The Canadian Mineralogist
Vol. 40, pp. 595-609 (2002)

PLATINUM-GROUP ELEMENTS IN BASALTS FROM MAUI, HAWAI'I: LOW ABUNDANCES IN ALKALI BASALTS

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ABSTRACT

Platinum-group elements (PGE) in alkali basalts from Maui, Hawai'i are as much as eight times lower than in tholeiitic basalts from Kīlauea. Correlations among different PGE and between abundances of PGE and base metals are entirely different in the alkaline and tholeiitic suites. The PGE-poor nature of the alkali basalts from the Haleakala and West Maui volcanoes is not readily explained by fractional crystallization of basic magma. There is very little evidence that the alkaline magmas were subject to scavenging of PGE by sulfide-rich liquids or that PGE were preferentially partitioned into ferromagnesian silicate or oxide phases such as chromite. Partial melting of mantle source-rocks would lead to slightly lower PGE contents in alkali melts by virtue of the lower degrees of partial melting associated with the generation of alkali melts. However, this effect is very small and not likely to account for the large differences in PGE abundances in alkaline and tholeiitic melts. As most models of melt generation in the Hawaiian plume require at least two lithological components of differing isotopic composition and abundances of incompatible trace elements, and as mixing of melts derived from these different components is required to generate the Maui alkaline suites, it is possible that the low-PGE alkaline melts result from mixing of larger proportions of melts from a PGE-poor source-rock such as MORB. Although acceptable as a low-PGE source, MORB is not a satisfactory plume component according to some recent models, which argues that recycled oceanic lithosphere is the only component required within the Hawaiian plume to account for the isotopic variations observed in Hawaiian lavas, and further that ophiolite sequences provide analogues of such lithosphere. A review of PGE abundances in ophiolites indicates that significant volumes of both low- and high-PGE lithologies are present in ophiolite sequences, which leads to the conclusion that from a PGE perspective, the concept that only recycled oceanic lithosphere is required in the Hawaiian plume is satisfactory. Hence, the lower PGE content of Maui alkaline lavas is considered to reflect a low degree of partial melting with a significant contribution from low-PGE lithologies in the source region.

Keywords: platinum-group elements, alkali basalt, Maui, Haleakala, Hawai'i, ophiolite.

SOMMAIRE

Les éléments du groupe du platine (EGP) sont jusqu'à huit fois moins abondants dans les basaltes alcalins provenant de Maui, à Hawai'i, que dans les basaltes tholéiitiques de Kīlauea. Dans les suites alcalines et tholéiitiques, les corrélations parmi les différents EGP, et entre l'abondance des EGP et des métaux de base, sont entièrement différents. La présence de basaltes alcalins appauvris en EGP dans les volcans Haleakala et Maui Ouest ne pourrait s'expliquer par cristallisation fractionnée d'un seul magma basique. De plus, il y a très peu d'évidence que les magmas alcalins soient sujets à un appauvrissement en EGP dû à une répartition dans un liquide sulfuré ou dans un minéral silicaté ferromagnésien ou un oxyde, tel la chromite. La fusion partielle d'une source dans le manteau pourrait bien expliquer les niveaux légèrement inférieurs des EGP dans les basaltes alcalins à cause du taux plus faible de fusion partielle responsable pour la génération des bains fondus alcalins. Toutefois, l'effet est secondaire, et ne saurait expliquer les différences importantes en abondances des EGP dans les suites alcalines et tholéiitiques. La plupart des modèles proposés pour expliquer la génération de magmas associés au panache hawaiien font appel à au moins deux composantes lithologiques différant en composition isotopique et en teneurs en éléments incompatibles. De plus, ces modèles impliquent un mélange de magmas dérivés de ces différentes composantes pour donner les suites alcalines de Maui. Il est donc possible que les magmas alcalins à faible teneur en EGP résultent d'un mélange d'une plus grande proportion de liquide provenent d'une source à faible teneur en EGP, par exemple une suite basaltique typique des rides médio-océanique (MORB). Quoiqu'un tel matériau soit acceptable comme source à faible teneur en EGP, les suites MORB ne sont pas à considérer comme une composante réaliste d'un panache, selon certains modèles récents. Ceux-ci proposent plutôt une lithosphère océanique recyclée comme seule composante requise pour expliquer les variations isotopiques observées dans les laves associées au panache hawaiien. De plus, il semble que les séquences ophiolitiques fournissent un analogue de telles séquences recyclées. Une évaluation des abondances des EGP dans les séquences ophiolitiques indique que des volumes importants de roches soit enrichies, soit appauvries en EGP sont présents; du point de vue des EGP, donc, l'idée que les suites océaniques recyclées expliqueraient la composition du panache

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hawaiien est attirante. Les teneurs faibles en EGP des laves alcalines de Maui résulteraient donc d'un faible taux de fusion partielle, avec une contribution importante de types de roches à faible teneur en EGP à la source.

(Traduit par la Rédaction)

Mots-clés: éléments du groupe du platine, basalte alcalin, Maui, Haleakala, Hawai'i, ophiolite.

INTRODUCTION

Abundances of platinum-group elements (PGE) have been determined in Hawaiian basaltic rocks and their associated volcanic gases and pyroclastic ejecta (Zoller *et al.* 1983, Finnegan *et al.* 1990, Crocket 2000). Most data pertain to tholeiitic basalts of the shield-building stage, and thus represent some 95 to 98% of Hawaiian lavas (Clague 1987). The terminal stage of Hawaiian volcanism typically includes post-shield eruptions whose lavas fill and spill over summit calderas and post-erosional lavas, which overlay erosional surfaces cut into these rocks (Clague 1987). This post-shield volcanism generates predominantly alkali basalts. Here, a comparison of abundances of PGE in post-shield alkali basalts and the main tholeiitic basalts of the shield-building stage is presented.

BACKGROUND

The most extensive PGE dataset for Hawaiian tholeiites pertains to the historic basalts of Kīlauea volcano. These rocks offer distinct advantages for trace-element studies. The volcanism is contemporary, so that very fresh, unaltered samples can be obtained. More significantly, the extensive geophysical, geochemical and petrological research on Kīlauea (Decker *et al.*1987) has yielded a relatively detailed understanding of the Kīlauea subsurface and processes of magma generation and crystallization. However, Kīlauea has not reached the post-shield eruptive stage, so that the terminal stage of PGE fractionation cannot be evaluated. The youngest major post-shield alkali volcanism exhibited by a Hawaiian volcano is represented by rocks of the Kula and Hana formations of the East Maui volcano, Haleakala.

In this study, data are presented on abundances of PGE (Os, Ir, Pt and Pd) and Au, along with data on Cu, Co, Cr, Ni and MgO for eight samples from Haleakala and four from the West Maui volcano, which also includes rocks from the alkali stage. The Haleakala data are compared with those for Kīlauea (Crocket 2000), and averages for oceanic basalts associated with the Reunion hot-spot trace (Greenough & Fryer 1990) and mid-ocean ridge basalts (MORB), as compiled from the literature. The principal goals are to assess the geochemical behavior of the PGE in the terminal stages of Hawaiian volcanism and the linkages between PGE in the main shield-building and terminal post-shield stages.

The study of PGE evolution by comparison of Kīlauea and Haleakala samples has the advantage of using rocks substantially free of postmagmatic subaerial alteration due to weathering. In addition, Kīlauea (Decker *et al.* 1987) and Haleakala (Macdonald 1978, Moore *et al.* 1980, Chen & Frey 1985, West & Leeman 1987, Chen *et al.* 1991) are two of the most thoroughly studied Hawaiian volcanoes. However, differences in isotopic and chemical composition of lavas from different shields at comparable stages of evolution (Lassiter *et al.* 1996, Stille *et al.* 1986) and in inferred constituents of the Hawaiian plume (Wyllie 1988) are well recognized and could potentially influence magmatic evolution. Even so, both Kīlauea and Haleakala are on the Kea trend volcanic lineament, and much of the isotopic and geochemical diversity of Hawaiian rocks seems to depend on the location of a volcano on either of the Kea or Loa lineaments. Lassiter *et al.* (1996) proposed that the Kea and Loa lineaments are generated by differences in magmatism originating near the center of the Hawaiian plume (Loa trend) or near the plume margin (Kea trend).

The principal finding of this study is that the alkali basalts of Maui (mainly Haleakala) are ~5 to 8 times lower in PGE–Ni–Cu than the tholeiites of Kīlauea. Using constraints on the evolution of Haleakala magmas from the work of Chen & Frey (1985) and Chen *et al.* (1991) and recent models of magma generation in Hawai'i, in particular that of Lassiter & Hauri (1998), possible explanations for the low contents of the PGE in the alkali lavas of Haleakala are presented.

SAMPLE DESCRIPTIONS

Samples were collected from the East and West volcanoes of Maui; sample locations are shown in Figure 1 on maps taken from Easton & Easton (1983). The East Maui or Haleakala suite includes specimens from the Honomanu (Ho 1, Ho 2), Kula (Ku 1 to Ku 3) and Hana (Ha 1 to Ha 3) formations. The two Honomanu samples were taken from outcrops along the Hana Highway (route 36, see Easton & Easton, 1983, Fig. 30). The Kula and Hana samples were collected using Macdonald's map I–1088 (Macdonald 1978), which focuses on the crater section of Haleakala National Park. The Kula rocks are from roadcuts along Highway 378 leading up to the Haleakala summit. Ku 1 and Ku 2 consist of hawaiite (unit ka), and Ku 3 consists of ankaramite (unit kpa, Map I–1088). Samples of the Hana Formation were collected from within Haleakala Crater near the south-

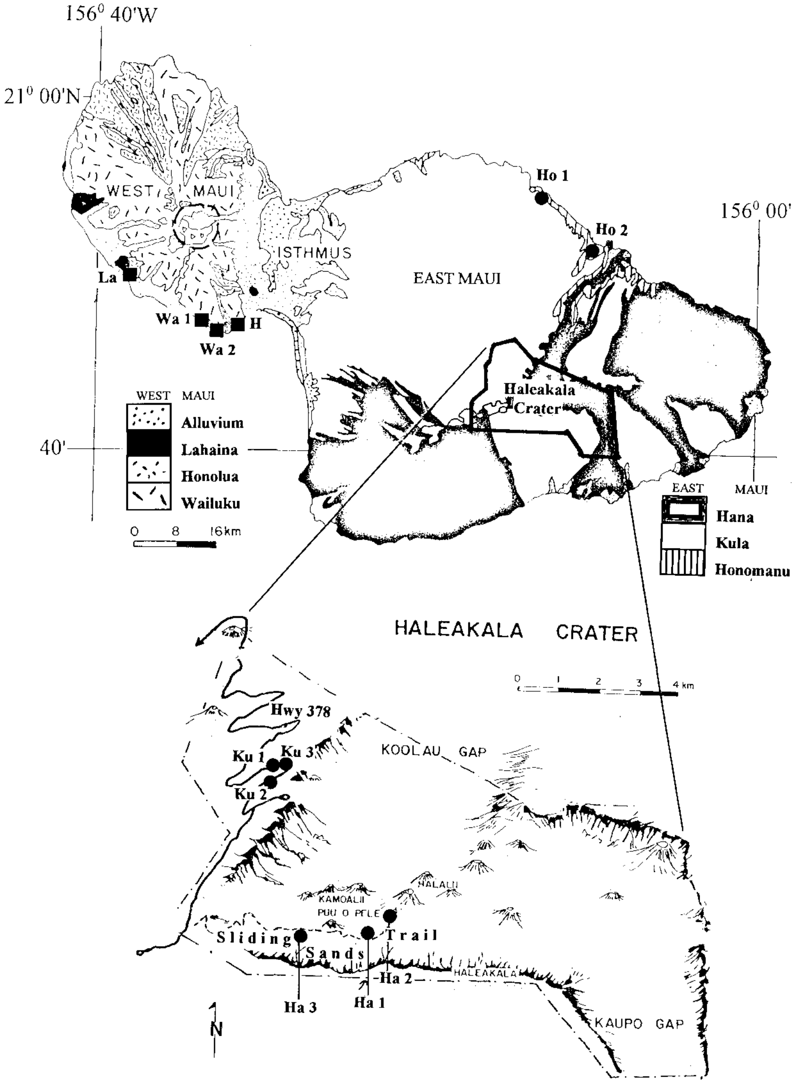

FIG. 1. Sample location map (adapted from Easton & Easton 1983).

ern wall and along the Sliding Sands Trail. These rocks are alkali olivine basalt flows (units hb and hk).

Samples of the West Maui volcano were collected along Highway 30 from the road log of Easton & Easton (1983). The suite includes two samples of the shield-building Wailuku Formation (Wa 1 and Wa 2), one sample of the post-caldera Honolua Formation (H) of the alkali stage, and one sample of the post-erosional Lahaina Formation (La). Sample WA 1 is a tholeiitic olivine basalt, and Wa 2 is a plagioclase-phyric basalt at the top of the Wailuku Formation. The Honolua Formation sample is an extremely alkali-rich trachyte, and the Lahaina Formation sample is a basanite.

Brief notes on the petrographic character of samples are presented in Appendix 1.Their classification in terms of the total alkali *versus* silica diagram of Le Maitre *et al.* (1989) is shown in Figure 2.

ANALYTICAL METHODS

Platinum-group elements

With the exception of Pt, the concentrations of the PGE were determined by radiochemical neutron-activation analysis (RNAA) as described by Crocket (2000), using the McMaster Nuclear Reactor as the source for

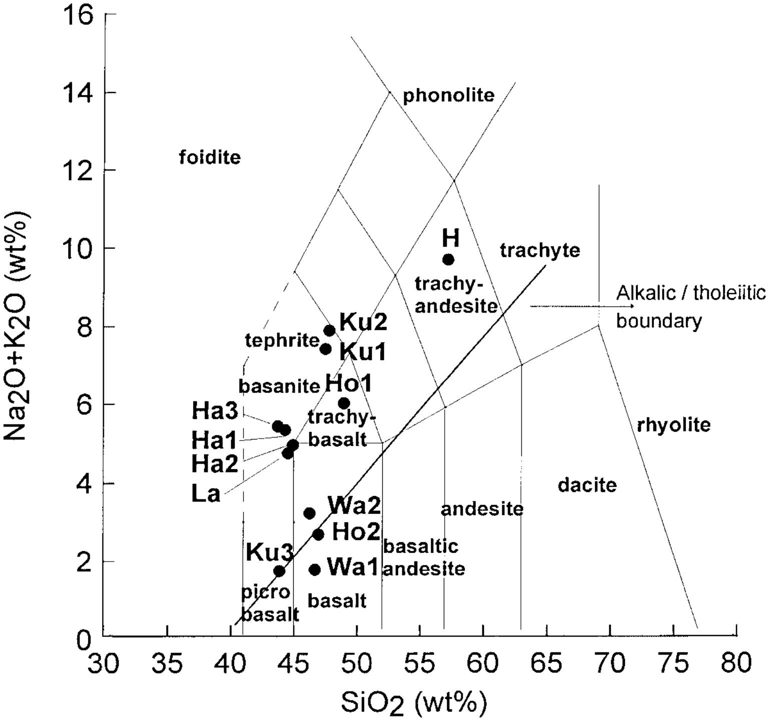

FIG. 2. Classification of Maui rocks according to SiO_2 *versus* total alkali content, with field boundaries after Le Maitre *et al.* (1989).

neutron irradiation. The average weight of sample analyzed was 200 mg of rock powder taken from a total mass of ~100 g, which had been ground to pass 100 mesh nylon sieving cloth. The analyses for Pt were done by Activation Laboratories Ltd., Ancaster, Ontario, using a fire assay on ~30 g samples of <200 mesh (74 μm) powder, followed by analysis by inductively coupled plasma – mass spectrometry (ICP–MS). The procedure has a minimum limit of detection of 0.1 ppb.

Major, minor and trace elements

The ten major elements were determined on ~5 g samples using a lithium metaborate–tetraborate fusion followed by ICP–MS analysis. The minimum detectable concentration for each element (expressed as an oxide) was 0.01 wt.%. In addition, a loss on ignition (LOI) was determined by heating samples to 800°C for 24 hrs.

A suite of minor and trace elements was determined to permit comparison of the alkali and tholeiitic basalts of Haleakala and Kīlauea, respectively, and to allow for more detailed characterization of PGE abundance profiles. Concentrations of these elements, also established by Activation Laboratories Ltd., were determined by: 1) instrumental neutron-activation analysis (INAA), 2) total acid (HF, $HClO_4$,HNO_3,HCl) digestion followed by ICP–MS, and 3) borate fusion – ICP–MS, along with the major elements. Those determined by INAA include As, Ba, Co, Cr, Hf, Mo, Rb, Sc, Ta, Th, U and the rare-earth elements (*REE*) La, Ce, Nd, Sm, Eu, Tb, Yb and Lu. The minimum detection-limits for these elements are 0.01–0.1 ppm for Co, Sc, Th, and U, >0.1–1 ppm for As, Cr, Cs, Hf, and Ta, and >1 to 10 ppm for Mo and Rb. The *REE* minimum limits of detection are 1 ppm for Ce and Nd, 0.05 ppm for La, Eu and Yb, and 0.01 ppm for Sm, Tb and Lu. Elements determined by total acid digestion followed by ICP–MS are Cu, Zn and Ni, all with a minimum detection-limit of 1 ppm. Those determined by borate fusion ICP–MS are Ba, Sr, Y and Zr, with minimum detection-limits of 1, 2, 1 and 4 ppm, respectively.

PETROGRAPHIC AND GEOCHEMICAL PROPERTIES OF MAUI BASALTS

Classification of Maui rocks: silica versus total alkalis

Figure 2 is a plot of the Maui suite in the SiO_2 *versus* (Na_2O + K_2O), in wt.%, with fields as defined by Le

Maitre *et al.* (1989) and includes the "tholeiitic *versus* alkaline" boundary of Macdonald (1968). In this plot, all the Maui rocks are classified as basalt except for West Maui sample H, which is classified as trachyandesite. Three samples, Wa 1, Ho 2 and Ku 3, are identified as being tholeiitic. Wa 1 and Ho 2 are from the Wailuku and Honomanu formations, and represent the main shield-building volcanism of the West and East Maui volcanoes, respectively. Both contain significant normative orthopyroxene. Ku 3, a post-caldera stage rock of the Kula Formation, with abundant pyroxene and olivine phenocrysts, is classified as picrobasalt. It could be termed ankaramite according to criteria noted by Macdonald (1978). This rock, as well as Ho 2, lies virtually on the alkali *versus* tholeiite field boundary. Three tholeiitic samples, Wa 1, Ho 2 and Ku 3, contain much more Ni, Cr and MgO than the alkaline rocks, consistent with their tholeiitic character. The major-element compositions of the Maui rocks are presented in Appendix 2. The data represent anhydrous compositions, having been normalized to zero LOI. The major-element data are supplemented by data for a suite of trace elements used to compare patterns of incompatible trace elements in Maui and Kīlauea rocks (Fig. 3).

PGE and selected trace elements in Maui rocks

Concentrations of PGE, transition metals (Co, Cr, Cu, Ni) and MgO in basalts from the Haleakala and West Maui volcanoes are presented in Table 1, along with an average for the alkaline rocks of Haleakala (samples Ku 1, Ku 2, Ha 1, Ha 2, Ha 3 and Ho 1). The average for the Maui alkaline rocks is compared with other basalts from oceanic settings including Kīlauea (Crocket 2000; except Pt, Appendix 3, this study), oceanic volcanoes associated with the Reunion hot-spot trace (Greenough & Fryer 1990) and MORB (literature survey, see footnotes to Table 2). The comparison is made in terms of mantle-normalized ratios (Fig. 4, Table 2) for PGE, Co, Ni and Cu plotted against individual elements ordered in the sequence Co, Ni, Os, Ir, Ru, Rh, Pt, Pd, Au, Cu (Barnes *et al.* 1988, Barnes & Maier 1999). Figure 4 shows that much lower PGE, Ni and Cu concentrations prevail in the Haleakala alkaline suite than in the tholeiitic Kīlauea suite. Ignoring Os for the moment, the Haleakala rocks are five to eight times lower in PGE–Ni–Cu than the Kīlauea samples on average. MORB is the most strongly PGE-depleted type of oceanic basalt. The Reunion rocks are characterized

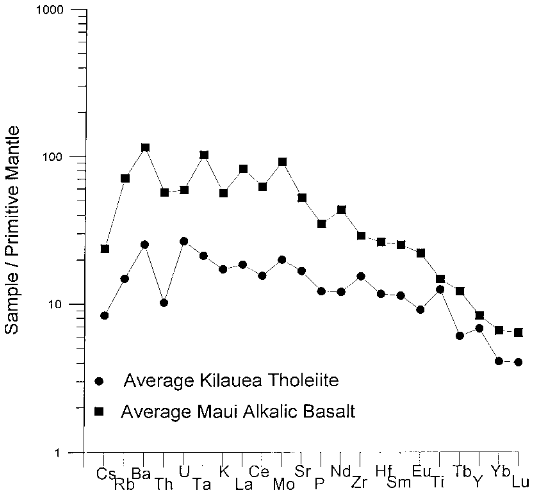

FIG. 3. Comparison of mantle-normalized abundances of incompatible trace elements in Maui alkaline basalts and Kīlauea tholeiites.

TABLE 1. CONCENTRATIONS OF THE PGE, SELECTED TRACE ELEMENTS AND MgO IN BASALTIC ROCKS FROM MAUI

Sample	Os	Ir	Pt	Pd	Au	Pd/Ir	Co	Cr	Cu	Ni	MgO
East Maui Volcano (Haleakala)											
Hana Formation											
Ha 1	0.13	0.106	24	0.30	0.5* 12	3	39	57	23	25	5.58
Ha 2	-	0.027	1.3	0.67	0.37	25	42	150	33	59	6.41
Ha 3	-	0.022	0.4	0.23	0.16	10	44	43	19	33	5.88
Kula Formation											
Ku 1	0.10	0.027	0.2	0.54	0.31	20	16	24	2	10	2.97
Ku 2	0.062	0.051	0.7	0.65	0.38	13	20	34	6	29	2.93
Ku 3	0.20	0.069	0.4	0.29	0.22	4	77	1050	32	375	16.20
Honomanu Formation											
Ho 1	0.16	-	0.2	0.42	0.40	-	21	17	2	6	4.44
Ho 2	0.069	0.29	0.3	<0.1	-	0.3	66	485	71	489	11.75
West Maui Volcano											
Lahaina Formation											
La	0.033	-	207	1.11	1.48	-	57	560	70	338	11.48
Honolua Formation											
H	-	0.365	1.7	0.82	0.80	2	3.3	12	2	7	1.83
Wailuku Formation											
Wa 1	0.049	0.026	0.80	2.18	1.43	84	66	749	77	536	14.32
Wa 2	0.089	0.078	0.20	0.33	1.60	4	47	148	69	100	6.18
Average Haleakala alkali basalts (excludes tholeiites Ku 3 and Ho 2)											
	0.11	0.047	0.56§	0.47	0.35§	14	30	54	14	27	4.7

Concentrations of the PGE are expressed in ppb; those of selected trace elements are given in ppm, and the proportion of MgO is quoted in wt.%.

* These two values of gold concentration were obtained on two different aliquots of Ha 1 powder and are considered to reflect significant heterogeneity of the sample in masses of ~250 mg.

§ The Pt and Au averages exclude the value of 24 ppb Pt and 12 ppb Au for Ha 1. The rocks sampled are mainly alkali basalts, except Ku3, Ho2, Wa1 and Wa 2, which are tholeiites. -: Not available.

TABLE 2. PGE AND TRANSITION METAL VALUES FOR PRIMITIVE MANTLE-NORMALIZED PLOTS

	Primitive[1] Mantle	Haleakala[2] Alkaline Rocks	Kīlauea[3]	MORB[4]	Reunion[5] Hot Spot Trace
Co ppm	105	30	47	50	-
Ni	1960	27	186	150	98
Os ppb	3.4	0.11	0.38	0.0034	-
Ir	3.2	0.047	0.38	0.032	0.11
Ru	5	-	-	0.0337	0.22
Rh	0.9	-	-	0.002	0.31
Pt	7.1	0.56	3.6	0.28	7.3
Pd	3.9	0.47	2.4	0.26	8.1
Au	1	0.35	1.78	1.2	3.2
Cu ppm	30	14	125	93	147

1. McDonough & Sun (1995). 2. Table 1, this work. 3. Crocket (2000), except Pt, for which see Appendix 3. 4. Data for Co and Ni are from Peach *et al.* (1990). The concentrations of Os, Ir and Pd are averages from literature compilations. The main sources of the data are: Os: Hertogen *et al.* (1980), Martin (1991); Ir: Gottfried & Greenland (1972), Laul *et al.* (1972), Keays & Scott (1976), Crocket & Teruta (1977), Hertogen *et al.* (1980), Plessen & Erzinger (1997) and Rehkämper *et al.* (1999); Pd: Crocket & Teruta (1977), Hertogen *et al.* (1980), Plessen & Erzinger (1997) and Rehkämper *et al.* (1999). Ru and Pt data are from Rehkämper *et al.* (1999). Data for Au are from Crocket (1991), and Cu is an average of values cited by Mathez (1999) and Rehkämper *et al.* (1999). 5. Greenough & Fryer (1990). Averages for Ni and Cu calculated from their data using the same sample groupings employed by them to obtain averages for the PGE.

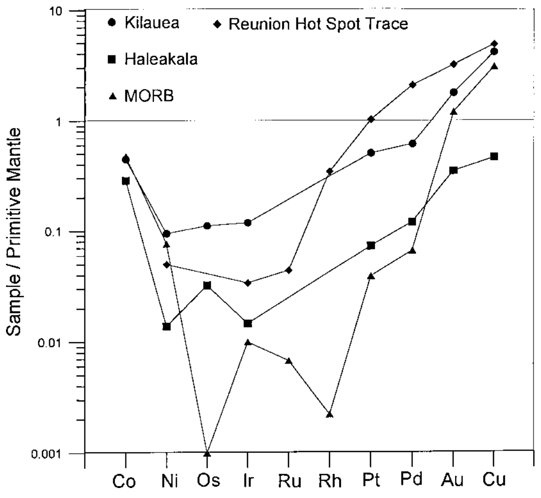

FIG. 4. Mantle-normalized PGE and transition metal ratios for ocean-basin basalts. Mantle values are those for primitive mantle, from McDonough & Sun (1995).

by moderate depletion and little fractionation in the Ni to Ru region, but increasing fractionation and enrichment relative to a mantle reservoir from Pt to Cu.

The detailed character of the mantle-normalized curves is somewhat constrained by analytical considerations. Analytical error associated with the Os average for Haleakala rocks is probably significant because Os abundances in these rocks are close (× 2) to the minimum level of detection for the analytical procedure (RNAA) used. Thus, the apparent positive anomaly of Os (relative to Ni and Ir) is not regarded as significant. Further, the very low contents of PGE in most MORB rocks cast some uncertainty on the MORB profile, especially in the Os-to-Rh region. The Os data for MORB, although very limited, were obtained by isotope dilution – thermal ionization mass spectrometry, a method with good sensitivity for Os down to the ppt range. The Ir data are primarily RNAA, taken from the literature, with about 25% of the values as upper limits of concentration only. Hence, the nearly order-of-magnitude difference in mantle-normalized Os and Ir in MORB may reflect significant analytical uncertainty in the Ir average. Nevertheless, the general character of the profiles of mantle-normalized oceanic basalt (positive slope with higher gradient in the Pt-to-Cu region relative to the Ni-to-Ir region) is a distinctive property of basalts from oceanic tectonic settings.

TABLE 3. COMPARISON OF CORRELATIONS FOR MAUI ALKALI OLIVINE BASALTS AND KĪLAUEA THOLEIITIC BASALTS

	Ir	Pt	Pd	Au	Co	Cr	Cu	Ni	MgO
				Maui Correlations					
Os	0.63	0.26	-0.81	0.27	0.32	-0.13	0.12	-0.64	0.72
Ir		0.07	0.54	0.07	-0.68	-0.42	-0.42	-0.56	-0.59
Pt			-0.41	**0.999**	0.34	0.10	0.39	0.04	0.31
Pd				-0.42	-0.66	0.08	-0.27	-0.02	-0.62
Au					0.32	0.07	0.36	0.00	0.3
Co						0.67	**0.88**	0.76	**0.96**
Cr							**0.89**	**0.92**	0.71
Cu								**0.89**	**0.87**
Ni									0.71
				Kīlauea Correlations					
Os	0.62	0.44	**0.66**	0.45	**0.48**	**0.88**	-0.61	**0.73**	**0.58**
Ir		**0.58**	**0.70**	0.53	**0.61**	**0.79**	-0.37	**0.61**	**0.72**
Pt			0.20	-0.60	0.76	0.44	-0.39	0.50	0.45
Pd				**0.64**	0.33	**0.68**	-0.49	0.49	**0.61**
Au					-0.01	0.45	-0.24	0.08	0.14
Co						**0.68**	-0.03	**0.83**	**0.75**
Cr							-0.57	**0.83**	**0.80**
Cu								-0.39	-0.30
Ni									**0.85**

Coefficients shown in bold-face font are significant at a 1% level or better.

Comparison of correlations in Maui and Kīlauea basalts

PGE and transition metal correlations (Pearson's r) for alkaline rocks from Maui (Ku 1, 2, Ha 1, 2, 3, Ho 1 and H) and Kīlauea tholeiites are presented in Table 3. The Maui correlations are based on seven samples, six of which are from Haleakala. Maui tholeiites are excluded, as are two alkaline samples, Wa 2 and La. Although sample La (West Maui, Laihana Formation) falls within the alkaline field in Figure 2, a relatively high MgO content (11.5%) and abundant clinopyroxene and olivine suggest that the rock does not represent the composition of a primary magma. Sample Wa 2 is at the stratigraphic top of the shield-building Wailuku Formation, a dominantly tholeiitic unit, and just within the alkaline field. Its exclusion is based on occurrence within the Wailuku Formation.

The correlation matrix for the Kīlauea data, taken from Crocket (2000), is based on 18 samples, except for Pt, where r values represent seven samples. Pt concentrations are listed in Appendix 3. As the two matrices represent suites of different numbers of samples, comparisons are made in terms of r values at the same level of significance. Thus, r values with significance levels of 1% or better, as tested against Student's t distribution, are the basis for comparison of the two suites. These data are shown in a boldface font.

The correlations for the alkali basalts from Maui differ from those of the tholeiitic Kīlauea suite with respect to both inter-PGE and PGE – transition metal correlations. For correlations among the PGE only, the Maui alkaline suite has just one significant correlation, an extremely high positive correlation between Pt and Au, with r = 0.99. In contrast, several significant positive inter-PGE correlations are found in the Kīlauea suite, and the correlation between Pt and Au concentrations is weak and negative. Correlations between PGE and transition metals also are different. There are no significant correlations between PGE and Co, Cr, Ni, Cu and MgO in the Maui alkaline suite, whereas the Kīlauea matrix shows strongly positive and significant correlations of Os, Ir and Pd with Cr, Ni and MgO. Further, correlations between Cu and other elements contrast sharply in the two sets. In the Kīlauea suite, Cu shows weak negative correlations with all PGE except Os, where the negative correlation is strong. In the Maui suite, there are very strong and significant correlations of Cu concentrations with some transition metals (Co, Cr and Ni) and MgO, and no significant correlation of Cu with PGE concentrations. The strongest positive correlations of Cu are with Pt and Au, and the strongest negative correlation is with Ir. In summary, the strong correlations of PGE such as Os and Ir with elements associated with the earliest phases to precipitate from the basaltic magma (*e.g.,* chromite) seen in Kīlauea tholeiites are not apparent in alkali basalts from Maui.

Comparison of incompatible trace elements in Maui and Kīlauea basalts

A comparison of data on incompatible trace elements in alkaline rocks from Maui (all Haleakala samples except Ku 3 and Ho 2 and including West Maui sample H) and Kīlauea tholeiites from Crocket (2000) is presented in Figure 3 in mantle-normalized format (McDonough & Sun 1995). The Maui alkaline suite is characterized by higher concentrations of incompatible elements over the entire suite, the larger differences being characteristic of the more incompatible elements. Among factors that could cause this difference, a lower degree of partial melting in the region of magma generation is probably very significant (Chen *et al.* 1991).

DISCUSSION

The main contrast in PGE–Ni–Cu abundances between the post-shield alkaline rocks of Haleakala and the shield-building tholeiites of Kīlauea is the much lower concentration of these elements in the Haleakala suite. With the exception of Os, where analytical uncertainty is significant, mantle-normalized PGE–Ni–Cu concentrations in the Haleakala rocks are, on average, some five to eight times lower than in the Kīlauea suite (Fig. 4). These differences could arise during magma generation or upon subsequent cooling and fractional crystallization of magma, or some combination of both processes. Magma generation in the Hawaiian plume can be constrained by petrological considerations, isotopic data and incompatible-trace-element abundances which, to a first approximation, require sources dominated by two isotopically (Sr, Nd and Pb) distinct components, one depleted or MORB-like, with $^{87}Sr/^{86}Sr \approx 0.7035$, and the other enriched, with $^{87}Sr/^{86}Sr \approx 0.7043$ (Lassiter & Hauri 1998). The shield-building basalts of the Honomanu Formation of Haleakala volcano are tholeiites except for the upper 200 m, where the stratigraphy consists of interbedded alkali and tholeiitic basalts. Chen & Frey (1985) and Chen *et al.* (1991) showed that basalts in these uppermost Honomanu rocks are isotopically indistinguishable with respect to Sr. In contrast, the overlying alkali basalts of the post-shield Kula and Hana formations are characterized by lower and stratigraphically controlled Sr-isotope ratios. These authors interpreted the Sr isotopic systematics to imply that the terminal upper 200 m of the shield-building Honomanu basalts are the products of variable degrees of partial melting of a homogeneous source-rock. The post-shield alkali basalts, by contrast, were considered the products of mixing of magma from at least two sources, with the proportion of isotopically enriched and depleted Sr changing with time. Further, Chen *et al.* (1991) showed that Haleakala lavas were generated from source rocks in which garnet was stable, and that the degree of partial melting required for the alkaline rocks was lower by a factor of about two relative to that required for generation of the tholeiitic magmas. Hence, the PGE contents of end-member components of Hawaiian plume and the extent of melting of PGE-bearing phases as a function of percentage of partial melting of source rocks are possible factors that may control the PGE content of derivative melts.

Significance of partial melting for concentrations of PGE in melts

The degree of partial melting will also be relevant to the melting behavior of the principal phases concentrating PGE in the source rocks. These phases include PGE alloys (Os–Ir, Os–Ir–Ru, Pt–Fe) and sulfides as high-efficiency but low-modal-proportion components, along with chromian spinel and perhaps olivine as moderate-to low-efficiency but high-volume carriers [see Barnes & Maier (1999), for a brief review, and also Keays *et al.* (1981) for data on lherzolites from Australia, Legendre & Augé (1986) for a review, and Corrivaux & Laflamme (1990) for data on Thetford Mines, Quebec]. Peach *et al.* (1990) developed a batch partial melting model that requires 20 to 23% melting of bulk upper mantle to generate silicate and sulfide melts with PGE contents similar to those of primitive MORB glasses and their associated sulfide globules. They noted that this model does not work well for Ir, and concluded that the behavior of Ir is controlled by phases other than sulfide. Keays (1995) calculated that a similar degree of partial melting, 25%, would dissolve all mantle sulfide. Depending on how sulfide melting is related to percentage partial melting of a source rock, differing degrees of partial melting could generate melts of differing PGE concentration. The melting of Pd and Ir sulfides as a function of percent partial melting of whole-rock mantle was modeled by Barnes & Maier (1999). Their results are reproduced here in Figure 5 (see caption for model parameters). This model indicates that the degree of partial melting could significantly affect the Pd, but not the Ir content, of derivative melts.

The estimates of Chen *et al.* (1991) for the degree of partial melting required to generate the alkaline and tholeiitic lavas of the upper 200 m of the Honomanu Formation are 6.5 to 8 and 10 to 15%, respectively, with much smaller percentages applicable to generation of the alkaline basalts of the Kula and Hana formations. The roughly linear increase of Ir *versus* the degree of partial melting (Fig. 5) will generate an increase in Ir of ~1.5 times over a range of 1 to 15% partial melting (upper limit for Haleakala tholeiites). Although the trend is toward increasing Ir with increasing degree of partial melting, the magnitude is much less than the factor of eight difference observed in average Ir content of the Haleakala alkali basalts (Table 1) and Kīlauea tholeiites (Crockett 2000). The case of Pd, however, is somewhat different in that the Pd content of melts begins to increase rapidly when partial melting exceeds ~10%, and will increase ~3 times as partial melting increases from 1 to 15%. Nevertheless, this increase is still much less than the factor of five difference in average Pd in alkali and tholeiitic basalts. Thus, an increasing degree of partial melting does increase both the Pd and Ir contents of mafic melts, with the effect more prominent for Pd. However, the magnitude of the increase for the degrees of partial melting appropriate to alkaline basaltic magmas as compared with tholeiitic magmas is substantially less than the observed differences in PGE between alkali and tholeiitic lavas. In summary, partial melting of sulfides should generate tholeiitic magmas with somewhat higher PGE contents than alkali basaltic magmas, but with differences much less than observed in the actual suites of rocks.

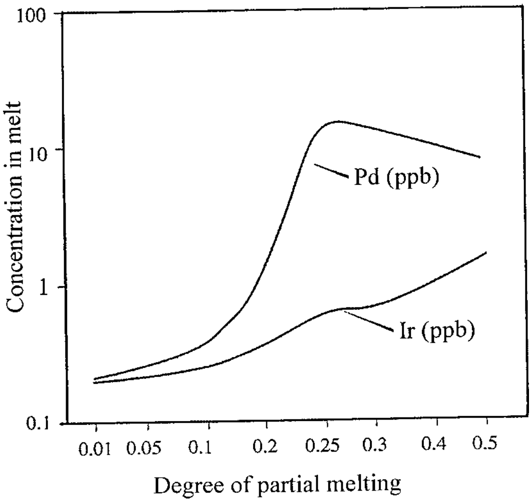

FIG. 5. Ir and Pd concentrations (ppb) in mafic melt as a function of degree of partial melting, after Barnes & Maier (1999), with slight modification. Model parameters are: sulfide/silicate partition coefficients, $D^{\mathrm{Sul/Sil}}(\mathrm{Ir}) = D^{\mathrm{Sul/Sil}}(\mathrm{Pd}) = 3 \times 10^4$; bulk partition-coefficient for Ir into the residuum = 10 for 1% partial melting, with a decrease to 5 at 25% partial melting.

The role of Os–Ir alloys as contributors to PGE in partial melts of mantle source-rocks is uncertain. Peach & Mathez (1996) showed experimentally that such alloys are unstable in natural silicate melts and concluded that they are not early-crystallizing phases. Bai *et al.* (2000) found that such alloys in chromitites from the Luobusa ophiolite are associated with a suite of minerals including diamond, silicon carbide and native silicon, which they interpreted as xenocrysts indicative of very high pressure and strongly reducing conditions. They suggested that the associated PGE alloys were xenocrysts derived from a deep mantle source and perhaps remnants of early stages of differentiation of the Earth, a scenario similar to a hypothesis on the origin PGE-bearing alloys put forward by Bird & Bassett (1980). Experimental partitioning of Os and Ir between solid metal and liquid metal sulfide yield $D^{\mathrm{Metal/Sulfide}}$ values that vary from approximately 2×10^2 to 6×10^3 at temperatures of 1000 to 1200°C (Fleet *et al.* 1999). These values imply minimal mobility of these metals should sulfide liquid be present during magma generation. Hence, the role of phases other than sulfides as contributors of PGE to mafic partial melts is uncertain. In summary, the limited available evidence suggests that partial melting to variable degrees will generate alkaline basaltic melts with lower PGE contents than tholeiitic melts, but with predicted differences substantially less than those observed in rock suites.

PGE loss during cooling of magma

Fractional crystallization accompanying magma cooling could potentially deplete melts in PGE owing to partition of these metals into immiscible sulfide liquid or chromian spinel or as inclusions of these phases in olivine. In ophiolites, Os, Ir and Ru abundances are directly related to modal proportion of chromite and Cr content, suggesting that these PGE are removed from the melt by fractional crystallization of chromite (Oshin & Crocket 1982, Talkington & Watkinson 1986). In Kīlauean tholeiites, positive and significant correlations of Os and Ir with Cr have been reported (Crocket 2000). However, such processes are unlikely to cause the large depletion of PGE seen in the alkaline rocks of Haleakala, as similar correlations are absent from the Maui suite.

Effective depletion of PGE by preferential partition into chromite or olivine (*i.e.*, $D^{\text{Crystal/Magma}} > 1$) should be enhanced if subsurface magma-holding chambers are components of conduit systems transporting magma to the surface. Clague (1987) noted that such storage zones are lacking in conduit systems of the post-erosional stage, although present at subcrustal depths in post-caldera systems. Accordingly, Kula lava might be expected to exhibit greater depletion in PGE than Hana lava owing to crystallization of chromite or olivine in subcrustal magma-chambers. Although only three samples are available, neither average Ir or Pd is significantly different in the Kula and Hana rocks (Table 1).

Sulfide saturation is the most efficient mechanism to remove PGE from basic melts (Naldrett 1989). Rare sulfides have been noted in Kīlauea rocks (Helz 1987), although modal abundances are probably much less than 0.01% (Stone & Fleet 1991). As discussed by Barnes *et al.* (1988) and Barnes & Maier (1999), scavenging by an immiscible sulfide will impart a distinctive trace-element signature if the PGE are compared with Cu and Ni on a mantle-normalized basis. Owing to the more efficient partition of PGE into a sulfide liquid [$D^{\text{Sul/Sil}}(\text{PGE}) \approx 1 \text{ to } 5 \times 10^4$: Fleet *et al.* 1999] relative to Cu and Ni [$D^{\text{Sul/Sil}}(\text{Cu}) \approx 10^3$ and $D^{\text{Sul/Sil}}(\text{Ni}) \approx 3 \times 10^2$: Francis 1990], magmas subject to sulfide saturation will be more strongly depleted in PGE. Thus, in a mantle-normalized plot such as Figure 4, the sample-to-mantle ratio will be greater for Ni than that of Ir and Os, and the ratio for Cu will be greater than that for Pd. Both Haleakala and Kīlauea rocks have similar mantle-normalized ratios in the Ni to Ir region, whereas mantle-normalized values of the Cu:Pd ratio are ~4 and ~7 in Haleakala and Kīlauea lavas, respectively. The Ni to Ir region then is not suggestive of sulfide scavenging as a cause of low PGE contents in the Haleakala rocks. Although Cu is greater than Pd on a mantle-normalized basis, the difference is much less than expected from sulfide immiscibility [~(25; $D(\text{Pd})/D(\text{Cu}) \approx 25000/1000$], and the effect is less for Haleakala than for Kīlauea rocks.

PGE behavior during magma generation

With minimal support for major PGE loss during crystallization of magma, the behavior of the PGE in the magma-generation process may be more relevant in understanding the low content of the PGE in alkaline lavas. Models for Hawaiian volcanism commonly invoke at least two end-member components, one representing an enriched and the other a depleted source, although the exact nature of such sources is uncertain. For example, Chen & Frey (1985) argued for an enriched plume-related source and a depleted MORB-related lithospheric source to explain the Sr isotope variation in the Haleakala alkaline units. The nature and origin of source rocks to Hawaiian volcanism were critically reviewed by Lassiter & Hauri (1998) in conjunction with their proposal that isotopic and chemical variations in Hawaiian rocks can be explained by melting and magma mixing within the Hawaiian plume and that reaction of the plume with enclosing lithosphere–asthenosphere is not a significant process in creating magma diversity. In their model, they recognized that both anomalously light and anomalously heavy oxygen isotope signatures are present in Hawaiian rocks, and that these oxygen isotope signatures correlate with osmium isotope compositions, such that high $\delta^{18}\text{O}$ is characteristic of rocks enriched in radiogenic Os, whereas low $\delta^{18}\text{O}$ correlates with unradiogenic Os. They proposed that ancient subducted oceanic lithosphere has been recycled into the Hawaiian plume. Such oceanic lithosphere, if previously subjected to hydrothermal alteration as observed in many ophiolites, will carry the requisite isotopic variations in oxygen and can generate the required Os isotopic compositions if sufficiently old (~10^9 yrs). To a first approximation, ophiolites represent the recycled lithosphere required in this model. From the PGE standpoint, models that include components of high and low contents of the PGE in the Hawaiian plume are advantageous in view of the large differences in PGE concentrations in alkaline and tholeiitic rocks.

PGE perspective on an ophiolite source model

As fractional crystallization and partial melting processes seem limited in their potential to generate lavas with large differences in PGE content (factors of 5 to 8), it is useful to consider whether these different levels of the PGE are inherited, in part, from the source rocks to plume magmatism. If this is the case, large variations in PGE abundances would be expected in the source rocks. As previously noted, at least two source-rock domains are commonly invoked to explain Hawaiian lavas, one a plume source (primitive mantle) and the other a depleted MORB-like source (*e.g.,* Chen *et al.* 1991). The question arises whether an ophiolite model as proposed by Lassiter & Hauri (1998), with recycling of oceanic lithosphere into the plume, is likely to result in lithological components of contrasting PGE content. With some limitations, literature data on ophiolites can be used to provide a qualitative assessment. The required data are PGE concentrations in the main rock-types of ophiolite complexes. There is an extensive PGE literature pertaining to the geochemistry and mineralogy of ophiolites, much of which is focused on chromite-rich facies. An example with partial PGE (Ir, Pt, Pd and Au) data on all major lithological facies is the ophiolite province of southern Quebec [see Laurent (1977) for geology, petrology and tectonic setting, Oshin & Crocket (1982, 1986b) for PGE geochemistry, Laurent & Hébert (1989) for geotectonic setting, Corrivaux & Laflamme (1990) for PGE mineralogy, and Tanguay *et al.* (1990) for additional PGE geochemistry on ultramafic cumulates].

Table 4 presents averages and ranges of PGE values from the Thetford Mines ophiolite complex from Oshin & Crocket (1982, 1986a), with estimates of the thickness represented by major units. The thickest unit in the Thetford Mines complex is tectonized harzburgite comprising some 75% of the total thickness of all ophiolite-related rocks. The PGE content of the harzburgite samples differs little from estimates of McDonough & Sun (1995) for primitive mantle. These rocks constitute the largest volume of the ophiolite complex; with their relatively high PGE concentrations (3 to 10 ppb), they are the main PGE reservoir of the Thetford Mines complex. The dunite and pyroxenite ultramafic cumulate units represent small-volume components that are somewhat higher in Pt and Pd and lower in Ir than the harzburgites. The thickest of the cumulate rocks are gabbros (~38% of the total cumulus plus volcanosedimentary units). These rocks show major depletions in Ir and Pd (160 and 9 times, respectively, relative to harzburgite), but little difference in Pt or Au. The volcanic rocks capping the sequence consist of lower and upper members separated by ophiolitic sediments (Oshin & Crocket 1986a). The lower member consists of intercalated flows strongly depleted in Ir but highly variable in Pd content, whereas the upper member shows moderate depletion in Ir (~10 times) and concentrations of the other PGE that are slightly higher than in the harzburgites.

A qualitative appraisal of these concentration–volume relations suggests that the cumulus gabbros could provide a "depleted" component (relative to harzburgite) within recycled oceanic crust – upper mantle, at least with respect to Ir and Pd. The lower volcanic units do consist of flows with very low Ir and Pd contents, but these rocks are intercalated with flows containing relatively high Pd values. Selective melting of the flows with very low levels of Ir and Pd seems extremely unlikely. A significant consideration in the assumption that obducted on-land ophiolites may model the trace-element geochemistry of recycled oceanic lithosphere is whether dehydration and melt extraction in the subduction process will generate a loss or gain of the metals of interest [see Bennett *et al.* (1996) and Lassiter & Hauri (1998) with respect to Pb]. The Pd/Ir ratio in the Thetford Mines gabbros is 20 compared with 10 (Table 1) for Haleakala alkali basalts. Whereas the trace-element signature of a single component is not expected to persist in magmas derived by multiple-component mixing processes, the sense of change in the Pd/Ir ratio in response to high-temperature dehydration of rocks is probably toward lower values, reflecting preferential Pd loss to fluids. For the very small concentrations of these metals in the low-sulfur (<100 ppm) Thetford Mines gabbros, compatible behavior is probable for Ir (inclusion in ferromagnesian minerals) and incompatible behavior for Pd (interstitial or grain-boundary sites). In this case, partition of Pd into fluids generated by dehydration is probably greater than for Ir, so that subduction may modify Pd/Ir toward values closer to those found in the Haleakala rocks.

PGE in basalts of the West Maui volcano

The West Maui suite consist of two samples of shield-building Wailuku basalt and one sample each of the post-shield alkaline lavas of the Honolua and Lahaina formations. Wa 1, a tholeiite with MgO, Ni, Cr, Cu Pd and Au contents significantly higher than Haleakala alkaline rocks, generally is comparable to

TABLE 4. CONCENTRATION OF THE PGE AND Au IN THE THETFORD MINES OPHIOLITE COMPLEX, QUEBEC

Unit	Approx. thickness km	% total section	Number of samples	Ir ppb	Pt ppb	Pd ppb	Au ppb
Harzburgite	5	-	7	3.2	10	3.8	1.5
Cumulus units							
Dunite	0.13	8	13	2.4	28	21	1.3
Pyroxenite	0.15	9	12	0.2	17	29	1.4
Gabbros	0.6	38	6	0	6	0.4	1.3
Harz/Gabbro	-	-	-	160	1.6	9	1.2
Volcanic units							
Lower unit	0.23	14	20	0	4-11	0.42-10	2.4
Sediment	0.13	8	4	0	11	5	2.4
Upper unit	0.35	22	8	0.2	11	8.5	2.2
Primitive mantle	-	-	-	3.2	7.1	3.9	1

Notes: a.. Metal concentrations are from Oshin & Crocket (1982, 1986a, their Tables 2 and 3, respectively) except for the Pd average for gabbros, which is calculated from their Table 2 and includes gabbro and hypabyssal metagabbro, but excludes samples more than three standard deviations from the mean.

b. Approximate thicknesses of cumulus-textured units are estimated from Oshin & Crocket (1982, their Fig.2). Total thickness of cumulus and volcanosedimentary units is approximately 1600 m.

c. Values for the primitive mantle are from McDonough & Sun (1995).

Kīlauea tholeiites, but with lower Ir and Os contents. Wa 2, at the stratigraphic top of the Wailuku, as recognized by Macdonald & Katsura (1964), actually plots just within the alkaline field in Figure 2. The post-erosional Lahaina sample is a basanite with abundant coarse-grained olivine and augite and occasional nepheline in the matrix. It has high Cr, Ni and MgO, but its most distinctive compositional feature is a Pt content of 207 ppb. On the basis of primitive isotopic compositions, high contents of incompatible elements and the presence of garnet-bearing lherzolite xenoliths, Clague (1987) argued that post-erosional alkaline lavas are generated at depths of ~100 km and rapidly erupted to the surface without subcrustal storage of magma. If this scenario is valid, the coarse-grained ferromagnesian minerals may be xenocrysts, with the high Pt content inherited directly from the source region. It is of interest that one other rock from Maui, Ha 1, is anomalous in Pt (and Au), with 24 ppb Pt (Table 1); Ha 1 is also a post-erosional rock (Hana Formation).

CONCLUSIONS

Post-shield-building magmatism is characterized by very small volumes of alkaline rocks with much lower concentrations of the PGE than the main shield-building tholeiites, as represented by Kīlauea. Much isotopic and compositional evidence indicates that the change from tholeiitic to alkaline volcanism involves a change in the composition of source rocks. The low PGE content of Haleakala alkaline rocks is probably not due to fractional crystallization of magma after segregation of melt in that a) depletion due to an immiscible sulfide liquid is not supported by comparison of base metal (Ni and Cu) and PGE abundances in mantle-normalized profiles, b) a genetic linkage of tholeiitic and alkaline rocks through fractional crystallization is not compatible with isotopic evidence (*e.g.,* Sr), and c) subcrustal magma-storage chambers where fractional crystallization would be effective are lacking in the case of post-erosional volcanism.

Magma generation is probably the stage at which the low PGE content of the alkaline magmas is established. Either partial melting or the nature of source rocks in the Hawaiian plume are possible fundamental causes of PGE depletion. Although low degrees of partial melting would generate alkaline magmas with lower PGE contents than tholeiitic magmas, the differences would probably be small, and the PGE content of source rocks is likely the more important factor. Relevant conclusions with respect to the significance of partial melting include the following: a) the liberation of sulfide-hosted PGE during melting varies little with respect to Ir over likely ranges in degree of partial melting; b) whereas the Pd content of partial melts increases rapidly above about 10% partial melting, the Haleakala alkaline basaltic melts are probably generated at much lower degrees of melting.

Finally, the lithological components of the source region seem to require at least one component of very low PGE content. Recycled oceanic lithosphere is a significant component of plumes and, as represented by ophiolites, seems to satisfy the requirement of both high- and low-PGE components, as well as isotopic and other trace-element constraints elaborated by Lassiter & Hauri (1998).

ACKNOWLEDGEMENTS

The study was funded by a research grant from the Natural Sciences and Engineering Research Council of Canada. Constructive comments from two reviewers significantly improved the manuscript. In addition, the editorial expertise of Prof. Robert F. Martin is sincerely appreciated. This article is dedicated to Dr. Louis J. Cabri in view of his contributions to the field of platinum-group-element mineralogy. There is not much platinum-metal mineralogy in this paper, but if it were not for the fundamental contributions of Louis to this field, many of the interesting questions about PGE geochemistry could not even be asked.

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Received October 28, 2000, revised manuscript accepted September 14, 2001.

APPENDIX 1. SAMPLE DESCRIPTIONS

Sample	Location	Classification[1]	Texture[2]	Mineralogy[3]
East Maui Volcano (Haleakala)				
Hana Formation				
Ha 1	Haleakala Crater SST[4]	Basanite	A; vesicular	Microphenocrysts: Ol Groundmass: Pl, Ol, Cpx, Opq[5]
Ha 2	Haleakala Crater SST	Basanite	P	Phenocrysts: Ol Microphenocrysts: Ol, Cpx, Pl Groundmass: Pl, Opq, Ol, Cpx
Ha 3	Haleakala Crater SST	Basanite	P; fragmental aa	Phenocrysts: Ol Microphenocrysts: Ol Groundmass: Pl, Opq, Ol, Cpx
Kula Formation				
Ku 1	Hwy 378	Tephrite	A; fragmental aa; trachytic	Microphenocrysts: Pl, Opq Groundmass: Pl, Opq, Ol
Ku 2	Hwy 378	Tephrite	A; trachytic	Microphenocrysts: Pl, Opq, Ol Groundmass: Pl, Opq, Ol, Cpx
Ku 3	Hwy 378	Picritic basalt	P	Phenocrysts: Cpx, Ol Microphenocrysts: Cpx, Ol Groundmass: Ol, Cpx, Pl, Opq
Honomanu Formation				
Ho 1	Hwy 60	Trachybasalt	A; trachytic	Microphenocrysts: Pl, Opq, Ol Groundmass: Pl, Opq, Kfs, Ol, Cpx
Ho 2	Hwy 60	Basalt, tholeiitic	P; vesicular pahoehoe	Phenocrysts: Ol, Cpx Microphenocrysts: Ol, Pl Groundmass: Pl, Cpx, Opq
West Maui Volcano				
Lahaina Formation				
La	Stop 86[6] Puu Kilea	Basanite	P	Phenocrysts: Ol, Cpx Microphenocrysts: Ol, Cpx Groundmass: Pl, Opq, Ksp, Ol, Cpx
Honolua Formation				
H	Stop 3 McGregor Point	Trachyandesite	A	Microphenocrysts: Pl, Opq, Cpx Groundmass: Pl, Cpx, Opq
Wailuku Formation				
Wa 1	Stop 6	Basalt, tholeiitic	P	Phenocrysts: Ol, Cpx Microphenocrysts: Ol, Cpx Groundmass: Pl, Ol, Cpx, Opq
Wa 2	Stop 4	Basalt	P; trachytic	Phenocrysts: Pl Microphenocrysts: Ol, Cpx Groundmass: Cpx, Pl, Ol

Notes: 1. On the basis of the SiO_2 *versus* $(Na_2O + K_2O)$ diagram, after LeMaitre *et al.* (1989). 2. A: Aphyric, P: Porphyritic. 3. Mineralogy: phenocrysts (>0.5 mm) and microphenocrysts (<0.5 mm) are differentiated in thin section according to size. 4. SST: Sliding Sands Trail. 5. The sections were examined in transmitted light only; the opaque phases are mainly magnetite and ilmenite. Symbols: Cpx clinopyroxene, Kfs K-feldspar, Ol olivine, Opq opaque phase, Pl plagioclase.

APPENDIX 2. COMPOSITION OF BASALTS FROM MAUI, HAWAI'I

	Haleakala								West Maui			
Sample	Ho 1	Ho 2	Ku 1	Ku 2	Ku 3	Ha 1	Ha 2	Ha 3	Wa 1	Wa 2	H	La
	AB	T	AB	AB	T	AB	AB	AB	T	T	AB	AB
SiO_2 wt.%	48.75	46.74	47.28	47.59	43.67	44.10	44.73	43.53	46.46	46.04	56.99	44.32
Al_2O_3	16.40	13.22	17.86	17.68	9.97	16.38	16.18	15.68	11.58	14.56	17.80	12.61
Fe_2O_3	11.27	12.90	12.99	12.64	13.23	15.37	13.84	15.82	13.57	15.72	8.24	14.05
MgO	4.44	11.75	2.97	2.93	16.20	5.58	6.41	5.88	14.32	6.18	1.83	11.48
CaO	8.59	9.28	7.67	7.64	12.44	8.65	9.77	9.13	9.54	9.75	3.35	9.83
Na_2O	4.50	2.31	5.69	5.99	1.35	4.00	3.81	4.12	1.64	2.73	7.14	3.62
K_2O	1.49	0.35	1.70	1.88	0.37	1.32	1.12	1.29	0.11	0.47	2.54	1.10
TiO_2	3.64	3.02	2.58	2.51	2.34	3.77	3.32	3.74	2.33	3.92	1.19	2.25
MnO	0.26	0.18	0.25	0.24	0.19	0.21	0.21	0.21	0.18	0.19	0.27	0.19
P_2O_5	0.64	0.25	1.02	0.89	0.24	0.63	0.60	0.61	0.26	0.45	0.66	0.54
Cs ppm	0.4	0.2	0.4	0.7	0.5	0.6	0.8	0.4	0.2	0.2	0.2	0.7
Rb	36	10	46	51	14	37	36	43	10	10	49	46
Ba	690	153	812	867	374	772	704	713	82	217	796	751
Th	3.4	1.0	6.9	5.4	1.8	3.4	3.5	3.4	1.0	1.3	5.7	1.8
U	1.1	0.1	1.8	1.6	0.6	0.7	0.7	0.7	0.1	0.3	1.8	1.8
Ta	3.3	0.9	5.0	4.3	1.4	3.2	3.0	3.0	0.7	1.4	4.8	5.2
Mo	5	4	3	5	<2	6	4	4	4	4	5	8
Sr	972	390	1119	1350	352	978	905	876	300	496	1047	820
Zr	280	148	259	305	115	189	175	211	134	221	701	208
Hf	6.8	4.3	7.7	6.2	3.3	5.3	5.1	5.3	3.7	6.1	15.3	4.8
Y	40	35	28	33	18	31	30	31	27	33	57	27
La	47.0	15.7	72.1	58.8	19.3	40.6	39.9	39.8	10.5	19.9	74.4	42.0
Ce	98	35	130	111	40	79	79	79	29	49	155	78
Nd	53	21	62	57	23	41	41	43	16	32	82	34
Sm	10.8	6.32	10.8	9.94	5.15	8.28	8.07	8.33	5.35	7.97	15.1	6.88
Eu	3.65	2.19	3.59	3.39	1.72	2.79	2.71	2.86	1.79	2.61	4.94	2.35
Tb	1.3	0.9	1.2	1.1	0.6	1.0	0.9	1.0	0.7	1.0	1.7	0.8
Yb	3.04	2.41	2.79	2.33	1.33	2.29	2.30	2.42	2.30	2.98	5.35	1.95
Lu	0.46	0.35	0.41	0.34	0.20	0.34	0.33	0.35	0.30	0.41	0.79	0.28

Samples from each volcano are listed according to stratigraphic level, lowest units to the left. The names of the formations are abbreviated as follows: Haleakala volcano: Ho: Honomanu, Ku: Kula, Ha: Hana; West Maui volcano: Wa: Wailuku, H: Honolua, La: Lahaina. Abbreviations to express magma type: AB: alkali basalt, T: tholeiitic basalt.

APPENDIX 3. CONCENTRATION OF Pt IN BASALT FROM KĪLAUEA

K2	K4	MU2	MU2B	P1	PO1	HP1	Average
4.4	3.9	3.6	3.3	2.6	2.4	4.7	3.56 ± 0.86

Concentration is expressed in parts per billion (ppb). See Crocket (2000, Table 1A, Fig. 1) for other PGE data and location for these samples.