

Geochemical variations in lavas from Kahoolawe volcano, Hawaii: evidence for open system evolution of plume-derived magmas

W.P. Leeman¹, D.C. Gerlach², M.O. Garcia³, and H.B. West⁴

¹ Keith Wiess Geological Laboratories, Rice University, Houston, TX 77251-1892, USA

² Charles Evans and Associates, 301 Chesapeake Dr., Redwood City, CA 94063, USA

³ Geology-Geophysics Department, University of Hawaii, Honolulu, HI 96822, USA

⁴ Hawaii Institute of Geophysics, University of Hawaii, Honolulu, HI 96822, USA

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Abstract. The Kahoolawe shield volcano produced pre-caldera and caldera-filling tholeiites and mildly alkaline post-caldera lavas that petrographically and compositionally resemble such lavas from other Hawaiian shield volcanoes. However, Kahoolawe tholeiites display wide ranges in incompatible trace element ratios (e.g., Nb/Th = 9–24, Th/Ta = 0.6–1.3), $^{87}\text{Sr}/^{86}\text{Sr}$ (0.70379–0.70440), $^{143}\text{Nd}/^{144}\text{Nd}$ (0.51273–0.51298), and $^{206}\text{Pb}/^{204}\text{Pb}$ (17.92–18.37). The isotopic variation exceeds that at any other Hawaiian shield volcano, and spans about half the range for all Hawaiian tholeiites. Quasi-cyclic temporal evolution of Kahoolawe tholeiites is consistent with combined fractional crystallization and periodic recharge by primitive magmas. Ratios of highly incompatible trace elements and Sr, Nd, and Pb isotopic ratios form coherent sub-trends that reflect recurrent interactions between variably evolved magmas and two other mantle components whose compositions are constrained by intersections between these trends. The most MgO-rich Kahoolawe tholeiites are partial melts of a high Nb/Th (~ 23.5) ascending plume, possibly comprising ancient subducted oceanic lithosphere. Slightly evolved tholeiites experienced combined crystal fractionation and assimilation (AFC) of material derived from a distinct reservoir (Nb/Th ~ 9) of asthenospheric derivation. The most evolved tholeiites display compositional shifts toward a third component, having mid ocean ridge basalt-like isotopic ratios but enriched OIB-like trace element ratios, representing part of the lithospheric mantle (or melts thereof). Periodic recurrence of all three magma variants suggests that eruptions may have tapped coeval reservoirs distributed over a large depth range. Kahoolawe provides new evidence concerning the nature of the Hawaiian plume, the distribution of compositional heterogeneities in the suboceanic mantle, and the processes by which Hawaiian tholeiites form and evolve.

Introduction

The Hawaiian Islands provide a spectacular example of “hot spot” or “mantle plume” volcanism. Detailed geochemical studies of Hawaiian volcanoes thus offer an opportunity to learn more about the composition of the deep mantle and the processes by which hot spots form and are sustained. In recent years, numerous papers have presented radiogenic isotope data (Sr, Nd, Pb, Hf) that collectively show that lavas of the main volcanic edifices (shield volcanoes) are distinct in isotopic composition from associated minor-volume post-shield lavas. In particular, shield tholeiites define isotopic mixing arrays that require incorporation of at least two distinct mantle end members (cf., Stille et al. 1986; West et al. 1987). These are referred to as “Koolau” and “Kilauea” type components, as tholeiitic lavas from these two volcanoes typify the extremes in Pb and Sr isotopic composition for shield-building lavas. In contrast, the post-shield lavas from several Hawaiian volcanoes define distinctive mixing arrays between shield type and isotopically depleted mid ocean ridge basalt (MORB)-like components (Chen and Frey 1985; West and Leeman 1987a; West et al. 1987).

Although lavas with Koolau-type isotopic compositions have been considered anomalous or unique among Hawaiian tholeiites (e.g., Stille et al. 1986), tholeiites having similar isotopic compositions are now recognized on Lanai and Kahoolawe (West et al. 1987). Furthermore, tholeiites transitional between the two extremes (“Koolau” and “Kilauea”) are found on Kahoolawe (West et al. 1987). In this paper, isotopic, trace element, and major element variations are examined for a stratigraphically controlled suite of tholeiites from Kahoolawe that collectively exhibit the widest intra-volcano isotopic variations so far observed in the Hawaiian Islands. Previous studies of stratigraphically controlled samples from other Hawaiian shields covered only brief time spans (e.g., Mauna Loa and Kilauea; Wright 1971; Hofmann et al. 1984) and/or showed limited isotopic variations (e.g., Kohala, Mauna Kea; Hofmann et al.

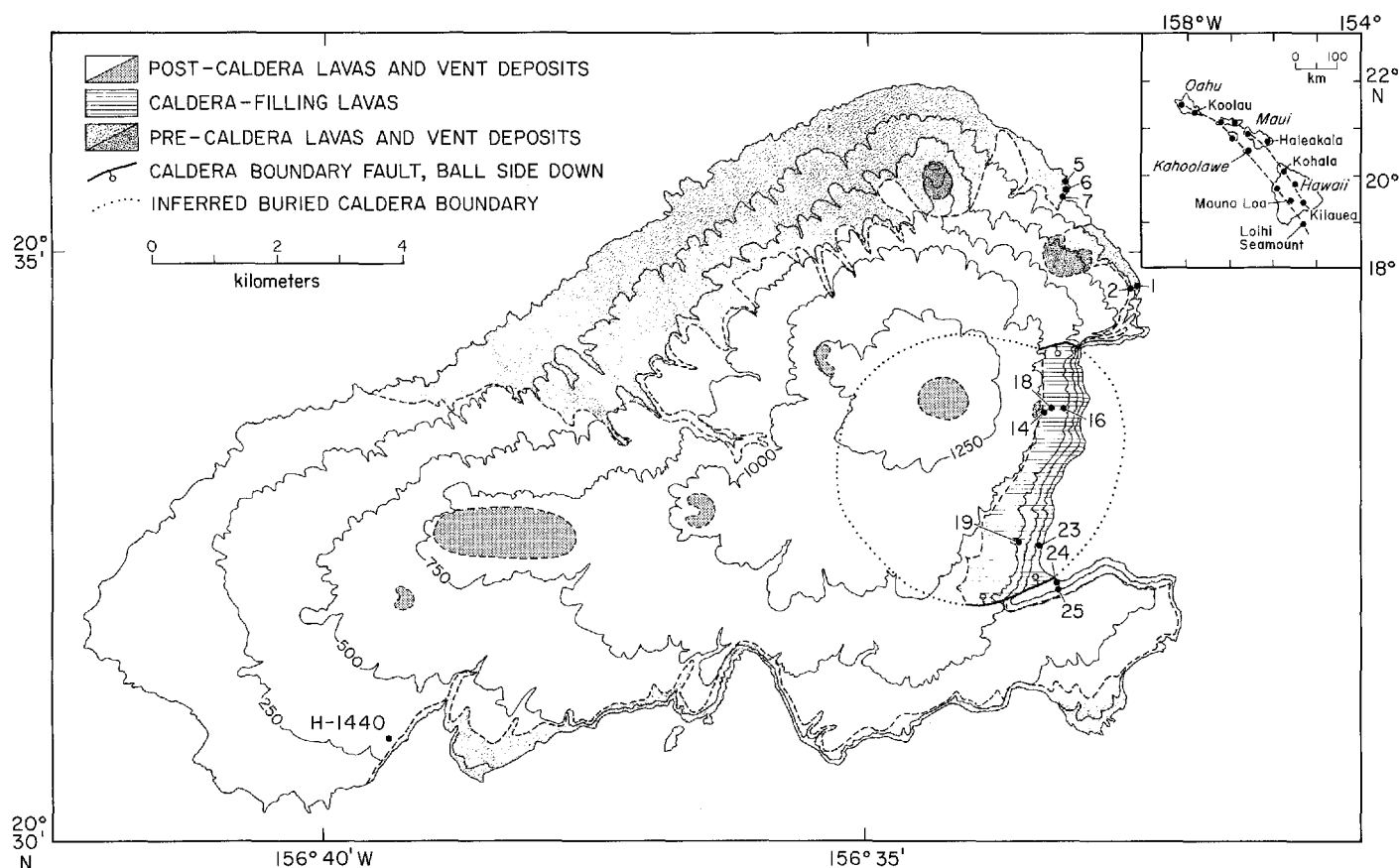


Fig. 1. Simplified geologic map of Kahoolawe Island (after Stearns 1940). Inset map shows location of Kahoolawe within the Hawaiian archipelago. Dots indicate localities of samples used in this study

1987; Lanphere and Frey 1987; Frey et al. 1990). Because of the relatively large isotopic variations at Kahoolawe, we can more readily (1) assess compositional variations as a function of relative age, and (2) discriminate between processes that potentially produced the tholeiite isotopic mixing array. This study also bears on geochemical implications concerning mantle geodynamics and magma ascent.

Geology of Kahoolawe volcano

Kahoolawe volcano lies on the westernmost of two subparallel volcanic lineaments ("Loa" to the west, "Kea" to the east) that together form the Hawaiian Islands (Fig. 1; cf., Jackson et al. 1972). The Loa trend includes West Molokai, Lanai, Kahoolawe, Hualalai, and Mauna Loa volcanoes, in order of decreasing age. Kahoolawe ($8.8 \times 10^3 \text{ km}^3$) is one of the smallest shield volcanoes in the Hawaiian Islands (Bargar and Jackson 1974). Unlike some of the larger volcanoes in this chain (e.g., Haleakala, Mauna Kea), it lacks a well developed post-shield alkalic suite. A generalized geologic map of the island is shown in Fig. 1.

Erosion on the east side of the island has exposed the interior of a caldera. Kahoolawe lavas are subdivided into three stratigraphic groups consisting of pre-caldera, caldera-filling, and post-caldera lavas (Stearns 1940). Pre-caldera lavas are thin bedded (typically 0.5–2 m thick) and dip moderately ($3\text{--}10^\circ$) away from the caldera. Caldera-filling lavas are thicker (3–30 m thick) and are virtually horizontal. The post-caldera lavas form a thin cap over

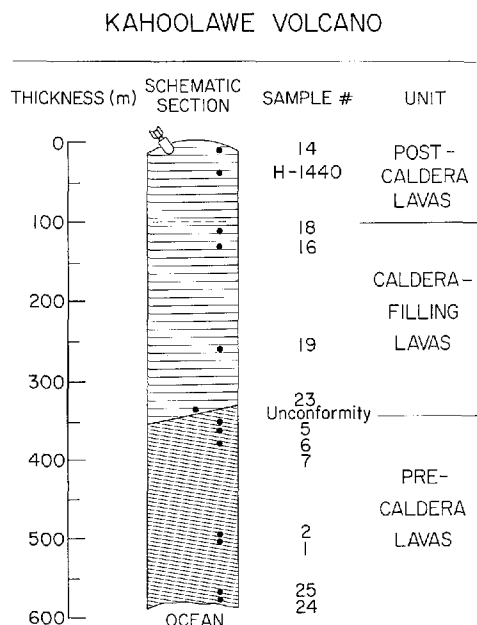


Fig. 2. Schematic composite stratigraphic column for the Kahoolawe lava suite used in this study. Note that pre-caldera samples were collected over a distance of $\sim 8 \text{ km}$. Their relative positions were estimated using attitudes of the lavas and assuming that the section is a homocline. An unconformity separates the thinly bedded, gently dipping pre-caldera lavas from the thicker, horizontal caldera-filling lavas. Bomb not to scale

the other units. Remnants of at least eight major post-caldera cinder cones are preserved.

The island of Kahoolawe has been a bombing target for U.S. military forces since World War II, and only recently has it been possible to collect samples. A total of 25 samples were collected along the eastern side of the island where an excellent section of pre-caldera and caldera-filling lavas is exposed; a subset of 13 samples was selected for geochemical analysis. Figure 2 presents a composite stratigraphic section of the nearly 600 m of exposed lavas, and shows the vertical position of all analyzed samples. These include seven pre-caldera, four caldera-filling, and two post-caldera lavas.

Naughton et al. (1980) published K–Ar dates of 1.02 and 1.05 Ma for two post-caldera lavas from Kahoolawe. New K–Ar dates (Fodor et al. 1992) range between 1.40–0.99 Ma for shield-building lavas and 1.20–1.08 Ma for post-caldera lavas. These ages are consistent with the observed age progression for Loa-trend volcanoes (Shaw et al. 1980) and with ages (1.21–1.46 Ma) of shield-building tholciites from nearby, but slightly older, Lanai volcano (Bonhommet et al. 1977). Several samples used in our study also were dated by the K/Ar method (Garcia, unpublished data); one pre-caldera sample (KW-5) yielded a 1.04 Ma age consistent with the published ages, but the remainder are anomalously young. Difficulties in dating young basaltic lavas are probably related to effects of minor tropical weathering or groundwater/hydrothermal alteration (cf., Chen and Frey 1985; Fodor et al. 1992; see below).

Petrography and previous work

Kahoolawe lavas petrographically resemble other Hawaiian lavas (Washington 1923; G. Macdonald, in Stearns 1940; Fodor et al. 1987, 1992). Our analyzed samples range from picritic (>15 vol.% olivine) and plagioclase-rich (>5 vol.%) basalts to aphyric hawaiites (Table 1). Olivine phenocrysts in most samples are euhedral and fresh, although some grains have thin iddingsite rims. Clinopyroxene and plagioclase phenocrysts are generally smaller (<1 mm) than olivine phenocrysts and also are euhedral. The groundmass varies from fine to coarse grained and consists of plagioclase, clinopyroxene, olivine, magnetite, and rare ilmenite and interstitial glass.

Picritic basalts (KW-24 and -25) from the basal section of the pre-caldera sequence contain euhedral olivine phenocrysts up to 2.5 mm in diameter as well as acicular olivines that are thought to form during rapid cooling of high-MgO melts (Donaldson 1979). The cores of the euhedral olivine phenocrysts range from Fo₈₈ to Fo₈₂, whereas the elongate grains are slightly more iron rich (Fo₈₆ to Fo₈₁); compositions of the olivine cores indicate that they equilibrated with less magnesian liquids – Mg# [Mg/(Mg+Fe)] ca. 0.66–0.60 – than the host lavas. Thus, the high MgO contents of these picrite basalts (16.7 and 15.3 wt%) in part reflect accumulation of olivine. One olivine-rich caldera-filling lava (KW-18, 12 wt% MgO) contains 12 vol.% subhedral olivine phenocrysts whose compositions (up to Fo₉₀) are too Mg-rich to have equilibrated with a liquid of the whole-rock composition. Because these olivines apparently crystallized from a relatively Mg-rich liquid, they are interpreted as xenocrysts. The host lava apparently represents a hybrid magma.

Analytical methods

Mineral analyses were carried out at the University of Hawaii using a Cameca electron microprobe. The XRF (X-ray fluorescence) analyses were conducted at the University of Leeds (trace elements) and the University of Edinburgh (major and trace elements). The INA (instrumental neutron activation) analyses were carried out at the Open University and at Oregon State University. Replicate analyses for major elements and selected trace elements were done by ICP (inductively coupled plasma) spectroscopy at Rice University. Because agreement was within a few percent for elements analyzed by more than one method, the averaged results are presented in Table 2, along with averaged analyses of the BHVO-1 standard. Analytical uncertainty is as follows: major elements (1–2%), all trace elements (<5–10%) except Rb and Nb (>20% in samples with less than 10 ppm). For all plots, the major element compositions are normalized to 100% on an anhydrous basis with all iron as FeO.

Isotopic analyses are summarized in Table 2 (cf., West et al. 1987). Analytical uncertainties are as follows: ²⁰⁶Pb/²⁰⁴Pb (±

Table 1. Modal mineralogy of Kahoolawe lavas

Samples	Olivine		Clinopyroxene		Plagioclase		Magnetite	Groundmass
	P	M	P	M	P	M	M	
Pre-caldera lavas								
KW-24 ^a	16.8	—	3.6	—	—	—	—	79.6
KW-25 ^a	20.6	—	5.8	—	—	—	0.1	73.5
KW-2	2.6	1.4	1.8	0.8	9.2	5.4	—	78.8
KW-2	4.4	0.8	0.4	0.4	3.0	3.3	—	87.7
KW-7	1.8	0.8	1.0	0.2	1.6	3.6	—	91.0
KW-6	2.0	1.2	—	—	—	5.8	—	91.0
KW-5	2.4	—	—	—	—	—	—	97.6
Caldera-filling lavas								
KW-23	—	27.8	—	18.2	—	46.6	7.4	—
KW-19	0.1	—	0.6	1.0	1.4	3.6	—	93.3
KW-16	1.2	1.8	1.8	0.4	2.0	1.4	—	91.4
KW-18 ^a	11.8	1.4	1.2	0.2	4.2	—	—	81.2
Post-caldera lavas								
H1440	0.1	0.5	—	—	0.1	0.2	0.4	98.7
KW-14	5.5	7.6	0.1	—	1.6	3.4	—	81.8

P, phenocryst (>0.5 mm); M, microphenocryst (0.1–0.5 mm); modes based on 500 points/sample

^a Sample contains acicular olivine

^b Sample has very coarse groundmass (0.2–0.5 mm); phenocrysts could not be identified readily

Table 2. Chemical and isotopic analyses of Kahoolawe lavas

Sample Elevation (m) Key/Fig. 8	Pre-caldera lavas				Caldera-filling lavas							Post-caldera lavas			Standard
	KW-24 20 1a	KW-25 30 2a	KW-1 95 3a	KW-2 105 4a	KW-7 220 5a	KW-6 240 6a	KW-5 250 7a	KW-23 260 1b	KW-19 340 2b	KW-16 465 3b	KW-18 485 4b	H-1440 560	KW-14 590	BHVO	
SiO ₂	47.87	49.69	50.75	49.53	48.83	50.07	49.35	49.98	50.25	48.85	49.5	49.43	51.96	50.16	
TiO ₂	1.79	1.84	2.44	2.60	3.35	2.38	2.53	1.77	2.51	3.03	1.97	3.02	2.47	2.78	
Al ₂ O ₃	11.01	10.75	14.09	14.94	14.91	14.45	14.44	12.04	14.09	14.81	12.48	16.32	14.91	13.88	
FeO*	12.07	11.87	11.76	12.02	12.96	11.91	11.33	12.42	11.10	12.46	11.14	10.62	10.5	11.01	
MnO	0.19	0.17	0.18	0.19	0.18	0.19	0.17	0.18	0.17	0.19	0.17	0.17	0.16	0.17	
MgO	16.71	15.27	6.82	6.18	5.80	7.23	7.52	11.15	6.13	5.94	12.04	4.62	5.93	7.15	
CaO	8.37	8.38	10.28	9.63	9.78	10.29	11.13	8.50	9.94	9.84	9.53	6.7	7.22	11.38	
Na ₂ O	1.82	1.83	2.45	2.55	2.88	2.25	2.54	2.18	2.62	2.87	1.98	4.65	3.55	2.35	
K ₂ O	0.12	0.34	0.31	0.15	0.15	0.12	0.28	0.39	0.56	0.19	0.08	1.83	1.38	0.53	
P ₂ O ₅	0.15	0.19	0.21	0.23	0.30	0.16	0.23	0.24	0.33	0.37	0.18	0.84	0.49	0.22	
LOI	0.05	ND	0.01	1.37	0.71	0.74	0.69	ND	ND	ND	ND	ND	ND	ND	
Total	100.15	100.33	99.30	99.39	99.85	99.79	100.21	98.85	97.70	98.55	99.07	98.20	98.57	99.63	
Mg#	71.2	69.6	50.8	47.8	44.4	52	54.2	61.5	49.6	45.9	65.8	43.7	50.2	53.7	
Ba (i,x) ^a	38	153	94	64	94	49	93	85	109	162	35	389	416	143	
Rb	1	4	4	BD	BD	BD	3	4	6	BD	BD	27	24	8	
Sr (i,x)	246	250	365	354	393	295	435	242	395	400	279	545	503	395	
Y (i,x)	14	18	24	24	33	23	23	29	31	35	25	53	100	25	
Zr (i,x)	98	106	156	164	232	146	167	142	180	216	118	420	301	178	
Nb	7	8	12	13	19	9	13	7	12	16	9	32	26	18	
Zn (i,x)	95	95	106	109	126	110	109	120	104	131	101	132	143	107	
Ni (i,x)	721	723	138	133	81	96	155	270	107	115	441	54	226	111	
Co	73	67	41	43	41	46	45	64	41	51	57	33	ND	44	
Cr (i,x,n)	937	853	211	209	186	313	304	641	150	118	641	30	249	271	
Se (i,x,n)	25	26	30	32	31	33	28	25	30	32	30	14	20	30.5	
V (i,x)	270	266	320	324	356	295	276	231	295	377	277	193	215	318	
La	5.5	7.6	9.5	10.7	15.7	8	10.9	9.7	13.7	16.2	8.9	38.2	51.1	15.0	
Ce	14	20	22.8	27.5	38	20.3	27.2	25.6	36.1	43.9	23.9	92.5	103.1	39.0	
Nd	ND	ND	ND	ND	ND	ND	ND	19.3	25.1	31	19.4	58.1	61.4	25.5	
Sm	2.86	3.87	4.67	5.1	6.94	4.34	5.29	5.24	6.41	7.49	4.87	13.3	13.4	6.37	
Eu	1.03	1.33	1.69	1.71	2.33	1.52	1.75	1.8	2.33	2.72	1.81	4.59	4.91	2.05	
Tb	0.45	0.63	0.74	0.79	1.02	0.71	0.8	0.97	1.14	1.26	0.88	1.99	2.46	0.91	
Yb	1.26	1.74	1.99	1.89	2.26	1.86	1.91	2.46	2.60	3.00	2.14	4.15	5.42	1.97	
Lu	0.18	0.26	0.28	0.29	0.32	0.27	0.29	0.36	0.39	0.45	0.30	0.61	0.83	0.28	
Hf	2.6	2.9	4.2	4.4	5.9	3.9	4.4	3.61	4.58	5.56	3.26	10.1	7.23	4.5	
Ta	0.55	0.62	0.87	0.96	1.5	0.77	1	0.68	0.91	1.44	0.67	2.35	2.53	1.3	
Th	0.31	0.34	0.57	0.75	1.43	0.48	0.73	0.63	1.02	1.67	0.58	3.11	2.50	1.0	
Sr87/86	0.70413	0.70416	0.70414	0.70421	0.70383	0.70404	0.70379	0.70409	0.70440	0.70415	0.70416	0.70426	0.70403	ND	
Nd143/144	0.512868	0.512848	0.512809	0.512784	0.512975	0.512887	0.512929	0.512901	0.512731	0.512897	0.512900	0.512864	0.512921	ND	
Pb206/204	18.025	18.047	17.954	17.921	18.337	18.120	18.369	18.036	17.946	18.149	18.092	18.005	18.027	ND	
Pb207/204	15.429	15.428	15.445	15.439	15.450	15.431	15.458	15.430	15.454	15.445	15.439	15.447	15.466	ND	
Pb208/204	37.759	37.786	37.805	37.733	37.990	37.842	38.030	37.800	37.836	37.845	37.866	37.817	37.770	ND	

ND, Not determined; BD, below detection limit

^a Letters indicate analytical methods averaged: i, ICP; x, XRF; h, INAA

0.017), $^{207}\text{Pb}/^{204}\text{Pb}$ (± 0.022), $^{208}\text{Pb}/^{204}\text{Pb}$ (± 0.05), $^{87}\text{Sr}/^{86}\text{Sr}$ (± 0.00003), and $^{143}\text{Nd}/^{144}\text{Nd}$ (± 0.00002). Ratios are corrected for mass fractionation relative to standards as follows: $^{87}\text{Sr}/^{86}\text{Sr}$ (0.71025, NBS 987), $^{143}\text{Nd}/^{144}\text{Nd}$ (0.51185, La Jolla Nd), and Pb isotopic ratios reported for NBS 981 (cf., Todt et al. 1984).

Results

Major and trace element variations

Kahoolawe pre-caldera and caldera-filling lavas are tholeiitic, whereas the two post-caldera lavas are classified as hawaiites based on their higher total alkali contents. Major element variations generally reflect the phenocryst contents. Magnesia-variation diagrams (not shown) are consistent with olivine control for the high-MgO basalts, and combined olivine, clinopyroxene, and plagioclase control in the more evolved tholeiites; the hawaiites are further evolved and exhibit effects of clinopyroxene, Fe—Ti oxide and apatite fractionation. The oxides TiO_2 , Al_2O_3 , Na_2O , and P_2O_5 generally increase with decreasing MgO in the tholeiites; CaO increases in the tholeiites but is relatively low in the hawaiites. Potash is poorly correlated with MgO as a consequence of variable K-loss during minor subaerial alteration (see Fig. 3).

Comparison of relative abundances of highly (Rb, Th, Ba, K, Nb, Ta, La, and Ce) and moderately (heavy REE, Zr, Hf, and Ti) incompatible elements in Kahoolawe lavas is facilitated using primordial mantle-normalized abundance profiles (Fig. 3). The sequence of elements shown is based on their predicted order of increasing compatibility in upper mantle lithologies (cf., Wood

et al. 1979). Patterns for most tholeiites are subparallel except for notable anomalies for Rb, Ba, and K_2O for some samples. Such anomalies apparently result from minor alteration and selective leaching or enrichment of the alkalis and alkaline earths. In fact, all but the two youngest tholeiites (KW-19, KW-23) display such anomalies. Thus, the data for K, Ba, and Rb (which also has relatively large analytical uncertainty) cannot be used reliably as petrogenetic indicators; these elements are not considered further in discussing the magmatic evolution of Kahoolawe lavas. In addition, Fodor et al. (1987, 1992) observed anomalous enrichment of Ba, REE, and Y in some of the Kahoolawe post-caldera lavas which they suggest resulted from weathering and soil formation. Such effects are apparent in only one of our samples (hawaiite KW-14).

Niobium was selected as an index element to monitor magmatic evolution because it is highly incompatible in the principal phenocryst phases, is insensitive to minor alteration, and has been emphasized in petrologic studies of ocean island basalts (OIBs) (e.g., Hofmann et al. 1986; Weaver 1991). Correlations of selected elements and element ratios with Nb are shown in Fig. 4. Contents of TiO_2 , P_2O_5 , V, and incompatible trace elements (Zr, Hf, Th, Ta, La) increase smoothly with Nb abundances in the tholeiites; Sc and Sr display inflections consistent with the onset of clinopyroxene and plagioclase removal. The hawaiites plot along these trends except for relative depletions in Sc, Ti, V, and P which reflect different phase assemblages crystallizing from these more evolved magmas. Systematic decreases of Mg#, Cr, Ni, and Co contents and $\text{CaO}/\text{Al}_2\text{O}_3$ ratios with increasing Nb in the tholeiites reflect fractionation

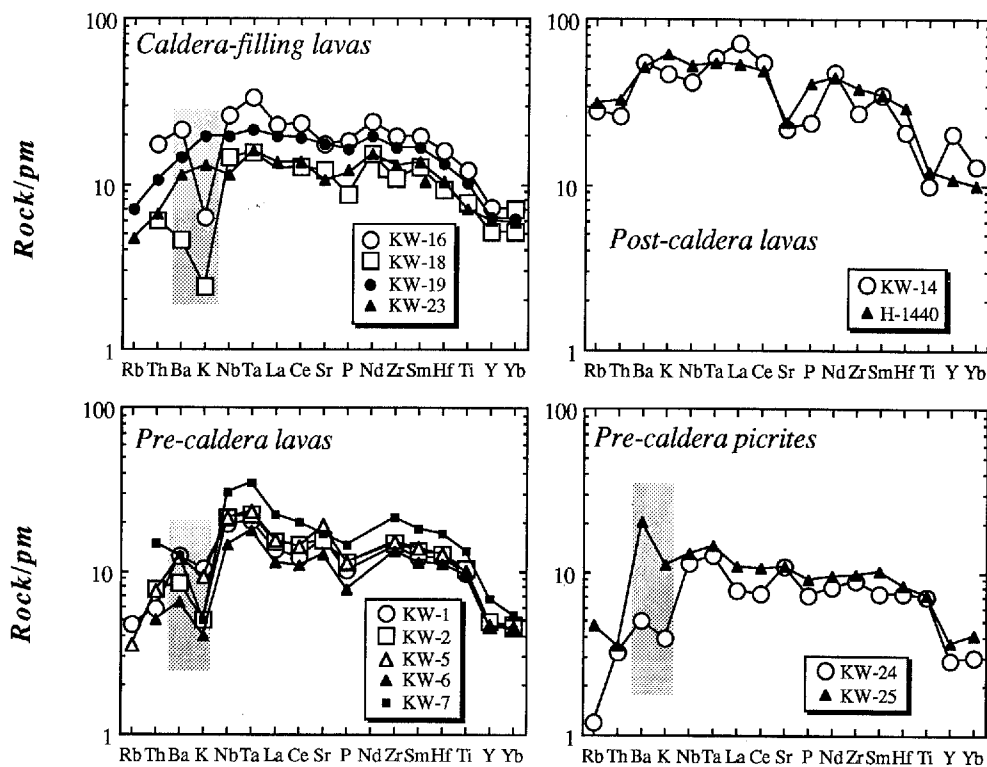


Fig. 3. Abundances of highly to moderately incompatible elements in pre-caldera tholeiites and picrites, caldera-filling tholeiites, and post-caldera hawaiites from Kahoolawe as normalized to primordial mantle, *pm*, abundances after Wood et al. (1979). Light stippled areas highlight the commonly erratic abundances of Ba and K, which apparently are affected by subaerial alteration

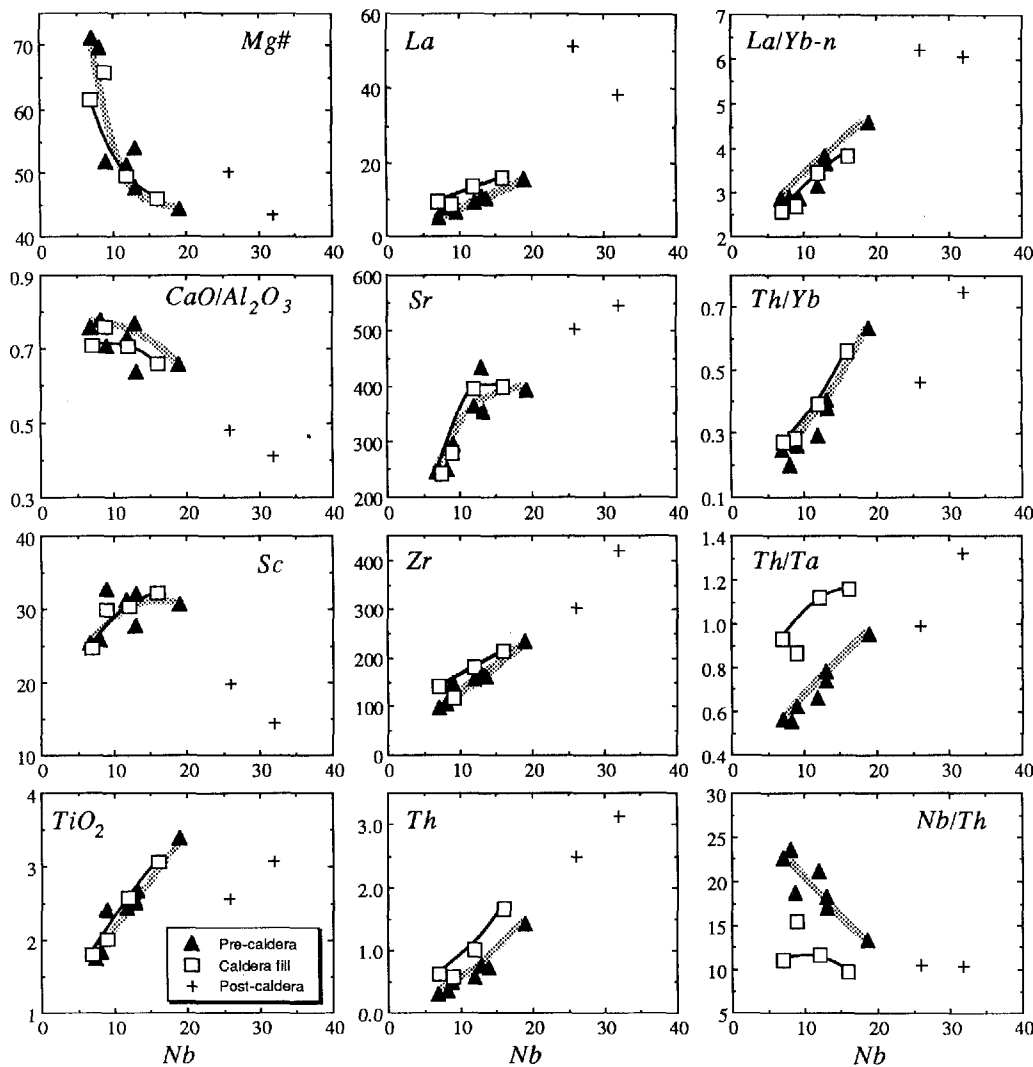


Fig. 4. Nb-variation diagrams. Nb serves as an index of differentiation against which other elements and ratios may be compared. *Stippled line* connects the lowermost pre-caldera lavas and *solid line* connects lowermost lavas of the caldera-filling sequence (groups "a" and "b" as discussed in the text)

of olivine and small amounts of clinopyroxene. The combination of high Nb and relatively low $\text{CaO}/\text{Al}_2\text{O}_3$, Sc, and V in the evolved hawaiites is consistent with fractionation of clinopyroxene and Fe—Ti oxides.

Despite strong correlations between Nb and other incompatible trace elements, ratios between such elements [e.g., chondrite normalized $(\text{La}/\text{Yb})_N$ (2.6–6.2), Th/Yb (0.25–0.73), Zr/Nb (12–20), and Hf/Ta (2.8–5.3)] vary significantly and coherently with increasing Nb (Fig. 4). Although most of these ratios normally are considered to be relatively invariant in closed magma systems (e.g., Hofmann et al. 1986; Weaver 1991), HREE, Y, Zr, and Hf are slightly less incompatible than Nb, Ta, and Th in clinopyroxenes and Fe—Ti oxides; thus, ratios involving these elements may vary slightly when the mineral assemblage includes these minerals (cf., Frey et al. 1990; Chen et al. 1990). Variations in other ratios [e.g., Nb/Th (9.6–23.5), Th/Ta (0.57–1.31)] in Kahoolawe tholeiites are surprisingly large considering that bulk distribution coefficients for these elements are probably small and quite similar. Ratios of Nb/Th measured in Hawaiian tholeiites are systematically higher than values reported for most oceanic island lavas from

the south Pacific and other regions (Fig. 5; cf., Weaver 1991). The range seen at Kahoolawe alone is comparable to that for tholeiites from the islands of Oahu, Maui, and Hawaii. These consistently high Nb/Th values have been observed in numerous studies (e.g., Staudigel et al. 1984; Lanphere and Frey 1987; Chen and Frey 1985; West et al. 1992; Leeman, unpublished data) and thus, are unlikely to represent analytical bias.

Isotopic variations

The wide variability of $^{87}\text{Sr}/^{86}\text{Sr}$ (0.70379–0.70440), $^{143}\text{Nd}/^{144}\text{Nd}$ (0.51273–0.51298), and $^{206}\text{Pb}/^{204}\text{Pb}$ (17.92–18.37) ratios in Kahoolawe tholeiites (Fig. 6) indicates that shield-building lavas do not always represent isotopically homogeneous melts, even in a specific volcano. The $^{87}\text{Sr}/^{86}\text{Sr}$ is negatively, and $^{143}\text{Nd}/^{144}\text{Nd}$ positively correlated with $^{206}\text{Pb}/^{204}\text{Pb}$; the near-linear arrays of isotopic data provide strong evidence for involvement of at least two isotopically distinct components in the genesis of Kahoolawe tholeiites. Thus, the lavas of this island mimic the isotopic variations seen for

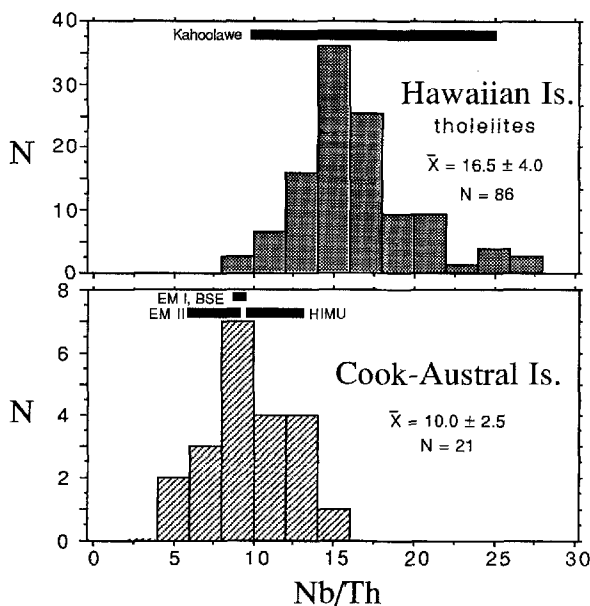


Fig. 5. Comparison of Nb/Th ratios in Hawaiian tholeiites versus basaltic rocks from the Cook-Austral Islands. *Solid bars* indicate the ranges in Nb/Th observed in Kahoolawe lavas and in averaged ratios for ocean island basalts corresponding to the different mantle reservoirs postulated by Zindler and Hart (1986). Data sources are: Hawaii (Chen and Frey 1985; Frey and Clague 1983; Lanphere and Frey 1987; Leeman and West, unpublished data); Cook-Austral Is. (Palacz and Saunders 1986; Dupuy et al. 1988, 1989; Leeman, unpublished data); EM I, EM II, HIMU, and BSE (cf., Weaver 1991; Anders and Ebihara 1982)

shield lavas in the archipelago in general (Stille et al. 1986; West et al. 1987). Despite the strong correlations between isotopic ratios, these data do not correlate in any simple way with trace element abundances or trace element ratios.

Temporal geochemical variations

The analyzed shield-building lavas are subdivided into stratigraphic groups separated by an angular unconformity as follows: group "a" includes the lower pre-caldera lavas (KW-24, KW-25, KW-1, KW-2, and KW-7), and group "b" includes the lower caldera-filling lavas (KW-23, KW-19, and KW-16). These groups were so defined because they display progressive decrease in Mg# upsection; higher lavas exhibit an increase in Mg# reflecting a perturbation of the normal fractionation sequence. Because the unconformity represents an unknown time interval, we have not attempted to project compositional trends across this boundary. Our analyses of post-caldera lavas are shown for comparison; they typify the data from Fodor et al. (1992). Subsequent discussion focuses on the tholeiitic succession owing to the sparse data for the alkalic lavas.

Within both groups a and b, there is a general upsection increase in TiO_2 , CaO, P_2O_5 , Na_2O , Th, Ta, Nb, Zr, Hf, V, Y, and REE, and decrease in MgO, Ni, Co, and Cr. Some major and trace elements (e.g., SiO_2 ,

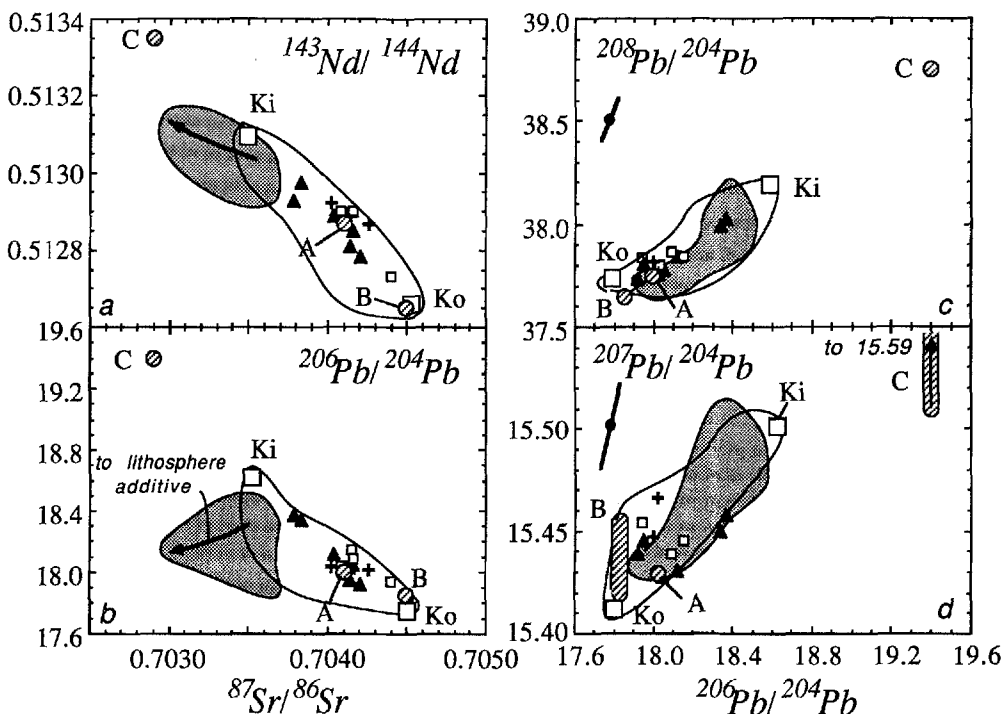


Fig. 6. Isotopic compositions of Kahoolawe lavas. Symbols are those used in Fig. 4. *Outlined fields* enclose data for shield-building tholeiites, *open*, and post-shield alkalic lavas, *stippled*, from other Hawaiian volcanoes; *arrows* in latter field (a, b) point to lithosphere component additive postulated for Haleakala (West and Leeman 1987a). Approximate compositions of Koolau ("Ko") and Kilauea ("Ki") end members, *open squares* are shown for reference (see text). *Diagonal ruled fields* indicate compositions of components

"A", "B", and "C" as inferred in this paper (see text and Fig. 8). Analytical uncertainty is either smaller than the symbols (a, b), or is indicated by 2σ error bars in upper left corner (c, d). Sources of data include Tatsumoto (1978), Sun (1980), White and Hofmann (1982), Stille et al. (1983, 1986), Roden et al. (1984), Staudigel et al. (1984), Chen and Frey (1985), Hegner et al. (1986), Park et al. (1988), West and Leeman (1987a), West et al. (1987), and Frey et al. (1990)

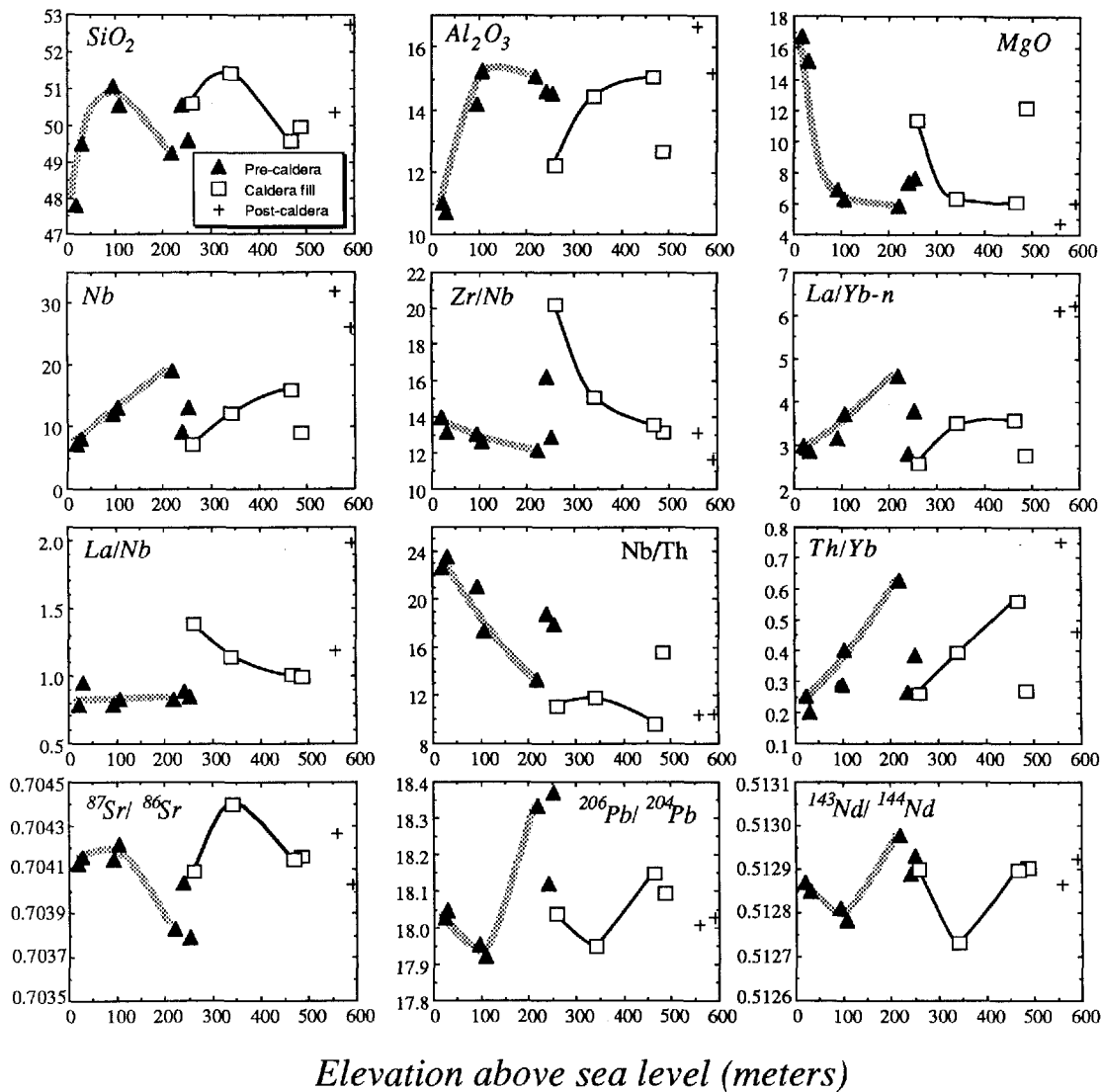


Fig. 7. Plots showing compositional variation of Kahoolawe lavas with respect to relative stratigraphic position (cf., Fig. 2). Note that an erosional unconformity separates the pre-caldera and cal-

dera-filling sequences. Symbols are those used in Fig. 4. *Stippled* and *solid* lines connect lower lava sequences of groups "a" and "b" as in Fig. 4

Al_2O_3 , Sc, Sr) display curved trends which correspond to changes in the crystallizing phase assemblage. With increasing stratigraphic height, $(\text{La}/\text{Yb})_N$ and Th/Yb (also Nb/Hf , Th/Ta , Th/Hf) ratios consistently increase, Nb/Th and Zr/Nb ratios decrease, whereas La/Nb is nearly constant (in group a) or decreases (in group b) with position upsection (Fig. 7). The uppermost samples in each sequence (KW-6 and KW-5; KW-18) show an increase in MgO and resemble the lowest samples in the respective sequences. The isotopic compositions of Kahoolawe lavas also fluctuate with time (Fig. 7). For example, $^{206}\text{Pb}/^{204}\text{Pb}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios initially decrease then increase within both of the "fractionation-controlled" sequences, whereas $^{87}\text{Sr}/^{86}\text{Sr}$ ratios display the reverse behavior. In each stratigraphic group, the uppermost lavas exhibit shifts toward isotopic compositions of the lowest lavas in much the same manner seen in the major and trace element data.

Discussion

Magmatic processes at Kahoolawe

These data indicate that the overall compositional evolution of Kahoolawe was complex and quite likely cyclic. The lack of complete flow-by-flow stratigraphic sampling precludes detailed interpretation of the temporal variations, but these are sufficiently large and systematic to indicate the nature and minimal extent of magmatic evolution. Within each group of shield-building lavas the systematic major and trace element variations are broadly consistent with progressive fractional crystallization upsection, but the uppermost lavas in each group exhibit reversals to less evolved compositions, suggesting periodic recharge of the Kahoolawe magma system with more primitive liquids. Additionally, variations in isotopic and incompatible element ratios suggest that fractionation must have been accompanied by AFC-like

open system behavior involving addition of compositionally distinct material to the evolving magma system from at least three (see below) isotopically and chemically distinct reservoirs. Detailed sampling of Haleakala post-shield lavas (West and Leeman 1987b; Chen et al. 1990) and Koolau tholeiites (Frey et al. 1993) reveals similar complexities.

The roles of fractional crystallization and open system behavior

Fractional crystallization alone obviously cannot explain many features of the Kahoolawe data, but this process clearly contributed to some of the compositional variation. For illustration, the magnitude of fractionation required to explain major element variations in the lowermost lavas of group a can be approximated using least-squares mass balance calculations. For this exercise, mineral compositions from other Hawaiian tholeiites of similar composition were used (e.g., Frey et al. 1990), and K_2O contents of altered samples were adjusted to yield K_2O/P_2O_5 ratios typical of fresh Hawaiian tholeiites (i.e., ca. 1.8; Wright 1971). Taking the picritic basalts (KW-24 and KW-25) as hypothetical parental magmas, formation of the more evolved tholeiites KW-1 and KW-2, respectively, requires crystallization of approximately 23 to 35% of olivine and minor clinopyroxene. Assuming a Rayleigh fractionation model and trace element distribution coefficients approaching zero, observed enrichments of Th suggest much higher degrees of crystallization (ca. 45–60%); Nb, Ta, and La enrichments indicate slightly lower degrees of crystallization (ca. 40–50%), suggesting that these elements are slightly less incompatible than Th. The magnitude of these estimates depends upon the choice of parental liquid, but the discrepancy between major and trace element modeling persists in all cases. Such behavior is characteristic of periodically tapped and replenished magma systems (cf., O'Hara and Matthews 1981).

The roles of source composition and partial melting

Some of the variation in isotopic and trace element ratios of Kahoolawe lavas could originate from source heterogeneities or varied melting conditions. For example, assuming a homogeneous source, large ranges in highly incompatible element ratios are more effectively produced by variations in degree of partial melting (particularly at low melt fractions) and/or by involvement of compositionally diverse magma sources than by fractional crystallization processes. Considering that both Nb and Th behave as highly incompatible elements in basaltic magmas (e.g., Hofmann 1988), the range of Nb/Th ratios seen in Kahoolawe lavas is unlikely to result from melting of a uniform source; this would require at least a two-fold difference between D_{Nb} and D_{Th} and very low melt fractions ($<0.1\%$; cf., Hickey et al. 1986). Similar restrictions also apply to dynamic melting models such as those proposed by McKenzie and O'Nions (1991) or Eggins (1992).

In contrast, the range in Nb/Th in Kahoolawe magmas and observed correlations of Nb/Th with isotopic ratios are plausibly attributed to the operation of one or more of the following processes: (1) melting of a heterogeneous source; (2) mixing of magmas derived from distinct sources; or (3) assimilation of material different in composition from the magma source(s). Similar conclusions have been drawn from correlations between isotopic and incompatible trace element ratios in post-shield-building alkalic lavas from Kauai (Feigenson 1984) and Haleakala (Chen and Frey 1985; West and Leeman 1987a). For example, compositions of Haleakala post-shield magmas appear to require mixing between plume and MORB source (i.e., depleted mantle) components (West and Leeman 1987a; Chen et al. 1990). However, source compositions and their relative contributions to magma evolution may be quite different for shield-building versus post-shield magmatism.

Nature of end members contributing to Kahoolawe tholeiitic magmatism

Identification of chemical reservoirs was approached using combined isotopic and incompatible trace element ratios, both of which are unlikely to vary significantly with fractional crystallization or during partial melting. The ratio Th/Ta was used in lieu of the analogous Nb/Th ratio because Ta was more precisely determined than Nb. The ratio Th/Ta correlates strongly with incompatible element contents (e.g., Nb), and fluctuates in a quasi-cyclic manner through the Kahoolawe section. The Th/Ta varies in complex fashion with isotopic compositions, but numerous sub-trends are apparent in Fig. 8, particularly when viewed in a relative temporal sequence. In detail, these variations appear to be controlled by three end member components defined by intersections of the sub-trends in the pre-caldera and caldera-filling sequences.

The first end member (denoted "A") corresponds to the lowest Th/Ta (or highest Nb/Th) lavas at the base of the section; these are also the most primitive magmas as noted earlier. The second pre-caldera lava from the top (KW-6), and the uppermost caldera-filling lava (identified as a hybrid based on olivine compositions), also define trends back toward component A, suggesting recharge of the magma system with liquids derived from this source. A second end member ("B") is indicated by progressive variations in the fractionation-controlled sequences (groups a and b). Admittedly, the $^{207}Pb/^{204}Pb$ ratio in end member B is poorly constrained because of relatively large analytical uncertainty; trends for the two tholeiite groups intersect at a $^{207}Pb/^{204}Pb$ value of 15.46 when anomalous sample KW-1 is ignored. Because it is evident only in the fractionated lavas containing less than 7% MgO, end member B may correspond to an assimilated component with which the evolving magmas interacted (cf., Leeman and Hawkesworth 1986); its Th/Ta ratio is ca. 1.2 as constrained by intersecting trends. Involvement of a third end member is indicated by variations in the upper-

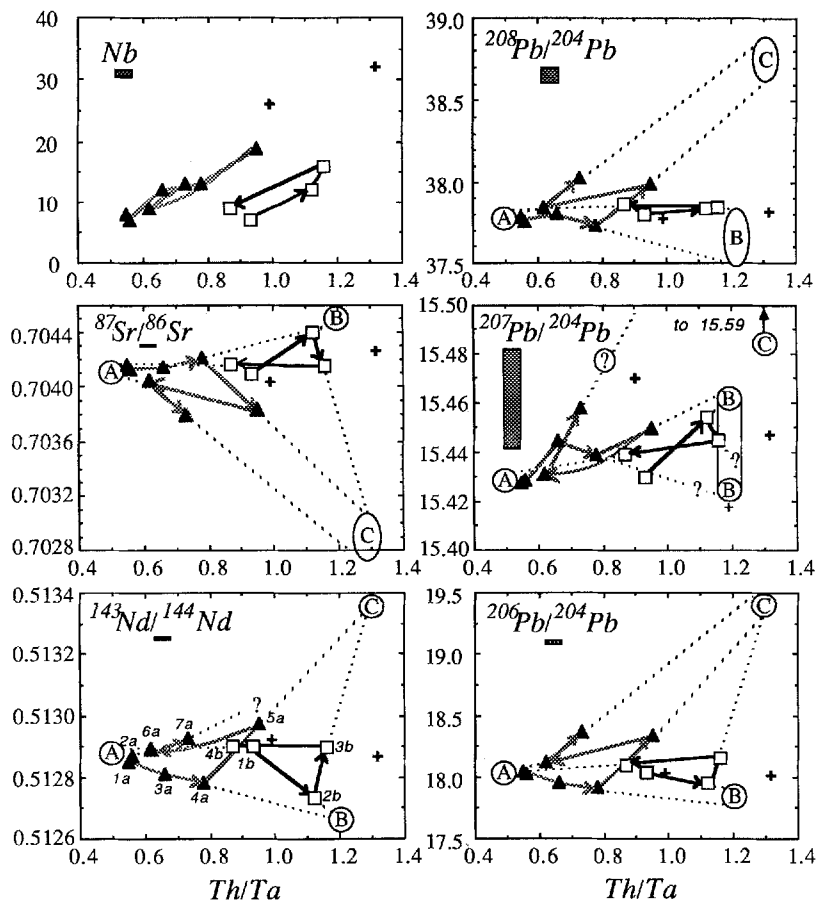


Fig. 8. Variations between Th/Ta versus isotope ratios in Kahoolawe lavas. Symbols are those used in Fig. 4. Pre-caldera and caldera-filling tholeiites are connected by *stippled* and *solid* lines, respectively; *arrows* indicate eruptive sequence, and representative examples are *numbered in relative stratigraphic position* for clarity (see key to samples in Table 2 and compare with evolutionary sequences in Fig. 7). *Dashed lines* show projections of sub-trends for the tholeiites; these tend to converge at three approximate end member compositions denoted by fields labelled "A", "B", and "C" at positions corresponding to Th/Ta ratios near 0.5, 1.2, and 1.3, respectively. Involvement of these end members accounts for most of the compositional variation seen in Kahoolawe tholeiites. Estimated analytical uncertainties are indicated by *shaded bars*

Table 3. Estimated compositions of reservoirs contributing to Hawaiian tholeiites

End member	A ^a	B ^a	C ^a	"Koolau" ^b	"Kilauea" ^b	N-MORB ^c
87Sr/86Sr	0.7041	0.7045	0.7027–0.7031	0.7045	0.7035	0.7032 ^d
143Nd/144Nd	0.51287	0.51265	0.5133–0.5134	0.51264	0.51307	0.5131 ^d
206Pb/204Pb	18.0	17.8–17.9	19.3–19.5	17.7	18.6	19.4
207Pb/204Pb	15.43	15.42–15.46	≥15.47–15.59	15.41	15.5	15.56 ^d
208Pb/204Pb	37.75	37.5–37.8	38.6–38.9	37.73	38.2	39.0 ^d
Th/Ta	0.5	1.2	1.3			0.91
Nb/Th	23.6	12	8.9			19.4
Th/Hf	0.11	0.25	0.45			0.059
Th/Yb	0.15	0.54	1.1			0.039
(La/Yb) _N	2.0–2.5	4.0–4.5	6.5–8.0			0.5

^a End members A–C estimated from compositional variations in Kahoolawe lavas (this paper)

^b "Koolau"- and "Kilauea"-types are based on Hawaiian tholeiite isotopic data (West et al. 1987)

^c N-MORB trace elements from Sun and McDonough (1989)

^d Isotopic ratios interpolated from correlations of White et al. (1987) assuming a 206Pb/204Pb ratio of 19.4

most, most evolved group a and group b lavas. Trends for these lavas project to a well-defined intersection point ("C") for every isotope ratio except $^{207}\text{Pb}/^{204}\text{Pb}$, for which analytical uncertainty again precludes accurate projections. The intersections are consistent with a Th/Ta ratio of ca. 1.3 in end member "C"; a tentative $^{207}\text{Pb}/^{204}\text{Pb}$ ratio as high as 15.59 is obtained by projecting the upper pre-caldera tholeiite trend to this Th/Ta value. Component C must represent a distinct reservoir with which only the most evolved tholeiites interacted to any significant extent.

Some aspects of this variation are seen at other Hawaiian volcanoes, but nowhere as completely developed as at Kahoolawe. For example, Koolau tholeiites (Frey et al. 1993) define Th/Ta versus Sr or Nd isotopic trends between ranging between end members A and B. On the other extreme, tholeiites and alkali basalts from the Loihi summit crater (Garcia et al. 1993) cluster near Kahoolawe sample KW-7 (i.e., toward end member C) in plots of Th/Ta versus Pb, Sr, and Nd isotopic ratios.

The nature of these three reservoirs may be further elucidated using plots of other incompatible trace ele-

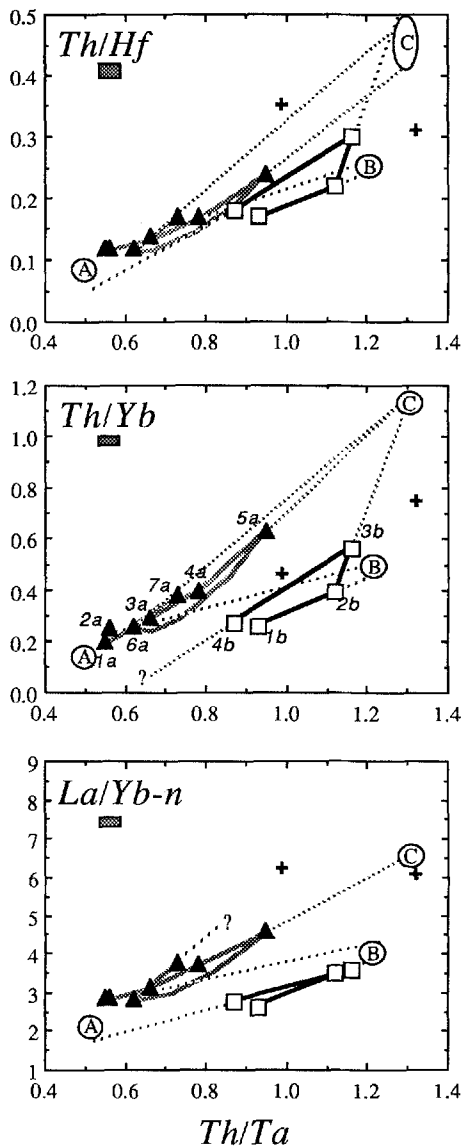


Fig. 9. Variations between Th/Ta versus selected trace element ratios in Kahoolawe lavas. Symbols, line styles, and end member designations are those used in Fig. 8. Estimated analytical uncertainties are indicated by shaded bars

ment ratios versus Th/Ta (Fig. 9). Projections using the same sample groupings and sequences as in Fig. 8 define approximate trace element ratios for components A, B, and C at Th/Ta values inferred from the isotope correlations. All incompatible trace element ratio trends, with exception of those involving Nb, and all isotopic ratio trends for the tholeiites in Figs. 8 and 9 are confined within triangles defined by these end members. Estimates of trace element ratios are summarized in Table 3; they are considered provisional because absolute trace element abundances are subject to larger analytical uncertainties than are the isotopic data, and in some cases may have been slightly modified by secondary processes. For example, La/Nb, Zr/Nb, and Nb/Th ratios in the primitive "recharge" magmas (KW-6 and KW-23) deviate from the estimated composition of A in a manner consistent with erroneously low Nb determinations in

these two low Nb samples (cf., their positions on correlation plots in Fig. 4). Until all samples are reanalyzed by more precise methods (work in progress) only the following brief comments are made concerning trace element compositions of Kahoolawe isotopic reservoirs.

End members A and B have similar, yet distinct Sr, Nd, and Pb compositions. Whereas A has isotopic ratios similar to the median values for Hawaiian lavas (see Fig. 6), B plots at one extreme of the Hawaiian isotope arrays and is essentially equivalent to the Koolau end member. Many of the inferred trace element ratios differ significantly between A and B, particularly those involving Th which appears to be enriched in B by at least a factor of two; this difference far exceeds analytical uncertainties. Higher $(\text{La/Yb})_N$ in B suggests that this end member is enriched in light REE. The other estimated ratios are not significantly distinct. Thus, combined isotopic and trace element characteristics imply that A and B represent distinct reservoirs in the mantle below Kahoolawe. Because A corresponds to the highest MgO lavas, and because such lavas recur periodically in the Kahoolawe section, it is reasonable to infer that it most closely approximates the principal magma source – presumably an ascending plume. The fact that component B seems involved only in moderately evolved tholeiites suggests that it may represent material with which the more primitive magmas interact following segregation from their source. Physically, B may represent upper asthenospheric mantle into which the plume and its derivative magmas are intruded. Although B could also be explained as a distinct component in a heterogeneous plume, in this scenario it is difficult to see why it appears only in fractionated lavas.

The most distinctive features of end member C are its low $^{87}\text{Sr}/^{86}\text{Sr}$ and high $^{143}\text{Nd}/^{144}\text{Nd}$ ratios, in which regard it resembles certain MORBs. As seen in Table 3, this similarity extends to the lead isotopic data when comparison is based on isotopic ratios interpolated from the MORB arrays (cf., White et al. 1987) at points corresponding to the inferred $^{206}\text{Pb}/^{204}\text{Pb}$ value for C (19.4). Relatively high Th/Ta, Nb/Hf, Th/La, Th/Hf, Th/Yb, Ta/Yb, $(\text{La/Yb})_N$, and low Nb/Th, Zr/Nb, and Y/Nb ratios estimated for C are significantly distinct from those in N (normal)-MORB; ratios in some OIBs produced by low degrees of melting provide a much closer match (cf., Sun and McDonough 1989). Thus, C could represent a low-degree partial melt of material with MORB-like isotopic ratios. Because the most fractionated Kahoolawe tholeiites display evidence for clinopyroxene fractionation (corresponding to crystallization at pressures > ca. 0.8 GPa; cf., Green and Ringwood 1967), component C likely is derived from lithospheric mantle (comprised of depleted MORB source?) rather than oceanic crust. The involvement of a MORB-like component in the formation of Haleakala post-shield lavas was proposed by Chen and Frey (1983), who postulated mixing between tholeiitic magma and melts derived by fusion of oceanic lithosphere. At Kahoolawe, this lithospheric additive (C) has much more radiogenic Pb than inferred from Haleakala lavas (see Fig. 6). Thus, the lithospheric mantle below the Hawaiian Islands ap-

pears to be isotopically heterogeneous with regards to Pb (cf., Chen 1987).

A possible scenario for evolution of Kahoolawe magmas

Our data imply that the Kahoolawe lavas contain varied contributions from at least three geochemically distinct reservoirs (cf., Staudigel et al. 1984). The processes responsible for the observed compositional variations and where they occur are important questions that bear on the geochemical structure of the mantle beneath the Hawaiian Islands. A plausible scenario that accounts for much of the geochemical evidence is presented briefly below and is illustrated schematically in Fig. 10. This model is largely suggested by the repetitive systematic evolution of Kahoolawe lavas, and includes reasonable albeit speculative assumptions concerning the distribution of components A, B, and C:

1. Primitive magmas presumably formed by partial melting of an ascending plume, which beneath Kahoolawe had a mean composition approximated by component A. Based on simple decompression melting models (e.g., McKenzie and Bickle 1988), melt segregation likely occurred at depths on the order of 100 km (cf., Liu and Chase 1991; Watson and McKenzie 1991).
2. Upon segregation from the plume source, ascending magmas passed through upper asthenospheric or thermally accreted lower lithospheric mantle where they assimilated material of composition B. Storage of primitive

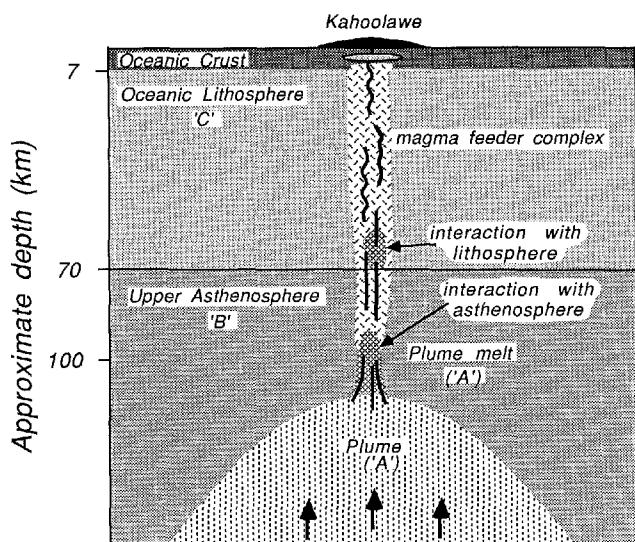


Fig. 10. Schematic cross section showing inferred spatial distribution of reservoirs corresponding to the inferred mantle end members that contributed to Kahoolawe magmatism. Depths of melt segregation (ca. 100 km) and thickness of the oceanic lithosphere (ca. 70 km) are taken from Watson and McKenzie (1991). The lithosphere is considered to be a coherent geochemical reservoir (e.g., mechanical boundary layer) that is isolated from the convecting upper asthenospheric mantle. Regions of inferred magma – wall rock interaction are shown schematically. Alternative scenarios, e.g., involving entrainment of upper asthenosphere (“B”) within ascending plume (“A”) material prior to melt segregation, are considered to be less consistent with overall compositional variations in Kahoolawe lavas. See text for details

tholeiitic magma at great depths, where assimilation of partial melts of asthenospheric mantle could occur, is a remote possibility suggested by the discovery of a garnet-bearing cumulate xenolith from Oahu that equilibrated near 3.0 GPa (ca. 90 km; cf., Sen and Jones 1990). However, a more likely scenario might involve entrainment of B-like material (or its partial melts) into ascending plume-derived magma. Mechanical incorporation of B-like material into the rising plume and subsequent melting of variably mixed (A + B) source material also might occur, but this process is unlikely to cause the coupled bulk chemical and isotopic variations observed in Kahoolawe tholeiites.

3. Assimilation of distinct C-type material likely occurred during subsequent ascent and storage of tholeiitic magmas at lithospheric depths (\leq ca. 70 km below the Hawaiian Islands; Watson and McKenzie 1991). Such additives could consist of low-degree partial melts of relatively fertile E(enriched)-MORB-source mantle, or possibly remelts of MORB-composition intrusive bodies within the lithosphere, formed in response to injection of hot plume-derived magmas. Seismic evidence indicates the presence of a complex magma conduit system that extends to at least 60 km below Kilauea (cf., Klein et al. 1987; Ryan 1988). The roots of such a system may extend to the depths of magma segregation from an ascending plume, in which case part of the compositional variation seen at Kahoolawe could reflect comingling of magmas stored at different levels within this dynamic system. However, the observed variations between isotopic compositions and incompatible trace element ratios for Kahoolawe lavas are not consistent with simple binary mixing between ascending magmas derived from two compositionally distinct sources.

The preservation of such unusual compositional diversity at Kahoolawe, in contrast to more uniform lavas at Kilauea for example (Leeman et al. 1980; Tilling et al. 1987), may be related to its relatively low magma supply rate and smaller size. The involvement of component B has been recognized exclusively in lavas from Kahoolawe, Lanai, and Koolau volcanoes, whereas this component has not been detected in other, usually larger, Hawaiian shield volcanoes. Several possible explanations could be related to plume dynamics. First, temporal variations in plume flux (cf., Skilbeck and Whitehead 1978; Davies 1992) could affect the depth of melt production and ascent. We envisage relatively vigorous upwelling beneath the larger Hawaiian volcanoes such that the plume impinges on the base of the lithosphere where melt segregation occurs – in this case, ascending magmas would “see” little if any asthenosphere. Slower or less voluminous upwelling beneath smaller volcanoes could result in melt segregation at greater (sub-lithospheric) depths and transit of magma through part of the upper asthenosphere. Secondly, efficiency of wall-rock interaction may be inversely related to total magma flux; e.g., beneath larger volcanoes relatively high mass ratios of magma to assimilated material could minimize compositional effects of magma-asthenosphere interaction. However, contributions of B component in lavas from the relatively large Koolau shield are not simply reconciled with

these processes. Finally, fluctuations in plume temperature could lead to varied interactions between mantle wall rocks and either the plume or its segregated melts; greater interaction would be expected during periods of elevated temperature, and vice versa.

Alternatively, the appearance of end member B in lavas from Kahoolawe could simply reflect heterogeneities in the asthenosphere, with a greater proportion of the B-like component in that vicinity. Mantle xenoliths provide direct evidence for this possibility; Okano et al. (1987) found a spinel lherzolite from Oahu to have B-like Sr and Nd isotopic compositions. Finally, if B residues in the lithosphere, this material must have been introduced into essentially MORB-like lithosphere prior to onset of Hawaiian volcanism – most likely from the underlying asthenosphere; possible mechanisms include magmatic transfer or accretion of asthenospheric mantle to the base of the cooling lithosphere.

Implications for compositional heterogeneities in suboceanic mantle

Figure 11 contrasts compositions of the Kahoolawe tholeiites with our inferred Kahoolawe end members, the previously postulated Koolau (essentially equivalent to B) and Kilauea “end members”, and the various mantle “reservoirs” suggested to explain global isotopic variations in OIB (cf., Zindler and Hart 1986); Nb/Th ratios for the latter end members are inferred from the compilations of Anders and Ebihara (1982), Sun and McDonough (1989), and Weaver (1991). The most primitive Kahoolawe tholeiites, whose compositions apparently are derived from a high Nb/Th source (end member A), do not correspond to any previously postulated mantle end members. Therefore, our data suggest that the Kahoolawe mantle plume is compositionally distinct from both lithospheric and shallow asthenospheric mantle beneath the Hawaiian Islands.

The end members defined here for Kahoolawe appear to be involved to varying degrees in the formation or evolution of other Hawaiian magmas. Isotopic variations in Kahoolawe lavas are confined within relatively narrow bounds between the Koolau and Kilauea end members, which themselves may represent either compositions of specific mantle domains beneath the Hawaiian Islands or mixtures between more extreme mantle variants. For comparison, tholeiitic and mildly alkalic lavas from Loihi seamount (Frey and Clague 1983) exhibit virtually the same span in Nb/Th observed at Kahoolawe, although their isotopic ratios are consistent with a slightly larger relative contribution of component C; this difference is more pronounced in lavas from Loihi summit crater (Garcia et al. 1993). If Loihi and Kilauea lavas are polluted by lithosphere material, they may be less representative of the ascending plume than the primitive Kahoolawe lavas. In other words, the “Ki” end member could be a mixture between our A and C components. An alternative view is that larger volcanoes are prone to produce magmas that more faithfully reflect their plume-source composition, whereas at smaller volcanoes such as Kahoolawe smaller volumes of plume-derived magma may be more susceptible to contamination by non-plume components. In this case, A could be a mixture between Ki (hypothetically, the true plume) and B. The latter scenario is least attractive because B and probably “Ki” both have higher Th/Ta or Th/Nb than A. Because A type characteristics are seen in some lavas from the relatively large Koolau shield, edifice size and inferred plume flux correlate poorly with chemistry.

If the high Nb/Th A end member always corresponds to the rising Hawaiian plume, then the small inter-volcano isotopic variations observed (Fig. 6) imply either that: (1) the plume is slightly heterogeneous on the volume scale of magma production (cf., Allegre and Turcotte 1986); or (2) it contains varied amounts of material entrained from compositionally distinct asthenospheric or lithospheric mantle reservoirs (Griffiths and Camp-

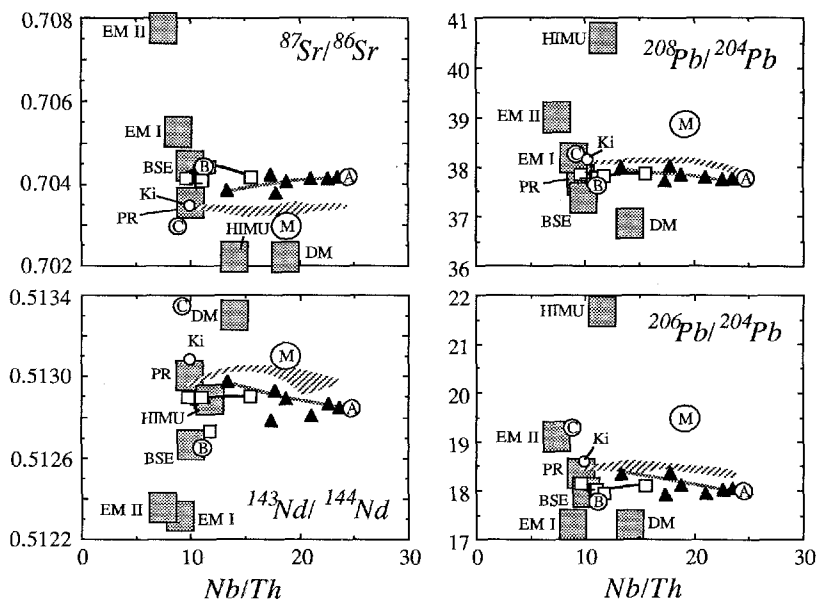


Fig. 11. Plot of Nb/Th versus Nb and isotopic ratios. Kahoolawe lavas are shown with symbols used in Fig. 4. Analytical errors are comparable to or smaller than symbol sizes. Compositions of postulated Kahoolawe end members “A”, “B”, and “C” are compared with the Kilauea (“Ki”; West et al. 1987), EM I, EM II, DM, PREMA, HIMU, and BSE end members proposed to explain variations in oceanic island basalts in general (Zindler and Hart 1986; Staudigel et al. 1991). Circled M indicates the composition of MORB as interpolated from isotope-isotope correlations at a $^{206}\text{Pb}/^{204}\text{Pb}$ value of 19.4 (White et al. 1987; Sun and McDonough 1989). The diagonally ruled field shows compositions of lavas from Loihi seamount (Frey and Clague 1983; Garcia et al. 1993); note that isotopic ratios in high Nb/Th Loihi samples differ slightly from those at Kahoolawe.

bell 1990). Isotopic diversity between Hawaiian volcanoes could reflect a non-uniform temporal and/or spatial distribution of such heterogeneities within the ascending plume material (variants of A) or within convecting upper mantle material (B) that becomes entrained in the rising plume or with which plume-derived magmas interact during ascent.

Finally, the high Nb/Th ratios inferred for the Kahoolawe mantle plume are consistent with Hofmann and White's (1982) suggestion that plumes may form by diapirism of subducted refractory ancient oceanic slabs from which low Nb/Th arc magmas were previously extracted. Differences in the age and in the metasomatism and melt extraction history of such material may lead to significant variations in the compositions of plumes world-wide, and within individual plumes feeding specific hot spot volcanic chains.

Summary and conclusions

Tholeiitic basalts from Kahoolawe volcano display wide ranges in Sr, Nd, and Pb isotopic ratios and in ratios between highly incompatible trace elements. These data are inconsistent with simple closed system magma evolution or partial melting of a single homogeneous mantle source. Compositional variations through the stratigraphic sequence are consistent with quasi-cyclic evolution of the magma system with periodic influx of relatively primitive high-MgO magmas interspersed with eruptions of variably fractionated and contaminated magmas. At least three distinct components evidently are required to explain the formation and evolution of Kahoolawe magmas. Combined isotopic and trace element characteristics deduced for these components have important implications for their spatial distribution below Kahoolawe.

The most magnesian Kahoolawe lavas are characterized by distinctive trace element ratios (e.g., high Nb/Th) and relatively uniform isotopic compositions which are interpreted as characteristics of an ascending mantle plume. Moderately evolved lavas display compositional shifts toward an isotopically and compositionally distinct (low Nb/Th) component that has been detected only in tholeiites from Koolau, Lanai, and Kahoolawe volcanoes (West et al. 1987). This component is interpreted to reside in, or be derived from, upper asthenospheric mantle. Finally, the most evolved tholeiitic lavas display compositional shifts toward another component that is isotopically similar to some mid-ocean ridge basalts and is inferred to reside in the oceanic lithosphere.

A simple physical model to explain the temporal variations seen at Kahoolawe involves segregation of primitive magmas from an ascending plume followed by assimilation of partial melts derived from upper asthenospheric or lithospheric mantle wall rocks. In this scenario, the eruptive sequence at Kahoolawe requires periodic eruption of magmas from different levels within the magma conduit system – possibly ranging from at least 60–70 km to depths exceeding 100 km (cf., Watson and McKenzie 1991).

The mantle components defined here for Kahoolawe may be involved to varying degrees in the formation and evolution of magmas from other Hawaiian volcanoes (e.g., Loihi seamount). However, differences in the isotopic composition of the high Nb/Th end member (i.e., plume source) at Kahoolawe and Loihi suggest that the ascending plume may be slightly heterogeneous. The identification of an upper asthenospheric mantle component in only a few Hawaiian volcanoes suggests either that the plume-derived magmas seldom interact with wall rocks at asthenospheric depths, or that the asthenosphere itself may be heterogeneous. Because the lithospheric end member identified at Kahoolawe has much more radiogenic Pb than that inferred for other Hawaiian volcanoes (e.g., Haleakala), this reservoir is also heterogeneous. A distinctive characteristic of the Hawaiian plume is its elevated Nb/Th ratio. This feature is consistent with the derivation of the plume from ancient subducted oceanic lithosphere from which low Nb/Th arc magmas were extracted previously (Hofmann and White 1982; Silver et al. 1987; Campbell and Griffiths 1992).

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