



## ORIGIN OF THE WEST MAUI VOLCANIC ROCKS INFERRED FROM Pb, Sr, AND Nd ISOTOPES AND A MULTICOMPONENT MODEL FOR OCEANIC BASALT

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### ABSTRACT

Isotope ratios of Pb, Sr, and Nd for West Maui volcanic rocks fall within the range of those from other volcanoes of the Kea trend. Shield-building and caldera-filling rocks have indistinguishable Sr and Nd isotope ratios, but the latter rocks have slightly more radiogenic Pb. Posterosional volcanic rocks have lower <sup>87</sup>Sr/<sup>86</sup>Sr ratios and less radiogenic Pb than underlying rocks. Isotope data for West Maui, as well as those for other Hawaiian volcanoes, are not adequately explained by binary mixing, but must be accounted for by mixing of at least three end members: the oceanic crust (Kilauea end member), the depleted upper mantle (posterosional end member), and a deep-mantle plume (Koolau end member). Isotope data thus far reported for oceanic basalt can also be described by a three-component mixing model: the depleted upper mantle (MORB source), Austral-St. Helena source, and recycled oceanic and continental crust (plumes?). However, data trends for oceanic basalt from individual areas suggest that mantle plumes may have only moderately radiogenic Pb, Sr, and Nd, requiring that such basalt was derived by mixing of these plumes with the chemically heterogeneous upper mantle (at least seven isotopically different components). Lithospheric contamination may also be involved.

### INTRODUCTION

The geology and geophysics of the Hawaiian Islands have been extensively studied in order to understand intraplate volcanism and the formation of linear island chains. It has been thought for some time that the Emperor-Hawaiian Chain was formed by the north-westward motion of the Pacific plate over a fixed mantle hot spot (Wilson, 1963). It follows from the hot-spot hypothesis that the volcanoes of the chain are progressively older northwestward from the currently active volcanoes of Kilauea, Mauna Loa, and Loihi. The progression of ages has been well documented by K-Ar age measurements of Hawaiian rocks (McDougall, 1964; Dalrymple and others, 1973; Dalrymple and Clague, 1976).

The loci of the Hawaiian volcanoes follow two parallel curves (fig. 26.1): the Loa (southwestern) and Kea (northeastern) trends (Jackson and others, 1972, and references therein). From youngest

to oldest, volcanoes on the Loa trend are Loihi, Mauna Loa (Hawaii), Hualalai (Hawaii), Kahoolawe, Lanai, West Molokai, Koolau (Oahu), Waiane (Oahu), and Kauai; those on the Kea trend are Kilauea (Hawaii), Mauna Kea (Hawaii), Kohala (Hawaii), Haleakala (East Maui), West Maui, and East Molokai.

Hawaiian volcanoes usually evolve in three stages: a shield-building stage, producing mostly tholeiite, a caldera-filling stage (alkalic rocks), and a posterosional stage (nephelinitic rocks) (Macdonald, 1968). However, recent studies of Loihi reveal a seamount stage in which basanite and alkali basalt with subordinate tholeiite are erupted before the shield-building tholeiitic stage (Moore and others, 1982). Although the earlier three-stage division may be too simple, subaerial Hawaiian rocks can be conveniently subdivided according to the three major stages. The two volcanoes on Maui, Haleakala in the east and the smaller West Maui Volcano, have developed to the posterosional stage. Those on the Island of Hawaii, however, 100 km to the southeast, have not yet erupted lava of posterosional type.

Strontium isotopes have been used for two decades to trace the origin of Hawaiian volcanic rocks (see, for example, Lessing and Catanzaro, 1964; Hamilton, 1965; Powell and Delong, 1966). From the Sr isotope difference between the Koolau Basalt (shield-building stage) and Honolulu Volcanics (posterosional alkalic-nephelinitic stage) on Oahu, Powell and Delong (1966) suggested that the two volcanic sequences were not derived from a common parent magma and that the upper mantle source regions are chemically and isotopically heterogeneous. Later isotope studies of Pb, Nd, Sr, and Hf (for example, Lanphere and Dalrymple, 1980; Stille and others, 1983; Roden and others, 1984) support this contention.

Chen and Frey (1983), in an extensive geochemical and isotopic study of Haleakala, found inverse correlations between <sup>143</sup>Nd/<sup>144</sup>Nd and Sm/Nd abundance ratios and between <sup>87</sup>Sr/<sup>86</sup>Sr and Rb/Sr abundance ratios. These findings are consistent with mixing of two kinds of melt, one derived from an enriched mantle and another formed by melting of a fraction (less than two percent of the total volume) of midocean-ridge basalt source.

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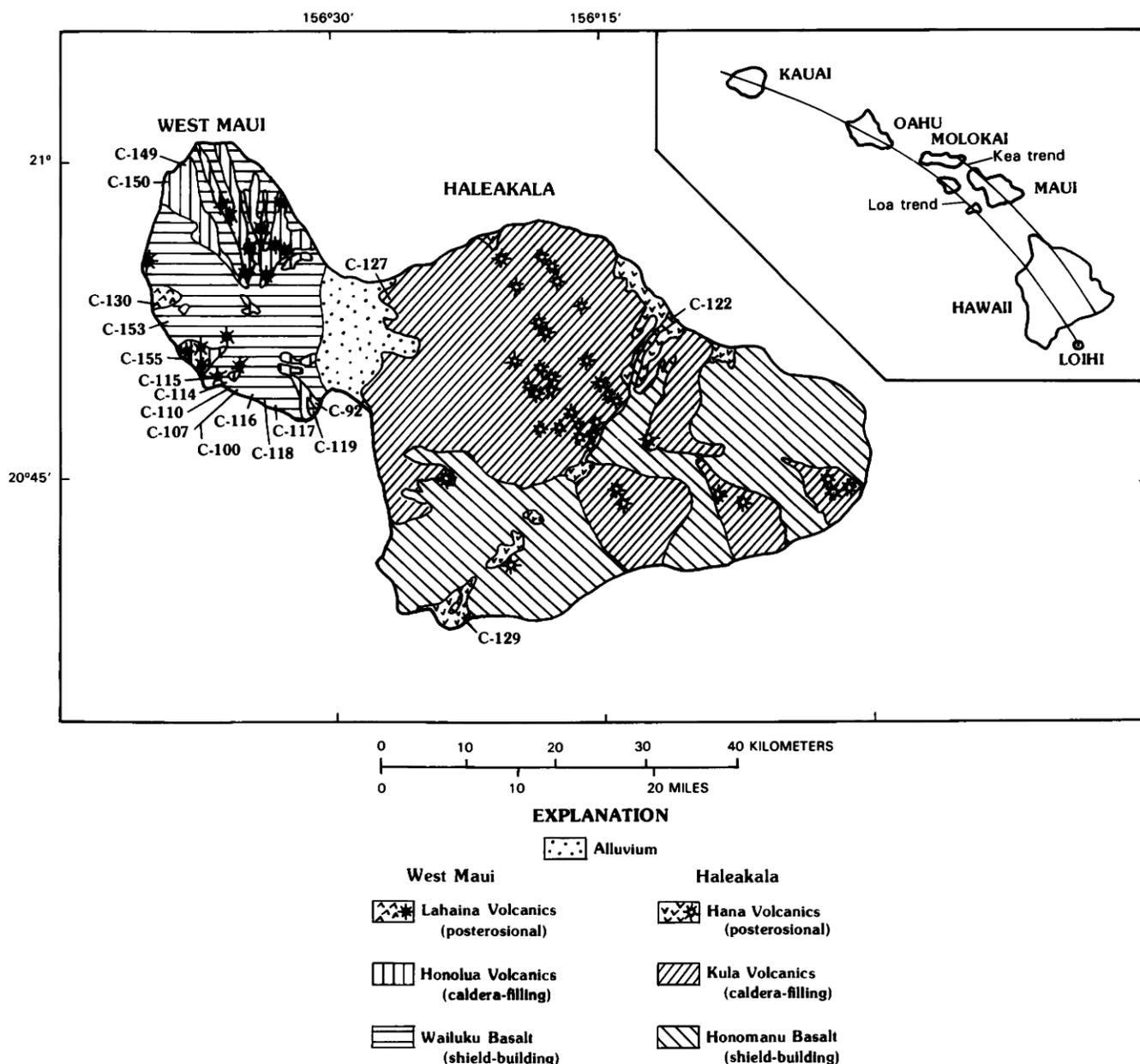


FIGURE 26.1.—Geologic map of the Island of Maui, simplified from Macdonald (1949). Sample locations are shown by numbers (for exact location, see Macdonald and Katsura, 1964, table 6 and 7; and this study). Inset shows location of Kea and Loa lineaments of volcanic centers in Hawaiian Islands.

Stille and others (1983) analyzed Pb, Sr, Nd, and Hf isotopes in volcanic rocks from Oahu and found that the isotope systematics can be best explained by a model involving mixing of three or more independent magma sources. Staudigel and others (1984) concurred with this view in an isotopic study of Loihi Seamount volcanic rocks. In order to place additional isotopic constraints on Hawaiian basalt petrogenesis, we measured Pb, Sr, and Nd isotopes in volcanic rocks from West Maui Volcano. The results were reported in brief previously (Hegner and others, 1986). In this paper we report concentrations of U, Th, Pb, Rb, Sr, Sm,

and Nd, as well as Pb, Sr, and Nd isotopic ratios, in Maui volcanic rocks, and we discuss the origins of both Hawaiian volcanic rocks and oceanic basalt in general.

Stratigraphic nomenclature used in this report is that from Langenheim and Clague (chapter I, part II).

#### ACKNOWLEDGMENTS

We are indebted to the late Dr. G. A. Macdonald for the C-series basalt samples. The supplemental samples were kindly

provided by Dr. J.A. Philpotts from the Macdonald collection. One of the authors (M.T.) thanks Drs. J.M. Sinton and Alan Zindler for their help with the field work.

## GEOLOGY AND SAMPLES

The geology of Maui has been described in Stearns and Macdonald (1942) and Macdonald (1949). The Island of Maui comprises two volcanoes, both of which lie on the Kea trend. These volcanoes are the youngest in which post-erosional volcanics have erupted (fig. 26.1). The West Maui Volcano is wholly prehistoric and about one-fourth the size of the Haleakala (East Maui) Volcano, which last erupted in A.D. 1790. The Wailuku Basalt (formerly Wailuku Volcanic Series) represents the shield-building stage of West Maui and is composed mostly of tholeiite and intercalated alkali basalts with minor ankaramite near the top. Caldera-filling lava of the Honolua Volcanics (formerly Honolua Volcanic Series) is dominantly differentiated alkalic rock ranging from mugearite to soda trachyte in composition. After cessation of the Honolua volcanism and prolonged erosion, basanitic lava of the Lahaina Volcanics (formerly Lahaina Volcanics Series) was erupted.

The shield-building lava of Haleakala has been designated the Honomanu Basalt, and the composition of exposed rocks ranges only from tholeiitic olivine basalt to oceanite. The Honomanu Basalt is overlain by the Kula Volcanics, which are dominantly composed of hawaiite. The Hana Volcanics represents Haleakala's posterosional volcanism; this unit is less silica-saturated and consists of alkalic olivine basalt and basanitoid in contrast to the basanite-dominated Lahaina Volcanics of West Maui.

We analyzed 6 samples of tholeiitic basalt and 2 of alkali olivine basalt collected from the Wailuku Basalt (shield building), 8 samples of alkalic differentiates from the Honolua Volcanics (caldera filling), and 3 samples of basanite from the Lahaina Volcanics (posterosional). Sample numbers prefixed by a "C" were collected by the late Dr. Gordon A. Macdonald. Rock names and locations of samples C-92 to C-130 are listed by Macdonald and Katsura (1964) and are shown in figure 26.1. Samples collected by Macdonald from other sites have the following identities and places of collection (G.A. Macdonald, written commun., 1966): C-149, mugearite, transitional toward trachyte, roadcut at top of grade on

west side of Pohakupule Gulch, West Maui, Honolua Volcanics; C-150, mugearite, transitional toward trachyte; dikelike body cutting mugearite flow in roadcut at berm where road starts to descend the west side of Honolua Gulch, West Maui, Honolua Volcanics; C-153, hawaiite, boulder in Kanaha Stream below Paupau Hill, just above Kelaweia Village on outskirts of Lahaina, West Maui, Honolua Volcanics; C-155, mugearite, south branch of Makamakaole Gulch, 25 feet southeast of culvert under road, West Maui, Honolua Volcanics.

Samples designated "HMT" series were collected by one of us (M.T.) under the guidance of Dr. J.M. Sinton, University of Hawaii. Sample localities are the same as those of the C-series of Macdonald and Katsura (1964, tables 6 and 7), except HMT-2b, which is an alkalic olivine basalt from above an ash bed separating it from tholeiite (HMT-2a). Three samples from Haleakala Volcano were analyzed for comparison with Chen and Frey's (1983) data; they include a tholeiite from the Honomanu Basalt (shield building), a hawaiite of the Kula Volcanics (caldera filling) and a basanitoid of the Hana Volcanics (posterosional). Haleakala samples are also from the late Dr. Macdonald's collection. K-Ar ages by McDougall (1964) are  $1.20 \pm 0.02$  Ma for olivine basalt from the upper part of the Wailuku Basalt,  $1.16 \pm 0.01$  Ma for mugearite and trachyte of the Honolua Volcanics, and  $0.44-0.86$  Ma for alkalic rocks of the Kula Volcanics of Haleakala. Therefore, the isotopic data obtained from these samples require no age correction.

## EXPERIMENTAL PROCEDURES

About 100 mg of each sample were dissolved in HF and HNO<sub>3</sub> with spike mixtures of <sup>233</sup>U-<sup>236</sup>U-<sup>230</sup>Th-<sup>205</sup>Pb, <sup>87</sup>Rb-<sup>84</sup>Sr, and <sup>150</sup>Nd-<sup>149</sup>Sm in a PFA Teflon vial. First, Pb was separated in a 1N HBr medium using a single anion-exchange resin column of 100 μL volume (Tatsumoto and Unruh, 1976). The effluent was converted into 7N HNO<sub>3</sub> solution. The U and Th were then separated using an anion-exchange resin column of 600 μL volume. The effluent was converted into HCl solution, and Rb, Sr, and rare-earth elements (REE) were separated using a cation-exchange resin column of 4 mL volume. The Sr was purified using a 400 μL cation-exchange resin column, and the Nd and Sm were separated using 2-methylactic acid in a 600 μL cation-exchange

TABLE 26.1.—Standard analyses for selected elements and isotope ratios  
[Uncertainties refer to last significant figure(s) and represent 95 percent confidence level]

Sample	<sup>143</sup> Nd/ <sup>144</sup> Nd	<sup>87</sup> Sr/ <sup>86</sup> Sr	Rb (ppm)	Sr (ppm)	Sm (ppm)	Nd (ppm)
La Jolla standard	0.511865±10	--	--	--	--	--
NBS SRm-987 standard	--	.710255±30	--	--	--	--
BCR-1	.512654±15	.705009±20	46.47	329.6	6.64	29.06

resin column (Nakamura and others, 1976).

Pb isotopes were measured using single Re filaments with a phosphoric acid-silica gel in a NBS tandem-type mass spectrometer equipped with a  $10^{11}\text{-}\Omega$  resistor. The beam intensities of  $^{206}\text{Pb}$  ranged from  $3 \times 10^{-12}$  to  $1 \times 10^{-10}$  A. A mass-fractionation correction of  $0.12 \pm 0.03$  percent per atomic mass unit was applied to the raw Pb data. The U, Th, and Rb were measured using Re triple filaments in the same tandem mass spectrometer. The Sr, Nd, and Sm isotopic analyses were done in a VG 54R mass spectrometer using oxidized Ta filaments for Sr and Re-Ta triple filaments for Nd and Sm (Nakamura and others, 1976). Sr and Nd

isotopic compositions of samples were deduced from the spiked runs by normalizing  $^{86}\text{Sr}/^{88}\text{Sr} = 0.1194$  and  $^{146}\text{Nd}/^{144}\text{Nd} = 0.7219$ , respectively, using a linear fractionation law (Dodson, 1970). The NBS Sr standard SRM-982 and La Jolla Nd standard were measured in runs of varying (normal  $^{86}\text{Sr}$ )/(spike  $^{84}\text{Sr}$ ) and (normal  $^{144}\text{Nd}$ )/(spike  $^{150}\text{Nd}$ ) ratios (up to unity), and we found that spiked runs agree with unspiked runs to within 0.005 percent. Four samples (powdered) that showed somewhat high  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios compared to other samples were leached with 6N HCl at  $\approx 80^\circ\text{C}$  for 10 h in order to eliminate alteration effects (O'Nions and others, 1977; Zindler and others, 1984). The leached samples were

TABLE 26.2.—Selected elemental concentration and elemental and isotope ratios in samples of Maui volcanic rocks

Sample		Rb (ppm)	Sr	Rb/Sr	Sm (ppm)	Nd	Sm/Nd	Pb	U (ppm)	Th	$\frac{^{238}\text{U}}{^{204}\text{Pb}}$	$\frac{^{232}\text{Th}}{^{238}\text{U}}$
<b>WEST MAUI</b>												
<b>Lahaina Volcanics</b>												
C-130	basanite	8.4	245	0.034	4.59	19.91	0.23	1.38	0.39	1.16	17.7	3.1
C-115	basanite	43.8	705	.062	4.82	25.94	.19	3.03	1.40	4.05	28.9	3.0
HMT 79-5	(C-115)	40.5	667	.061	7.97	39.42	.20	4.51	1.54	5.60	21.4	3.7
<b>Honolua Volcanics</b>												
C-155	mugearite	49.8	1,331	.037	17.62	87.05	.20	4.27	1.22	5.56	18.1	4.7
	--	--	--	--	--	--	--	1.18	5.33	--	4.7	--
C-153	hawaiite	8.7	499	.017	9.06	40.33	.22	1.59	.46	1.51	18.6	3.4
C-150	mugearite	44.1	1,149	.038	15.15	71.00	.21	4.23	1.66	78.05	24.7	4.4
C-149	mugearite	48.2	1,041	.046	14.95	64.91	.23	5.04	2.07	6.14	25.9	3.1
C-128	trachyte	59.1	400	.148	14.63	78.24	.19	7.27	1.84	7.24	16.0	4.1
C-116	trachyte	64.5	432	.149	20.33	98.84	.21	6.15	2.35	7.45	24.3	3.0
	--	--	--	--	--	--	--	2.55	8.42	--	3.4	--
C-92	trachyte	50.0	1,040	.048	15.67	81.33	.19	4.56	1.86	4.84	25.7	2.7
<sup>1</sup> HMT79-2b	alkali basalt	3.4	525	.006	8.58	31.05	.28	2.04	.29	1.46	8.9	5.3
<b>Wailuku Basalt</b>												
HMT 79-3	tholeiitic olivine basalt (C-120)	.6	360	.002	5.51	20.51	.27	.95	.11	.70	7.3	6.1
C-119	tholeiitic olivine basalt	7.7	489	.016	8.10	25.94	.31	1.16	.35	1.15	19.1	3.4
C-118	alkalic olivine basalt	1.4	254	.006	4.20	16.49	.25	1.20	.23	.61	11.8	2.8
C-117	tholeiitic olivine basalt	3.6	382	.009	4.14	16.96	.24	1.14	.22	.76	12.1	3.6
<sup>1</sup> C-114	alkalic olivine basalt	3.1	739	.004	6.88	27.41	.25	1.37	.27	1.23	12.5	4.7
C-110	tholeiite	4.2	319	.013	4.93	17.92	.27	2.85	.23	.62	5.0	2.8
<sup>1</sup> C-107	tholeiite	5.1	356	.014	6.97	25.99	.27	.96	.27	.77	17.8	2.9
C-100	tholeiite	3.1	346	.009	3.45	13.07	.26	.87	.20	.68	14.5	3.5
<b>EAST MAUI (HALEAKALA)</b>												
<b>Hana Volcanics</b>												
C-129	basanitoid	23.9	523	.046	6.16	27.58	.22	1.63	.61	2.09	23.6	3.5
<sup>2</sup> H65-11	alkalic basalt	18.8	582	.032	6.01	26.1	.23	--	--	2.6	--	--
<b>Kula Volcanics</b>												
C-127	hawaiite	29.8	831	.036	7.53	37.40	.20	2.27	.97	3.05	26.8	3.3
<sup>2</sup> H65-4	alkalic basalt	19.9	616	.032	6.83	30.7	.22	--	--	2.1	--	--
<b>Honomanu Basalt</b>												
C-122	tholeiitic olivine basalt	4.72	225	.021	4.23	14.71	.29	1.08	.20	.51	11.6	2.6
<sup>2</sup> C-122		4.11	207	.020	3.73	12.3	.30	--	--	.8	--	--

<sup>1</sup>Analyses from Unruh and others (oral commun., 1985).

<sup>2</sup>Chen and Frey (1983).

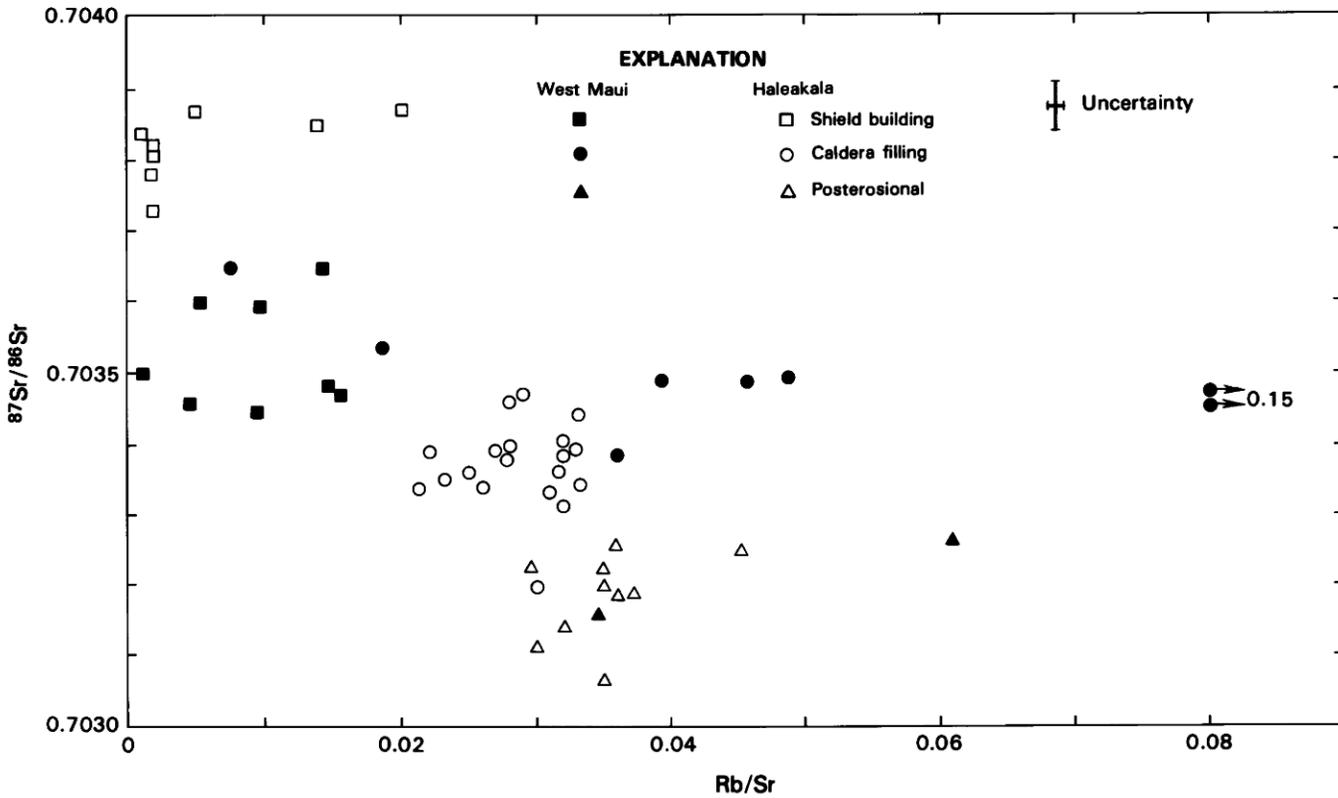


FIGURE 26.2.—Plot of  $^{87}\text{Sr}/^{86}\text{Sr}$  ratio against Rb/Sr ratio for samples of volcanic rocks from Maui. Data for Haleakala from Chen and Frey (1983) included. Uncertainty bars apply to all data.

analyzed only for Sr isotopic composition. The standard runs during this study averaged  $^{143}\text{Nd}/^{144}\text{Nd} = 0.511865 \pm 0.000010$  for the La Jolla Nd, and  $^{87}\text{Sr}/^{86}\text{Sr} = 0.710265 \pm 0.000030$  for NBS SRM-987 (table 26.1). No instrumental bias corrections were applied to the isotope data in table 26.3. The USGS rock standard BCR-1 was analyzed once during the course of this study, and the results are listed in table 26.1. Blanks for the analyses are in ng: Pb, 0.1–0.15; U, <0.01; Th, <0.01; Rb, 0.1–0.2; Sr, 0.5–1; Sm, <0.03; Nd, <0.1.

**RESULTS**

**TRACE-ELEMENT CONCENTRATIONS**

The elemental concentration data for volcanic rocks from West Maui and Haleakala are listed in table 26.2. Data for Haleakala samples from Chen and Frey (1983) are also included. The trace-element abundances of Rb, Sr, Sm, Nd, Pb, U, and Th of West Maui and Haleakala are similar to those in other subaerial Hawaiian basalts (Tatsumoto, 1978; Clague and Frey, 1982; Roden and others, 1984).

Plots of the ratios  $^{87}\text{Sr}/^{86}\text{Sr}$  versus Rb/Sr and  $^{143}\text{Nd}/^{144}\text{Nd}$  versus Sm/Nd for West Maui and Haleakala samples are shown in figures 26.2 and 26.3, which include data from Haleakala obtained by Chen and Frey (1983). The variation in Rb/Sr ratios of West

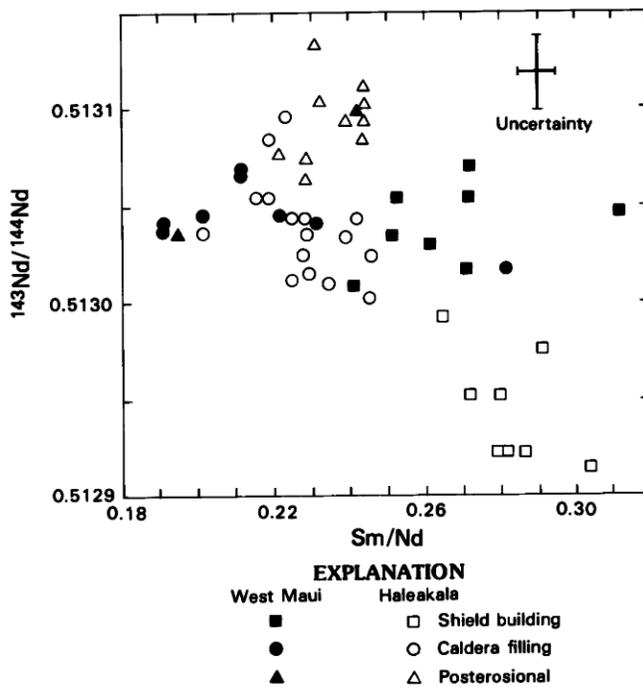


FIGURE 26.3.—Plot of  $^{143}\text{Nd}/^{144}\text{Nd}$  ratio against Sm/Nd ratio for samples of volcanic rock from Maui. Uncertainty bars apply to all data.

Maui volcanic rocks (fig. 26.2) is large compared to that of the Haleakala volcanic rocks, because West Maui samples contain highly fractionated alkalic rocks such as trachyte that do not occur at Haleakala. Sm/Nd ratios (fig. 26.3) of West Maui vary from 0.31 to 0.19 and are similar to those of the Haleakala volcanic rocks.

The lowest values are for trachyte and basanite, and Rb/Sr ratios of these rocks are correspondingly high. In contrast,  $^{143}\text{Nd}/^{144}\text{Nd}$  values of West Maui volcanics are almost constant. Chen and Frey (1983) discovered profound negative correlations between parent/daughter abundance ratios and radiogenic isotope abundances in the

TABLE 26.3.—Isotope data for Sr, Nd, and Pb in samples of Maui volcanic rocks

[Uncertainties correspond to last significant figure(s) and are 95 percent confidence level. Uncertainties for lead are limited to a few values because of correction uncertainties for mass fractionation]

Sample	Rock type	$^{87}\text{Sr}/^{86}\text{Sr}$	$\epsilon_{\text{Sr}}$	$^{143}\text{Nd}/^{144}\text{Nd}$	$\epsilon_{\text{Nd}}$	$^{206}\text{Pb}/^{204}\text{Pb}$	$^{207}\text{Pb}/^{204}\text{Pb}$	$^{208}\text{Pb}/^{204}\text{Pb}$
<b>WEST MAUI</b>								
<b>Lahaina Volcanics</b>								
C-130	basanite	0.70315±3	-22.0	0.513086±28 .513107±10	8.7	18.169±11	15.452±14	37.750±49
C-115	basanite	.70326±2	-20.4	.513034±20	7.7	18.191±11	15.460±14	37.833±49
HMT79-5	(C-115)	.70325±4 .70327±3	--	.513031±18	--			
<b>Honolua Volcanics</b>								
C-155	mugearite	.70343±4	-18.0	.513043±8	7.9	18.513±11	15.494±14	38.050±49
C-153	hawaiite	.70353±3	-16.6	.513041±7	7.8	18.447±11	15.456±14	37.917±49
C-150	mugearite	.70349±3	-17.2	.513065±9	8.3	18.502±11	15.483±14	38.033±49
C-149	mugearite	.70349±3	-17.2	.513037±64	7.7	18.509±11	15.489±14	38.030±49
C-128	trachyte	.70347±3	-17.5	.513034±20	7.7	18.492±11	15.476±14	38.008±49
C-116	trachyte	.70348±5	-17.3	.513064±7	8.3	18.523±11	15.471±14	37.961±49
C-92	trachyte	.70350±3	-17.0	.513035±16	7.7	18.482±11	15.466±14	37.957±49
HMT79-1a	(C-92)	.70355±4	--	.513033±20	--	18.487±11	15.462±14	37.947±49
HMT79-2b	alkali basalt	.70354±3	--	.513033±20	7.3	18.506±11	15.560±14	38.040±49
<sup>1</sup> HMT79-2b		.70352±4	--	.513005±16	--	18.474±16	15.535±16	37.974±49
<b>Mailuku Basalt</b>								
HMT79-3	tholeiitic olivine basalt (C-120)	.70349±4	-17.2	.513014±19	7.3	18.462±11	15.456±14	37.925±50
C-119	tholeiitic olivine basalt	.70347±4	-17.5	.513044±28	7.9	18.426±11	15.472±14	37.904±49
HMT79-2A	(C-119)	.70356±3	--	.513025±20		18.462±11	15.454±14	37.899±49
C-118	alkalic olivine basalt	.70364±4 <sup>3</sup> .70359±2	-- -15.8	.513049±12	8.0	18.258±11	15.471±14	37.762±49
C-117	tholeiitic olivine basalt	.70359±3 .70361±2	-15.8	.513007±15	7.2	18.404±15	15.429±14	37.940±49
<sup>1</sup> C-114	alkalic olivine basalt	.70346±4	-17.6	.513032±22	8.9	18.412±11	15.472±14	37.906±49
C-110	tholeiite	.70363±3 <sup>3</sup> .70365±3	-- -14.9	.513053±13	8.1	18.309±11	15.476±14	37.874±49
<sup>1</sup> C-107	tholeiite	.70348±3	-17.3	.513070±12	8.1	18.438±11	15.499±14	37.948±49
C-100	tholeiite	.70361±4 <sup>3</sup> .70344±3	-- -17.9	.513027±13	7.5	18.365±11	15.470±14	37.877±49
<b>EAST MAUI (HALEAKALA)</b>								
<b>Hana Volcanics</b>								
C-129	basanitoid	.70325±3	-20.6	.513070±13	8.4	18.259±11	15.453±14	37.828±49
<sup>2</sup> H65-1	alkalic basalt	.70314±4	-22.1	.51310±2	9.0	--	--	--
<b>Kula Volcanics</b>								
C-127	hawaiite	.70339±4	-18.6	.513030±15	7.6	18.351±11	15.468±14	37.938±49
<sup>2</sup> H65-4	alkalic basalt	.70334±3	-19.3	.51302±1	7.4	--	--	--
<b>Honomau Basalt</b>								
C-122	tholeiite olivine basalt	.70378±3 .70374±4 .70374±3	-13.1	.512976±16	6.6	18.303±11	15.452±14	37.924±49
<sup>2</sup> C-122		.70387±4	-11.8	.51291±2	5.3	--	--	--

<sup>1</sup>Analyses from Stille and others (1985).

<sup>2</sup>Chen and Frey (1983).

<sup>3</sup>On acid leached samples (6N HCl, 10 hours).

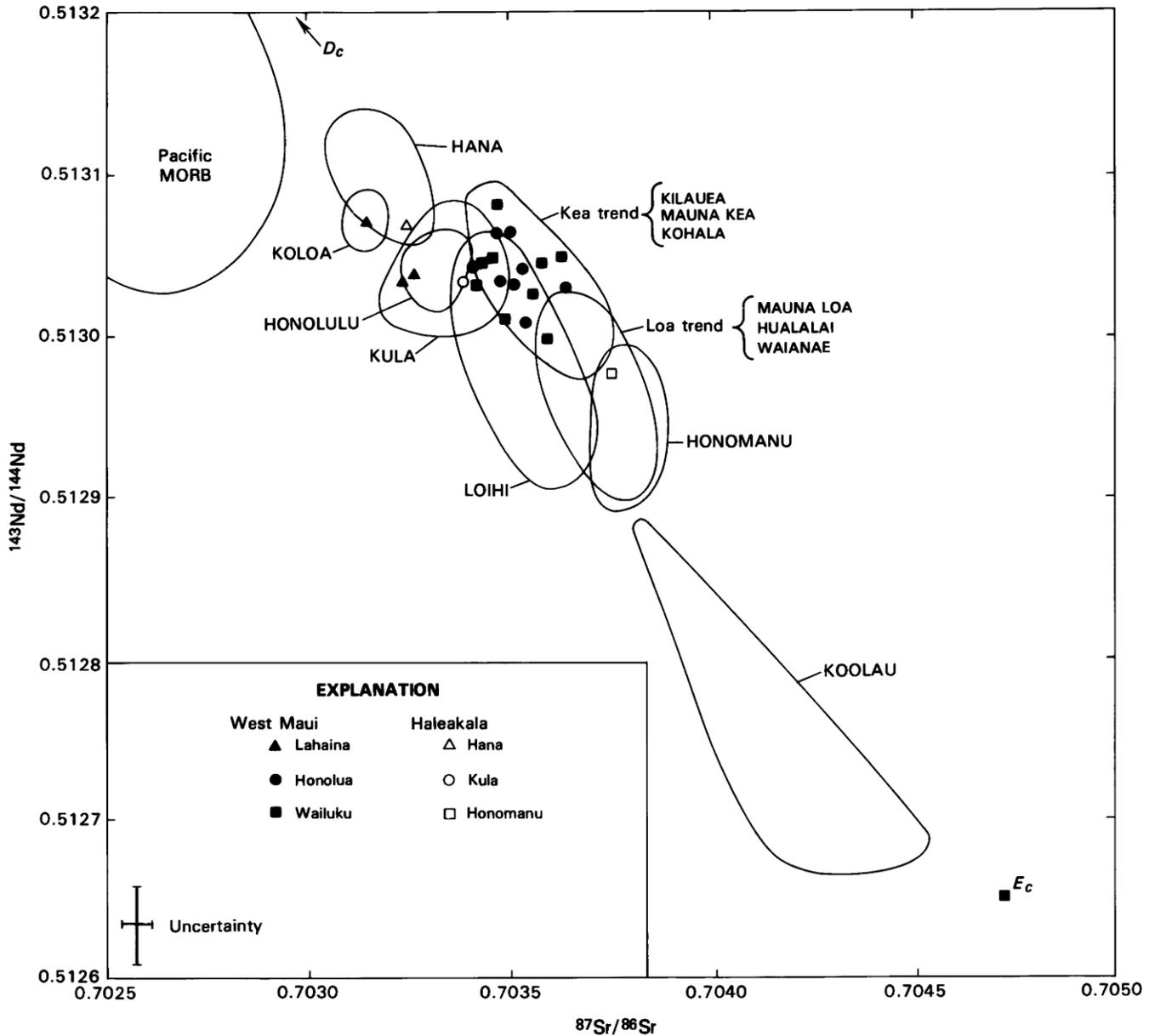


FIGURE 26.4.—Plot of  $^{87}\text{Sr}/^{86}\text{Sr}$  ratio against  $^{143}\text{Nd}/^{144}\text{Nd}$  ratio for Hawaiian basalt. Data for samples from Maui are shown by symbols. Data for Honomanu Basalt, Kula Volcanics, and Hana Volcanics of Haleakala from Chen and Frey (1983) are also shown. Also shown are fields for Koolau Basalt and Waianae Volcanics, Oahu; (Stille and others, 1983; Roden and others, 1984); Loihi (Staudigel and others, 1984); Koloa Volcanics, Kauai, (O'Nions and others, 1977); and volcanics from Island of Hawaii (Stille and others, in press). Data for posterosional basalt appears to be shifted slightly to left from main Hawaiian trend and close to East Pacific Ridge basalt (Pacific MORB) field.  $D_c$  and  $E_c$ , possible depleted and enriched members, respectively, in hypothetical two-component mixing model. Uncertainty bars apply to all individual symbols.

Rb-Sr and Sm-Nd systems. Although such negative correlations for West Maui samples are somewhat indistinct, our results clearly do not support the mantle-isochron concept of Brooks and others (1976), who proposed that a positive correlation of Rb/Sr with  $^{87}\text{Sr}/^{86}\text{Sr}$  corresponds to an age of mantle differentiation of 1.8 Ga. The trace-element concentrations of Maui tholeiite (this study; Chen and Frey, 1983) are distinctly higher than those of normal mid-ocean-ridge basalt (N-MORB) but are comparable to those in

enriched MORB (P-MORB) (see, for example, Tatsumoto and others, 1965; Cohen and O'Nions, 1982).

For seven samples from the Wailuku Basalt, West Maui, Rb and Sr concentrations and Rb/Sr ratios are similar to those of the Koolau Basalt, Oahu (Roden and others, 1984) (sample HMT 79-3 is omitted because of its weathered nature). However, the Rb concentrations and Rb/Sr ratios of Wailuku rocks appear to be distinctly lower than those of Loihi tholeiite (Frey and Clague,

1983; Staudigel and others, 1984), although their  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios are similar.

### ISOTOPIC COMPOSITIONS

The Sr, Nd, and Pb isotopic data for West Maui and Haleakala are listed in table 26.3. The Sr and Nd data of Haleakala by Chen and Frey (1983) are also included in the table. The Nd-Sr data of West Maui basalt are plotted on a Nd-Sr correlation diagram in figure 26.4.

The C-series samples generally appear to be fresh; however, a  $^{87}\text{Sr}/^{86}\text{Sr}$  value higher than 0.7036 in one of the three Wailuku samples proved to be related to alteration. Sample HMT 79-2b also shows slight alteration and a higher  $^{87}\text{Sr}/^{86}\text{Sr}$  value.

The  $^{87}\text{Sr}/^{86}\text{Sr}$  ratio of all samples from West Maui varies from 0.70315 to 0.70364, which correspond to an  $\epsilon_{\text{Sr}}$  range of 7.0 (0.07 percent) and is well within the range (0.7031–0.7046;  $\epsilon_{\text{Sr}}$  range, 21.3) of Hawaiian volcanic rocks so far reported (Lanphere and others, 1980; Chen and Frey, 1983; Stille and others, 1983, in press; Roden and others, 1984). The  $\epsilon_{\text{Sr}}$  is the deviation in parts per ten thousand from the present day bulk earth value of  $^{87}\text{Sr}/^{86}\text{Sr}=0.7047$  (O'Nions and others, 1977). Although the bulk-earth value has been debated (for example, Zindler and others, 1982), we use the  $\epsilon_{\text{Sr}}$  just for convenience to compare  $^{87}\text{Sr}/^{86}\text{Sr}$  variations in the present paper. The variation is smaller than that for the three Haleakala volcanic sequences (0.70307–0.70387;  $\epsilon_{\text{Sr}}$  range, 11.4) and the tholeiitic series of Koolau (0.70367–0.70458;  $\epsilon_{\text{Sr}}$  range, 12.9), but it is similar to that for the Loihi Seamount volcanics (0.7033–0.7037;  $\epsilon_{\text{Sr}}$  range, 5.7, Staudigel and others, 1984). Similar  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios (0.70346–0.70370) for West Maui volcanic rocks are also reported by Feigenson and others (1984). The shield-building Wailuku Basalt and the late-stage Honolua alkalic rocks from West Maui have an almost constant  $^{87}\text{Sr}/^{86}\text{Sr}$  ratio of  $0.70353 \pm 0.00010$ , corresponding to an  $\epsilon_{\text{Sr}}$  range of only 1.4. The  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios are distinctly smaller than those of Honomanu Basalt but slightly higher than those of Kula alkalic rocks from Haleakala. The  $^{87}\text{Sr}/^{86}\text{Sr}$  ratio of Lahaina basanite from West Maui is distinctly smaller than that of volcanic rocks of the earlier stages, but similar to those in the Hana Volcanics at Haleakala (fig. 26.2) and in the Honolulu Volcanics on Oahu (Stille and others, 1983).

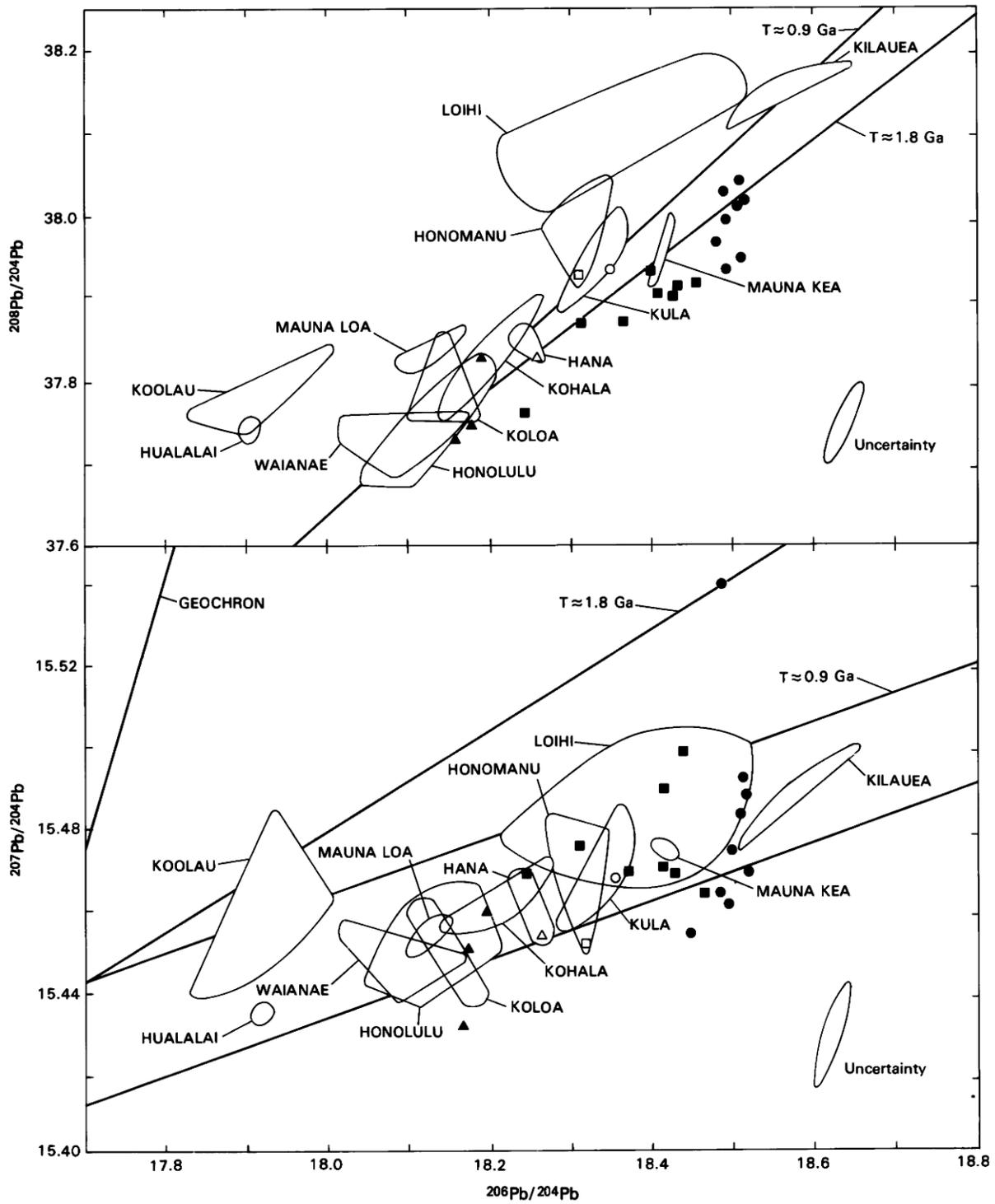
Chen and Frey (1983, fig. 1) reported that the  $^{143}\text{Nd}/^{144}\text{Nd}$  ratio in Haleakala rocks varies from 0.51313 to 0.51291. This variation corresponds to an  $\epsilon_{\text{Nd}}$  variation of 4.2, and Nd isotopic ratios are negatively correlated with Sm/Nd ratios. Like  $\epsilon_{\text{Sr}}$ ,  $\epsilon_{\text{Nd}}$  is the deviation in parts per ten thousand from the present day bulk-earth value of  $^{143}\text{Nd}/^{144}\text{Nd}=0.51264$  (DePaolo and Wasserburg, 1976). In contrast to Haleakala, the  $^{143}\text{Nd}/^{144}\text{Nd}$  ratio of all samples analyzed from West Maui exhibits almost constant values (0.51301 to 0.51309), which correspond to an  $\epsilon_{\text{Nd}}$  range of only 1.6. Furthermore,  $^{143}\text{Nd}/^{144}\text{Nd}$  ratios for shield-building Wailuku Basalt and alkalic Honolua Volcanics vary from only 0.51301 to 0.51307, these values correspond to a narrow range of  $\epsilon_{\text{Nd}}$  of 1.2.

The Sr and Nd ratios of West Maui volcanic rocks are inversely correlated and plot at the upper end of the Hawaiian data array shown in figure 26.4 (Stille and others, 1983, 1985; Roden and others, 1984; Staudigel and others, 1984). The Lahaina Volcanics (posterosional) has similar  $^{143}\text{Nd}/^{144}\text{Nd}$  ratios to the underlying rocks but smaller  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios. Consequently, the Lahaina data are shifted slightly to the left, close to the MORB field, and plot in the fields of Hana (Haleakala) and Honolulu (Oahu) data.

The Pb isotopic compositions of West Maui samples are shown in a diagram plotting  $^{206}\text{Pb}/^{204}\text{Pb}$  against  $^{207}\text{Pb}/^{204}\text{Pb}$  and  $^{208}\text{Pb}/^{204}\text{Pb}$  in figure 26.5. Note that the scale for  $^{207}\text{Pb}/^{204}\text{Pb}$  is expanded five times relative to those for  $^{206}\text{Pb}/^{204}\text{Pb}$  and  $^{208}\text{Pb}/^{204}\text{Pb}$ . The variation in Pb isotopic compositions (table 26.3) is partly caused by measurement errors associated with mass fractionation. However, the ratios of West Maui samples vary beyond measurement errors.

The Pb isotopic compositions of West Maui volcanic rocks fall within the field for Hawaiian volcanic rocks between Waianae (Oahu) and Kilauea (Hawaii); in contrast to the Sr and Nd isotopes; however, the West Maui rocks display the largest range in Pb isotopic composition of any single Hawaiian volcano so far determined except Loihi. On the other hand, Pb isotopic compositions of Honomanu, Kula, and Hana rocks from Haleakala show smaller variations compared to West Maui, although Nd and Sr show larger variations (Chen and Frey, 1983). However, only three samples from Haleakala were analyzed in the present study. The Pb data for Haleakala measured by Chen and Frey are not published and were compiled from Staudigel and others (1984, fig. 5), and the actual sample population is not known to us. The least radiogenic Pb in Wailuku samples is similar to those in the Haleakala samples, but Pb in upper Wailuku samples is more radiogenic. The Pb in the alkalic late-stage Honolua Volcanics is more radiogenic than that of the Wailuku samples showing oceanic-crust assimilation as discussed later. Alkali basalt sample HMT 79-2b has a significantly high  $^{207}\text{Pb}/^{204}\text{Pb}$  ratio compared to other samples (fig. 26.5). The data fall on the "1.8-Ga MORB trend" instead of the "0.9-Ga Hawaiian trend" in the plot of  $^{206}\text{Pb}/^{204}\text{Pb}$  versus  $^{207}\text{Pb}/^{204}\text{Pb}$  (Tatsumoto, 1978) and indicate a Pb-isotopic diversity in Hawaiian

FIGURE 26.5.—Plots of  $^{206}\text{Pb}/^{204}\text{Pb}$  ratio against  $^{207}\text{Pb}/^{204}\text{Pb}$  and  $^{208}\text{Pb}/^{204}\text{Pb}$  ratios for Hawaiian volcanic rocks. Data from Maui (this study), Island of Hawaii (Tatsumoto, 1978), Oahu (Stille and others, 1983), Koloa Volcanics, Kauai (Sun, 1980), and Loihi (Staudigel and others, 1984). Data of shield building and caldera-filling rocks from West Maui plot in large area similar to Loihi data but with lower  $^{208}\text{Pb}/^{204}\text{Pb}$ , whereas posterosional rocks are less radiogenic and plot close to Hana Volcanics (Haleakala) and Honolulu Volcanics (Oahu). Hawaiian basalt data fall on apparent isochron of  $\sim 0.9$  Ga, whereas some data plot near 1.8-Ga MORB trend. In upper diagram, reference lines with  $K(^{232}\text{Tl}/^{238}\text{U})=3.0$  are shown for ages of 0.9 and 1.8 Ga. Uncertainty ellipses apply to data of this study.



**EXPLANATION**

■	West Maui	□	Haleakala	Shield building
●		○		Caldera filling
▲		△		Posterosional

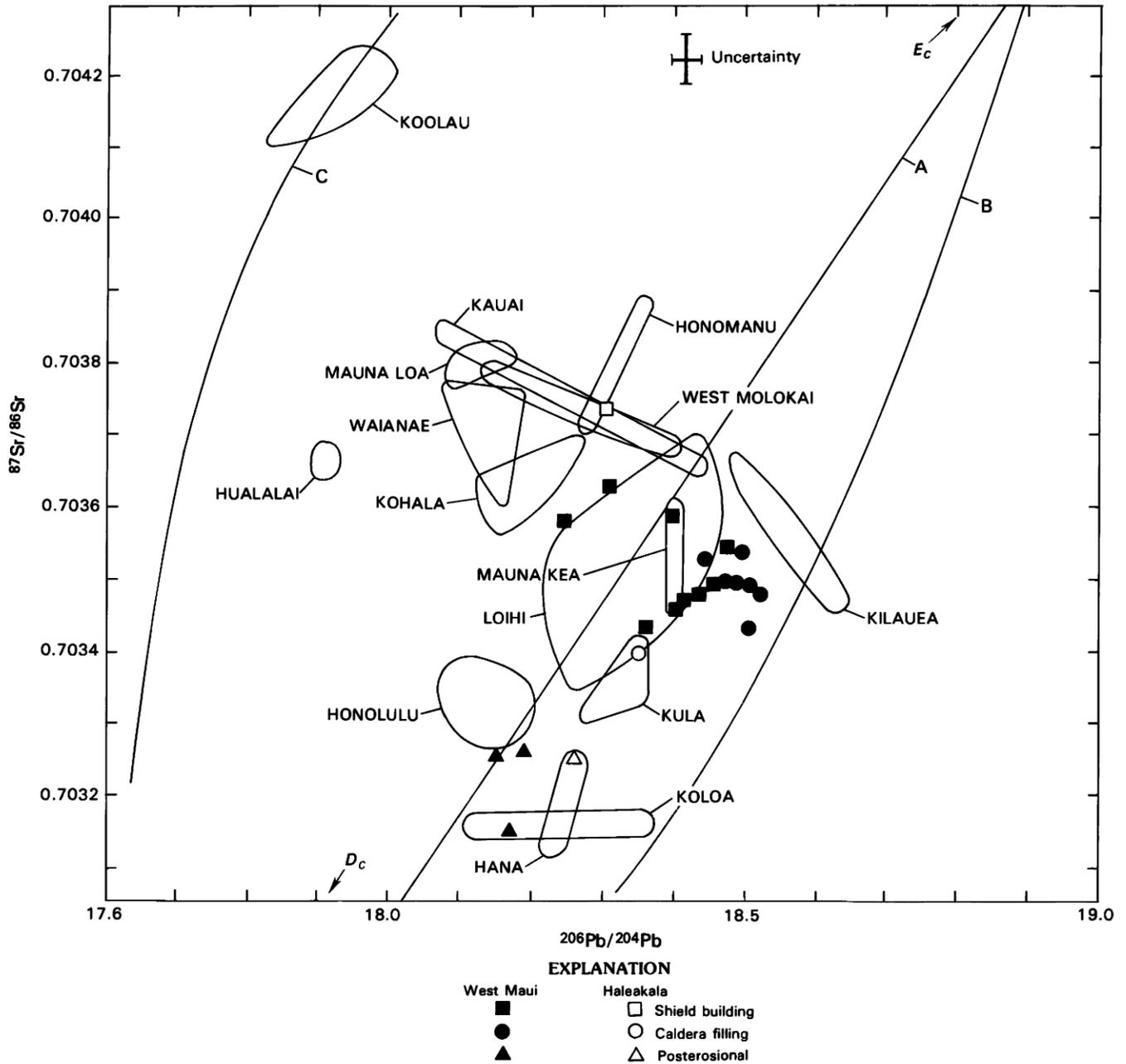


FIGURE 26.6.—Plot of  $^{206}\text{Pb}/^{204}\text{Pb}$  ratio against  $^{87}\text{Sr}/^{86}\text{Sr}$  ratio for Hawaiian rocks. Sources of data as in figures 26.4 and 26.5. Shield-building and caldera-filling rocks from West Maui plot in and near field for Loihi, whereas posterosional rocks plot to less radiogenic side of trend. Also shown are hypothetical mixing curves between depleted and enriched sources: A,  $Q_1=1$ ; B,  $Q_1=0.5$ ; C,  $Q_1=10$ . See discussion in text.  $D_c$  and  $E_c$ , inferred depleted and enriched components, respectively.

basalt. Although these peculiar data need confirmation, it is conceivable that the high  $^{207}\text{Pb}/^{204}\text{Pb}$  ratio is the reflection of oceanic lower crust involvement; the basalt locally contains 1-cm gabbroic clots. Basanite of the Lahaina sequence has less radiogenic Pb than Wailuku tholeiite, indicating a different origin from the underlying volcanic rocks.

The Pb, Sr, and Nd data for Hawaiian volcanic rocks are shown in plots of  $^{206}\text{Pb}/^{204}\text{Pb}$  versus  $^{87}\text{Sr}/^{86}\text{Sr}$  and  $^{143}\text{Nd}/^{144}\text{Nd}$  diagrams in figures 26.6 and 26.7. Differences between the posterosional Lahaina Volcanics and the underlying rocks are more evident on the Sr-Pb diagram. The Pb becomes slightly more radiogenic, and the Sr becomes slightly less radiogenic moving from

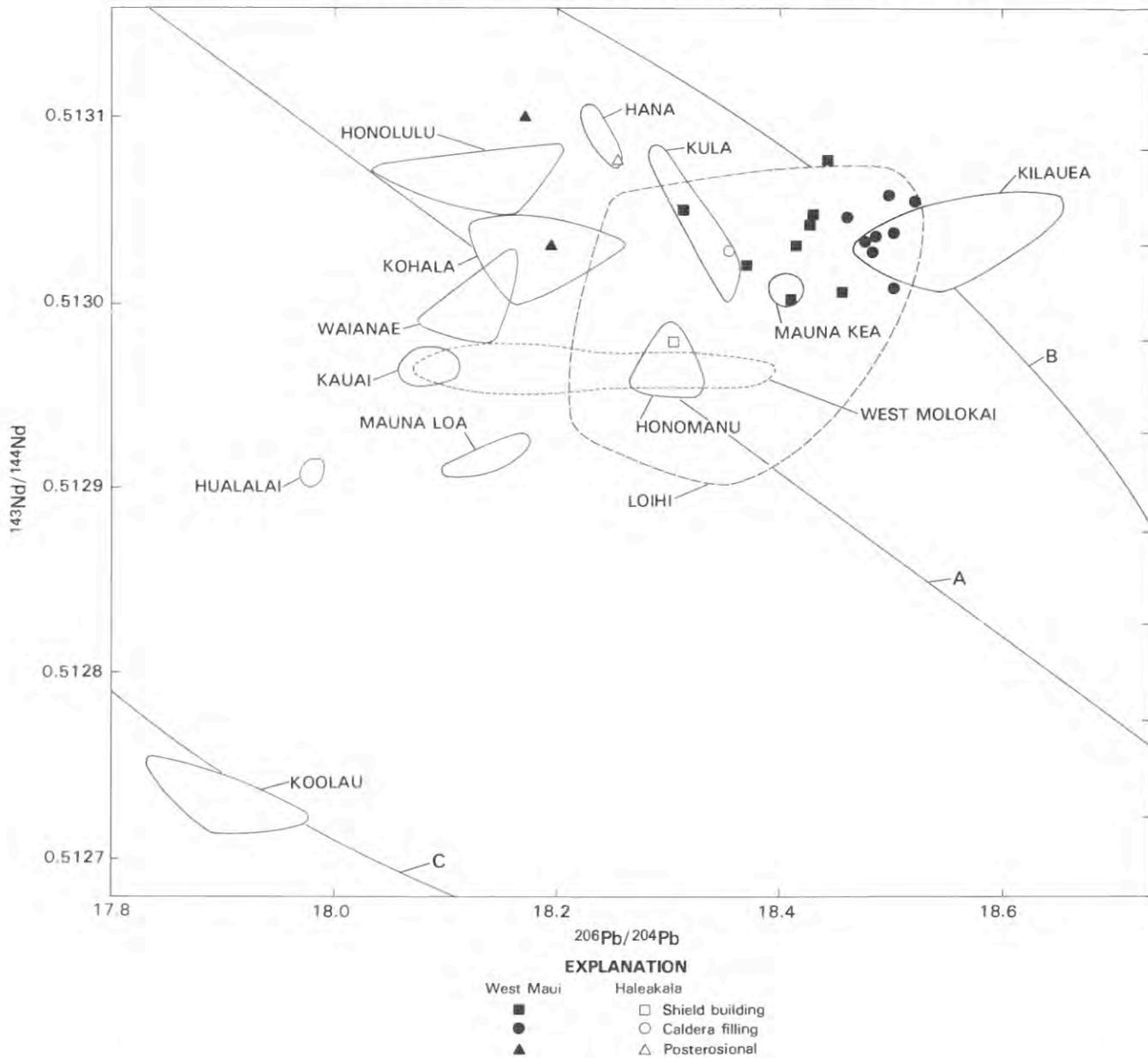


FIGURE 26.7.—Plot of  $^{206}\text{Pb}/^{204}\text{Pb}$  ratio against  $^{143}\text{Nd}/^{144}\text{Nd}$  ratio for Hawaiian rocks. Sources of data as in figures 26.4 and 26.5. Plot can be regarded as upside-down image of figure 26.6. Hypothetical mixing curves calculated as in figure 26.6: A,  $Q_2=1$ ; B,  $Q_2=0.5$ ; C,  $Q_2=10$ .

tholeiite to caldera-filling basalt, and the data define a crude negative correlation. The negative correlation is similar to the main Hawaiian trend that supports a model in which two-component mixing occurs between a mantle plume and depleted mantle or lithosphere. The mixing with depleted mantle progressively increases as the volcano evolves (Chen and Frey, 1983). However, the data for posterosional basalt do not appear to be related to the main trend and indicate the possible presence of a second depleted source with relatively non-radiogenic Pb and Sr.

## DISCUSSION

Results of high-pressure experiments show that alkali olivine basalt can be produced from olivine tholeiite magma by orthopyroxene and olivine fractionation at pressures greater than 1.2 GPa (12 kbar) (see, for example, Kushiro, 1969; Presnall and others, 1979). No significant differences in Pb isotopes between tholeiite and alkali basalt were found for Kohala and Mauna Kea (Tatsumoto, 1978) nor for Pb, Sr, Nd, and Hf isotopic composition between the two

rock types from the Waianae Volcano (Stille and others, 1983). These isotopic similarities between tholeiite and alkali basalt appear to be consistent with results of high-pressure experiments. Distinct differences in  $^{87}\text{Sr}/^{86}\text{Sr}$  ratio between shield-building Koolau tholeiite and nephelinitic Honolulu Volcanics from Oahu have been observed for about two decades (Powell and DeLong, 1966). Recent Sr-isotope measurements (Lanphere and Dalrymple, 1980) as well as studies of Pb, Nd, and Hf isotopes in Oahu volcanic rocks (Stille and others, 1983; Roden and others, 1984) have confirmed the isotopic differences between the two rock series. Thus Tatsumoto (1978) and Stille and others (1983) suggested a close genetic relationship exists between tholeiite and late-stage alkali basalt, but that nephelinitic rocks are of independent origin. This suggestion derived from isotopic studies is consistent with Jackson and Wright's (1970) view; from a study of xenoliths they concluded that the Honolulu Volcanics originated from garnet lherzolite at greater depths than the Koolau tholeiite magma.

In contrast, Chen and Frey (1983) found that Sr and Nd isotopic compositions of Honomanu Basalt (shield-building stage), Kula Volcanics (alkalic basalt), and Hana Volcanics from Haleakala are distinctly different among the three sequences and are inversely correlated with Rb/Sr and Sm/Nd ratios. Chen and Frey invoked a two-component melt-mixing model to account for the isotopic and incompatible-element relationships. Present Nd and Sr isotopic data for West Maui volcanics do not show such clear distinctions as Chen and Frey observed for the east. These results may indicate that Haleakala late-stage magma had crustal con-

tamination along differentiation and changed its isotopic signature, while West Maui magma had little assimilation along differentiation.

Staudigel and others (1984) stated that the heterogeneous isotopic character of Loihi Seamount volcanic rocks may attest to Loihi's inability to homogenize lava before eruption, whereas parental magma of more mature and active volcanoes may be well mixed in established magma chambers. Waianae (Stille and others, 1983) and West Maui data seem to support such a hypothesis.

Staudigel and others (1984) argued that Honolulu and Koolau data define two extreme positions in the  $^{143}\text{Nd}/^{144}\text{Nd}$ - $^{87}\text{Sr}/^{86}\text{Sr}$  correlation diagram, with Kilauea data falling in the center, but that the Honolulu and Koolau data fall close to one end of the  $^{206}\text{Pb}/^{204}\text{Pb}$ - $^{207}\text{Pb}/^{204}\text{Pb}$  trend (fig. 26.5). They used this inconsistency to preclude simple mixing of two end members. We observe, as shown in the previous section, that on the plot of  $^{87}\text{Sr}/^{86}\text{Sr}$  versus  $^{143}\text{Nd}/^{144}\text{Nd}$  (fig. 26.4) Kilauea and Koolau are two extremes (Kilauea does not fall in the center) that define the general trend (tholeiite and late-stage samples) of the Hawaiian volcanic rocks. Posterosional volcanic rocks of Honolulu, Hana, and Lahaina deviate to the left of the general trend, implying a three-component system.

On the  $^{206}\text{Pb}/^{204}\text{Pb}$ - $^{207}\text{Pb}/^{204}\text{Pb}$  plot of figure 26.5, Koolau deviates slightly from the general trend of Waianae, Honolulu, and Kilauea. This deviation is even more pronounced in the plot of  $^{206}\text{Pb}/^{204}\text{Pb}$  versus  $^{208}\text{Pb}/^{204}\text{Pb}$ . In these Pb-Pb diagrams, the denominator of the ratios on each axis is the same ( $^{204}\text{Pb}$ ); therefore, data representing two-component mixing should

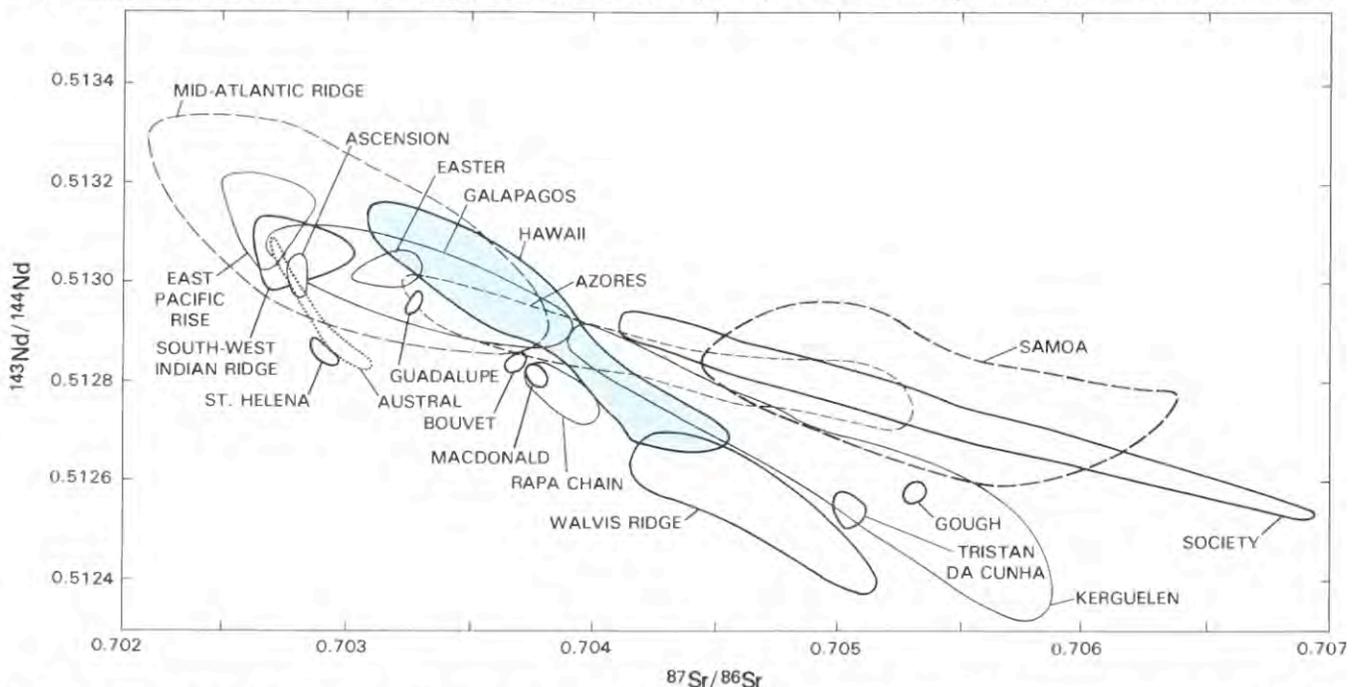


FIGURE 26.8.—Plot of  $^{87}\text{Sr}/^{86}\text{Sr}$  ratio against  $^{143}\text{Nd}/^{144}\text{Nd}$  ratio for oceanic basalt. All data can possibly be explained by three-component mixing model with end-member isotopic compositions close to N-MORB, Austral-St. Helena, and Kerguelen-Society Islands. Data from literature as summarized by Zindler and others (1982), Hart (1984), and White (1985). Data sources for Indian MORB and Rapa-Macdonald chain are from Hamelin and Allegre (1985) and Tatsumoto and others (1984), respectively.

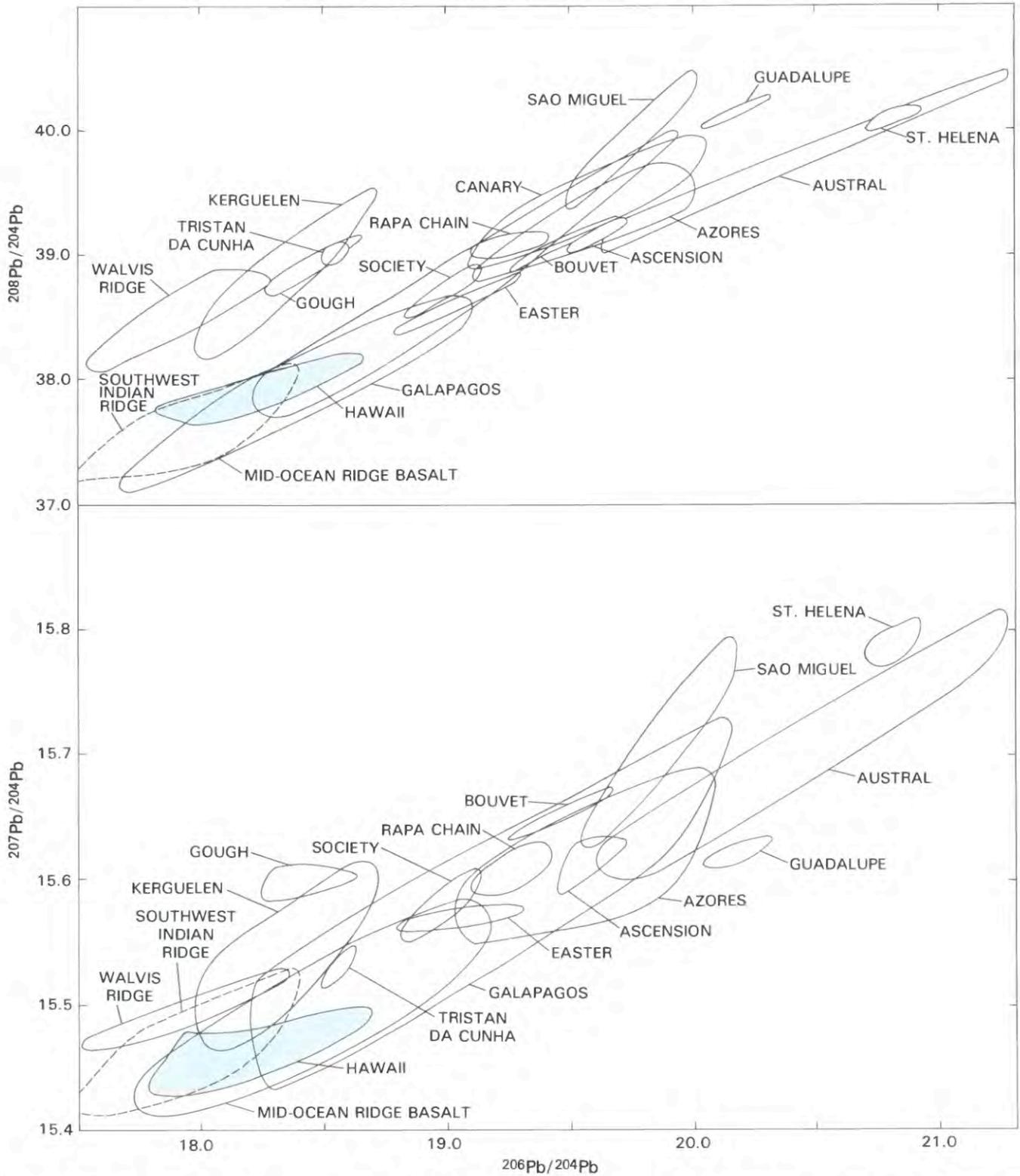


FIGURE 26.9.—Plot of  $^{206}\text{Pb}/^{204}\text{Pb}$  ratio against  $^{207}\text{Pb}/^{204}\text{Pb}$  and  $^{208}\text{Pb}/^{204}\text{Pb}$  ratios for oceanic basalt. Data from Tatsumoto (1978), Sun (1980), Zindler and others (1982), Dupré and Allègre (1983), Stille and others (1983, in press), and Staudigel and others (1984).

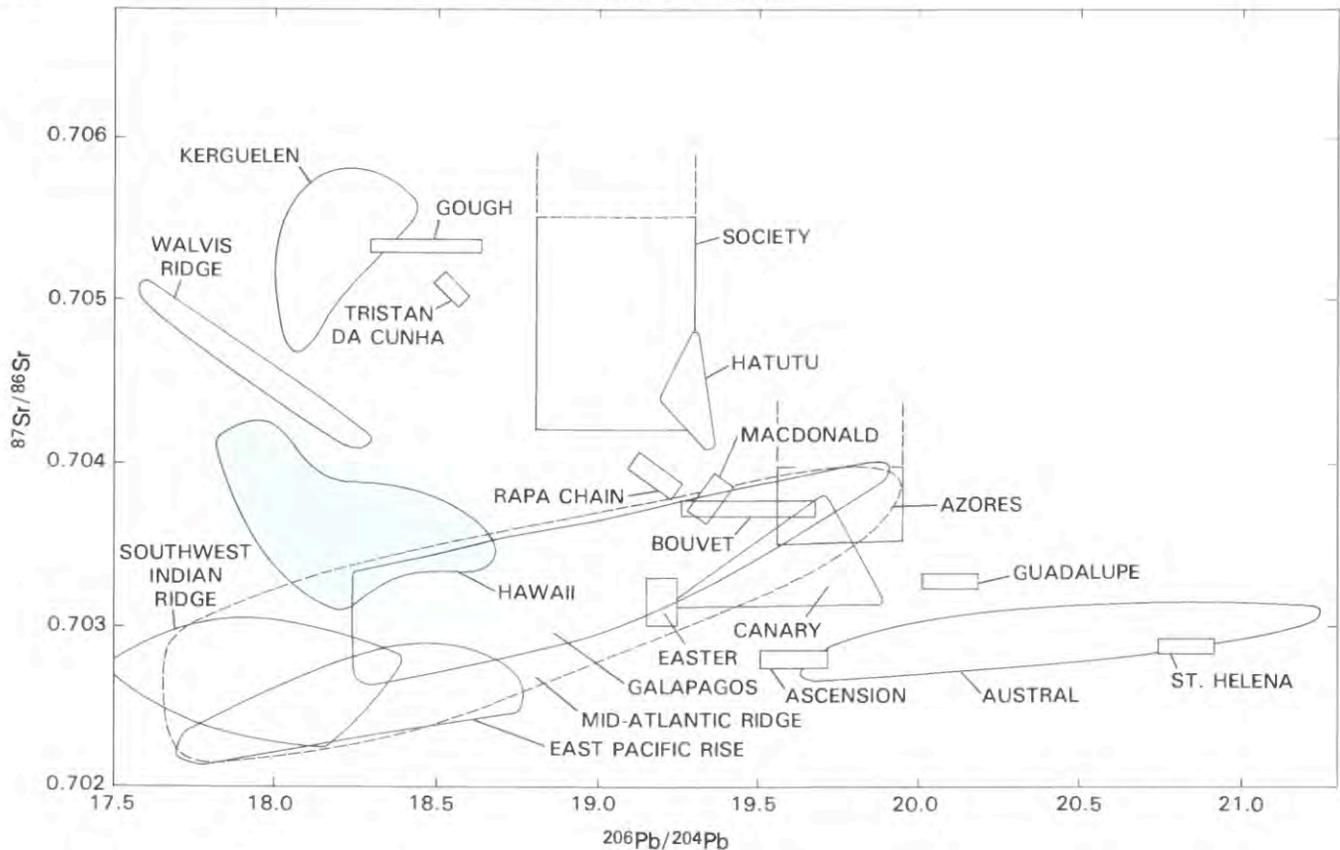


FIGURE 26.10.—Plot of  $^{206}\text{Pb}/^{204}\text{Pb}$  ratio against  $^{87}\text{Sr}/^{86}\text{Sr}$  ratio for oceanic basalt. Data from Tatsumoto (1978), Sun (1980), Zindler and others (1982), Dupré and Allègre (1983), Stille and others (1983, in press), and Staudigel and others (1984).

define a linear correlation. It is evident from the Pb data that Hawaiian volcanics must be described by a model using at least three components.

Evidence for at least three components can also be obtained from the Pb-Sr and Pb-Nd relations among Hawaiian basalt, as shown in figures 26.6 and 26.7. Relative positions of data points for Hawaiian basalt on the  $^{206}\text{Pb}/^{204}\text{Pb}$ - $^{87}\text{Sr}/^{86}\text{Sr}$  diagram (fig. 26.6) are very similar to those on the  $^{206}\text{Pb}/^{204}\text{Pb}$ - $^{207}\text{Pb}/^{206}\text{Pb}$  diagram (fig. 26.5), and the former resembles an expanded (*y*-axis) plot of the latter. This resemblance is observable not only for Hawaiian rocks but also on a global scale for all oceanic-island basalt (OIB) (see figs. 26.8 and 26.9). The plot  $^{206}\text{Pb}/^{204}\text{Pb}$  versus  $^{143}\text{Nd}/^{144}\text{Nd}$  is an upside-down image of the plot of  $^{206}\text{Pb}/^{204}\text{Pb}$  versus  $^{87}\text{Sr}/^{86}\text{Sr}$  (see figs. 26.10 and 26.11). The shield-building Wailuku Basalt and the late-stage Honolulu Volcanics lie close to the Koolau-Kilauea trend, but the posterosional Lahaina Volcanics plots near Hana Volcanics data on both diagrams (figs. 26.6 and 26.7).

In our previous paper (Hegner and others, 1986), in order to test Chen and Frey's two-component mixing model, we followed their premise that the enriched component has isotopic compositions close to those of bulk-earth Sr and Nd (similar to Koolau) and the depleted source has isotopic compositions similar to MORB or to

Kilauea. However, the Pb isotopic compositions of the depleted MORB and Kilauea are not similar, and we use the depleted-MORB-like Pb for the present discussion instead of Kilauea-like Pb used in our previous paper. Posterosional alkali basalt of the Lahaina (West Maui), Hana (Haleakala), Honolulu (Oahu), and Koloa (Kauai) sequences all have similar Pb values near  $^{206}\text{Pb}/^{204}\text{Pb} = 18.2$ , which is higher than that ( $\sim 17.9$ ) of Koolau, and these rocks cannot be one of the end members in the two-component mixing model, even though their Sr-Nd characteristics are closest to MORB among Hawaiian basalt. In figures 26.6 and 26.7, we set the following values of ratios for the two hypothetical end members in order to demonstrate the shortcomings of a two-component mixing model:  $^{206}\text{Pb}/^{204}\text{Pb} = 17.3, 19$ ;  $^{87}\text{Sr}/^{86}\text{Sr} = 0.702, 0.7045$ ;  $^{143}\text{Nd}/^{144}\text{Nd} = 0.5134, 0.51264$ —for the depleted and enriched components, respectively.

Chen and Frey's (1983) model proposed that a melt produced from the enriched end member (mantle plume) mixes with melt formed by low degrees of melting ( $<2$  percent) of a depleted MORB source. The melting degree of MORB progressively decreases (thereby abundances of incompatible elements increase) but mixing amounts progressively increase from tholeiite through late-stage alkalic rocks to posterosional volcanic rocks. According to this model, elements with stronger incompatibility in mantle minerals

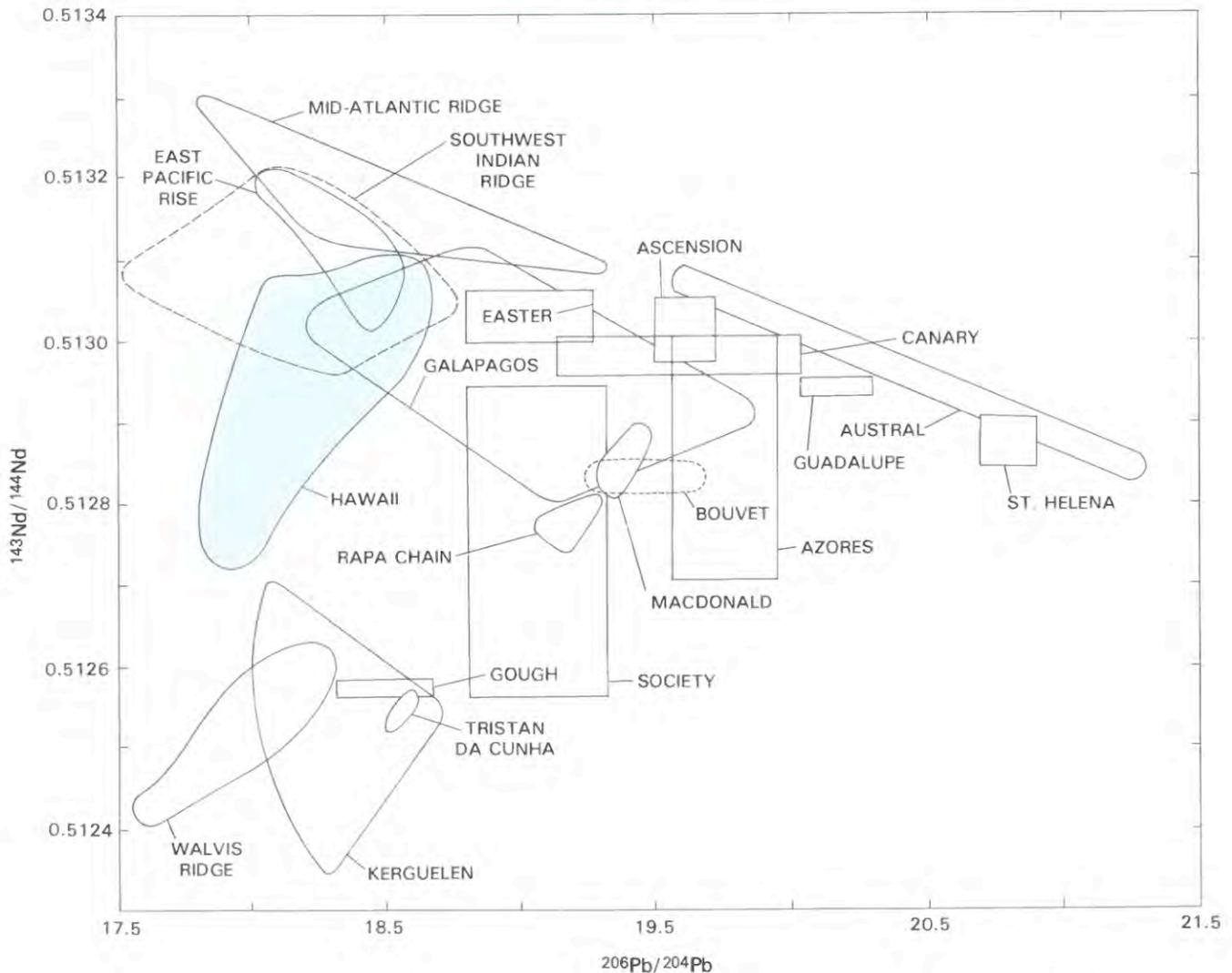


FIGURE 26.11.—Plot of  $^{206}\text{Pb}/^{204}\text{Pb}$  ratio against  $^{143}\text{Nd}/^{144}\text{Nd}$  ratio for oceanic basalt. Data are from Tatsumoto (1978), Sun (1980), Zindler and others (1982), Dupré and Allègre (1983), Stille and others (1983, in press), and Staudigel and others (1984).

are more enriched in later stage volcanic products. Because isotopic compositions of Pb, Sr, and Nd do not change by melting, these elements must have more MORB characteristics in posterosional volcanic rocks than in underlying volcanic rocks. Consequently, parent-daughter elemental abundances and isotopic compositions of Sr and Nd are negatively correlated in Haleakala volcanic rocks.

General equations for two-component mixing have been discussed by Vollmer (1976) and Langmuir and others (1978). Mixing curves of ratio-to-ratio plots of  $^{206}\text{Pb}/^{207}\text{Pb}$  versus  $^{87}\text{Sr}/^{86}\text{Sr}$  (fig. 26.6) and  $^{206}\text{Pb}/^{204}\text{Pb}$  versus  $^{143}\text{Nd}/^{144}\text{Nd}$  (fig. 26.7) are hyperbolae, whose curvature is controlled by the ratio of elemental abundance ratios of the two elements in a MORB melt relative to that in a melt of enriched component (that is,  $Q_1 = (\text{Pb}/\text{Sr})_{\text{MORB}}/(\text{Pb}/\text{Sr})_{\text{enriched}}$  or  $Q_2 = (\text{Pb}/\text{Nd})_{\text{MORB}}/$

$(\text{Pb}/\text{Nd})_{\text{enriched}}$ . When  $Q_1$  or  $Q_2$  is in unity, the mixing curve will be a straight line. Partition coefficients of common mantle minerals are listed in table 26.4. The values for Rb, Sr, La, Ce, Nd, and Sm are from Chen and Frey (1984) and Clague and Frey (1982). For setting the Th values, those reported by Clague and Frey (1982), Tatsumoto (1978), and Sun (1980) are used. The values for U and Pb are selected rather arbitrarily from the consideration of U/Pb and U/Th ratio changes in basaltic rocks. Incompatibility of Pb is assumed to be less than U, but higher than Sr, considering that the U/Pb ratio in a melt increases from that in the source during partial melting. The actual Pb partition coefficient for mantle silicate minerals may be smaller than the value estimated. However, Pb is one of the chalcophile elements, and the bulk partition coefficient for the mantle may not be much different from the value estimated from

TABLE 26.4.—Partition coefficients of orthopyroxene (opx), olivine, clinopyroxene (cpx), and garnet for selected elements

[Coefficients other than for Pb, U, and Th are from Chen and Frey (1983)]

Element	Opx	Olivine	Cpx	Garnet
Rb	0.00010	0.0001	0.0080	0.0010
Sr	.00020	.0005	.0720	.0070
Nd	.00190	.0013	.0900	.0184
Sm	.00280	.0019	.1400	.0823
Pb	.00015	.0002	.0500	.0050
U	.00010	.0001	.0180	.0015
Th	.00010	.0001	.0150	.0010
La	.00050	.0005	.0200	.0010
Ce	.00090	.0008	.0400	.0033

## U/Pb ratios in basalt.

The initial concentrations of the relevant elements for MORB and enriched sources are listed in table 26.5. The values, except U, Th, and Pb, are from Chen and Frey (1983). The values for U, Th, and Pb were chosen on the basis of the concentration data reported in this study and previously reported (for example, Tatsumoto, 1978; Sun, 1980) and on the basis of partition coefficients listed in table 26.4.

Modal compositions of the MORB and enriched sources, melt compositions, and partition-coefficient values for Sr and Nd are from Chen and Frey (1983). The bulk solid/liquid partition coefficients of Th, U, Pb, and Sr calculated for a basaltic liquid in equilibrium with a solid composed of 60 percent olivine, 25 percent orthopyroxene, 10 percent clinopyroxene, and 5 percent garnet (Chen and Frey, 1983) are 0.00164, 0.00196, 0.00541, and 0.00790, respectively. We assumed that the melts composed of olivine (10 percent), orthopyroxene (10 percent), clinopyroxene (40 percent), and garnet (40 percent). If these partition coefficients and concentrations in the two end members of table 26.5 are used,  $Q_1$  and  $Q_2$  must both be about 10 in order to cover Koolau data on the hyperbolae. However, if the values in tables 26.4 and 26.5 are used, the maximum calculated  $Q_1$  and  $Q_2$  are only 1.23 and 1.34, respectively, even for an extreme mixing case of 0.001 percent melt of the MORB source with 20 percent melt of the enriched source.

Tests were made using varying  $D_{Pb}$  in the range of 0.0025–0.010 and Pb concentrations of the enriched mantle source in the range of 0.08–1.7 ppm. However, we could not obtain  $Q_1$  and  $Q_2$  larger than 4 in any conceivable combination. Therefore, we conclude that the two-component mixing model is not adequate to explain Hawaiian isotope data and that at least three isotopically distinct sources are required to account for the observed data. Stille and others (in press) reached the same conclusion by using depleted and enriched end members similar to those listed above but with the Pb isotopic compositions of the two end members reversed.

Recent isotope studies of Hawaiian volcanic rocks also imply multicomponent hypotheses (for example, Stille and others, 1983, in press; Roden and others, 1984; Staudigel and others, 1984; Hegner and others, 1986), but some differences exist between studies in the assignment of the isotopic components to specific mantle sources. From the preceding discussion on the Pb-Sr-Nd isotopic characteristics, it is essential that Koolau, Kilauea, and the

TABLE 26.5.—Elemental compositions of the inferred MORB and enriched mantle sources of West Maui volcanic rocks

Component	MORB	Enriched mantle
	Initial concentration (ppm)	
Rb	0.1	0.73
Sr	13.2	23.7
Nd	.86	1.2
Sm	.32	.39
Pb	.08	.17
U	.009	.02
Th	.028	.077
La	.31	.71
Ce	.95	1.9
Isotope ratios		
$^{238}\text{U}/^{204}\text{Pb}$	7.08	8.12
$^{232}\text{Th}/^{238}\text{U}$	3.21	3.98

posterosional volcanic rocks (Lahaina, Hana, and Honolulu) are the most extreme in terms of isotopic composition, and all other Hawaiian data fall in a field bounded by these end members. In the following discussion on the origin of Hawaiian basalt, we will refer to these as the Koolau, Kilauea, and PE (posterosional) components (Stille and others, in press), although we recognize that the pure end members are probably more extreme in composition than the Koolau, Kilauea, and PE. A rigorous test using three-dimensional regressions of the Hawaiian data was made in our previous paper (Stille and others, in press) and will not be repeated here.

Staudigel and others (1984) proposed that the Koolau component represents an enriched portion of the oceanic lithosphere, Kilauea the mantle plume, and PE the uppermost lithosphere or crust. We have previously proposed (Tatsumoto, 1978; Stille and others, 1983, in press; Hegner and others, 1986) that the main tholeiite and late-stage alkalic trend is predominantly a plume-lithosphere mixing trend with Koolau having the largest proportion of plume and Kilauea having the largest proportion of oceanic crust or lithosphere. In our previous model, the PE component was derived from the depleted upper mantle (MORB source or material residual from MORB production). The reasoning behind this interpretation is presented below.

The Koolau component has low  $^{206}\text{Pb}/^{204}\text{Pb}$  and  $^{143}\text{Nd}/^{144}\text{Nd}$  ratios, but high  $^{207}\text{Pb}/^{204}\text{Pb}$ ,  $^{208}\text{Pb}/^{204}\text{Pb}$ , and  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios; Koolau data therefore plot distinctly below the mantle plane of Zindler and others (1982). Such isotopic characteristics might be assigned to recycled oceanic lithosphere (Hofmann and White, 1982). However, as will be discussed later (fig. 26.9), Koolau and Walvis Ridge data fall on the left side of the mantle array in a Nd-Sr correlation diagram with Kerguelen as the enriched member, whereas Samoa and Society Island data, which are considered to present the most convincing evidence for having a recycled-crust component (White and Hofmann, 1982), plot on the opposite side of the mantle array.

Staudigel and others (1984) alternatively proposed that the Koolau component may represent an enriched part of the (lower)

oceanic lithosphere, which originated from ancient metasomatized mantle and accreted to the lithosphere at or near the ridge. Such an interpretation may be possible, but the proposed physical process should be tested as more OIB data become available. The process of metasomatism is not well understood, and the chemical composition of the hydrothermal fluids proposed to be involved in metasomatism appear to vary widely depending on the type of process postulated. This obscurity is, we believe, largely caused by a tendency to arbitrarily attribute unexplainable enrichments in incompatible elements to metasomatism.

Wright (1984) postulates, from rigorous comparison of chemical composition between MORB and Hawaiian tholeiite, that the latter originated from depleted mantle that has been metasomatized by a liquid with a composition of nephelinite plus amphibole (plus some minor phases); hence such metasomatism is characterized by H<sub>2</sub>O and alkali-rich fluids (Hawkesworth and others, 1979). The metasomatized rocks would have high Rb/Sr ratios and eventually high <sup>87</sup>Sr/<sup>86</sup>Sr ratios. Consequently, Nd-Sr data for these rocks would plot on the right-hand side of the mantle array on a Nd-Sr diagram. Most Nd-Sr data of the Hawaiian tholeiite plot is indeed on the right-hand side and support Wright's postulation. However, Koolau data plot on the lower left side of the array. Furthermore, because Koolau Pb-Sr data appear to be unique among all OIB (see next section), we interpret Koolau as a unique end member. If Koolau originated from metasomatized mantle, the proposed metasomatizing fluids would also have to have been unique in their enrichment in alkali and light REE without enrichment in U relative to Pb.

In contrast, we have postulated that ancient metasomatism would slightly decrease the Rb/Sr ratio and increase U/Pb if the metasomatizing fluid is carbonatite-like (Tatsumoto and others, 1984) so that fairly low <sup>87</sup>Sr/<sup>86</sup>Sr but high <sup>206</sup>Pb and <sup>207</sup>Pb values would result from the process. Koolau data plot on the lower left of the mantle array on the Nd-Sr correlation diagram but show very low <sup>206</sup>Pb values. Thus, Koolau isotope data are not consistent with a hypothesis of such a metasomatized source. Kilauea tholeiite, however, could be related to metasomatism because those data plot on the right-hand side of the Nd-Sr mantle array and show a high <sup>206</sup>Pb/<sup>204</sup>Pb ratio (although in this paper we interpret the Kilauea end member to represent lithosphere contamination instead of metasomatism).

Stille and others (1983, in press) have shown that the isotopic characteristics of Koolau samples are consistent with an origin from primitive lower mantle and that the Koolau component may be unique among oceanic basalt. Alternatively, the Koolau component may represent a less depleted or possibly even enriched upper mantle material that exists as a distinct body (see, for example, Davies, 1984) or is preserved in the center of a concentrically zoned convection cell (see, for example, Tatsumoto, 1978).

The Honolulu, Hana, and Lahaina Volcanics all have  $^{206}\text{Pb}/^{204}\text{Pb} = 18.2$ ,  $^{87}\text{Sr}/^{86}\text{Sr} = 0.71031$ , and  $^{143}\text{Nd}/^{144}\text{Nd} = 0.51315$ . In the Nd-Sr correlation diagram (fig. 26.4) it is evident that these PE volcanic rocks are most depleted and are isotopically closest to MORB among Hawaiian rocks.

Because basanite may originate, in the presence of garnet, at greater depth than tholeiite (Presnall and others, 1979; Jackson and Wright, 1970), the PE component may represent the depleted upper mantle.

The Kilauea component has moderately radiogenic Pb, but MORB-like isotopic compositions of Sr, Nd, and Hf (Stille and others, in press). The Pb isotopic compositions in volcanics from the Kea trend are generally more radiogenic than those from the Loa trend (Tatsumoto, 1978). The isotopic difference was thought to reflect contamination of the ~80-Ma lithosphere (MORB) under Hawaii, and Kilauea was thought to represent a lithosphere component. The half-life of <sup>238</sup>U ( $4.47 \times 10^9$  yr) is short compared to those of <sup>87</sup>Rb ( $4.9 \times 10^{10}$  yr) and <sup>147</sup>Sm ( $1.06 \times 10^{11}$  yr), and so an increase in the <sup>206</sup>Pb/<sup>204</sup>Pb ratio may be detected in ~100-Ma lithosphere (a <sup>207</sup>Pb/<sup>204</sup>Pb increase is not detectable because of the small amount of <sup>235</sup>U). Recent detailed studies of Nd, Sr, and Hf isotopes support this view (Stille and others, in press). However, Staudigel and others (1984) assigned the Kilauea component to the plume, and the differences between Loa and Kea isotope trends were considered to have been generated by mixing of plume material with the sublithospheric Koolau component. While such a model may be formally allowed, we maintain that the isotopic differences observed between the presently active Kilauea and Mauna Loa would be extremely difficult to produce if mixing occurs below the lithosphere. Furthermore, the isotopic characteristics of the Kilauea component are not those considered typical for mantle plumes but are actually quite MORB-like. Thus we maintain our earlier conclusions that the Kilauea component represents MORB-related old oceanic lithosphere and that the isotopic differences among volcanoes of the Loa and Kea trends are best explained if mixing occurs within the lithosphere.

#### COMPARISON OF HAWAIIAN DATA WITH THOSE OF OTHER OCEANIC BASALT

In this section we compare isotopic data of Hawaiian basalt with those of other oceanic basalt and discuss the genesis of oceanic basalt.

The Nd-Sr, Pb-Pb, Pb-Sr, and Pb-Nd isotopic correlation diagrams for selected OIB and MORB from the literature are shown in figures 26.8–26.11 and an explanatory combined diagram in figure 26.12. The diagrams contain data for Walvis Ridge basalt that are not corrected for in-place growth of radiogenic isotope during the 70–80 m.y. of their existence (Richardson and others, 1982). However, the isotopic compositions are very similar to those of Koolau Basalt, and in-place growth of radiogenic isotope should not change the overall picture.

Recently, several researchers (for example, Allègre, 1982; Hart, 1984) have argued that the isotopic characteristics of oceanic basalt could be correlated geographically. Allègre (1982) contends that systematic variations occur within each ocean basin and are probably related to convection units in the mantle. Hart (1984) emphasizes that isotopic differences can be observed between the northern and southern hemispheres. However, for the sake of

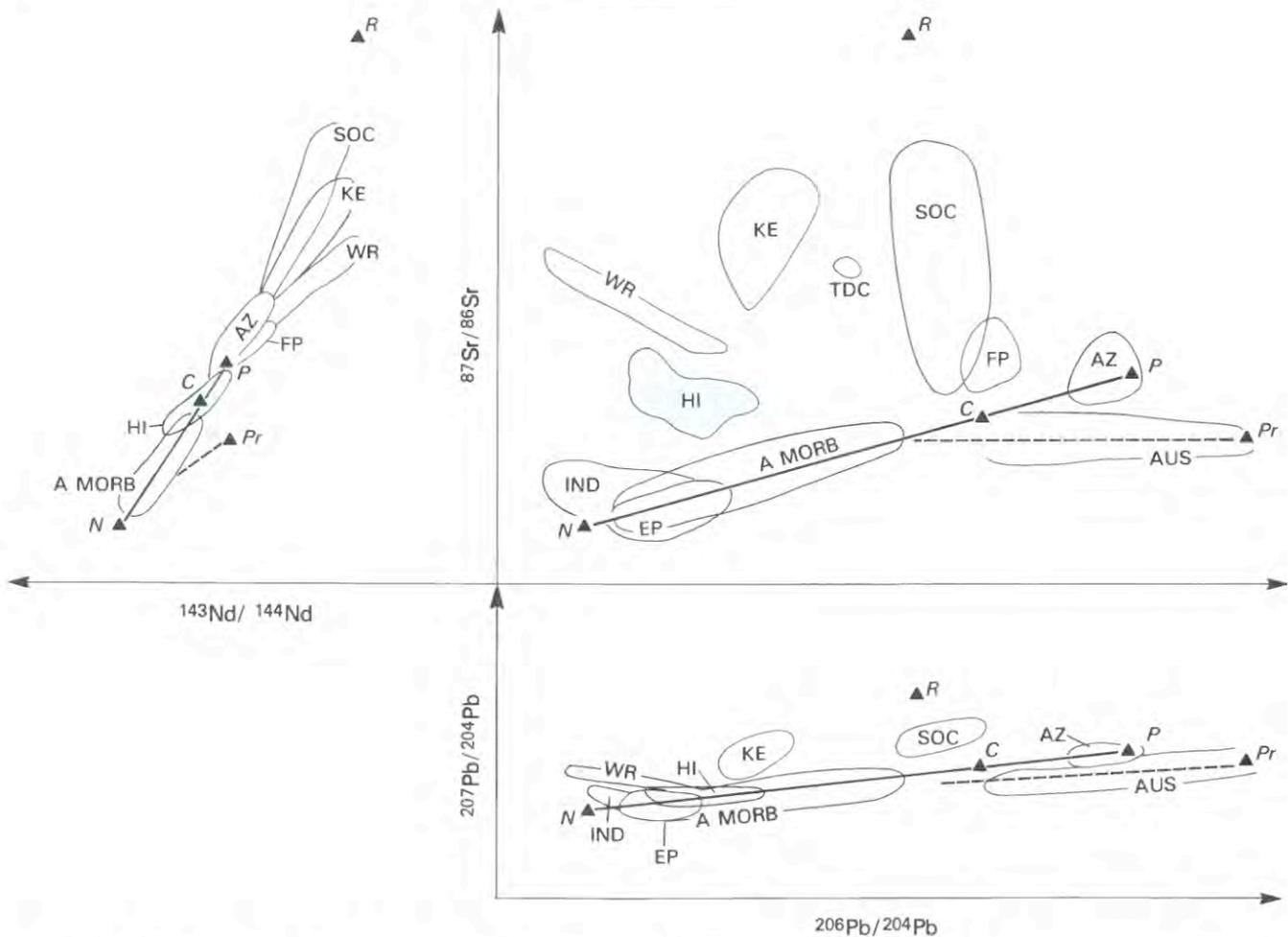


FIGURE 26.12.—Explanatory diagram for oceanic-basalt isotope data. Combined diagram consists of plots of  $^{206}\text{Pb}/^{204}\text{Pb}$  versus  $^{207}\text{Pb}/^{204}\text{Pb}$ ,  $^{206}\text{Pb}/^{204}\text{Pb}$  versus  $^{87}\text{Sr}/^{86}\text{Sr}$ , and  $^{87}\text{Sr}/^{86}\text{Sr}$  versus  $^{143}\text{Nd}/^{144}\text{Nd}$ . Triangles, hypothetical parent materials in possible three-component mixing of *N* (depleted MORB), *Pr* (enriched source represented by Austral-St. Helena), and *R* (recycled crust). Data for islands on or near ridges fall on *N*-*P* line (depleted-enriched MORB) or its extension. In alternative multicomponent mixing model, plume (stars labeled *C*) ascending from lower mantle mixes with heterogeneous upper mantle and oceanic crust. Fields represent Mid-Atlantic Ridge (A MORB), East Pacific Rise (EP), Indian Ocean Ridge (IND), Hawaiian Islands (HI), Walvis Ridge (WR), Kerguelen (KE), Tristan da Cunha (TDC), Society Islands (SOC), French Polynesia islands other than Society and Austral (FP), Azores (AZ), and Austral (AUS). For scales on plots and sources of data, see figures 26.8–26.10.

simplicity, we will proceed with our discussion on the basis of oceanic basalt in general, and later we will demonstrate that the systematic geographic variations that occur are probably caused by upper mantle heterogeneity.

It is widely accepted from isotope data of oceanic basalt that their sources are chemically heterogeneous and that such heterogeneity has existed for a long time (>1 b.y.). The variety of trace-element and isotope signatures in oceanic basalt has often been explained by a layered-mantle model involving mixing of partial melts from a depleted upper mantle and an undepleted lower mantle (see for example, Schilling, 1973; Sun and Hansen, 1975; Allègre and others, 1983). When Nd isotopic data became available (for example, DePaolo and Wasserburg, 1976), the layered-mantle hypothesis appeared to be substantiated. Mass balance, computed

with Sm-Nd and Rb-Sr data, indicated that only 30–50 percent of the whole mantle could be involved in continental crust formation (Jacobsen and Wasserburg, 1979; O'Nions and others, 1979; Allègre and others, 1980; DePaolo, 1980). In these computations, the boundary between depleted and enriched mantle was set at a depth of 700 km, which coincides with the seismic discontinuity.

The mantle array observed for MORB and OIB in the Nd-Sr correlation diagram was often interpreted as a mixing line between the depleted mantle and the mantle plumes. However, as more data for oceanic basalt were accumulated, more exceptions that did not plot on a linear mantle array were found (fig. 26.8). In the Nd-Sr diagram, one can easily see that data trends for oceanic basalt do not form a linear mantle array. McCulloch and others (1983) and Anderson (1986) argued that, if some extreme isotopic

values are assigned to depleted and enriched reservoirs, a two-component mixing model may be sufficient to explain all basalt data. However, as discussed above for Hawaiian data, the Pb data plotted in figure 26.9 cannot be explained by a two-component mixing model because the denominator is the same and two-component mixing must therefore define a straight line. The negative correlations of Hawaiian and Walvis Ridge data in the Pb-Sr plot (fig. 26.10) are also difficult to explain by a two-component mixing model.

The diversity of isotope data for oceanic basalt may be adequately explained by a three-component model. The depleted MORB component (N-type) has the least radiogenic Pb and Sr and the most radiogenic Nd and occupies one corner in the Pb-Sr-Nd correlation diagrams (figs. 26.8–26.11). N-MORB and trace-element-rich P-MORB form a single trend that corresponds to a 1.8-Ga isochron on the Pb-Pb diagram. Data for oceanic islands on or near ridges, such as the Iceland-Reykjanes Ridge, Galapagos Islands, Easter Island, and the Azores, also lie on this line (defined as N–P in figure 26.12). This line has been interpreted as a mixing line between a depleted component and mantle plumes instead of a differentiation trend, and the Pb isotopic composition of the plume as defined from Pb-Sr-Nd relations is  $^{206}\text{Pb}/^{204}\text{Pb} = 20.6$ ,  $^{207}\text{Pb}/^{204}\text{Pb} = 15.75$ ,  $^{208}\text{Pb}/^{204}\text{Pb} = 40.8$  (Tatsumoto and others, 1984). A similar composition was defined by Dupré and Allègre (1980) as the “Planetary Lead Atlantic” value, although they did not specify that this could represent the plume component. Very radiogenic Pb from St. Helena and Ascension (Sun, 1980) appears to lie on an extension of the N–P trends in the Pb-Pb diagram (fig. 26.9), whereas radiogenic Pb from the Australs plots on a slightly shallower trend (Tatsumoto and others, 1984; Vidal and others, 1984). In the Pb-Sr diagram, however, data from St. Helena, the Australs, and Ascension form a trend that is distinctly shallower than the N–P trend. We define the component that shows extremely radiogenic Pb (but less radiogenic Sr and Nd) as Pr in figure 26.12.

Enrichment of incompatible elements or metasomatism in some part of the depleted mantle may have occurred by mixing or contamination with carbonatite-like fluid (high in  $\text{CO}_2$ , and U/Pb and Sr/Rb ratios and slightly enriched in light REE), which originated somewhere else in the depleted mantle within the last ~1.8 b.y. Thus, when time passed, this contaminated mantle (Pr) source has extremely radiogenic Pb, moderately radiogenic Nd, but less radiogenic Sr. Consequently, the Pr component plots on a distinctly lower position from the mantle array (Tatsumoto and others, 1984; White, 1985). The Austral Pb may represent a more recently contaminated mantle source because it plots below the N–P line in the diagram  $^{206}\text{Pb}/^{204}\text{Pb}$  versus  $^{207}\text{Pb}/^{204}\text{Pb}$ .

As can be seen in figure 26.12, all oceanic data can be explained by a three-component mixing model with N, Pr, and another component (R) that has high  $^{87}\text{Sr}/^{86}\text{Sr}$  (>0.706), moderate  $^{206}\text{Pb}/^{204}\text{Pb}$  (~18.2), and very low  $^{143}\text{Nd}/^{144}\text{Nd}$  (<0.5123) ratios. Such a component is most likely recycled crust. If so, then the Koolau and Walvis Ridge data that plot above the N–P line on the Pb-Sr diagram might indicate involvement of a small amount of recycled oceanic crust, and data for Kerguelen and Society Islands

indicate involvement of large amounts of this crust component.

According to this interpretation, Koolau may contain the largest proportion of the recycled crust among Hawaiian volcanoes because its Sr is the most radiogenic. Therefore, if the subducted oceanic crust penetrates down to the mantle-core boundary, as proposed by Hofmann and White (1982), then Koolau could represent a recycled crustal plume at Hawaii. If this hypothesis is correct, all OIB that plot above the N–P line in the Pb-Sr diagram contain a component of recycled oceanic crust (R) whose isotope ratios vary regionally. Those OIB that plot on the N–P or N–Pr trends contain little or no R component.

On the Nd-Sr correlation diagram (fig. 26.8), the so-called mantle array (N–P line) extends to Kerguelen. Society Islands data plot above the extension of the N–P line, but Koolau and Walvis Ridge plot below the line. This positioning is not consistent with that observed above on the Pb-Sr diagram.

Dupré and Allègre (1983) and Hamelin and Allègre (1985) found that the depleted end member of Indian MORB, which they considered to represent the asthenosphere, is characterized by particularly nonradiogenic  $^{206}\text{Pb}/^{204}\text{Pb}$ , but higher  $^{207}\text{Pb}/^{204}\text{Pb}$ ,  $^{208}\text{Pb}/^{204}\text{Pb}$ , and  $^{87}\text{Sr}/^{86}\text{Sr}$  and lower  $^{143}\text{Nd}/^{144}\text{Nd}$  ratios compared to Atlantic and Pacific MORB. This means that the asthenosphere is heterogeneous but there is still a regional similarity that may be related to each convection unit. Thus, in the same manner, the higher  $^{207}\text{Pb}/^{204}\text{Pb}$ ,  $^{208}\text{Pb}/^{204}\text{Pb}$ , and  $^{87}\text{Sr}/^{86}\text{Sr}$  and low  $^{143}\text{Nd}/^{144}\text{Nd}$  ratios of Koolau and Walvis Ridge basalt may be interpreted as due to regional upper mantle heterogeneities instead of a contribution from the recycled oceanic crust.

In any case, Koolau basalt is unique with respect to its relatively primitive isotopic composition (Stille and others, in press). Hawaiian basalt contains the highest  $^3\text{He}/^4\text{He}$  ratios thus far reported for oceanic basalt (Lupton and Craig, 1975; Kaneoka and Takaoka, 1978; Kurz and others, 1982; Allègre and others, 1983; Rison and Craig, 1983). If our argument is correct, Koolau would have had the highest  $^3\text{He}/^4\text{He}$  value among Hawaiian basalt (Stille and others, in press). Conversely, the Hawaiian basalt with a high  $^3\text{He}/^4\text{He}$  and relatively primitive Pb, Nd, and Sr isotopic composition would not support the assumption of radiogenic plume lead proposed by Tatsumoto and others (1984).

White (1985) argued that at least five trends can be observed in Pb-Sr-Nd isotopic systematics: MORB, St. Helena, Kerguelen, Society Islands, Hawaii (see figs. 26.10 and 26.12). Extrapolations of the latter four trends all intersect the central part (C) of the N–P trend, but they do not trend toward P or Pr. The intersection corresponds to  $^{206}\text{Pb}/^{204}\text{Pb} \approx 18.5\text{--}19.5$ ,  $^{87}\text{Sr}/^{86}\text{Sr} \approx 0.7030\text{--}0.7035$ , and  $^{143}\text{Nd}/^{144}\text{Nd} \approx 0.5129\text{--}0.5130$ . It is conceivable that this point represents the plume component instead of P in figure 26.12. According to this interpretation, more than five trends are easily observable, because each hot spot creates its own trend by mixing C with upper mantle of variable isotopic composition.

In this context, the Galapagos and Azores trends should be separated from the N–C trend if the radiogenic components do not lie on the N–C trend but on the extension. Walvis Ridge and Rapa-Macdonald may be separated from Hawaii and Society

Islands trends, respectively, because the Walvis Ridge and Hawaii may not plot on the same trends in the Pb-Sr and Pb-Nd diagrams and Rapa Macdonald data plot apart from Society Islands data on the Nd-Sr diagram. This means that as more data become available, more trends will show up as regional variations. The fact that C cannot be defined as a unique point but has a small range in isotopic compositions may indicate that small regional variations also exist for the lower mantle values.

It should be noted that Austral-St. Helena data do not appear to converge to C on the Nd-Sr plot. Their Pb is extremely radiogenic and indicates extremely enriched sources, while their slightly radiogenic Sr indicates a moderately depleted source. We retain our earlier interpretation that these data reflect an old metasomatized component enriched in U and Sr from carbonatite-like fluids derived from a previously depleted (MORB) source (Tatsumoto and others, 1984). Thus the Austral-St. Helena-Ascension trend is predominantly an upper mantle trend.

We previously suggested that Kilauea contains the most MORB component derived from the oceanic crust because of the depleted nature of its Sr and Nd isotopes. Alternatively, Kilauea might be considered to contain the most plume component since Kilauea occupies the position closest to C (Staudigel and others, 1984). However, as previously mentioned, this interpretation does not adequately account for the observed isotopic differences among simultaneously active volcanoes: Kilauea and Mauna Loa.

The important role of subduction of oceanic crust and sediments for isotope evolution of the crust-mantle system was first suggested by Armstrong (1968), and influences of subducted material on Pb isotopes were detected for volcanics from the Japanese Islands (Tatsumoto, 1969) and the Aleutian Island arc (Kay and others, 1978) but not detected for those from the Mariana arc (Majner, 1976). Extremely low  $^{238}\text{U}/^{204}\text{Pb}$  ratios (as low as 2.1) in Japanese arc volcanics (Tatsumoto, 1960; Tatsumoto and Knight, 1969) indicate that volatile Pb is preferentially extracted from the subducted material by arc volcanism, whereas  $^{238}\text{U}/^{204}\text{Pb}$  ratios of the Mariana arc volcanics are about 7. Thus, plumes generated from the recycled crust are likely to have variations in isotope ratios depending on their storage times and environments; this implies the existence of a heterogeneous lower mantle (see for example, Davies, 1984) providing that the subducted crust penetrates into the lower mantle.

Several seismic studies have presented evidence that subducting slabs penetrate into the lower mantle (Jordan, 1977; Creager and Jordan, 1984). Although some workers suggest that the penetration is only a few hundred kilometers below the seismic boundary at a depth of 670 km (see for example, Christensen and Yuen, 1984), others consider that the subducted slabs reach the core-mantle boundary (Hofmann and White, 1982; Davies, 1984; Loper, 1985). Loper (1985) suggested that numerous narrow axisymmetric plumes rise up into the upper mantle. Thus, it may be possible that numerous plumes with various isotope signatures rise from the lower mantle. However, with the presently available isotopic evidence, we prefer a convective layered mantle model with a chemically heterogeneous upper mantle and a nearly homogeneous (less differentiated)

lower mantle because (1) the upper mantle is strongly differentiated by magma extraction and (2) no high  $^{87}\text{Sr}/^{86}\text{Sr}$  signature of subducted material is found for volcanics from Hawaii where the volcanism is most likely related to a mantle plume as evidenced by high  $^3\text{He}/^4\text{He}$  ratios. Contamination of magmas as they rise through the oceanic lithosphere makes the isotopic signatures more complicated.

## CONCLUSIONS

Pb, Sr, and Nd isotope data of Hawaiian basalt provide strong evidence for the involvement of three distinct sources in the production of these rocks. The volcanoes or groups of rocks that most closely approach the three end members are Koolau (representing the mantle plume), Kilauea (the oceanic crust), and the Honolulu-Hana-Lahaina posterosional volcanic rocks (the depleted mantle). Pb, Sr, and Nd relationships are not compatible with a two-component mixing model for Hawaiian basalt genesis.

Pb, Sr, and Nd isotope characteristics of oceanic basalt from other areas are also compatible with a three-component mixing model. The three main sources are the depleted upper mantle, enriched lower or upper mantle, and recycled oceanic crust. However, the oceanic basalt data form several trends that intersect in the middle of the three-component field. These trends may indicate multicomponent mixing between a heterogeneous upper mantle and a nearly homogeneous lower mantle, plus contamination by the lithosphere.

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