with time, ^{40}Ar/^{36}Ar ratios will be high and variable. Recognizing that recycled material may be a mixture of ^{3}He-rich IDPs plus pelagic sediments makes it possible to understand the decoupling of He and Ne isotopes from Sr, Nd, and heavy noble gas isotopes; the nonprimordial Sr, Nd, and Pb isotopic signatures of high ^{3}He/^{4}He ratio material; and the highly variable ^{40}Ar/^{36}Ar ratio of hotspot basalts.

REFERENCES AND NOTES

1. J. E. Lupton and H. Craig, *Science* **214**, 13 (1981). Basalts with high ^{3}He/^{4}He ratios are called "primitive" or "primordial" because of the presumed excess of ^{3}He, a so-called primordial element ^{3}He is actually less primordial than other stable isotopes such as ^{86}Sr, ^{144}Nd, or ^{204}Pb. It is constantly added to the Earth by ET debris and cosmic rays. Variations in ^{3}He/^{4}He ratios of Earth materials reflect variations in initial ^{3}He/^{4}He ratios, production of ^{4}He from decay of U and Th, and contamination with ET and cosmogenic ^{3}He. Small amounts of ^{3}He are generated in and on the Earth by spallation and cosmogenic reactions. Although the hypothetical undegassed primitive mantle must have high integrated ^{3}He/(U+Th) (implying high initial ^{3}He and low U + Th or young age), hotspot basalts have nonchondritic ratios of other isotopes such as ^{87}Sr/^{86}Sr and ^{206}Pb/^{204}Pb. The conditions required to incorporate and preserve significant ^{3}He and other noble gases in the proto-Earth during its initial accretion, and to isolate a primordial reservoir for $>4 \times 10^9$ years, are unclear.
2. H. Craig *et al., Earth Planet. Sci. Lett.* **26**, 125 (1975).
3. G. Davies, *ibid.*, p. 125, D. Stevenson, *Lunar Planet. Sci.* **16**, 821 (1985); T. Ahrens, in *Origin of the Earth*, J. Jones and H. Newsom, Eds. (Lunar and Planetary Institute, Houston, TX, 1990). Recently it has been suggested that noble gas partitioning into the core is negligible [J. Matsuda *et al., Science* **259**, 788 (1993)].
4. T. Staudacher and P. Sarda, *Pour Sci.* **168**, 44 (1991); M. Honda, J. Reynolds, E. Roedder, S. Epstein, *J. Geophys. Res.* **92**, 12507 (1987), P. Sarda, T. Staudacher, C. Allègre, *Chem. Geol.* **70**, 40 (1988); *Terra Abstr.* **1**, 333 (1989); *ibid.* **3**, 492 (1991); R. Carlson, *Rev. Geophys.* **25**, 1011 (1987).
5. J. Lupton and H. Craig, *Earth Planet. Sci. Lett.* **26**, 133 (1975); M. D. Kurz, W. J. Jenkins, S. R. Hart, D. Clague, *ibid.* **66**, 388 (1983); J. E. Lupton, *Annu. Rev. Earth Planet. Sci.* **11**, 371 (1983); D. Graham, S. E. Humphris, W. J. Jenkins, M. D. Kurz, *Earth Planet. Sci. Lett.* **110**, 121 (1992); R. Poreda *et al., ibid.* **78**, 1 (1986). Although many hotspots do not have high ^{3}He/^{4}He ratios [H. Craig, *Eos* **71**, 1669 (1990)], locations with ratios in excess of MORB values are usually defined as hotspots or are assumed to be influenced by hotspots or deep mantle plumes.
6. A "depleted" mantle reservoir means that it has low concentrations of elements that are removed by melts or fluids and low isotopic ratios such as ^{86}Sr/^{86}Sr. "Enriched" means the opposite and often implies the presence of crust-like or sedimentary material.
7. F. T. Kyte and J. T. Wasson, *Science* **232**, 1225 (1986).
8. B. Esser and K. Turekian, *Geochim. Cosmochim. Acta* **52**, 1383 (1988).
9. E. Anders, *Nature* **342**, 255 (1989).
10. E. M. Shoemaker, *Annu. Rev. Earth Planet. Sci.* **11**, 461 (1983).
11. C. Merrihue, *Ann. N.Y. Acad. Sci.* **119**, 351 (1964).
12. S. Amari and M. Ozima, *Nature* **317**, 520 (1985); J. Matsuda, M. Murota, K. Nagao, *J. Geophys. Res* **95**, 7111 (1990).
13. F. P. Fanale and J. L. Kulp, *Econ. Geol.* **578**, 735 (1962); Futagami, M. Ozima, Y. Nakamura, *Earth Planet. Sci. Lett.* **101**, 63 (1990).
14. M. Ozima, M. Takayanagi, S. Zashu, and S. Amari [*Nature* **311**, 448 (1984)] measured ^{3}He/^{4}He ratios in ten piston cores of Quaternary and late Tertiary age sediments in the Pacific and two cores of Cretaceous age sediments from the Atlantic. The ^{3}He/^{4}He ratios were $>5 \times 10^{-5}$ at six sites, well above the values generally found in terrestrial materials, which they attributed to >0.1 ppm of ET debris in the sediments. The highest ^{3}He/^{4}He ratios were found at Pacific sites with low sedimentation rate. The Cretaceous sediments in Atlantic cores have very low ^{3}He/^{4}He ratios partly because sedimentation rates were high associated with a narrow ocean and partly because of possible diffusive loss of ^{3}He. Dust inputs from the Sahara into the Atlantic are at least 10^3 greater than the global input of ET debris. Conversely, ^{3}He/^{4}He ratios and ^{3}He contents are high in regions of low sedimentation and, presumably, during times and places of low oceanic productivity The high ^{3}He/^{4}He ratios found deep in the core show that ^{4}He build-up by radioactive decay has not eliminated the ET signal.
15. C. Merrihue (*11*) reported ^{3}He/^{4}He ratios of $>10^{-4}$ in ferromagnetic separates from deep ocean sediments and concluded that the high ratio was due to ET debris; A. Ya. Krylov *et al., Geochem Int* **10**, 202 (1973)
16. Total ^{3}He is the concentration in sediments (5×10^{-11} cm^3 g^{-1}) times average sediment thickness (300 m) times density (2 g cm^{-3}) times the area of oceans (3×10^{18} km^2) = 10^{13} cm^3 of ^{3}He Episodic cometary and meteoritic showers and the ^{3}He in large particles, not sampled by ocean cores, may considerably increase the total. The present atmospheric inventory of ^{3}He is about 3×10^{13} cm^3. Some of the sedimentary ^{3}He may leak into the ocean and be mistaken for mantle outgassing. The rest recycles into the mantle at subduction zones or accretes to the edges of continents.
17. M Takayanagi and M Ozima, *J. Geophys Res* **92**, 12,531 (1987).
18. A O. Nier, D. J. Schlutter, D E Brownlee, *Geochim Cosmochim Acta* **54**, 173 (1990)
19. Icecap micrometeorites include many unmelted large grains (>100 μm) and unmelted chondritic fragments [M. Maurette *et al., Nature* **351**, 44 (1991)]. These observations plus the short timescale of the high-temperature entry suggest that much more ^{3}He-rich material may be delivered to the Earth's surface than commonly assumed.
20. B. Marty and M. F. Le Cloarec, *J. Volcanol. Geotherm. Res* **53**, 67 (1992)
21. The rate of outgassing by volcanism has varied considerably. The Cretaceous was a time of extensive volcanism. Pre-Cretaceous carbonate accumulated in shallow water and pelagic carbonate accumulation was minor. Since then there has been a gradual shift from shallow water to deep water accumulation such that the recent rate of pelagic carbonate accumulation, and IDP dilution, is the highest of any past epoch [B. H. Wilkinson and J. G C. Walker, *Am J. Sci.* **289**, 525 (1992)]. The increased subduction of carbonates increases the volcanic CO_2 flux and rate of outgassing [K. Caldeira, *Nature* **357**, 578 (1992)], including, presumably, noble gases
22. B. L. Weaver, *Earth Planet. Sci. Lett.* **104**, 381 (1991); *Geology* **19**, 123 (1991)
23. C Chase, *Earth Planet. Sci. Lett.* **52**, 277 (1981). A. W Hofmann and W. M. White [*ibid.* **57**, 1 (1982)] proposed that hotspots were formed from deep mantle plumes that involved ancient recycled oceanic crust and sediment. They speculated that higher concentrations of the heat-producing elements U and Th might be the cause of plumes. This would also generate ^{4}He, which would lower the ^{3}He/^{4}He ratio with time. Magmas with high ^{3}He/^{4}He ratios magmas probably come from mantle sources with low integrated U and Th abundances compared with the MORB source.
24. R. von Huene and D. Scholl, *Rev. Geophys.* **29**, 279 (1991).
25. S. M. Peacock, *Science* **248**, 329 (1990).
26. H. Hiyagon and M. Ozima, *Geochim Cosmochim. Acta* **50**, 2045 (1986).
27. I. N. Tolstikhin, V. Dokuchaeva, I. Kamensky, Yu Amelin, *ibid.* **56**, 987 (1992); S. Azuma, M. Ozima, H Hiyagon, *Earth Planet. Sci Lett.* **114**, 341 (1993).
28. C. L. Broadhurst, M. J. Drake, B. Hagee, T.

Bernatowicz, *Geochim. Cosmochim. Acta* **56**, 709 (1992).
29. The magnetic separates with high ^{3}He ratios are thought to be secondary magnetite, formed upon atmospheric entry of IDP (*12*). ^{10}Be in IABs suggests that sediments are able to subduct to depths of the order of 100 km. The temperatures at the top of the slab at such depths are of the order of 500°C for rapidly subducting slabs (*25*). Modern seawater and pelagic sediments are relatively high in U. Thus, ^{4}He is produced in the sediments before and after subduction and, if retained, will lower the ^{3}He/^{4}He of the sediments. The radiogenic ^{4}He, however, is sited differently than the trapped IDP He and may be more mobile or more prone to flushing out by hydrous fluids (*27*). Many ancient rocks appear to have retained their trapped ^{3}He but lost the radiogenic helium.
30. W. Fyfe, in *The Handbook of Environmental Chemistry*, O. Hutzinger, Ed. (Springer-Verlag, Berlin, 1992), vol. 1, pp. 1–26.
31. J. Morris, W. Leeman, F. Tera, *Nature* **344**, 31 (1990).
32. M. Cloos and R. Shreve, *Palaeograph. Palaeoclimatol. Palaeoecol.* **128**, 501 (1988). Most western Pacific margins are not fronted by accretionary prisms, indicating that the oceanic sediments are not scrapped off but are subducted beneath the overriding plate. Modern examples of sediment-starved trenches include Mariana, Philippine, Tonga, South Sandwich, Kermadec, and Northern Chile.
33. D. Vance, J. Stone, R. O'Nions, *Earth Planet. Sci. Lett.* **96**, 147 (1989).
34. C. F. You *et al.*, *Geology* **21**, 207 (1993).
35. R. J. Poreda and H. Craig, *Nature* **338**, 473 (1989). IABs have ^{3}He/^{4}He ratios comparable to or slightly less than those of MORBs. The present influx rate of IDPs is so small and the present sedimentation rate is so high that subduction of modern sediments does not significantly perturb the noble gas content of modern island arc magmas. Nevertheless, even today, some ^{3}He is entering the mantle by way of IDPs.
36. A. Moran, V. Sisson, W. Leeman, *Earth Planet. Sci. Lett.* **111**, 331 (1992); O. Navon and E. Stolper, *J. Geol.* **95**, 285 (1987).
37. R. C. Maury, M. J. Defant, J.-L. Joron, *Nature* **360**, 661 (1992).
38. R. J. Poreda and K. A. Farley, *Earth Planet Sci. Lett.* **113**, 129 (1992).
39. W. Rison and H. Craig, *Eos* **63**, 1144 (1982).
40. R. Armstrong, *Aust. J. Earth Sci.* **38**, 613 (1991).
41. Basalts are commonly thought to represent up to about 10% melting of a mantle source. If pelagic sediments are mixed into a mantle reservoir they will contribute a disproportionate amount of volatile and incompatible elements to the initial melts. In this case a smaller sediment and IDP content will suffice to explain the high ^{3}He/^{4}He ratio of basalts. In the limit, the sediment mixed into a refractory peridotitic upper mantle can be treated as a low–melting temperature component, and the highly incompatible elements that are added are quantitatively removed in the first melt fraction and are diluted by subsequent melting. The ^{3}He/^{4}He ratio of the mixture will decrease with time as a result of the radioactive decay of U and Th, which produces ^{4}He, depending on the U and Th content of the sediments. The radioactively generated excess ^{4}He may be sited differently than the IDP ^{3}He and may be removed or redistributed by hydrous fluids (*27, 33*).
42. The isotopic composition of most deep oceanic sediments (authigenic and carbonate) is controlled by the properties of seawater for Nd, Sr, and Pb isotopes. Present-day seawater and sediments are dominated by the continental component and are therefore isotopically enriched. Recycling of such sedient into the mantle introduces an enriched component, but this has not always been the case. During periods of low continental runoff or periods having a low cratonic component in river water or periods of high hydrothermal activity, sediments can inherit a depleted isotopic signature. Modern seawater is particularly enriched and Precambrian seawater was particularly depleted; see S. B. Jacobsen, *Geochim. Cosmochim. Acta* **52**, 1341 (1988); *Earth Planet Sci. Lett.* **90**, 315 (1988).
43. A. Derry and S. B. Jacobsen, *Geochim. Cosmochim. Acta* **54**, 2965 (1990).
44. D. J. Des Marais, H. Strauss, R. E. Summons, J. M. Hayes, *Nature* **359**, 605 (1992).
45. K. A. Farley, J. H. Natland, H. Craig, *Earth Planet. Sci. Lett.* **111**, 183 (1992); S. R. Hart, E. H. Hauri, L. A. Oschmann, J. A. Whitehead, *Science* **256**, 517 (1992). The isotopic characteristics of the high ^{3}He/^{4}He component can be estimated to fall in the ranges ^{87}Sr/^{86}Sr (0.7040 to 0.7055), ^{143}Nd/^{144}Nd (0.5126 to 0.5128), ^{206}Pb/^{204}Pb (18.4 to 19.2), ^{207}Pb/^{204}Pb (15.50 to 15.57), ^{208}Pb/^{204}Pb (38.4 to 39.2), and ^{3}He/^{4}He (35 R_a to >50 R_a). Except for He, these characteristics are similar to IABs and metasomatic fluids in the subarc mantle (*37*), which undoubtedly contain a recycled component. This mantle component has been designated FOZO, PHEM, or EH. I suggest that the noble gas–rich component be called ET. See also J. Barling and S. Goldstein, *Nature* **348**, 59 (1990).
46. The survival time of a shallow mantle enriched reservoir is probably related to the migration rate of ridges, which flushes out chemical inhomogeneities, as they sweep across the mantle. Consider a 40,000-km-long ridge system migrating at 1 to 2 cm year^{-1}. Each part of the mantle will then be sampled every 1 to 2 billion years. This scenario may explain why basalts with high ^{3}He/^{4}He ratios are usually early in the eruption sequence. The presence of high ^{3}He/^{4}He material at ridge-straddling hotspots (Iceland) and long-lived midplate hotspots (Hawaii) is probably due to entrainment of the shallow mantle by the plate, bringing fresh shallow mantle toward the locus of magmatism. Eventually, the shallow mantle is replaced by deeper depleted mantle.
47. M. Kurz and D. Krammer, *Earth Planet. Sci. Lett.* **103**, 257 (1991); M. Kurz *et al.*, *ibid.* **66**, 388 (1983).
48. C. Martin, *Geochim. Cosmochim. Acta* **55**, 1421 (1991); J. Greenough and J. V. Owen, *Chem. Geol.* **98**, 203 (1992); Also komatiites: N. Arndt, *Tectonophysics* **187**, 411 (1991); G. Brugmann *et al.*, *Geochim. Cosmochim. Acta* **51**, 2159 (1987).
49. H. Craig and J. Lupton, *Earth Planet. Sci. Lett.* **31**, 369 (1976).
50. H. Fukumoto, K. Nagao, J. Matsuda, *Geochim. Cosmochim. Acta* **50**, 2245 (1986); S. Amari and M. Ozima, *ibid.* **52**, 1087 (1988).
51. The He and Ne isotopes in IDPs do not appear to be of solar wind origin. Nier *et al.* (*18*) noted the similarity of IDP gases to those in carbonaceous chondrite and gas-rich meteorites. There may be some solar flare He and Ne seen in the high-temperature extractions.
52. A. O. Nier and D. Schlutter, *Meteoritics* **25**, 263 (1990).
53. M. Honda, I. McDougall, D. P. Patterson, A. Doulgeris, D. A. Clague, *Nature* **349**, 149 (1991).
54. H. Staudigel *et al.*, *Earth Planet. Sci. Lett.* **69**, 13 (1984). See also (*5, 38*).
55. Midocean ridge basalts contain about 0.5 to ~5 × 10^{12} cm^3 g^{-1} of ^{21}Ne [H. Hiyagon and B. Kennedy, *Geochim. Cosmochim. Acta* **56**, 1569 (1992)]. A typical MORB U content of 0.014 ppm yields 1.3 × 10^{12} cm^3 g^{-1} of ^{21}Ne in 4.5 × 10^9 years, within the range of observed values. For this U content there will be 1.4 × 10^{-5} cm^3 g^{-1} of ^{4}He generated compared with typical MORB values of 1.2 × 10^{-5} cm^3 g^{-1}. Unless the MORB reservoir was completely outgassed, there will also be some original ^{21}Ne and ^{4}He in the MORB reservoir and, of course, some gas may have been lost before eruption. Nevertheless, the agreement is remarkable and suggests that the MORB reservoir may have been ancient and has been protected from outgassing and contamination for a long period of time.
56. By OIB reservoir I mean the trace element and isotopic components that represent recycled material or material transferred by fluids or melts from the slab to the mantle or from one part of the mantle to another. This reservoir may be infertile, that is, lacking a basaltic component, and refractory (dunite or harzburgite). Just as a subducted slab may lose its surficial and volatile components at shallow depth, deeper depleted magmas may acquire these components as they rise.
57. T. Staudacher and C. Allègre, *Earth Planet. Sci. Lett.* **89**, 173 (1988).
58. W. Rison and H. Craig, *ibid.* **66**, 407 (1983).
59. E. R. Oxburgh and R. K. O'Nions, *Science* **237**, 1583 (1987).
60. M. Ozima and F. Podosek, *Noble Gas Geochemistry* (Cambridge Univ. Press, Cambridge, 1983).
61. N. H. Sleep, K. J. Zahnle, J. F. Kasting, H. J. Morowitz, *Nature* **342**, 139 (1989).
62. C. F. Chyba, P. J. Thomas, L. Brookshaw, C. Sagan, *Science* **249**, 366 (1990).
63. E. Anders and T. Owen, *ibid.* **198**, 453 (1977).
64. C. L. Chou, *Proc. Lunar Planet Sci. Conf.* **9**, 219 (1978); J. W. Morgan, J. Wandless, R. Petrie, A. J. Irving, *Tectonophysics* **75**, 47 (1981). If the mantle were ever in equilibrium with the core or with molten iron, it should have negligible siderophile concentrations. The excess siderophiles in mantle rocks motivated the idea of a late veneer (post-core formation). Mantle regions with excess siderophiles are presumably not involved in whole mantle convection and are not in contact with the core.
65. C. Chou, D. Shaw, J. Crocket, *J. Geophys. Res* **88**, A507 (1983).
66. C. Scrivner and D. L. Anderson, *Geophys. Res. Lett.* **19**, 1053 (1992); D. L. Anderson, T. Tanimoto, Y.-S. Zhang, *Science* **256**, 1645 (1992).
67. D. L. Anderson, *Theory of the Earth* (Blackwell Scientific, Boston, 1989).
68. ______, *Science* **213**, 82 (1981); *Earth Planet. Sci. Lett.* **57**, 1 (1982).
69. W. M. White and A. W. Hofmann, *Nature* **296**, 821 (1982).
70. M. Gaffery, in *Origin of the Earth*, N. E. Newsom and J. Jones, Eds. (Oxford Univ. Press, New York, 1990), p. 17; S. Taylor and M. Norman, in *ibid.*, p. 29.
71. D. L. Anderson, *Nature* **318**, 145 (1985).
72. R. Poreda, *Earth Planet. Sci. Lett.* **73**, 244 (1985).
73. P. Sarda, T. Staudacher, C. Allègre, *ibid.* **91**, 73 (1988).
74. M. Ozima and S. Zashu, *Science* **219**, 1067 (1983).
75. R. Poreda, J.-G. Schilling, H. Craig, *Earth Planet. Sci. Lett.* **78**, 1 (1986).
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