

CONSTRAINTS ON THE CHARACTERISTICS OF MAGMA SOURCES FOR HAWAIIAN VOLCANOES BASED ON NOBLE-GAS SYSTEMATICS

By Ichiro Kaneoka¹

ABSTRACT

The implications of noble-gas-isotope signatures of Hawaiian volcanic rocks and ultramafic nodules for the characteristics of Hawaiian volcanism are discussed. Hawaiian volcanic rocks have higher $^3\text{He}/^4\text{He}$ ratios and lower $^{40}\text{Ar}/^{36}\text{Ar}$ ratios than MORB, strongly indicating that they derived from a source enriched in volatile and primordial components. That source is probably located in the deeper part of the Earth's interior (lower mantle?). These observations are compatible with the hot-spot hypothesis.

The variations in the observed $^3\text{He}/^4\text{He}$ and $^{40}\text{Ar}/^{36}\text{Ar}$ ratios for Hawaiian volcanic rocks and nodules can be explained in terms of varying degrees of mixing among noble-gas components of three types: *P*(plume), *M*(MORB), and *A*(atmosphere). Mixing of *P* and *M* components occurs during ascent of a diapir through the asthenosphere and lithosphere, followed by introduction of the *A* component during eruption.

Systematic variations in the $^3\text{He}/^4\text{He}$ ratios with the volume of individual volcanoes, together with other radiogenic-isotope signatures, indicate a possible model for the evolution of Hawaiian volcanism based on the rise of volatile-enriched diapirs from the Earth's deep interior. However, the amount of noble-gas (volatile) components would vary at each stage of a volcano's evolution. The model suggests that the partially melted zone plays an important role during the evolution of Hawaiian volcanism. It is further pointed out that the voluminous outpourings of Hawaiian volcanoes do not always reflect the direct material supply from the hot-spot source but may reflect the combined effects of material transfer and heat supply.

INTRODUCTION

Because the Hawaiian Islands have formed far from any lithospheric plate boundary, they provide a good opportunity to study the characteristics of intraplate volcanism, especially of hot-spot areas as identified by Wilson (1963) and Morgan (1971).

Noble gases are chemically inert and include both radiogenic and stable isotopes. Furthermore, their isotopic ratios are rather insensitive to the physical and chemical processes occurring during magmatic fractionation. Hence, the isotopic ratios of noble gases can give us unique information concerning the characteristics of source materials. In this paper, I summarize the noble-gas data so far obtained for Hawaiian samples and discuss the characteristics of source materials for Hawaiian rocks based on those noble-gas data.

Dalrymple and Moore (1968) identified that recent submarine pillow basalts erupted on the east rift zone of Kilauea showed excess ^{40}Ar . Similar results were obtained for the same kind of samples by Noble and Naughton (1968). It has been realized that magmatic Ar is trapped in glassy rims because of rapid cooling under a high hydrostatic pressure of sea water, which hinders the degassing. Compared to the atmospheric values of 295.5, the $^{40}\text{Ar}/^{36}\text{Ar}$ ratio in these samples is as high as about 1,200. However, these ratios are not so high as those observed in some glassy pillow samples of midocean ridge basalt (MORB), which reach as high as about 10,000 (Funkhouser and others, 1968).

Funkhouser and Naughton (1968) examined radiogenic He and Ar in ultramafic inclusions from Hualalai Volcano, Hawaii, and Salt Lake Crater, Oahu, and concluded that they include magmatic radiogenic components in fluid and gaseous inclusions in the minerals of the nodules. This confirmed the previous results by Lovering and Richards (1964), who suggested the presence of excess ^{40}Ar in clinopyroxene from a pyroxenite nodule erupted at Salt Lake Crater that showed an apparently very old K-Ar age of 1.4 Ga. The radiogenic components in these ultramafic nodules were further investigated by Gramlich and Naughton (1972), who inferred, on the basis of diffusion experiments of He and Ar, that the $^4\text{He}/^{40}\text{Ar}$ ratio in the environment where the minerals of ilherzolite nodules were concentrated at Salt Lake Crater had a value of 1.2 ± 0.2 . These investigations demonstrate the occurrence of radiogenic noble gases such as ^4He and ^{40}Ar of magmatic origin in pillow glass and ultramafic nodules. At that time, these data were interpreted to show the characteristics of noble-gas isotopes in the Earth's interior in a general sense, and it was premature to discuss the characteristics of noble-gas isotopes that might be unique for Hawaiian samples.

The value of noble-gas isotope studies on Hawaiian samples was illustrated by Hennecke and Manuel (1975) when they found excess ^{129}Xe in ultramafic nodules. Furthermore, Craig and Lupton (1976) found a higher $^3\text{He}/^4\text{He}$ ratio in Kilauea fumaroles than in those of MORB, which is known to be relatively uniform on a global scale. Such data are significant because they might characterize the uniqueness of Hawaiian magma sources. The occurrence of high $^3\text{He}/^4\text{He}$ ratios and excess ^{129}Xe in Hawaiian samples was confirmed in both lava flows and ultramafic nodules (Kaneoka and

¹Geophysical Institute, University of Tokyo, Bunkyo-ku, Tokyo 113, Japan.

Takaoka, 1978; Kaneoka and others, 1978). On the basis of systematic differences in $^3\text{He}/^4\text{He}$ and $^{40}\text{Ar}/^{36}\text{Ar}$ ratios, Kaneoka and Takaoka (1980) discussed the difference of magmatic sources between lava flows and ultramafic nodules. Further data on these samples were accumulated, discussed, and compared with similar data from other areas (Kyser and Rison, 1982). Detailed studies on samples from Loihi Volcano have revealed that high $^3\text{He}/^4\text{He}$ ratios for Hawaiian magma sources are real but rather variable even between different eruptions of the same volcano (Kaneoka and others, 1983; Kurz and others, 1983; Rison and Craig, 1983). However, no excess ^{129}Xe has been found in samples of Loihi pillow glass (Allègre and others, 1983; Kaneoka and others, 1983). Furthermore, large variation in the $^3\text{He}/^4\text{He}$ ratio is found even for different volcanoes on the same island (Kurz and others, 1983). These data could indicate more complexity in magmatic sources for Hawaiian magmatism, which however still remain distinct from those of MORB.

ACKNOWLEDGMENTS

For my experimental work on noble-gas systematics of Hawaiian samples, I have benefited much from using a mass spectrometer developed by N. Takaoka of Yamagata University. R. W. Decker of the U.S. Geological Survey and K. Nakamura of the Earthquake Research Institute, University of Tokyo, have given me the chance to write this review paper and gave encouragement through the work. The constructive reviews by G. B. Dalrymple and T. L. Wright of U.S. Geological Survey were very helpful for improving the manuscript.

This study was financially supported in part by a Grant-in-Aid for Scientific Research from the Ministry of Education, Science and Culture, Japan, under contracts No. 58540220 and No. 59065001.

ABUNDANCES AND TRAPPING SITES OF NOBLE GASES IN HAWAIIAN ROCK SAMPLES

Noble-gas abundances in a sample reflect both the characteristics of the source and the sample's subsequent history (see, for example, Dymond and Hogan, 1978; Kaneoka and Takaoka, 1985). If we can evaluate the latter contribution properly, it is possible to get some information on the characteristics of the sample source, though it is generally not easy to do so. Noble gases in selected samples of different kinds from the Kilauea region are compared in table 27.1. Since these samples can be regarded as almost recent, it is not necessary to consider the effect of radiogenic components formed after extrusion on the surface. Hence, these samples should show similar isotopic ratios and characteristic abundances of noble gases reflecting the processes involved in their magmatic history. Judging from the $^{40}\text{Ar}/^{36}\text{Ar}$ ratio and relative noble-gas abundance patterns, Kapoho lava seems to have lost most magmatic gas and to have been equilibrated with atmospheric noble gases. The observed amount of ^{40}Ar in this sample (2.0×10^{-7} cm³/g at standard temperature and pressure) is just the value expected in a molten magma in equilibrium with the Earth's

atmosphere as inferred from the solubility data of Ar in melt of tholeiitic basalt (Hayatsu and Waboso, 1982). However, there still remain small amounts of magmatic noble gases in this sample, which is shown by the relative enrichment of ^4He and ^{132}Xe compared with Ar (see the data for lava in figure 27.1).

On the other hand, olivine phenocrysts (diameter >1 mm) in this sample contain lesser amounts of noble gases (except for ^4He) than lava, but they show definitely higher $^3\text{He}/^4\text{He}$ and $^{40}\text{Ar}/^{36}\text{Ar}$ ratios than the atmosphere. These phenocrysts are considered to have been formed in the magma reservoir at depth without atmospheric contaminations. Hence, these values can be considered to approximate the values for the magmatic noble gases, or at least for their isotopic ratios.

It has been known that the chilled margins of submarine basalt pillows contain magmatic noble gases (see, for example, Dalrymple and Moore, 1968; Funkhouser and others, 1968; Noble and Naughton, 1968). However, the efficiency of trapping of noble gases is dependent on both the cooling rate and the ambient hydrostatic pressure. As shown in table 27.1, tholeiite from the east rift zone reflects such tendencies (Kyser and Rison, 1982). The sample from the derived depth of 2,960 m clearly shows the evidence of magmatic noble gases, but that from the depth of 30 m does not, though the latter seems to retain a small amount of ^4He .

In a subaerial sample (glassy spatter), the observed amounts of noble gases are quite large, but the $^{40}\text{Ar}/^{36}\text{Ar}$ ratio is atmospheric and the noble-gas abundance pattern is similar to that of the atmosphere (fig. 27.1). Hence, it is not likely that these gases truly represent those in the original magma. Many microscopic gas pores are observed in this sample; atmospheric noble gases were probably trapped in such pores when it erupted.

The high $^3\text{He}/^4\text{He}$ ratio reported in Kilauea gas samples (Craig and Lupton, 1976) is similar, within their analytical uncertainties, to those observed in phenocrysts and pillow glass from this region. However, we have no guarantee that the observed values for fumarolic gases approximate those of the magmatic gases for heavier noble gases, since they are much more easily contaminated from the atmosphere (see, for example, Ozima and Podosek, 1983; Kaneoka and Takaoka, 1985).

Thus it is essential in noble-gas analyses to select appropriate samples to infer the conditions of magmatic noble gases. For this purpose, pillow glass and large olivine and pyroxene phenocrysts have been regarded as suitable. Ultramafic nodules also contain magmatic noble gases, since they are formed at depth without atmospheric contaminations. For $^3\text{He}/^4\text{He}$ analyses, fumarolic gas and even water samples can be used, because the effect of atmospheric contamination is generally quite small for He compared with the abundance of magmatic He in these samples.

Noble-gas abundances in different kinds of samples from the Hawaiian region are compared in figure 27.1. To exclude the ambiguity from the addition of secondary atmospheric components, low-temperature (500–700 °C) and high-temperature (1,700–1,800 °C) fractions were taken for most samples, and the values for the higher temperature fraction are shown in figure 27.1. These samples include some from Loihi Seamount and from Hawaii, Maui, and Oahu Islands. Except for some quite different

TABLE 27.1—Comparison of noble gases in different kinds of samples from Kilauea Volcano

[Values for gas contents are at standard temperature and pressure; atmospheric values calculated from total amount in atmosphere divided by mass of the Earth; R/R_A given by $(^3\text{He}/^4\text{He})_{\text{sample}}/(^3\text{He}/^4\text{He})_{\text{atm}}$; —, not determined]

Sample	Noble-gas content					Ratios			Reference
	$(\text{cm}^3/\text{g} \times 10^{-8})$			$(\text{cm}^3/\text{g} \times 10^{-10})$		$^3\text{He}/^4\text{He}$ ($\times 10^{-6}$)	R/R_A	$^{40}\text{Ar}/^{36}\text{Ar}$	
	^4He	^{20}Ne	^{36}Ar	^{84}Kr	^{132}Xe				
Kapoho lava063	.0111	.0674	.256	.0796	<226	<161	297.0±1.6	1
Olivine phenocrysts in Kapoho lava	4.43	.0100	.0245	.0476	.0058	25.1±3.4	17.9±2.4	732.0±2.2	2
Glassy splatter on Kilauea caldera floor, erupted November 29, 1975	2.46	4.25	9.76	26.9	.682	—	—	294.0±1.4	3
Pillow glass Tholeiite of east rift zone, depth=30 m	1.9	.0103	.039	.084	.0058	0±14	0±10	296±22	4
Tholeiite of east rift zone, depth=2,960 m	32.6	.129	.213	.473	.0245	17±2	12±1	1175±22	4
Kilauea gas	—	—	—	—	—	20.9±1.1	14.9±0.8	—	5
Atmosphere347	1.09	2.08	4.31	.152	1.399	1	295.5	

1. Kaneoka and Takaoka (unpublished data).

2. Kaneoka and Takaoka (1978).

3. Kaneoka and others (1978).

4. Kyser and Rison (1982).

5. Craig and Lupton (1976).

samples (lava and glassy spatter in table 27.1), the samples show generally similar noble-gas abundance patterns, in which high enrichment of He is significant. Among different kinds of samples, lighter noble gases (He, Ne) are generally more enriched in ultramafic nodules and glassy samples than in phenocrysts. Nodule samples include dunite, ilherzolite, and pyroxenite; however, their noble-gas abundance patterns are similar.

Noble gases are generally more abundant in dunite nodules than in olivine phenocrysts. Numerous microscopic gas or liquid inclusions (probably CO_2) are observed in such ultramafic nodules, and most noble gases are expected to have been entrapped in such inclusions. This inference has been supported by crushing or decrepitation of samples (see, for example, Funkhouser and Naughton, 1968), which showed that rather large amounts of He and radiogenic ^{40}Ar were released. Furthermore, it has been demonstrated that most He is included in vesicles and not dissolved in the glass itself in pillow-glass samples from Loihi Seamount (Kurz and others, 1983; Rison and Craig, 1983). Hence, the noble-gas abundances in these samples would correspond to those in inclusions or vesicles.

Phenocrysts seem to contain fewer inclusions than nodules and glassy samples, and this probably affects the noble-gas abundances. Olivine phenocrysts from Hawaii contain amounts of noble gases comparable to those in samples from Reunion Island, whereas samples from Iceland contain lesser amounts of the lighter noble gases than those from Hawaii and Reunion (Kaneoka and Takaoka,

1985). This difference may reflect different tectonic environments for extrusion of magma; the samples from Hawaii may be closer to the original abundances of noble gases in the magma reservoir than those from Iceland.

Noble-gas abundances in MORB are also shown in figure 27.1 for comparison (plotted to the right of the Hawaii data). MORB samples are typically enriched in lighter noble gases, especially He. There is no reason to believe that U(+Th) is more abundant in the MORB source than in that of Hawaii, and such enrichment probably reflects secondary processes that occurred during the formation and transportation of MORB magma to the surface.

NOBLE-GAS ISOTOPIC CHARACTERISTICS OF HAWAIIAN SAMPLES

HELIUM ISOTOPES

As noted in the previous section, the concentrations of noble gases in a sample are susceptible to secondary processes and it is therefore useful to evaluate such processes. To infer the characteristics of magma sources, however, isotopic ratios are more useful than elemental concentrations, since they are much less affected by the processes of formation, transportation, and extrusion of a magma, except for the effects of mass fractionation and mixing among different sources. Among the many aspects of noble-gas isotope systematics, the $^3\text{He}/^4\text{He}$ ratio ($=R$) has been most

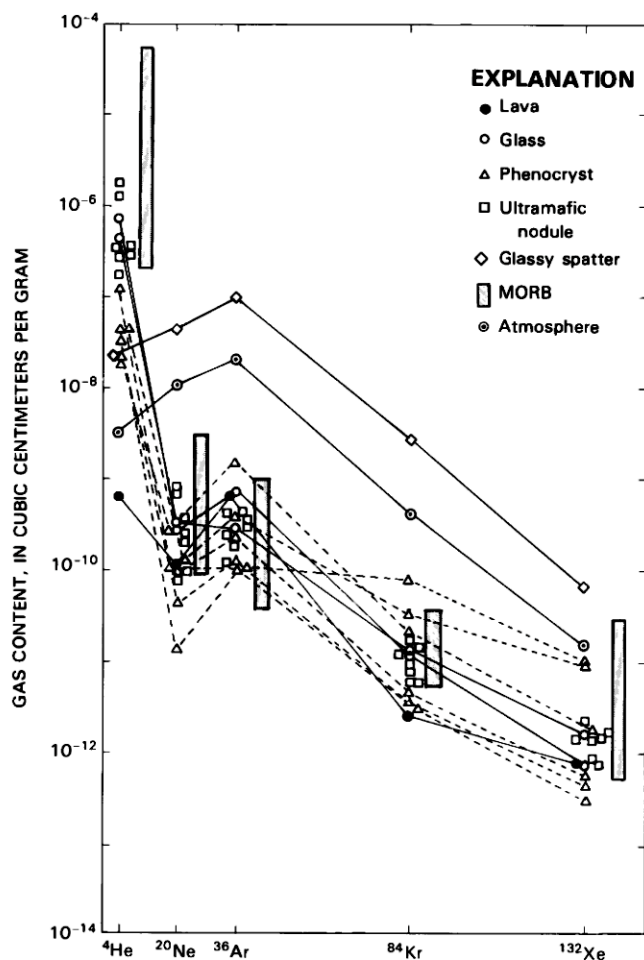


FIGURE 27.1.—Noble-gas abundances in Hawaiian rock samples. Values plotted are total amounts of noble gases in the samples adjusted to standard temperature and pressure. Values for atmosphere derived from total amounts of noble gases in the atmosphere divided by mass of the Earth. Data sources for Hawaiian samples: Hennecke and Manuel (1975); Kaneoka and Takaoka (1978, 1980); Kaneoka and others (1983). Data sources for MORB: Kyser and Rison (1982); Allègre and others (1983); Ozima and Zashu (1983).

successfully used so far because He is least contaminated by the atmosphere.

It has been revealed that MORB shows a rather uniform $^3\text{He}/^4\text{He}$ ratio that varies from 1.1×10^{-5} to 1.4×10^{-5} ($R = 8-10R_A$, where R_A is the ratio of $^3\text{He}/^4\text{He}$ for air) irrespective of regional differences and elemental fractionation, whereas other samples show more variability and some systematic differences. For example, crustal gases generally show a lower $^3\text{He}/^4\text{He}$ ratio than that of atmospheric He ($R_A = 1.4 \times 10^{-6}$). Examples are shown in figure 27.2, where the $^3\text{He}/^4\text{He}$ ratio is plotted against the $^4\text{He}/^{20}\text{Ne}$ ratio for each sample. The $^3\text{He}/^4\text{He}$ ratio for MORB is rather uniform, though the $^4\text{He}/^{20}\text{Ne}$ ratio varies by a factor of more than 10^4 . Such uniformity in the $^3\text{He}/^4\text{He}$ ratio for MORB surely reflects a characteristic of the magma source.

Hawaiian samples show rather different characteristics from those of MORB in figure 27.2. Although most ultramafic nodules show similar $^3\text{He}/^4\text{He}$ ratios to that of MORB, which has an $^4\text{He}/^{20}\text{Ne}$ ratio of $0.5 \times 10^3-5 \times 10^3$, phenocryst samples have higher $^3\text{He}/^4\text{He}$ ratios. This indicates that Hawaiian magma sources are isotopically different from that of MORB.

Detailed studies of the $^3\text{He}/^4\text{He}$ ratio have revealed that it varies among different volcanoes on each Hawaiian island and even within the individual volcanoes (Kaneoka and Takaoka, 1980; Kurz and others, 1983; Rison and Craig, 1983). The geographical variation of the $^3\text{He}/^4\text{He}$ ratio for Hawaiian samples is summarized in figure 27.3.

Such variations in the $^3\text{He}/^4\text{He}$ ratio surely reflects the conditions in the magma sources for these volcanoes. Furthermore, the variation in the $^3\text{He}/^4\text{He}$ ratio seems to be related to the rock type and the sequence of volcano formation. For example, tholeiite from Loihi Seamount shows higher $^3\text{He}/^4\text{He}$ ratios than alkali basalt (Kurz and others, 1983; Rison and Craig, 1983). Kurz and others (1983) further reported that the $^3\text{He}/^4\text{He}$ ratio for basalt on the Island of Hawaii seems to decrease with either increase in the volume of a volcano or decrease in the eruption age. Such observations are significant for interpreting Hawaiian volcanism and will be discussed later in more detail.

ARGON ISOTOPES

It has been known that ultramafic nodules from Hawaii, such as those from Hualalai and the Salt Lake Crater region, contain large amounts of excess ^{40}Ar , resulting in high $^{40}\text{Ar}/^{36}\text{Ar}$ ratios of more than a few thousand (see, for example, Funkhouser and Naughton, 1968). Since these ultramafic nodules are mantle derived (they are mostly cumulates with minor amounts of mantle materials), such values have sometimes been considered to represent the $^{40}\text{Ar}/^{36}\text{Ar}$ value in the mantle. Although submarine glassy pillows from the east rift zone of Kilauea show $^{40}\text{Ar}/^{36}\text{Ar}$ ratios as high as about 1,200 (see, for example, Dalrymple and Moore, 1968; Kyser and Rison, 1982), these $^{40}\text{Ar}/^{36}\text{Ar}$ ratios are still much lower than those observed in MORB. Furthermore, Kaneoka and Takaoka (1980) have demonstrated that higher $^3\text{He}/^4\text{He}$ ratios that exceed the MORB value seem to be associated with lower $^{40}\text{Ar}/^{36}\text{Ar}$ ratios. Hence, there is a possibility that such relatively low $^{40}\text{Ar}/^{36}\text{Ar}$ ratios represent the characteristics of the source materials for hot-spot areas.

The most serious problem in the measurement of the $^{40}\text{Ar}/^{36}\text{Ar}$ ratio is that the value is easily modified by the addition of atmospheric components, since the atmosphere (and sea water) contains fairly large amount of Ar. For this reason, it is often argued that the observed low $^{40}\text{Ar}/^{36}\text{Ar}$ ratios that are closer to the atmospheric value were caused by atmospheric contamination (see for example, Fisher, 1983). Such effects should be properly evaluated to infer the significance of the Ar isotopic ratios. One solution is to make comparison with other isotopic ratios that are not seriously affected by the atmospheric contamination. Another solution is to check the observed data in a plot of $^{40}\text{Ar}/^{36}\text{Ar}$ against $1/^{36}\text{Ar}$, in which samples contaminated with atmospheric component should be

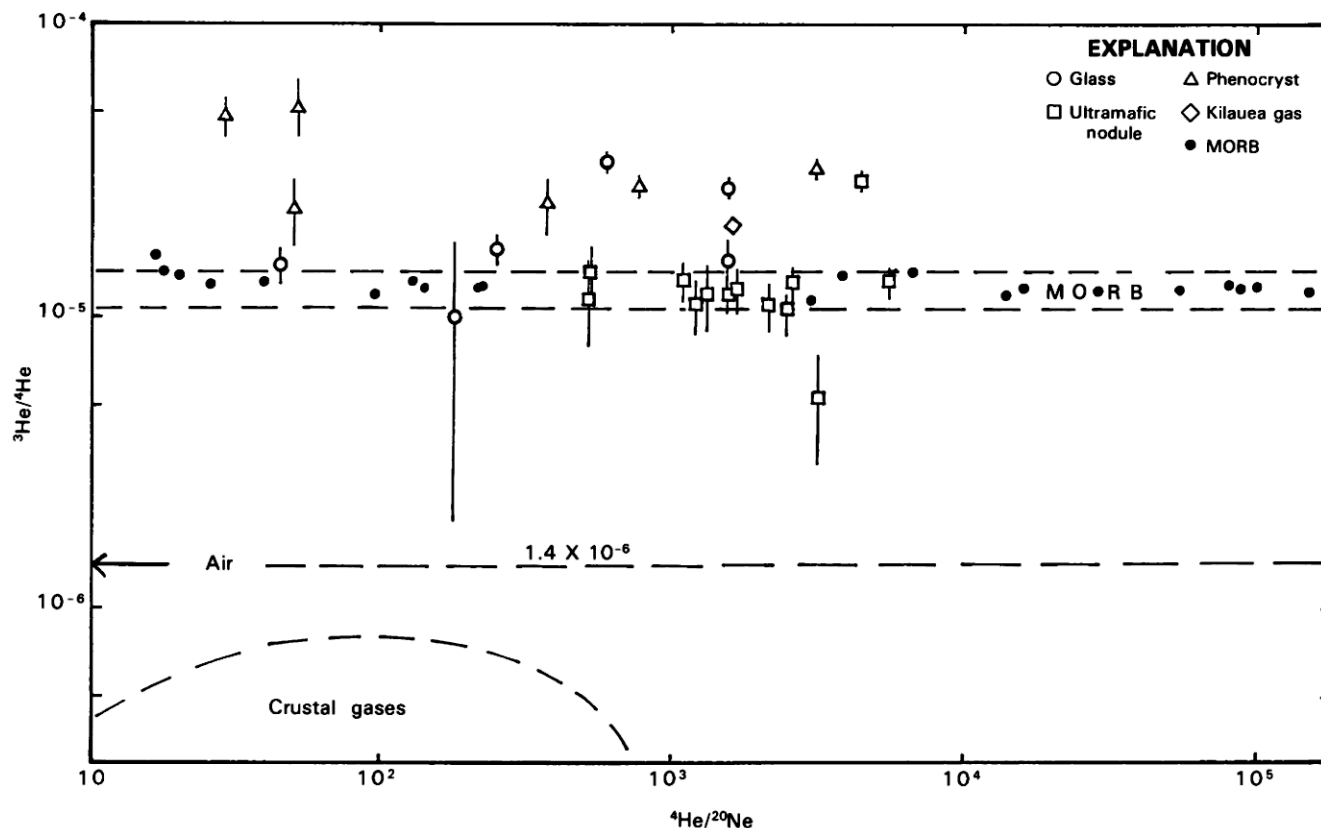


FIGURE 27.2.—Plot of $^3\text{He}/^4\text{He}$ ratio against $^4\text{He}/^{20}\text{Ne}$ ratio for Hawaiian samples and MORB. Vertical lines show uncertainty in the $^3\text{He}/^4\text{He}$ ratio measured. In spite of large variation in $^4\text{He}/^{20}\text{Ne}$ ratio for MORB, its $^3\text{He}/^4\text{He}$ ratio is rather uniform (as shown by bracketing dashed lines), but $^3\text{He}/^4\text{He}$ ratios for Hawaiian samples are generally higher than those of MORB. Data sources: Lupton and Craig (1975); Craig and Lupton (1976); Kaneoka and Takaoka (1978, 1980); Kyser and Rison (1982); Kaneoka and others (1983); Ozima and Zashu (1983).

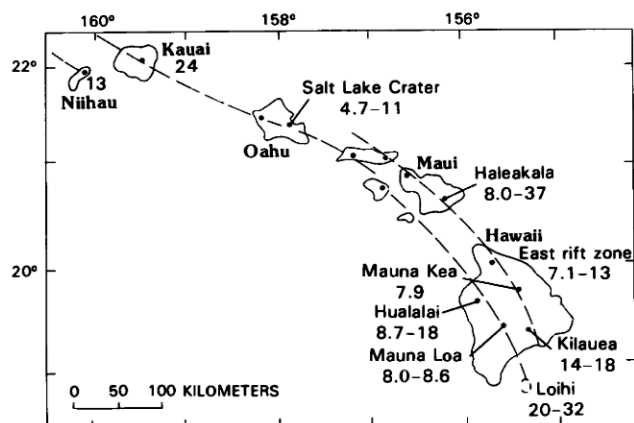


FIGURE 27.3.—Distribution of $^3\text{He}/^4\text{He}$ ratio for rock and gas samples from the Hawaiian Islands. Each number corresponds to the value or range of values of R/R_A , where R is the $^3\text{He}/^4\text{He}$ ratio in the sample(s) and R_A indicates the atmospheric value (1.4×10^{-6}). Dashed lines mark the two lines of volcanic centers (dots). Data sources: Craig and Lupton (1976); Kaneoka and Takaoka (1978, 1980); Kyser and Rison (1982); Kaneoka and others (1983); Kurz and others (1983); Lupton (1983); Rison and Craig (1983).

on a line that connects the atmospheric $^{40}\text{Ar}/^{36}\text{Ar}$ ratios (295.5) on the ordinate and the point for uncontaminated sample, as long as the samples contain the same amount of ^{36}Ar (see, for example, Dymond and Hogan, 1978). The data for Hawaiian samples are plotted on such a diagram in figure 27.4; in this diagram, the data scatter and do not define any single line. However, it is worthwhile to point out that most ultramafic nodules show systematically higher $^{40}\text{Ar}/^{36}\text{Ar}$ ratios of more than 1,000, whereas glass samples and phenocrysts show mostly values of less than 1,000, irrespective of the amount of ^{36}Ar . These submarine glass samples are not different in their eruptive conditions from MORB. Hence there is no reason why they should be much more contaminated with atmospheric components through sea water than MORB. In effect, these glass samples retain magmatic He. Hence, even if the relatively low $^{40}\text{Ar}/^{36}\text{Ar}$ ratios observed are affected by the atmospheric contamination for some samples, it does not preclude the possibility of a low $^{40}\text{Ar}/^{36}\text{Ar}$ ratio for magmatic Ar in Hawaii. This point will be discussed again later in discussing the $^3\text{He}/^4\text{He}$ - $^{40}\text{Ar}/^{36}\text{Ar}$ diagram.

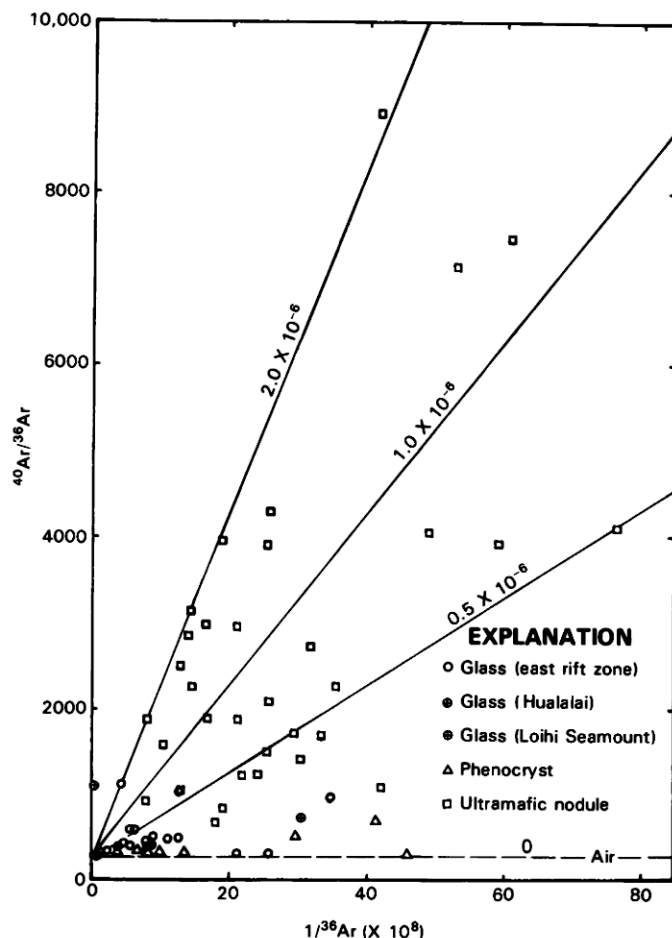


FIGURE 27.4.—Plot of $^{40}\text{Ar}/^{36}\text{Ar}$ ratio against $1/^{36}\text{Ar}$ for Hawaiian rock samples. Value of $^{40}\text{Ar}/^{36}\text{Ar}$ in air indicated by dashed line. The lines in the figure indicate reference lines which correspond to the mixing line between the original magmatic Ar and the atmospheric Ar, where the concentration of magmatic ^{40}Ar in each sample is assumed as the numerical figures on each line in the unit of cubic centimeters at standard temperature and pressure for one gram sample. Data sources: Dalrymple and Moore (1968); Funkhouser and Naughton (1968); Noble and Naughton (1968); Hennecke and Manuel (1975); Kaneoka and Takaoka (1978, 1980); Kyser and Rison (1982); Allègre and others (1983); Fisher (1983); Kaneoka and others (1983).

OTHER NOBLE GAS ISOTOPES

In Kilauea gas and in MORB, a higher $^{20}\text{Ne}/^{22}\text{Ne}$ ratio (10.3) than the atmospheric value (9.80) was reported by Craig and Lupton (1976), who considered it as evidence that the primordial Ne component of solar type may be still retained in the Earth's deep interior. However, this value and the corresponding $^{21}\text{Ne}/^{22}\text{Ne}$ ratio can also be explained by mass fractionation from the atmospheric value (Kaneoka, 1980), and the occurrence of primordial Ne component has not yet been confirmed.

On the other hand, the occurrence of higher $^{129}\text{Xe}/^{130}\text{Xe}$ ratios than the atmospheric value has been confirmed in ultramafic nodules from Hualalai (Hennecke and Manuel, 1975; Kaneoka

and others, 1978; Kyser and Rison, 1982). Even olivine phenocrysts of Kapoho lava, Kilauea, seem to contain excess ^{129}Xe (Kaneoka and Takaoka, 1978). However no excess ^{129}Xe has been confirmed in Loihi samples (Allègre and others, 1983; Kaneoka and others, 1983). Since excess ^{129}Xe has been regarded as the decay product of extinct ^{129}I (half-life, 17 m.y.), its occurrence indicates that the magma source was not totally equilibrated with the surrounding part for more than 4 b.y. Allègre and others (1983) reported the occurrence of excess ^{129}Xe in MORB.

The $^{129}\text{Xe}/^{130}\text{Xe}$ ratio is plotted in figure 27.5 against the $^{40}\text{Ar}/^{36}\text{Ar}$ ratio for Hawaiian samples and some MORB samples. Although the uncertainty in the $^{129}\text{Xe}/^{130}\text{Xe}$ ratio is fairly large, the array of samples seems to indicate on a mixing line of $r \geq 0.5$, where r is defined as the ratio of $^{130}\text{Xe}/^{36}\text{Ar}$ for air divided by $^{130}\text{Xe}/^{36}\text{Ar}$ for the MORB source and the value of $^{130}\text{Xe}/^{36}\text{Ar}$ for air is 1.13×10^{-4} . For $^{130}\text{Xe}/^{36}\text{Ar}$ of the MORB source, if we adopt the value observed for a pillow glass that showed the highest excess ^{129}Xe in MORB (Allègre and others, 1983), it becomes 11.6×10^{-4} . Hence, mixing between this component and the atmosphere indicates a value of r close to about 0.1. The data array observed in figure 27.5 do not seem to be explained by the addition of an atmospheric component alone. This may imply that the Hawaiian magma source, which may be a typical hot-spot area, shows an atmospheric $^{129}\text{Xe}/^{130}\text{Xe}$ ratio and a rather low $^{40}\text{Ar}/^{36}\text{Ar}$ ratio close to the atmospheric value. Large analytical

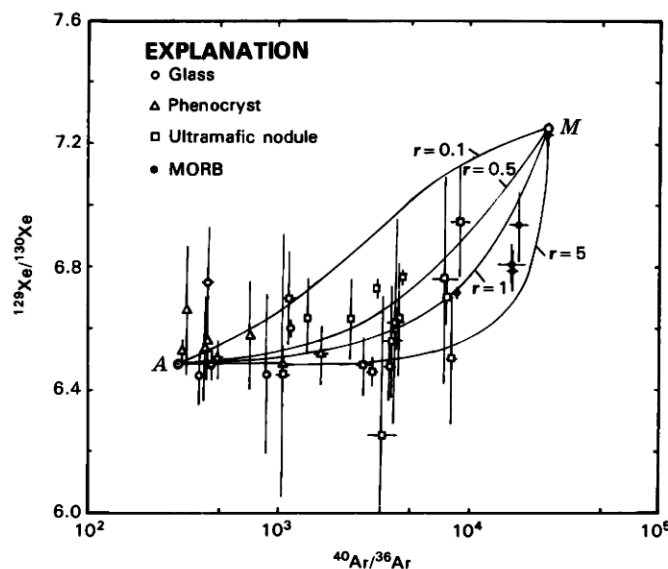


FIGURE 27.5.—Plot of $^{129}\text{Xe}/^{130}\text{Xe}$ ratio against $^{40}\text{Ar}/^{36}\text{Ar}$ ratio for Hawaiian rock samples and a few MORB samples. Mixing lines between the atmospheric component (A: $^{40}\text{Ar}/^{36}\text{Ar} = 295.5$; $^{129}\text{Xe}/^{130}\text{Xe} = 6.48$) and an assumed MORB source component (M: $^{40}\text{Ar}/^{36}\text{Ar} = 25,000$; $^{129}\text{Xe}/^{130}\text{Xe} = 7.25$) are drawn with a function of r , where r is $(^{130}\text{Xe}/^{36}\text{Ar})_{\text{air}} / (^{130}\text{Xe}/^{36}\text{Ar})_{\text{MORB}}$ source. Vertical and horizontal lines indicate uncertainty. Data sources: Hennecke and Manuel (1975); Kaneoka and Takaoka (1978, 1980); Kyser and Rison (1982); Allègre and others (1983); Kaneoka and others (1983).

uncertainties remain, however, concerning the $^{129}\text{Xe}/^{130}\text{Xe}$ ratio for most samples from Hawaii. Hence more precise data are necessary to draw a definite conclusion.

THE $^3\text{He}/^4\text{He}$ - $^{40}\text{Ar}/^{36}\text{Ar}$ SYSTEM

A magma source would be more clearly characterized by an appropriate combination of isotopic ratios than by a single one. For this purpose, the $^3\text{He}/^4\text{He}$ - $^{40}\text{Ar}/^{36}\text{Ar}$ system is one of the most effective ones among possible combinations of noble-gas isotopic ratios. The merits of using this system have been discussed in detail elsewhere (Kaneoka and Takaoka, 1985).

All the available data for samples from Hawaiian volcanoes are plotted on the $^3\text{He}/^4\text{He}$ - $^{40}\text{Ar}/^{36}\text{Ar}$ diagram in figure 27.6. Except for a dunite nodule from Loihi Seamount, higher $^3\text{He}/^4\text{He}$ ratios seem to be accompanied by lower $^{40}\text{Ar}/^{36}\text{Ar}$ ratios. Furthermore, ultramafic nodules are located in the region defined by MORB. Such relations for Hawaiian samples were first identified by Kaneoka and Takaoka (1980), who argued for different magma sources for Hawaiian magma and ultramafic nodules. As pointed out previously, the accumulation of data on the $^3\text{He}/^4\text{He}$ ratio in lava phases has suggested that Hawaiian volcanism is characterized by some complexity of processes in each locality, but this would not change the main scheme fundamentally from that stated above. This point will be discussed later in more detail.

However, the plot in figure 27.6 strongly suggests at least that a magma source enriched in primordial noble-gas components exists under the Island of Hawaii. If we extend such data to other areas,

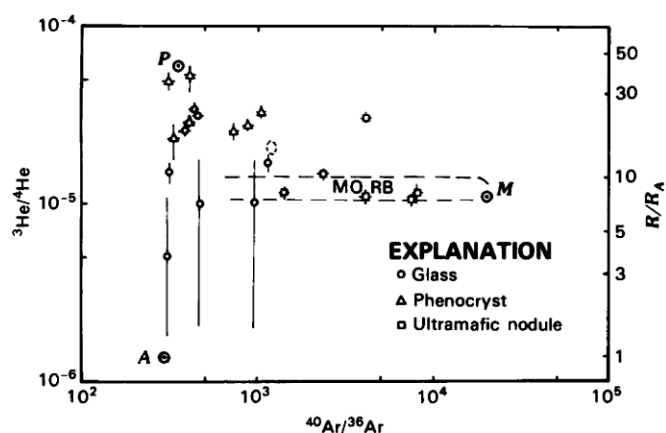


FIGURE 27.6.—Plot of $^3\text{He}/^4\text{He}$ ratio against $^{40}\text{Ar}/^{36}\text{Ar}$ ratio for Hawaiian rock samples. For most samples, only the results from high-temperature fractions are shown here. A, atmosphere; P, plume-type source; M, MORB source; dashed lines, approximate range of MORB samples. Vertical and horizontal lines indicate analytical uncertainties for each sample. Dashed circle indicates a combined value for samples for Kilauea, where $^3\text{He}/^4\text{He}$ ratios were obtained for fumarolic gases and $^{40}\text{Ar}/^{36}\text{Ar}$ ratios for pillow glasses from east rift zone. Data sources: Kaneoka and Takaoka (1978, 1980); Kyser and Rison (1982); Allegre and others (1983); Kaneoka and others (1983).

the same tendencies as observed for Hawaiian samples are also identified (fig. 27.7). The most remarkable feature of figure 27.7 is that higher $^3\text{He}/^4\text{He}$ ratios than the MORB value are always associated with relatively lower $^{40}\text{Ar}/^{36}\text{Ar}$ ratios, except for a few cases, and such high $^3\text{He}/^4\text{He}$ ratios are observed in typical hot-spot areas. This does not mean that samples from all hot-spot areas show such a trend. For example, volcanic rocks from the Tristan da Cunha and Gough Islands show lower $^3\text{He}/^4\text{He}$ ratios than that of MORB, but the ratios are still higher than that of the atmospheric component (Kurz and others, 1982). Such a trend may be explained by the addition of a crustal component to the pure plume-type component that has a high $^3\text{He}/^4\text{He}$ ratio (see, for example, Kurz and others, 1982; Kaneoka and Takaoka, 1985).

On the basis of the $^3\text{He}/^4\text{He}$ - $^{40}\text{Ar}/^{36}\text{Ar}$ systematics, we can identify at least four end members as shown in figure 27.7: P(plume-type), M(MORB-type), A(atmosphere-type), and C(crust-type) (Kaneoka, 1983; Kaneoka and Takaoka, 1985). C components are multi-divided further depending on the local and regional inhomogeneities including time-integrated effects. Although P and probably M components can be subdivided on a local scale, we need not do so to discuss general trends on rather global scales. The observed noble-gas data can be explained by mixing among these components, including the data observed at subduction zones, which may be mixtures of M, A, and C components.

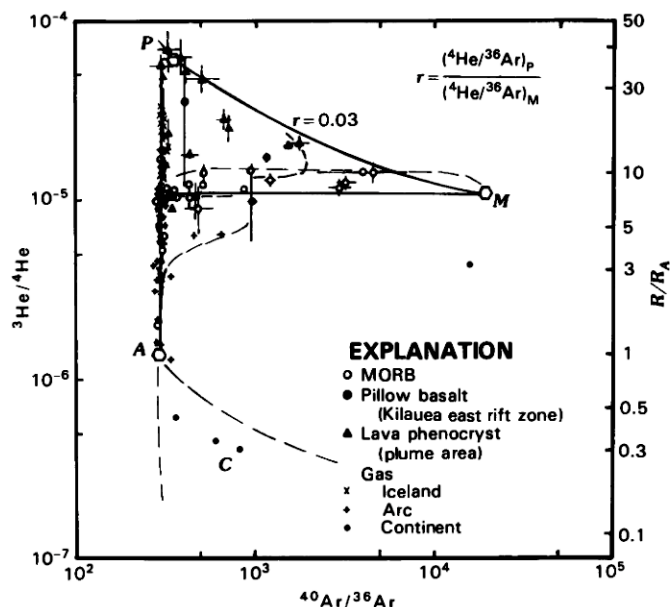


FIGURE 27.7.—Plot of $^3\text{He}/^4\text{He}$ ratio against $^{40}\text{Ar}/^{36}\text{Ar}$ ratio for samples that represent the recent state of the Earth's interior. P, M, A, and C represent assumed typical values for source materials of plume, MORB, atmosphere and crust types, respectively. Uncertainties are indicated by short vertical and horizontal lines. Solid lines show calculated mixing lines between sources. Dashed lines indicate regions where each component is dominant (though arc-type source material is not discussed in the present study) (after Kaneoka and Takaoka, 1985).

In this context, it is interesting to examine the data for Hawaiian samples in figure 27.7. Most data for Hawaiian samples are located in a region bordered by the three mixing lines, $P-M$, $M-A$, and $P-A$. Since A component is composed of atmospheric noble gases from the atmosphere, sea water, and ground water, those data located away from the mixing line $P-M$ would have been secondarily contaminated on the surface or at a relatively shallow depth. On the other hand, some Hawaiian samples seem to lie on the $P-M$ mixing line. Since the fit for mixing line is fairly good, the variation in the observed data has been interpreted as showing the mixing between these probably mantle-derived components (Kaneoka and Takaoka, 1980, 1985; Kyser and Rison, 1982; Kaneoka, 1983).

Because this diagram is composed of two kinds of noble-gas isotopic ratios, the variation in it can be explained rather simply as long as He and Ar behave similarly in the main process concerned. On the other hand, if we take a set of elements of different kind, such as He and Sr (see, for example, Kurz and others, 1982), the system is more easily disturbed because of the different transfer of elements caused by the difference in the chemical behavior. Hence, the occurrence of completely different components (P and M) in the $^3\text{He}/^4\text{He}$ - $^{40}\text{Ar}/^{36}\text{Ar}$ systematics is of great significance in inferring chemical characteristics of the Earth's interior. The component P would surely represent the source material of less depleted or fertile mantle that lies under the M -type (MORB) source, which is considered to have been much depleted in incompatible elements (Kyser and Rison, 1982; Kaneoka, 1983; Kaneoka and Takaoka, 1985). Although some investigators argue for the occurrence of fertile mantle as a result of enrichment process in the asthenosphere (for example, Anderson, 1982), this seems unlikely on the basis of noble-gas systematics (Kaneoka, 1983).

Thus, noble-gas isotopes for Hawaiian samples clearly suggest that the Hawaiian rocks are derived from fertile mantle, which is richer in primitive noble gases and incompatible elements such as K and U than the MORB source. Although similar inferences have been made on the basis of Nd-Sr systematics (see, for example, Jacobsen and Wasserburg, 1979; DePaolo, 1981), the identification of the typical component for the fertile mantle has remained hypothetical and could be interpreted in a different way in the $^{143}\text{Nd}/^{144}\text{Nd}$ - $^{87}\text{Sr}/^{86}\text{Sr}$ diagram. In the $^3\text{He}/^4\text{He}$ - $^{40}\text{Ar}/^{36}\text{Ar}$ diagram, however, we can identify it clearly as the P component. This component seems to be found in samples from typical hot-spot areas such as Hawaii, Iceland, and Reunion. Thus, noble-gas data give clear evidence for the inference that Hawaiian volcanism has been caused by a mantle plume or plumes from a hot spot that lies at a relatively deeper level (lower mantle?) in the Earth's interior (see Kaneoka, 1983, for further discussion).

NOBLE GAS ISOTOPES AS CONSTRAINTS TO INFER THE GENETIC RELATIONS FOR HAWAIIAN ROCKS

In attempts to reveal the origin of Hawaiian volcanic rocks by petrological, mineralogical, and geochemical studies, it has often been assumed that the ultramafic nodules included in Hawaiian

volcanic rocks represent either the source materials of these volcanic rocks or cumulates from them (for example, Jackson and Wright, 1970). On the basis of $^3\text{He}/^4\text{He}$ and $^{40}\text{Ar}/^{36}\text{Ar}$ ratios, however, it has been shown that phenocrysts in Hawaiian volcanic rocks (tholeiite and alkalic basalt) and ultramafic nodules would have been derived from different sources (Kaneoka and Takaoka, 1978, 1980). The source for ultramafic nodules seems to be similar to that for MORB. Such results have been supported by subsequent work (Kyser and Rison, 1982). It has been known that $^{87}\text{Sr}/^{86}\text{Sr}$ ratios for ultramafic nodules (0.7031–0.7032) (Shimizu, 1975) are slightly lower than those of Hawaiian tholeiite (0.7034–0.7038) (see, for example, O'Nions and others, 1977). However, the difference in the $^3\text{He}/^4\text{He}$ and $^{40}\text{Ar}/^{36}\text{Ar}$ ratios is so large between phenocrysts and ultramafic nodules (see fig. 27.6) that we cannot regard them as derived from similar sources. Recent summarized studies on the origin of Hawaiian tholeiite have also suggested that most ultramafic nodules are not directly related to the origin of Hawaiian volcanic rocks (Wright, 1984).

As discussed in the previous section, Hawaiian ultramafic nodules show similar $^3\text{He}/^4\text{He}$ and $^{40}\text{Ar}/^{36}\text{Ar}$ ratios to those of MORB. These analyzed nodules were mostly samples from Hualalai Volcano and Salt Lake Crater, and their rock types include dunite, spinel lherzolite, and garnet pyroxenite. As far as noble-gas data are concerned, these ultramafic nodules include gas components that were derived from the same source or a quite similar source to that of MORB. Hence, these nodules might have been the source rock or cumulates of MORB, or they might represent rocks formed in the asthenosphere to constitute the lower part of the lithosphere. More strictly, if the accompanying volcanic rocks also show M -type noble-gas isotopes, we cannot preclude the possibility of a genetic relation between such ultramafic nodules and the volcanic-rock source. However, dredged tholeiite samples from Hualalai show definitely higher $^3\text{He}/^4\text{He}$ ratios (14–18 R_A) than those of ultramafic nodules (8.6–9.2 R_A) (Kurz and others, 1983). Hence they have no direct genetic relation between them.

For the samples from Salt Lake Crater, we unfortunately have no noble-gas data on volcanic rocks. Since volcanic rocks of the Honolulu Volcanics are reported to have rather lower $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (about 0.7033) than those of the Koolau Basalt² (about 0.7039) (Lanphere and Dalrymple, 1980), there is a possibility that volcanic rocks of the Honolulu Volcanics may have M -type noble-gas isotopes. If this is so, then we can regard those ultramafic nodules as related to some source rocks or cumulates that also were a source for volcanic rocks of the Honolulu Volcanics.

In a detailed study of $^3\text{He}/^4\text{He}$ ratios for volcanic rocks from the Island of Hawaii, Kurz and others (1983) have revealed that each volcano shows different $^3\text{He}/^4\text{He}$ ratios and that the ratio seems to decrease with increasing volume of a volcano and with decreasing age of the volcano. In their results, it is noteworthy that tholeiite samples from Mauna Loa and an alkali basalt from Mauna Kea show very similar $^3\text{He}/^4\text{He}$ ratios (7.9–8.6 R_A) to that of

²Stratigraphic nomenclature used in chapter is from Langenheim and Clague (chapter 1, part II).

MORB. Rison and Craig (1983) also report a similar value ($8.5 R_A$) for one Mauna Loa sample from analyzing olivine phenocrysts. Unfortunately, we have no heavier noble-gas data for these samples. However, even these data give us important information to infer the processes that are occurring underneath the Island of Hawaii.

With reference to He isotopes, the He sources should be similar to that of MORB. Hence in the present frame of ideas the He source should be either in the asthenosphere or in the lithosphere itself, since *P*-type noble gases are assumed to be derived from the deeper part of the Earth's interior.

Samples from Loihi give us a unique chance to study the relations of rock types and nodule gas isotopes. Because Loihi Seamount is regarded as the most recent volcano in the evolution of the Hawaiian-Emperor Chain, from it we can get information on an earlier state of a volcano's evolution along the chain. Furthermore, the dredged rocks from the seamount show diverse basalt types, including tholeiitic, transitional, and alkali basalt and basanite, including dunite nodules (Moore and others, 1982).

The He isotopic measurements have shown very high $^3\text{He}/^4\text{He}$ ratios (up to $32 R_A$) and tholeiite samples show generally higher $^3\text{He}/^4\text{He}$ ratios than those of alkali rocks (Kurz and others, 1983; Rison and Craig, 1983). Since dunite nodules show $^3\text{He}/^4\text{He}$ ratios similar to that of alkalic lava that included nodules, Kurz and others (1983) have conjectured that the apparent lower $^3\text{He}/^4\text{He}$ ratios for alkalic rocks might be due to contamination by He included in xenocrysts of nodules. On the other hand, Kaneoka and others (1983), who examined all noble gases in both tholeiitic and alkalic rocks, found that the $^{40}\text{Ar}/^{36}\text{Ar}$ ratios for lava are not so high as observed in MORB and are still located in the region defined by the three kinds of mixing lines among *P*, *M*, and *A* components in the $^3\text{He}/^4\text{He}$ - $^{40}\text{Ar}/^{36}\text{Ar}$ diagram. Only a dunite nodule from Loihi Seamount seems to be an exception. Although this nodule shows a high $^3\text{He}/^4\text{He}$ ratio ($22 R_A$), its $^{40}\text{Ar}/^{36}\text{Ar}$ ratio is also high (4,200). To explain this, Kaneoka and others (1983) have conjectured that only He was equilibrated with the surrounding magma and the nodule had originally *M*-type noble gases.

In Loihi samples, fairly large variation in the $^3\text{He}/^4\text{He}$ ratio has been observed. This may be a characteristic of the initial state of a volcano's evolution caused by different degrees of contamination with lithospheric and (or) asthenospheric components that are similar to that of MORB.

On the other hand, apparent uniformity in the observed value of noble gases may simply reflect the sampling problem, because we have data only on lava flows of relatively limited time periods of young active volcanoes such as Mauna Loa and Kilauea. In fact, for Haleakala Volcano, we have observed very high $^3\text{He}/^4\text{He}$ ratios for olivine and augite phenocrysts of ankaramite ($32\text{--}36 R_A$) from the vent known as White Hill Cone (Kaneoka and Takaoka, 1980; Rison and Craig, 1983), which probably belongs to the Pleistocene Kula Volcanics (hawaiite). However, both pyroxene phenocrysts from the A.D. 1790 ankaramite flow in the post erosional Hana Volcanics and olivine phenocrysts from the tholeiitic shield lava of the Honomanu Basalt show *M*-type $^3\text{He}/^4\text{He}$ ratios of $8.1 R_A$ and $8.0 R_A$, respectively (Rison and Craig, 1983). Except for the Hana Volcanics, such tendency is at variance with that predicted

from some trace-element data and from Sr and Nd isotopic ratios for Haleakala Volcano (Chen and Frey, 1983). This means that noble gases (and probably some volatile elements, too) may not always behave similarly with other nonvolatile elements, even in a mantle plume.

HAWAIIAN VOLCANISM AS INFERRED FROM NOBLE-GAS DATA

As discussed previously, Hawaiian volcanism seems to be strongly related to a mantle plume or plumes of deep origin. Noble-gas data require a source material that still retains more of primitive components than does that of MORB (see, for example, Kyser and Rison, 1982; Kaneoka, 1983). If Hawaiian volcanism was initiated by propagation of a tensional fracture through the lithosphere, the magma source should be located in the uppermost part of the asthenosphere. However, it seems difficult to keep such a source in a relatively shallow part of the Earth's interior for a long time, because the source material is inferred to retain primitive components, including He, which would show one of the largest mobilities in the Earth's interior (Kaneoka, 1983).

The $^3\text{He}/^4\text{He}$ ratios for each volcano in the Hawaiian Islands are plotted against the volume of each volcano in figure 27.8. If we consider only the Island of Hawaii, volcanoes with small volume show high $^3\text{He}/^4\text{He}$ ratios, and volcanoes with large volume show $^3\text{He}/^4\text{He}$ ratios similar to that of MORB (for example, Mauna Kea and Mauna Loa). This suggests that the contribution of *P*-type source material is larger in the early stage of a volcano's evolution relative to that of *M*-type source material. Such a relation has been shown by Kurz and others (1983). If we include other volcanoes from different islands, however, such systematics seem to be disturbed, as shown in figure 27.8. Volcanic rocks from Haleakala on Maui Island and those from Kauai Island indicate relatively high $^3\text{He}/^4\text{He}$ ratios in spite of the large volumes of their volcanoes. From Haleakala Volcano, volcanic rocks with $^3\text{He}/^4\text{He}$ ratios similar to that of MORB have also been reported (Rison and Craig, 1983). Ultramafic nodules from Salt Lake Crater, Oahu Island, show $^3\text{He}/^4\text{He}$ ratios fairly similar to that of MORB. Hence, the situation may be not so simple as reported initially for samples from the Island of Hawaii. Relatively high $^3\text{He}/^4\text{He}$ ratios observed for samples from Haleakala Volcano indicate that the *P*-type component of noble gases could have been added even at a later stage in a volcano's evolution.

On the basis of the systematic changes in elemental ratios and $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios with eruption age for volcanic rocks of Haleakala Volcano, Chen and Frey (1983) have predicted a systematic decrease over time in the rate of contribution of plume-derived material to the formation of a magma. Basically their model seems to explain well the variations of elemental and isotopic data for Hawaiian volcanism. However, the data in figure 27.8 do not always follow the model. As suggested by Rison and Craig (1983), the Honomanu Basalt (shield-building tholeiite) of Haleakala Volcano shows *M*-type He component, whereas *P*-type He component would be expected in the model. Although higher $^3\text{He}/^4\text{He}$ ratios seem to be accompanied by higher $^{87}\text{Sr}/^{86}\text{Sr}$ ratios or lower

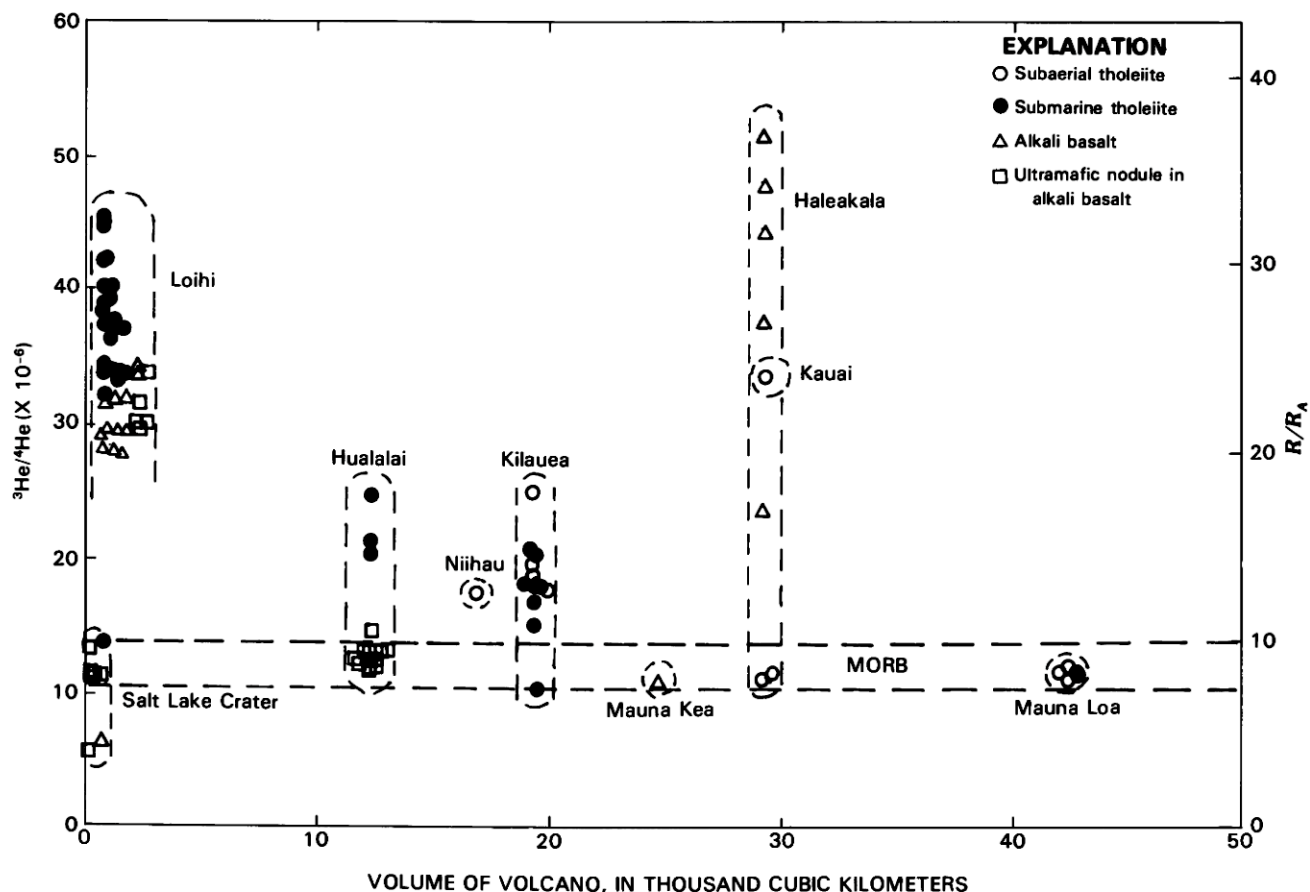


FIGURE 27.8.—Plot of $^3\text{He}/^4\text{He}$ ratio against volcano volume for Hawaiian volcanoes. Volcano volumes taken from Bargar and Jackson (1974). Typical range of $^3\text{He}/^4\text{He}$ values for MORB indicated by horizontal dashed lines. Data sources: Craig and Lupton (1976); Kaneoka and Takaoka (1978, 1980); Kyser and Rison (1982); Kurz and others (1982, 1983); Kaneoka and others (1983); Lupton (1983); Rison and Craig (1983).

$^{143}\text{Nd}/^{144}\text{Nd}$ ratios for Loihi samples (Kurz and others, 1983; Staudigel and others, 1984), such trends are not always retained when the data from other volcanoes are included. Furthermore, although tholeiitic shield lava from the Hawaiian Islands shows generally slightly higher $^{87}\text{Sr}/^{86}\text{Sr}$ and lower $^{143}\text{Nd}/^{144}\text{Nd}$ ratios than those of MORB (see, for example, Staudigel and others, 1984), the $^3\text{He}/^4\text{He}$ ratio is quite similar to the MORB value, at least for Mauna Kea and Mauna Loa. Thus, noble-gas isotopes seem to be not always correlated with the other radiogenic isotopes such as Sr, Nd, and Pb. Such differences probably reflect the different geochemical behavior of noble gases from the other non-volatile elements. This implies that noble gases would be transported in varying degrees together with other volatile components during the different stages of evolution of Hawaiian volcanoes. Furthermore, the $^3\text{He}/^4\text{He}$ ratios for Hawaiian rocks have not always varied smoothly. Such noble gases may have been transported in a mantle diapir rather than in a mantle plume.

So far, a number of models for Hawaiian volcanism have been proposed (see, for example, Macdonald, 1968; Chen and Frey,

1983; Sen, 1983; Staudigel and others, 1984). However, those do not take properly into account the noble-gas-isotope signature. As discussed in previous sections, the most significant feature of this signature is the occurrence in Hawaiian lava of much higher $^3\text{He}/^4\text{He}$ ratios than that of MORB, which strongly suggests the involvement of a source material that contains fairly primordial components and is located in the deeper part of the Earth's interior (lower mantle?). We cannot deny the involvement of oceanic lithosphere during the formation of a magma. As far as the noble gases (and probably volatile components) are concerned, however, the contribution from the oceanic lithosphere would be not so significant as has been suggested for nonvolatile elements (Staudigel and others, 1984). If the age of the oceanic lithosphere is great enough to have increased the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio, the $^3\text{He}/^4\text{He}$ ratio would have been lowered by the addition of radiogenic ^4He , resulting in a lower $^3\text{He}/^4\text{He}$ ratio than that of MORB. When the $^3\text{He}/\text{U}$ ratio can be postulated to be similar to that of the MORB source, the $^3\text{He}/^4\text{He}$ ratio would have to be similar also. However, it is very difficult to conjecture that the $^3\text{He}/\text{U}$ ratio is maintained

higher than that of the MORB source, because He would be easily lost during formation of the lithosphere around the ridge, resulting in lowering the $^3\text{He}/^4\text{He}$ ratio. If the lower part of the oceanic lithosphere is formed by the addition of the components of the asthenosphere, its isotopic composition would be similar to that of MORB. At least some Hawaiian ultramafic nodules might have been derived from such lower lithosphere.

Volcanic rocks from Loihi Seamount show definitely higher $^3\text{He}/^4\text{He}$ ratios than those of Kilauea (fig. 27.8). This implies that the contribution of *P*-type component is larger for Loihi samples than for Kilauea samples as indicated by the He data. Large variations in the observed values can be explained by varying degrees of addition of *M*-type component during the ascent of volatile-enriched mantle diapirs. It is true that magma chemistry as determined by the last depth of equilibration would have little bearing on the initial depth of origin for source diapirs (Staudigel and others, 1984). Even the voluminous outpourings are not always good indicators for characterizing the source material of a plume, because we cannot exclude the possibility of interaction during the ascent of a mantle diapir with *M*-type source materials. The $^3\text{He}/^4\text{He}$ ratio may be used more efficiently to indicate the plume component for noble-gas (volatiles) components.

From the noble-gas data together with those of other radiogenic isotopes, we can summarize a model for Hawaiian volcanism as shown schematically in figure 27.9.

The initial (infant) stage of Hawaiian volcanism would be exemplified by Loihi Seamount (fig. 27.9A). When the relatively cold lithosphere approaches the hot spot, volcanism would be initiated by the rise of volatile-enriched mantle diapirs with *P*-type noble gases derived from the deeper part of the Earth's interior

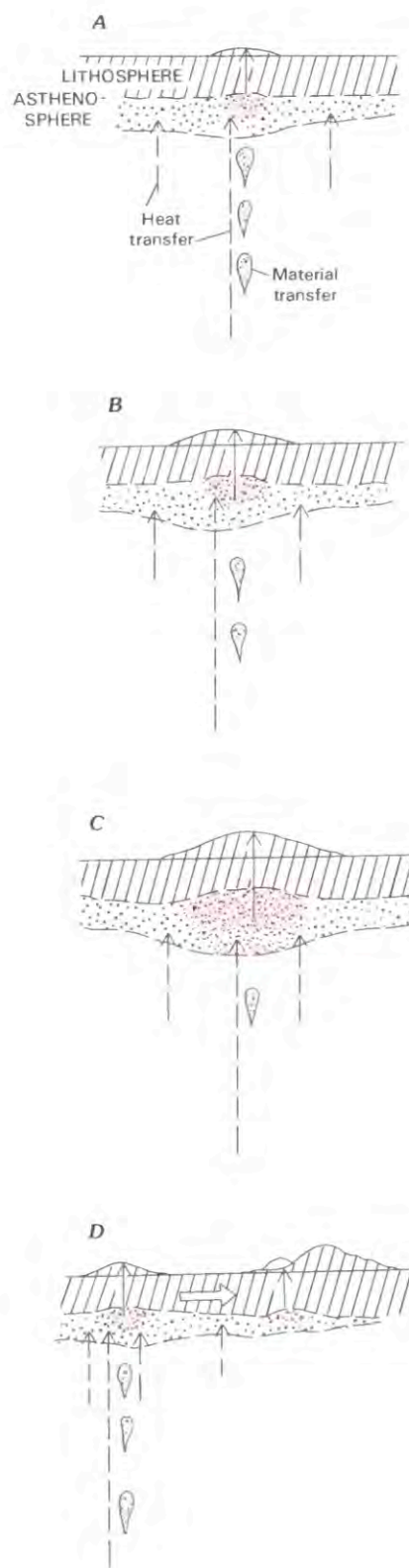


FIGURE 27.9.—Schematic model to show the evolution of Hawaiian volcanism.

Red color shows regions of intensive partial melting. **A**, Infant stage: volcanism is initiated by rise of diapir(s) from hot-spot source in Earth's deep interior (lower mantle?). Diapir will be enriched in volatile components. Possible example of this stage is Loihi Seamount. **B**, Early part of tholeiitic shield-building stage: diapirs still rise through asthenosphere and lithosphere. Partially melted zone will be increased because of accumulation of heat energy through diapirs and conduction, causing downward swelling and thinning of lithosphere over hot spot. Plume-type source material will mix with surrounding MORB-type source material during rise of diapirs(s) through asthenosphere and lithosphere (example, Kilauea Volcano). **C**, Later part of tholeiitic shield-building stage: contribution from diapir(s) probably much reduced. Instead, most melt may be derived from partially melted zone in uppermost part of asthenosphere. Most MORB-type He might be supplied from this portion (example, Mauna Loa). Although no example is shown here, in a volcano at the postcaldera stage, no significant contribution from diapirs can be expected. Diminishing heat and material supplies from hot-spot source cause eruption frequency of a volcano to decrease. Degrees of partial melting also decrease, resulting in production of alkalic basalt (example, Kula Volcanics, Haleakala Volcano). **D**, Posterosional stage: when a volcano migrates far away from hot-spot area, some melt would still remain at bottom of lithosphere. Addition of conductive heat from initial thermal anomaly and release of latent heat from solidification of melt may form some melt pockets under the volcano and produce alkalic and (or) nephelinitic rocks. Source material for such rocks is quite similar to that of MORB (example, Honolulu Volcanics, Oahu Island).

(lower mantle?). During ascent to the surface, such diapirs will interact with surrounding materials at various depths both in the asthenosphere and the lithosphere. Such interaction would have caused large variation in both isotopic and chemical signatures of magmas. We assume the existence under the lithosphere of a partially melted zone enriched with *M*-type noble gases that is indicated seismically as the low-velocity zone. The occurrence of such a zone may help the interaction between the diapirs and the surrounding materials.

The second stage is that of tholeiitic shield-building (fig. 27.9B, C). During this stage, melting in the partially melted zone of the asthenosphere would be increased because of the addition of heat from the ascending diapirs and by conduction. This may cause swelling of the partially melted zone and thinning of the lithosphere over the hot-spot area. During such processes, *M*-type noble gases might be enriched in the partially melted zone under the lithosphere. At Kilauea Volcano (fig. 27.9B), the contribution from mantle diapirs will still be significant, but the mixing rate of *M*-type noble gas will increase, resulting in lower $^3\text{He}/^4\text{He}$ and $^{87}\text{Sr}/^{86}\text{Sr}$ ratios and higher $^{143}\text{Nd}/^{144}\text{Nd}$ ratios compared with those of the infant stage.

On the other hand, for Mauna Kea and Mauna Loa Volcanoes (fig. 27.9C), the role of diapirs may be much less than for Kilauea Volcano. As shown in figure 27.8, we observe no difference in the $^3\text{He}/^4\text{He}$ ratio between these volcanoes and MORB. This implies that even when voluminous out-pourings occur from these volcanoes, it does not always mean a contribution from mantle diapirs as far as He isotopes are concerned. On the other hand, $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios are still different for these volcanoes from those of MORB (see, for example, Staudigel and others, 1984). This suggests that most He was derived from the partially melted zone in the uppermost asthenosphere and that the voluminous outpourings of lava were also caused by the expansion of the partially melted zone compared with the infant stage. At this stage, the hot spot may act mainly as a heat source, but not the main source for adding materials, though some small diapirs will still rise. The isotopic ratios for Sr and Nd may be explained by the interaction of melts formed in the asthenosphere with the oceanic lithosphere. The occurrence of mantle metasomatism is possible (Wright, 1984).

Another possibility is that the contribution from diapirs still continues but the diapirs are less enriched in volatile components than those of the infant stage. Hence, even if isotopic ratios for nonvolatile elements are still different from those of MORB, the $^3\text{He}/^4\text{He}$ ratio might become similar to that of MORB because of mixing with the volatile-enriched *M*-type source in the partially melted zone. At present, we have no good evidence to select one or the other possibility. In both cases, however, most He would have been derived from the partially melted zone in the uppermost asthenosphere, and the occurrence of such a zone is significant for Hawaiian volcanism at this stage.

When the volcano migrates away from the hot spot, the supply of heat and materials will be diminished. The rate of melting in the partially melted zone also decreases, resulting in the decrease of eruption frequency of a volcano. Total amounts of melt produced

would also decrease. Since the degree of partial melting would also decrease, alkalic basalt would be dominant. Isotopic ratios for both nonvolatile elements and noble gases may approach those of MORB. However, relatively primitive ratios may still be observed in some cases, since the addition of diapir components would dominate the isotopic ratios in the melt because of the mass balance between melts from diapirs and the surrounding materials. The very high $^3\text{He}/^4\text{He}$ ratios for the Kula Volcanics of Haleakala Volcano (Kaneoka and Takaoka, 1980; Rison and Craig, 1983) might have originated in such a way.

The final stage of Hawaiian volcanism is identified as the posterosional stage, which occurs after a period of volcanic quiescence. In this stage, no contribution from diapirs is expected because the volcano has migrated far away from the hot spot. At the bottom of the oceanic lithosphere, some melt remains after producing alkalic basalt. The addition of conductive heat from the initial thermal anomaly together with the release of some latent heat during solidification of melt may form some melt pockets under a volcano that produce alkalic and or nephelinitic rocks. In this case, the source materials would be mostly the *M*-type material in the uppermost asthenosphere. Thus, it is reasonable to conjecture that such volcanic rocks would show values for isotopic ratios, including $^3\text{He}/^4\text{He}$ ratios, very similar to those of MORB. The Honolulu Volcanics on Oahu Island and the Hana Volcanics at Haleakala Volcano, Maui Island, are good examples of such late volcanism (fig. 27.9D).

This model is rather qualitative, and the major scheme is not seriously different from those which have been proposed before. Furthermore, the present scheme shows only a general trend; small variations would occur on a local scale. However, it is worth noting that voluminous outpourings of Hawaiian volcanoes do not always reflect direct material supply from the hot-spot source but may rather reflect the combined effects of material transfer plus heat supply. The occurrence of a partially melted zone under the lithosphere would play an important role in controlling the activity and characteristics of Hawaiian volcanoes. The occurrence of such a partially melted zone might be related to the conductive transfer of heat from the deeper part of the Earth together with the existence of the overlying, less conductive lithosphere acting as an insulator.

To clarify further and in greater detail the evolutionary scheme for Hawaiian volcanism, more systematic isotopic surveys, both of the noble gases and of other elements are required.

REFERENCES CITED

- Allègre, C.J., Staudacher, T., Sarda, P. and Kruz, M., 1983, Constraints on evolution of Earth's mantle from rare gas systematics: *Nature*, v. 303, no. 5920, p. 762-766.
- Anderson, D.L., 1982, Isotopic evolution of the mantle: a model: *Earth and Planetary Science Letters*, v. 57, no. 1, p. 13-24.
- Barger, K.E., and Jackson, E.D., 1974, Calculated volumes of individual shield volcanoes along the Hawaiian-Emperor Chain: *Journal of Research of the U.S. Geological Survey*, v. 2, p. 545-550.
- Chen, C.-Y., and Frey, F.A., 1983, Origin of Hawaiian tholeiite and alkalic basalt: *Nature*, v. 302, no. 5911, p. 785-789.

- Craig, H., and Lupton, J.E., 1976, Primordial neon, helium and hydrogen in oceanic basalts: *Earth and Planetary Science Letters*, v. 31, no. 3, p. 369-385.
- Dalrymple, G.B., and Moore, J.G., 1968, Argon-40: excess in submarine pillow basalts from Kilauea Volcano, Hawaii: *Science*, v. 161, no. 3846, p. 1132-1135.
- DePaolo, D.J., 1981, Nd isotopic studies: some new perspectives on earth structure and evolution: *Eos, Transactions, American Geophysical Union*, v. 62, no. 14, p. 137-140.
- Dymond, J., and Hogan, L., 1978, Factors controlling the noble gas abundance patterns of deep sea basalts: *Earth and Planetary Science Letters*, v. 38, no. 1, p. 117-128.
- Fisher, D.E., 1983, Rare gases from the undepleted mantle?: *Nature*, v. 305, no. 5932, p. 298-300.
- Funkhouser, J.G., Fisher, D.E., and Bonatti, E., 1968, Excess argon in deep-sea rocks: *Earth and Planetary Science Letters*, v. 5, no. 2, p. 95-100.
- Funkhouser, J.G., and Naughton, J.J., 1968, Radiogenic helium and argon in ultramafic inclusions from Hawaii: *Journal of Geophysical Research*, v. 73, no. 14, p. 4601-4607.
- Gramlich, J.W., and Naughton, J.J., 1972, Nature of source material for ultramafic minerals from Salt Lake Crater, Hawaii, from measurement of helium and argon diffusion: *Journal of Geophysical Research*, v. 77, no. 17, p. 3032-3042.
- Hayatsu, A., and Waboso, C.E., 1982, The solubility of rare gases in rock melts: implications for K-Ar dating and Earth degassing: in *Abstracts for the Fifth International Conference on Geochronology, Cosmochronology and Isotope Geology*, p. 139.
- Hennecke, E.W., and Manuel, O.K., 1975, Noble gases in Hawaiian xenoliths: *Nature*, v. 257, no. 5529, p. 778-780.
- Jackson, E.D., and Wright, T.L., 1970, Xenoliths in the Honolulu Volcanic Series, Hawaii: *Journal of Petrology*, v. 11, part 2, p. 405-430.
- Jacobsen, S.B., and Wasserburg, G.J., 1979, The mean age of mantle and crustal reservoirs: *Journal of Geophysical Research*, v. 84, no. B13, p. 7411-7427.
- Kaneoka, I., 1980, Rare gas isotopes and mass fractionation: an indicator of gas transport into or from a magma: *Earth and Planetary Science Letters*, v. 48, no. 2, p. 284-292.
- , 1983, Noble gas constraints on the layered structure of the mantle: *Nature*, v. 302, no. 5910, p. 689-700.
- Kaneoka, I., and Takaoka, N., 1978, Excess ^{129}Xe and high $^3\text{He}/^4\text{He}$ ratios in olivine phenocrysts of Kapuho lava and xenolithic dunites from Hawaii: *Earth and Planetary Science Letters*, v. 39, no. 3, p. 382-386.
- , 1980, Rare gas isotopes in the Hawaiian ultramafic nodules and volcanic rocks: constraints on genetic relationships: *Science*, v. 208, no. 4450, p. 1366-1368.
- , 1985, Noble-gas state in the Earth's interior—some constraints on the present state: *Chemical Geology*, v. 52, no. 1, p. 75-95.
- Kaneoka, I., and Takaoka, N., and Aoki, K., 1978, Rare gases in mantle-derived rocks and minerals, in *Alexander, E.C., Jr., and Ozima, M., eds., Terrestrial Rare Gases*: Tokyo, Center for Academic Publications, p. 71-84.
- Kaneoka, I., and Takaoka, N., and Clague, D.A., 1983, Noble gas systematics for coexisting glass and olivine crystals in basalts and dunite xenoliths from Loihi Seamount: *Earth and Planetary Science Letters*, v. 66, p. 427-437.
- Kurz, M.D., Jenkins, W.J., and Hart, S.R., 1982, Helium isotopic systematics of oceanic islands and mantle heterogeneity: *Nature*, v. 297, no. 5861, p. 43-47.
- Kurz, M.D., Jenkins, W.J., Hart, S.R., and Clague, D., 1983, Helium isotopic variations in volcanic rocks from Loihi Seamount and the Island of Hawaii: *Earth and Planetary Science Letters*, v. 66, p. 388-406.
- Kyser, T.K., and Rison, W., 1982, Systematics of rare gas isotopes in basic lavas and ultramafic xenoliths: *Journal of Geophysical Research*, v. 87, no. B7, p. 5611-5630.
- Lanphere, M.A., and Dalrymple, G.B., 1980, Age and strontium isotopic composition of the Honolulu Volcanic Series, Oahu, Hawaii: *American Journal of Science*, v. 280-A, part 2, p. 736-751.
- Lovering, J.F., and Richards, J.R., 1964, Potassium-argon age study of possible lower-crust and upper-mantle inclusions in deep-seated intrusions: *Journal of Geophysical Research*, v. 69, no. 22, p. 4895-4901.
- Lupton, J.E., 1983, Terrestrial inert gases: isotope tracer studies and clues to primordial components in the mantle: *Annual Review of the Earth and Planetary Sciences*, v. 11, p. 371-414.
- Lupton, J.E., and Craig, H., 1975, Excess ^3He in oceanic basalts: evidence for terrestrial primordial helium: *Earth and Planetary Science Letters*, v. 26, no. 2, p. 133-139.
- Macdonald, G.A., 1968, Composition and origin of the Hawaiian lavas: *Memoir of the Geological Society of America*, no. 116, p. 477-522.
- Moore, J.G., Clague, D.A., and Normark, W.R., 1982, Diverse basalt types from Loihi Seamount, Hawaii: *Geology*, v. 10, no. 2, p. 88-92.
- Morgan, W.J., 1971, Convection plumes in the lower mantle: *Nature*, v. 230, no. 5288, p. 42-43.
- Noble, C.S., and Naughton, J.J., 1968, Deep-ocean basalts. Inert gas content and uncertainties in age dating: *Science*, v. 162, no. 3850, p. 265-269.
- O'Nions, R.K., Hamilton, P.J., and Evensen, N.M., 1977, Variations in $^{143}\text{Nd}/^{144}\text{Nd}$ and $^{87}\text{Sr}/^{86}\text{Sr}$ ratios in oceanic basalts: *Earth and Planetary Science Letters*, v. 34, no. 1, p. 13-22.
- Ozima, M., and Podosek, F.A., 1983, Noble gas geochemistry: Cambridge University Press, 367 p.
- Ozima, M., and Zashu, S., 1983, Noble gases in submarine pillow volcanic glasses: *Earth and Planetary Science Letters*, v. 62, no. 1, p. 24-40.
- Rison, W., and Craig, H., 1983, Helium isotopes and mantle volatiles in Loihi Seamount and Hawaiian Island basalts and xenoliths: *Earth and Planetary Science Letters*, v. 66, p. 407-426.
- Sen, G., 1983, A petrologic model for the constitution of the upper mantle and crust of the Koolau shield, Oahu, Hawaii, and Hawaiian magmatism: *Earth and Planetary Science Letters*, v. 62, no. 2, p. 215-228.
- Shimizu, N., 1975, Geochemistry of ultramafic inclusions from Salt Lake Crater, Hawaii and from Southern African kimberlites: *Physical Chemistry of the Earth*, v. 9, p. 655-669.
- Staudigel, H., Zindler, A., Hart, S.R., Leslie, T., Chen, C.-Y., and Clague, D.A., 1984, The isotopic systematics of a juvenile intraplate volcano: Pb, Nd, and Sr isotope ratios of basalts from Loihi Seamount, Hawaii: *Earth and Planetary Science Letters*, v. 69, no. 1, p. 13-19.
- Wilson, J.T., 1963, A possible origin of the Hawaiian Islands: *Canadian Journal of Physics*, v. 41, p. 863-870.
- Wright, T.L., 1984, Origin of Hawaiian tholeiite: a metasomatic model: *Journal of Geophysical Research*, v. 89, no. B5, p. 3233-3252.