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The Incompatible Element Characteristics of an Ancient Subducted Sedimentary Component in Ocean Island Basalts from French Polynesia

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The concentrations of incompatible elements in basaltic rocks from the Cook–Austral and Pitcairn–Gambier island chains of French Polynesia were determined to refine geochemical models for the origin of EM I-type ocean island basalts. It has been proposed previously that the addition of pelagic sediment to a HIMU source could yield mantle with EMI-like characteristics. The very high abundances of many large ion lithophile elements (e.g. Pb) in pelagic sediments would control incompatible element ratios in the proposed EMI source. Calculations indicate that partial melts of this HIMU–pelagic sediment mix would have incompatible element ratios that do not overlap with those of EMI basalts. An investigation of the effect of subduction on a pelagic sediment composition shows that this is a critical process: the sediment undergoes significant loss of large ion lithophile elements during sub-arc devolatilization and metamorphism. Incompatible element ratios in partial melts of a model EMI source which includes this residual metasediment component are shown to be consistent with the range of EMI-type lavas from French Polynesia. In addition, the EMI source may include a smaller ratio of altered oceanic crust to depleted mantle peridotite than is present in a HIMU source.

KEY WORDS: *EMI; HIMU; Polynesia; sediments; subduction*

INTRODUCTION

Variations in the isotopic and trace element composition of ocean island basalts (OIBs) are generally interpreted to reflect mixing of mantle source components with distinct compositions and histories (e.g. Zindler & Hart, 1986; Weaver, 1991; Hofmann, 1997). At least four end-member mantle components have been identified based on the radiogenic isotopic compositions of OIBs: DM [depleted MORB (mid-ocean ridge basalt) source mantle], EMI and EMII (enriched mantle I and II), and HIMU (high μ, i.e. high $^{238}U/^{204}Pb$). The origin and nature of these end-member components are matters of considerable controversy, in particular the depth in the mantle at which the components reside (McKenzie & O'Nions, 1983; Zindler & Hart, 1986; Nakamura & Tatsumoto, 1988; Cousens & Ludden, 1991; Weaver, 1991; Anderson *et al.*, 1992; Halliday *et al.*, 1995). There is, however, considerable evidence that OIBs originate by partial melting within ascending mantle plumes (summarized by Woodhouse & Dziewonski, 1984; Castillo, 1988; Zhang & Tanimoto, 1992; Nataf & VanDecar, 1993; White & Duncan, 1996). The isotopic characteristics of the end-member mantle components are proposed to be derived by (1) ancient subduction of oceanic crust, (2) delamination and subsequent sinking of subcontinental lithosphere deep into the mantle, or (3) mantle metasomatism (Chase, 1981; White & Hofmann,

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1982; McKenzie & O'Nions, 1983; Nakamura & Tatsumoto, 1988). An alternative hypothesis is that all or some of the isotopic characteristics of OIB might be derived from a shallow source in the lithosphere (Halliday *et al.*, 1995; Eiler *et al.*, 1996).

Some of the best examples of OIB volcanoes displaying isotopic heterogeneities are in the islands of French Polynesia located in the South Pacific Ocean (Fig. 1). Within French Polynesia, volcanic rocks from particular islands commonly have the isotopic characteristics of one of the geochemically defined mantle end-members (Fig. 2): HIMU (islands of Tubuai, Mangaia and the older generation of volcanic rocks on Rurutu in the Austral–Cook Archipelago), EMI (islands of Rarotonga in the Austral–Cook Archipelago and Pitcairn in the Pitcairn–Gambier Islands) and EMII (Society Islands) (Palacz & Saunders, 1986; Weaver, 1991; Chauvel *et al.*, 1992; White & Duncan, 1996). Chauvel *et al.* (1992) have modelled the source of HIMU-type basalts from Tubuai as a mixture of ancient subducted ocean crust (AOC) and DM. Based on incompatible element ratios, such as Ce/Pb, and Pb isotopic compositions, they further proposed that EMI and EMII basalts required the inclusion of pelagic and terrigenous sediments (respectively) in a HIMU-like source (White & Hofmann, 1982; Patchett *et al.*, 1984; Weaver, 1991). A similar conclusion has been drawn for basaltic rocks from the Society Islands (White & Duncan, 1996).

The geographic proximity of islands composed of either HIMU- or EMI-type basalts in French Polynesia may support the proposal that the materials carrying the isotopic signatures of these OIBs are also intimately related, such as subducted oceanic crust and sediments. It is this geographic link that leads us to further investigate the proposed sedimentary origin of the EMI mantle component, although other origins have been proposed. In most attempts to model the incompatible element ratios in EMI basalts, however, the composition of modern pelagic sediments has been used as a mixing component, even though it is generally acknowledged that any sediments that descend through a subduction zone must be modified chemically by dehydration, partial melting and metamorphic reactions (Woodhead & McCulloch, 1989; Philippot & Selverstone, 1991; Bebout & Barton, 1993; Cousens *et al.*, 1994). The purpose of this paper is to suggest what the effects of subduction zone 'processing' might be on the incompatible element concentrations in pelagic sediments, contrasting mixing models for the source of EMI-type OIB that rely on modern vs 'processed' sediment trace element ratios. We then use combined isotope-trace element systematics to discuss the origin of the EMI mantle source. In other words, we will investigate the possible geochemical effects of a shallow level process—subduction—on ancient sediments recycled into the mantle source of OIB.

GEOLOGICAL SETTING AND ANALYTICAL METHODS

French Polynesia consists of several NW–SE-trending archipelagos that are scattered over an area of about 1×10^6 km^2. Most prominent are the Austral–Cook, Society, Pitcairn–Gambier and Marquesas chains (Fig. 1). Volcanic activity that produced these islands has been related to several different mantle plumes, although the regional magmatism has been attributed to the presence of a large-scale mantle upwelling or 'superswell' under Polynesia (McNutt & Fischer, 1987; Castillo, 1988). In the case of French Polynesia, both the HIMU and EM components probably reside in mantle plumes. Although mantle-derived magmas may be modified during ascent through the lithosphere, this is not considered to be the process by which basalts from French Polynesia acquire their extreme isotopic characteristics. The youth of the lithosphere (50–55 Ma) does not allow enough time for low-degree melts trapped in the lithosphere during the formation of the crust to attain the high Pb isotope ratios of HIMU or the high ^{87}Sr/^{86}Sr of EM-type OIB unless Rb/Sr and U/Pb ratios are extraordinarily high. In addition, the non-radiogenic Pb isotopic characteristics of EMI-type OIB compared with other oceanic basalts cannot be produced by this mechanism.

The basaltic samples analysed in this study are from the Austral–Cook (Tubuai, MacDonald, Mangaia, Rarotonga, Aitutaki, Atiu, Marotiri, Rapa, Raivavae islands) and Pitcairn–Gambier (Gambier, Mururoa and Fangataufa islands) chains. The volcanoes of French Polynesia are dominated by variably evolved basaltic rocks, which are typically silica undersaturated and belong to the alkali basalt–basanite series. Subordinate tholeiitic basalts are also present in several islands. Ninety samples were analysed for Ce, Pb, Th and U. Many of these samples were also reanalysed for La, Ba and Nb (Table 1). All the trace element analyses were performed by inductively coupled plasma mass spectrometry (ICP-MS) at Centre Géologique et Géophysique, Université Montpellier II, following the procedure of Ionov *et al.* (1992). As a result, we have an internally consistent trace element data set for all of the islands in the study. The precision of the ICP-MS data is 2–4%. The major and some trace element abundances as well as Sr, Nd and Pb isotopic ratios of the analysed samples have been reported previously (Dostal *et al.*, 1983, 1996; Dupuy *et al.*, 1987, 1988, 1989, 1993).

TRACE ELEMENT GEOCHEMISTRY

Our study focuses on the distinction between HIMU (Tubuai, Mangaia) and EMI (Rarotonga, Pitcairn) types of OIB, which are the most common types in French

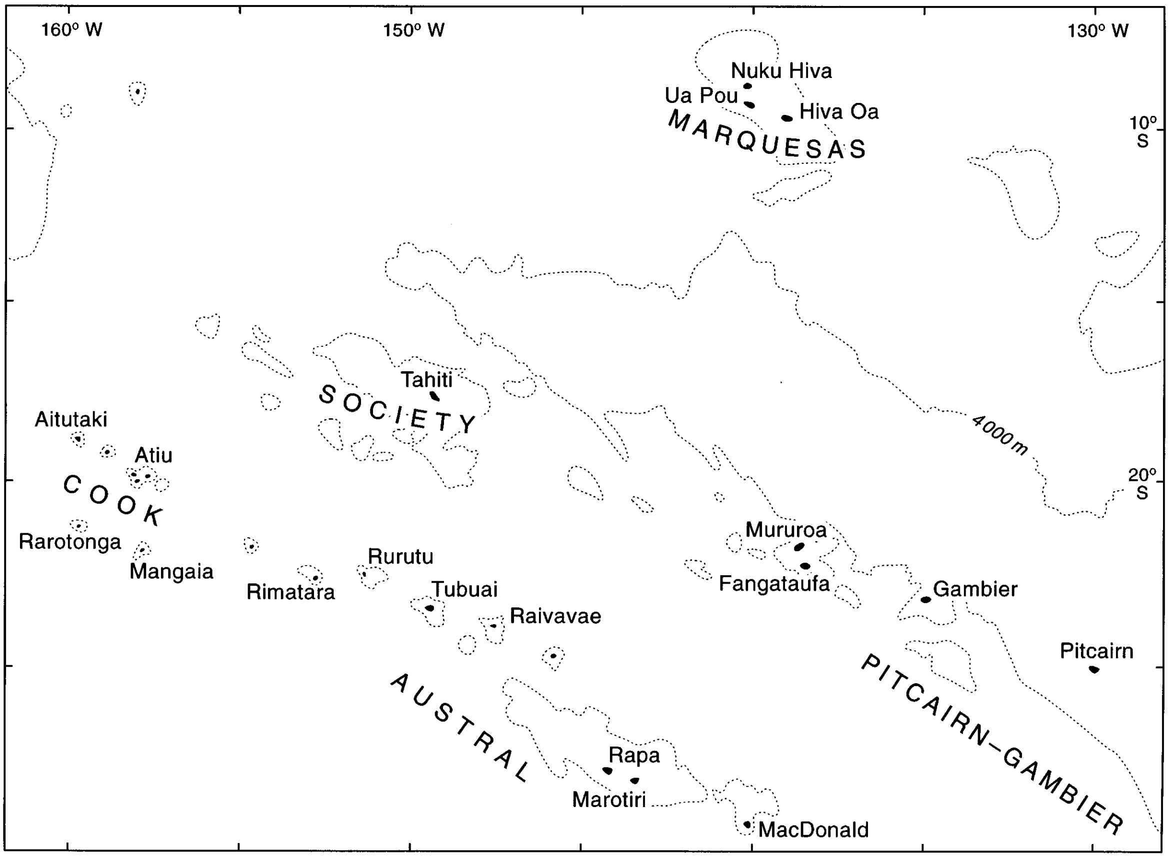

Fig. 1. Generalized map of French Polynesia showing the locations of the archipelagos and islands.

Polynesia. Pb, Th and U contents of the fresh Polynesian OIBs fall within the range typical of other OIBs and are significantly higher than those of MORB (e.g. Hofmann *et al.*, 1986; Sun & McDonough, 1989). In our basaltic samples, Pb contents typically vary from 1 to 6 ppm, Th from 2 to 8 ppm and U from <1 to 2·5 ppm. As typical incompatible trace elements, their abundances increase with differentiation and with the degree of silica undersaturation. Figure 3 shows the variation in abundance of several incompatible elements with respect to Ce in our samples from Tubuai, Mangaia, Rarotonga, and other Cook–Austral and Pitcairn–Gambier islands, as well as previously published data from these islands. EMI basaltic rocks from Rarotonga are enriched in Pb and Ba, but depleted in Nb, at a given Ce content compared with HIMU basalts from Tubuai and Mangaia. Rare earth element (REE; e.g. La) abundances and chondrite-normalized REE patterns are similar in both OIB types (e.g. Dupuy *et al.*, 1988; Sun & McDonough, 1989). The strong relative enrichment of Pb in Rarotonga basalts (and other Cook–Austral island volcanic rocks) is not accompanied by an equivalent Th or U enrichment, although EMI-type basalts have higher Th/U ratios than HIMU-type basalts (see following section).

RATIOS OF INCOMPATIBLE ELEMENTS IN HIMU VS EMI OIB

It has been shown that ratios of some incompatible elements are different in EMI- and HIMU-type OIBs (Weaver, 1991). In a previous study (Dostal *et al.*, 1996) we demonstrated that in moderately altered OIBs from French Polynesia, REE, Nb, Th and Pb are not noticeably redistributed relative to the equivalent fresh rocks. Consequently, the ratios of Ce/Pb, Th/Ce, Th/Pb and Nb/Th can be used confidently to constrain petrogenetic processes. Barium is only moderately mobile during alteration processes (Dostal *et al.*, 1996) and the differences in the trace element ratios involving Ba between EMI and HIMU volcanic rocks do not represent an effect of alteration. As previously suggested by Hart & Staudigel (1982), uranium in OIB does show some mobility during alteration, as exemplified in Fig. 4 by fresh and altered

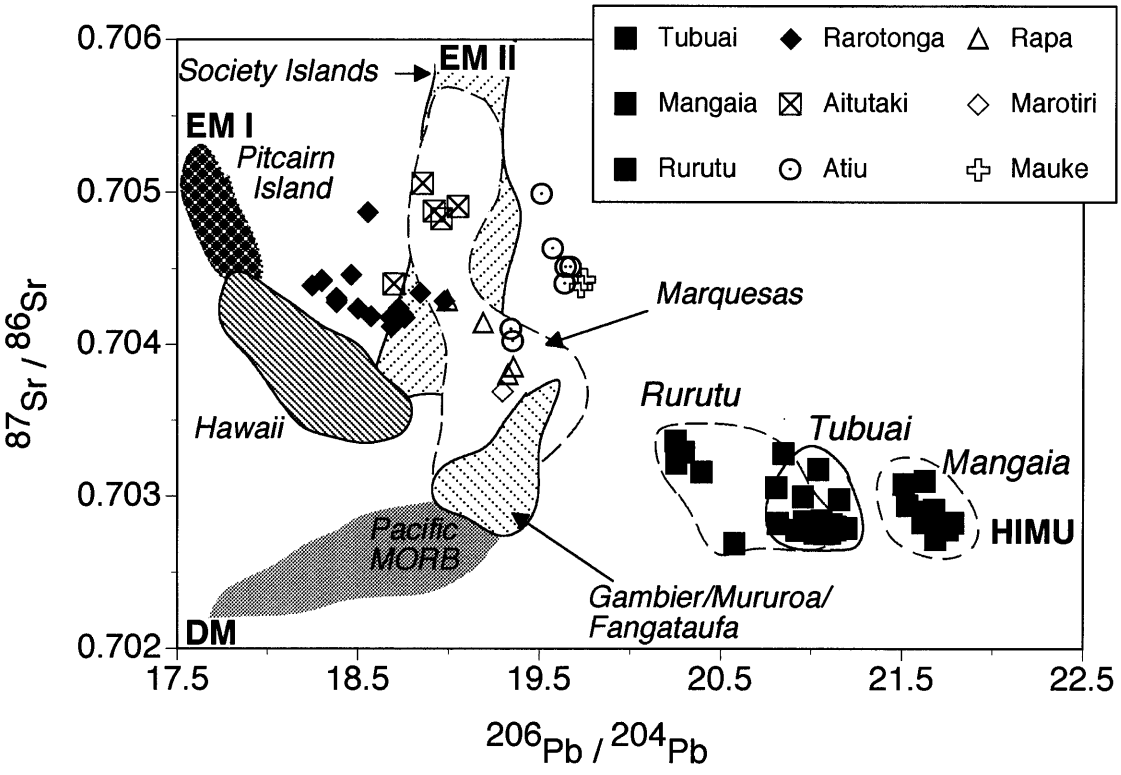

Fig. 2. $^{87}Sr/^{86}Sr$ vs $^{206}Pb/^{204}Pb$ in volcanic rocks from French Polynesia. Data sources: Palacz & Saunders (1986), Nakamura & Tatsumoto (1988), McDonough & Chauvel (1991), Chauvel *et al.* (1992), Hauri & Hart (1993); Hawaii: Yang *et al.* (1994); Pitcairn: Woodhead & McCulloch (1989); Marquesas: Vidal *et al.* (1984), Duncan *et al.* (1986), Desonie *et al.* (1993); Gambier Group: Dupuy *et al.* (1993); MORB: White *et al.* (1987). Mantle end-members DM, EMI, EMII and HIMU from Chauvel *et al.* (1992).

basalts from Mururoa and Fangataufa (Dostal *et al.*, 1996). Only petrographically fresh samples are considered in the following sections.

The differences in U/Pb and Th/Pb ratios in basaltic rocks from several Polynesian islands are illustrated in Fig. 4. These ratios are lowest in tholeiitic basalts and increase with increasing degree of SiO_2 undersaturation of the basaltic rocks. For Tubuai, the highest values are encountered in nephelinites. At each island, the differences of these ratios are probably the result of variable degrees of partial melting, based on the bulk partition coefficient (*D*) of Pb, which is higher than those of Th and U (e.g. Chauvel *et al.*, 1992). As shown in Fig. 4, the Th/U ratio varies geographically, from island to island. At Tubuai, the average value (3·7) is the same as the time-integrated Th/U ratio (3·65–3·75) calculated by Chauvel *et al.* (1992) on the basis of Pb isotope ratios for basalts from this island. The average Th/U ratio in Rarotonga alkali basalts and basanites, which contain the EMI mantle component, is 4·5. A comparison of fresh basanites from several islands of the Austral–Cook Archipelago shows that these volcanic rocks commonly have higher Th/U than HIMU volcanic rocks (Nakamura & Tatsumoto, 1988). Lavas from Pitcairn Island (Woodhead & McCulloch, 1989), which have the strongest EMI isotopic signatures in French Polynesia, also have the highest average Th/U ratio (>10).

The Ce/Pb ratio, which is not significantly modified from that of the mantle source during partial melting, has been used to characterize OIB mantle sources (Hofmann *et al.*, 1986; Newsom *et al.*, 1986; Chauvel *et al.*, 1992). The lowest values of Ce/Pb (20 ± 4) are found in the basalts of Rarotonga and the nephelinites from Aitutaki, which both have isotopic signatures dominated by the EMI component (Palacz & Saunders, 1986; Nakamura & Tatsumoto, 1988; Chauvel *et al.*, 1992). Higher ratios (29 ± 4) are encountered in the HIMU-type basaltic rocks of Tubuai and Mangaia and the first (older) generation of basalts from Rurutu (Vidal *et al.*, 1984; Palacz & Saunders, 1986; Nakamura & Tatsumoto, 1988; Weaver, 1991).

Besides Ce and Pb, the elements Ba and Nb serve to distinguish EMI-type from HIMU-type OIB. Unlike Ce/Pb, however, the range of ratios such as Nb/Pb, Ba/Pb

Table 1: Trace element abundances in basaltic rocks from French Polynesia

Sample	Type	Th	U	Pb	La	Ce	Nb	Ba
COOK–AUSTRAL								
Mangaia								
7931	B	3·03	0·89	1·88	25·9	56·1	35	166
7922	B	4·39	1·34	3·90	39·3	87·2	52	270
7926	B	2·95	0·91	1·91	27·5	61·8	37	174
7933	B	3·04	0·92	2·42	27·7	60·8	36	210
7934	B	6·48	2·04	4·71	55·8	123	79	399
Rurutu (older)								
RTT-B21	B	2·15	0·60	1·86	21·1	47·2	27	132
7303	B	2·24	0·70	1·98	22·1	49·9	30	154
7327	B	4·84	1·75	3·32	48·4	107	59	355
7314	B	2·19	0·59	1·44	21·0	45·8	28	125
7329	B	3·58	1·07	2·59	33·2	73·0	45	225
7317	B	2·75	0·81	1·75	26·5	59·7	34	184
Rurutu (younger)								
7325	B	6·21	1·70	3·22	—	131	—	—
7310	B	6·24	1·71	2·82	—	133	—	—
7309	B	6·55	1·93	3·86	*68·9*	156	*84*	*442*
RRT-30	B	4·33	1·17	2·44	—	106	—	—
7307	B	6·95	2·01	3·76	—	160	—	—
7323	B	5·66	1·60	3·34	—	125	—	—
Tubuai								
7288	B	3·46	1·01	2·16	30·4	65·2	47·0	226
9012	B	4·22	1·24	2·49	36·7	77·1	57·0	270
8903	B	2·65	0·68	1·58	45·9	52·2	30·7	169
8891	B	5·92	1·54	3·62	49·3	103	75·0	327
8905	B	8·34	2·29	5·28	73·5	154	105	489
7291	B	4·96	1·33	3·14	45·9	98·5	68·7	313
9548	B	4·81	1·28	3·64	42·1	88·4	60·0	266
TB-35	B	3·87	0·97	2·29	33·8	70·6	48·4	215
7282	T	7·58	0·93	8·65	—	118	—	—
9551	N	20·7	5·32	11·7	—	271	—	—
9552	N	19·8	5·24	9·18	—	264	—	—
7294	N	19·4	5·13	10·2	—	270	—	—
7285	N	11·1	3·08	7·20	—	179	—	—
9553	N	19·3	5·45	10·6	—	244	—	—
Rarotonga								
103	B	4·81	1·10	3·46	32·5	66·8	39·0	371
104	B	4·27	0·94	3·55	29·0	59·6	30·4	335
106	B	12·0	2·30	7·82	81·6	161	81·1	985
108	B	9·19	2·08	6·52	68·2	137	85·8	788
110	B	7·81	1·79	5·69	55·0	110	70·1	628
114	B	2·89	0·64	2·21	23·8	47·7	24·9	291
115	B	4·15	0·92	2·88	28·5	58·4	35·9	322
116	B	8·49	1·95	5·29	58·9	120	76·8	677
118	B	8·16	1·77	6·68	62·5	128	77·9	824
130	B	7·78	1·61	5·06	58·2	110	62·5	635

Table 1: continued

Sample	Type	Th	U	Pb	La	Ce	Nb	Ba
Rarotonga								
132	B	4·81	1·10	3·62	37·1	75·9	47·9	413
140	B	8·49	1·96	6·26	65·1	136	66·7	757
141	B	6·74	1·59	5·36	51·6	107	52·0	554
147	B	3·76	0·85	3·06	29·3	61·9	32·0	336
152	B	8·25	1·89	5·95	62·0	125	66·5	686
161	B	7·03	1·56	5·13	53·0	107	56·8	630
176	B	12·5	2·59	9·37	86·6	168	87·9	1055
177	B	9·35	2·17	6·84	70·6	140	76·1	817
180	B	7·26	1·62	5·35	53·7	110	62·6	610
182	B	4·71	1·11	3·29	39·3	81·6	43·1	439
186	B	8·72	2·01	5·32	61·1	119	71·6	698
188	B	8·30	1·80	6·92	62·9	126	68·6	750
188A	B	8·13	1·83	6·58	61·5	121	66·2	735
190	B	7·96	1·68	5·74	57·9	114	63·9	642
192	B	7·88	1·81	6·02	60·9	124	65·7	687
193	B	7·25	1·62	5·30	54·0	109	55·1	634
Aitutaki								
7987	B	5·49	1·57	4·31	—	106	—	—
7990	N	15·2	3·35	8·35	—	158	—	—
7996	N	18·6	3·66	9·35	—	179	—	—
8001	B	7·81	1·64	5·75	—	83·9	—	—
8004	B	7·77	1·62	5·63	—	85·1	—	—
Atiu								
7939	B	3·59	1·01	2·41	—	64·1	—	—
7940	B	5·78	1·69	3·73	—	105	—	—
7943	B	3·47	0·98	2·35	—	63·0	—	—
Raivavae								
7268	B	3·86	1·02	2·42	—	73·1	—	—
7275	T	2·03	0·42	1·44	—	41·6	—	—
7277	T	2·5	0·36	1·63	*21·9*	48·6	*34*	*193*
Rapa								
7258	B	3·57	0·93	2·94	—	73·9	—	—
7263	B	2·91	0·80	2·29	*26·4*	59·1	*39*	*242*
7266	B	3·63	1·01	2·95	*34·0*	76·4	*48*	*330*
Marotiri								
6988	B	5·04	1·43	3·35	—	104	—	—
6991	B	3·13	0·69	2·22	*25·4*	54·0	*43*	*492*
6996	B	2·99	2·16	5·13	*27·2*	60·4	*42*	*242*
MacDonald								
6974	B	3·26	0·95	2·76	*31·2*	69·0	*44*	*284*
6983	B	1·94	0·86	1·99	*14·0*	40·0	*27*	*151*
6984	T	2·39	0·57	2·43	*23·0*	48·2	*31*	*155*

Sample	Type	Th	U	Pb	La	Ce	Nb	Ba
PITCAIRN–GAMBIER								
Gambier								
MG10E	T	2·33	0·56	2·10	*30·0*	53·1	—	*215*
TR14S	T	2·45	0·79	1·99	*24·7*	51·4	—	*200*
OG29D	T	2·27	0·74	1·80	—	50·6	—	—
AK34H	T	2·09	0·53	1·97	*20·1*	41·4	—	*175*
KM7L	T	1·91	0·63	1·84	*19·9*	45·5	—	*155*
Fangataufa								
Fa608	T	2·40	0·74	2·31	*24·0*	53·4	—	*172*
Fa682·7	T	1·46	0·43	1·33	*14·8*	33·5	—	*102*
Fa808	T	1·49	0·51	1·71	*16·6*	37·8	—	*125*
Mururoa								
V3592	B	5·23	1·66	3·93	*57·7*	130	—	*431*
V3970	B	3·11	0·85	2·50	*29·1*	67·5	—	*211*
V4792	B	3·61	1·19	3·19	*37·4*	79·3	—	*268*
V8310	B	5·06	1·57	3·62	*53·0*	121	—	*409*
V8502	B	4·88	1·48	3·44	*52·3*	117	—	*394*

All concentrations in weight parts per million. Values in italics are XRF or INAA data; dash indicates no analysis made. Rock types: B, basanite or alkali basalt; N, nephelinite; T, tholeiite.

and Nb/La in EMI and HIMU basalts overlap. In addition, Th/Pb ratios in the basalts from the different islands of this study overlap almost completely (Fig. 4), yet EMI and HIMU basalts have very different ^{208}Pb/^{204}Pb ratios (Palacz & Saunders, 1986; Nakamura & Tatsumoto, 1988; Chauvel *et al.*, 1992). Using the data analysis computer program DataDesk 5.0, we have constructed three-axis rotating plots of key trace element ratios to determine if the overlap in certain trace element ratios is an artefact of plotting in two dimensions. In all of the 3-D plots (e.g. Fig. 5), Rarotonga lavas form a very homogeneous group compared with the Tubuai–Mangaia group. The most striking feature of all of the plots is that the EMI and HIMU data points cluster into distinct, elongate arrays that do not overlap. This is particularly evident in a plot of Nb/La–Nb/Pb–Ba/La (Fig. 5). Individual arrays may be the product of variable degrees of partial melting of a fairly homogeneous mantle source, and each array requires a chemically distinct source.

DEEPLY SUBDUCTED SEDIMENTS AND MELTING OF A MIXED SOURCE

Incompatible elements

Based primarily on Sr–Nd–Pb isotopic data and Ce/Pb ratios, the EMI source has been postulated to be a mixture of AOC, DM and small amounts of ancient (1–2 Ga) subducted pelagic sediment (Weaver, 1991; Chauvel *et al.*, 1992). This assumes that much of the sedimentary section is subducted along with the crust–lithosphere slab, similar to what occurs at several modern subduction zones (e.g. Hole *et al.*, 1984; Von Huene & Scholl, 1991). Geochemical models for the evolution of EMI-type OIB based on this scenario invariably use the measured trace element composition of modern pelagic sediments for the sedimentary component. Any subducted sediment that is present in a modern OIB source will, however, have been modified chemically by sub-arc dehydration reactions and metamorphism (Woodhead & McCulloch, 1989; Philippot & Selverstone, 1991; Bebout & Barton, 1993; Cousens *et al.*, 1994, 1996). In this section, we evaluate the effects of sub-arc reactions on subducted sediments and then model partial melting of a mantle source containing such residual sedimentary material.

Some back-arc basalts from the Japan Sea include an enriched component, inferred to be derived from recently subducted sediments that had been modified during subduction (Cousens *et al.*, 1994). By comparing the incompatible element character of the apparently sediment-derived enriched component in the basalts with that of pelagic sediments drilled in the northwestern Pacific, Cousens *et al.* (1994) estimated the relative losses of incompatible elements in sediments as a result of sub-arc processing (devolatilization and/or melting?). Relative

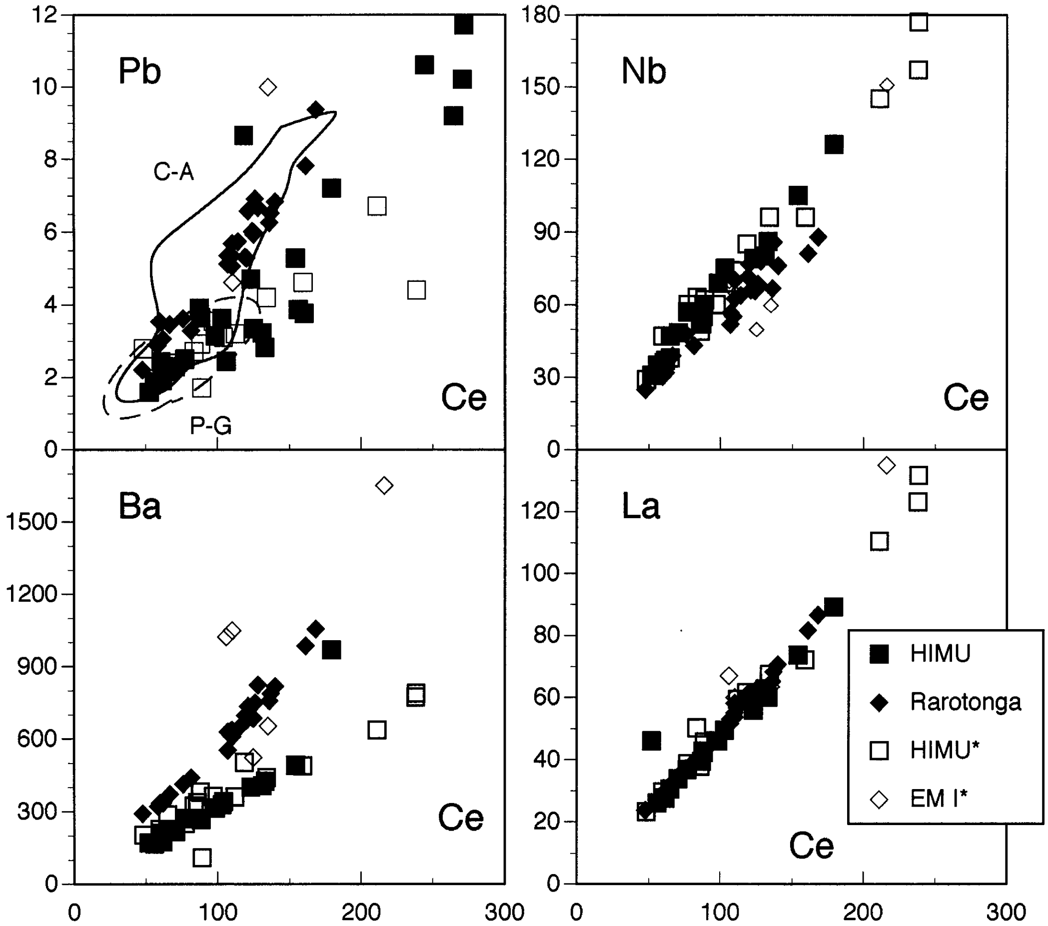

Fig. 3. Pb, Nb, Ba and La abundances (in ppm) plotted vs Ce in basaltic rocks from French Polynesia. HIMU (Tubuai, Mangaia, Rurutu), Rarotonga (EMI), Cook–Austral (C-A) and Pitcairn–Gambier (P-G) data are from this study, HIMU* and EMI* are data from the literature (Palacz & Saunders, 1986; Chauvel *et al.*, 1992).

to the REE, large ion lithophile elements (LILE) are preferentially depleted in the residual sediment, whereas the high field strength elements (HFSE) are enriched (Fig. 6). The largest losses are in Rb, Ba, Sr and Pb. The residual sediment has a lower Rb/Sr but higher U/Pb and Th/Pb than unmodified pelagic sediment. The dramatic loss of Pb from the sediments and consequent higher U/Pb in the residue are consistent with the complementary low U/Pb ratio in island arc basalts relative to MORB and OIB (Tatsumoto & Knight, 1969). The inferred high U/Pb of the sediment residue is contrary to the supposition of Woodhead & McCulloch (1989), who assumed a low U/Pb in subducted sediments to model the Pb isotope evolution of the source of the Pitcairn Island EMI source. Table 2 presents estimates of key trace element concentrations and ratios in devolatilized (residual) pelagic sediments.

In three-axis plots of incompatible element ratios including Th/Pb, Ba/Pb, Nb/Pb, Ce/Pb, Ba/La and Nb/La (e.g. Fig. 5), the Rarotonga and Tubuai–Mangaia arrays are roughly linear and are distinct from each other. Figure 7 presents two-dimensional scatterplots including (a) Ba/Pb vs Nb/Pb and (b) Ba/La vs Nb/La. A noteworthy feature is the factor of 1·5–2 range in these ratios for both islands, which probably results from variable degrees of partial melting. There is some overlap in Ba/Pb and Nb/Pb between HIMU and EMI OIBs, but the data arrays for each OIB type do not overlap. Also shown are batch melting curves (Table 2, Shaw, 1978) for (1) DM, (2) a model HIMU source that is a mixture of 2 Ga AOC and DM in the proportions 25:75 (following Chauvel *et al.*, 1992), and (3) the same model HIMU source mixed with a small proportion of sediment. The sediment composition 'RS' is a 'residual sediment' after sub-arc devolatilization (Cousens *et al.*, 1994), which has a significantly higher U/Pb but lower Ba/La than modern pelagic sediment ('PS', Cousens *et al.*, 1994). The mineralogy of the model HIMU source

Table 2: Trace element and isotopic characteristics of mantle components

	Depleted mantle	2·0 Ga altered oceanic crust	HIMU source	10% AOC/ 90% DM	Pelagic sediment	Residual sediment
Rb	0·12	1·81	0·54	0·29	122	39
Ba	1·39	11·56	3·93	2·41	988	464
Th	0·02	0·16	0·05	0·03	17·5	12·1
U	0·009	0·066	0·023	0·014	1·87	1·48
Nb	0·42	2·92	1·05	0·67	13	26
La	0·47	3·25	1·16	0·75	47·0	42·3
Pb	0·058	0·230	0·101	0·075	53	17
Sr	12·0	94·3	32·6	20·2	163	80
Nd	1·30	9·32	3·31	2·10	49·0	44·1
Sm	0·45	2·96	1·08	0·70	13·3	12·0
Rb/Sr	0·010	0·019	0·016	0·014	0·748	0·487
Sm/Nd	0·346	0·317	0·326	0·333	0·271	0·272
U/Pb	0·155	0·287	0·229	0·187	0·035	0·087
Th/Pb	0·345	0·696	0·495	0·400	0·330	0·712
^{87}Sr/^{86}Sr	0·70221	0·70351	0·70314	0·70281	0·76559	0·74446
^{143}Nd/^{144}Nd	0·51312	0·51285	0·51290	0·51300	0·51148	0·51148
^{206}Pb/^{204}Pb	17·94	22·67	21·20	19·39	16·97	17·37
^{207}Pb/^{204}Pb	15·46	16·04	15·77	15·64	15·48	15·52
^{208}Pb/^{204}Pb	37·61	41·38	40·30	38·77	39·05	39·35

All trace element contents in weight parts per million. All isotope ratios are present-day compositions. Depleted mantle (DM), altered ocean crust (AOC), and HIMU (25% AOC, 75% DM) source are after Chauvel *et al.* (1992), except that Rb/Sr is assumed to be higher in AOC. Pelagic sediment taken from Cousens *et al.* (1994), and residual sediment is calculated from Cousens *et al.* (1994) assuming 10% Nd loss during devolatilization. Initial ratios 2 Ga ago for both sediment types are: ^{87}Sr/^{86}Sr = 0·70500, ^{143}Nd/^{144}Nd = 0·50960, ^{206}Pb/^{204}Pb = 15·30, ^{207}Pb/^{204}Pb = 15·27, ^{208}Pb/^{204}Pb = 34·60.

is presented in Table 3 along with the mineral–liquid partition coefficients used in the modelling. In the mixed HIMU plus sediment models, the sediment is treated as if it were a mineral phase, for which bulk partition coefficients are based on the estimates of Irifune *et al.* (1994) at pressures <6 GPa [constrained by final equilibration of melt and sediment residue at depths where eclogite and peridotite are also melting to form a basaltic liquid (Takahashi, 1986; Kato *et al.*, 1988)]. Melting of the 'HIMU plus sediment' sources yields melting arrays with steeper slopes than the HIMU melting curve, parallel to the Rarotonga data array. Partial melts of a mixed HIMU–PS source never attain the highest Th/Pb and Nb/Pb of EMI basalts, whereas partial melts of mixed HIMU–RS sources do. The reason for this is that pelagic sediments are too enriched in Pb relative to the HIMU source for melting of a mixed HIMU–PS source to produce the high Ba/Pb in EMI basalts. In Figs 5 and 7, the HIMU–RS melting curves do not always overlap the Rarotonga data arrays, but the direction of elongation of the arrays is consistently parallel to the melting curves. It should be noted that the HIMU source would better model the Tubuai basalts in Fig. 7a and b if it had higher Ba/Pb and Th/Pb. Such a HIMU source would also shift the HIMU + RS melting curves to higher Ba/Pb such that they would pass closer to or through the Rarotonga data. From Fig. 7, the Rarotonga data are consistent with a source containing 1–2% residual sediment.

Radiogenic isotopes

In the following, we model and compare the isotopic evolution of mixed HIMU–PS and HIMU–RS sources. The initial compositions of the sources are listed in Table 2. Because any sedimentary component in an OIB source must be ancient (1–2 Ga) to evolve to significantly higher ^{207}Pb/^{204}Pb than MORB, initial isotope ratios are chosen for average upper crust at 2·0 Ga (Stacey & Kramers, 1975; Zartman & Haines, 1988). The trace element composition of the HIMU source is taken from

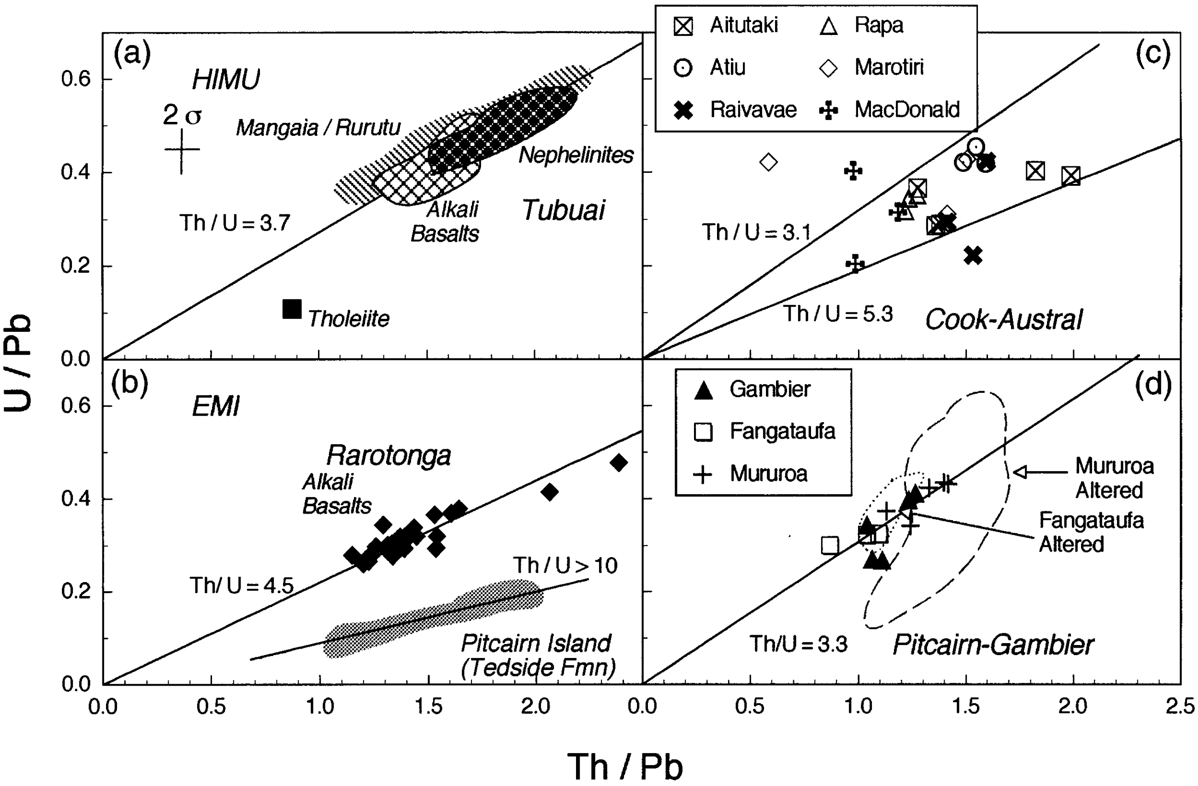

Fig. 4. Variations of U/Pb vs Th/Pb in basaltic rocks from (a) HIMU islands, (b) EMI islands [Pitcairn Island: Woodhead & McCulloch (1989)], (c) other Cook–Austral Islands, and (d) other Pitcairn–Gambier islands, including altered (dotted and dashed fields) and fresh samples from Fangataufa and Mururoa. The cross corresponds to 2σ analytical error.

Chauvel *et al.* (1992), except that the Rb content of AOC is assumed to be higher than DM [based on the results of Hart & Staudigel (1982)]. Parent/daughter ratios in PS are taken from Cousens *et al.* (1994); pelagic clay samples only), which differ from those measured in surface sediments by Ben Othman *et al.* (1989) in part because of diagenesis of the DSDP Leg 86 clays. Parent/daughter ratios in RS are calculated from Cousens *et al.* (1994).

Sr–Nd isotope bulk mixing curves between HIMU and 2·0 Ga PS and RS are shown in Fig. 8a. Devolatilization lowers Rb/Sr in the residual sediments but does not change Sm/Nd, and thus the $^{87}Sr/^{86}Sr$ in RS is lower than that in PS. It should be noted that the Rb/Sr value for PS (Table 2) is significantly higher than that assumed by Chauvel *et al.* (1992) and thus the HIMU–PS mixing curve trends towards EMII compositions rather than EMI. The HIMU–RS curve passes close to, but above, the Pitcairn Island array, suggesting that the crustal component in the Pitcairn source may have had an even lower Rb/Sr than RS. Unlike the Pitcairn lavas, the Rarotonga basalts do not form a curvilinear array that would indicate simple binary mixing with a sedimentary component. Instead, the Rarotonga data plot as a cluster between the HIMU–PS and HIMU–RS mixing curves, suggesting that either the HIMU or the sedimentary component is heterogeneous.

Devolatilization increases U/Pb in the subducted sediment, and thus mixing lines between HIMU and a 2·0 Ga residual sediment plot above a HIMU–PS mixing line in Fig. 8b. However, both the Pitcairn and Rarotonga data plot below the HIMU–PS line, indicating that the sedimentary component has a lower U/Pb than either modern pelagic sediments [as suggested by Woodhead & McCulloch (1989)] or residual sediments. Thus, if subducted sediments are the non-radiogenic Pb components in EMI mantle sources, some process(es) must further modify the residual sediments before their incorporation in the EMI source. Cousens *et al.* (1996) proposed that subsequent to sub-arc devolatilization, the residual sediments undergo further metamorphism at greater depths which lowers Rb/Sr, U/Pb and, to a more variable extent, Th/Pb in the residual sediments. Metamorphism within the granulite facies is known to preferentially drive off U (and Th?) relative to Pb, as

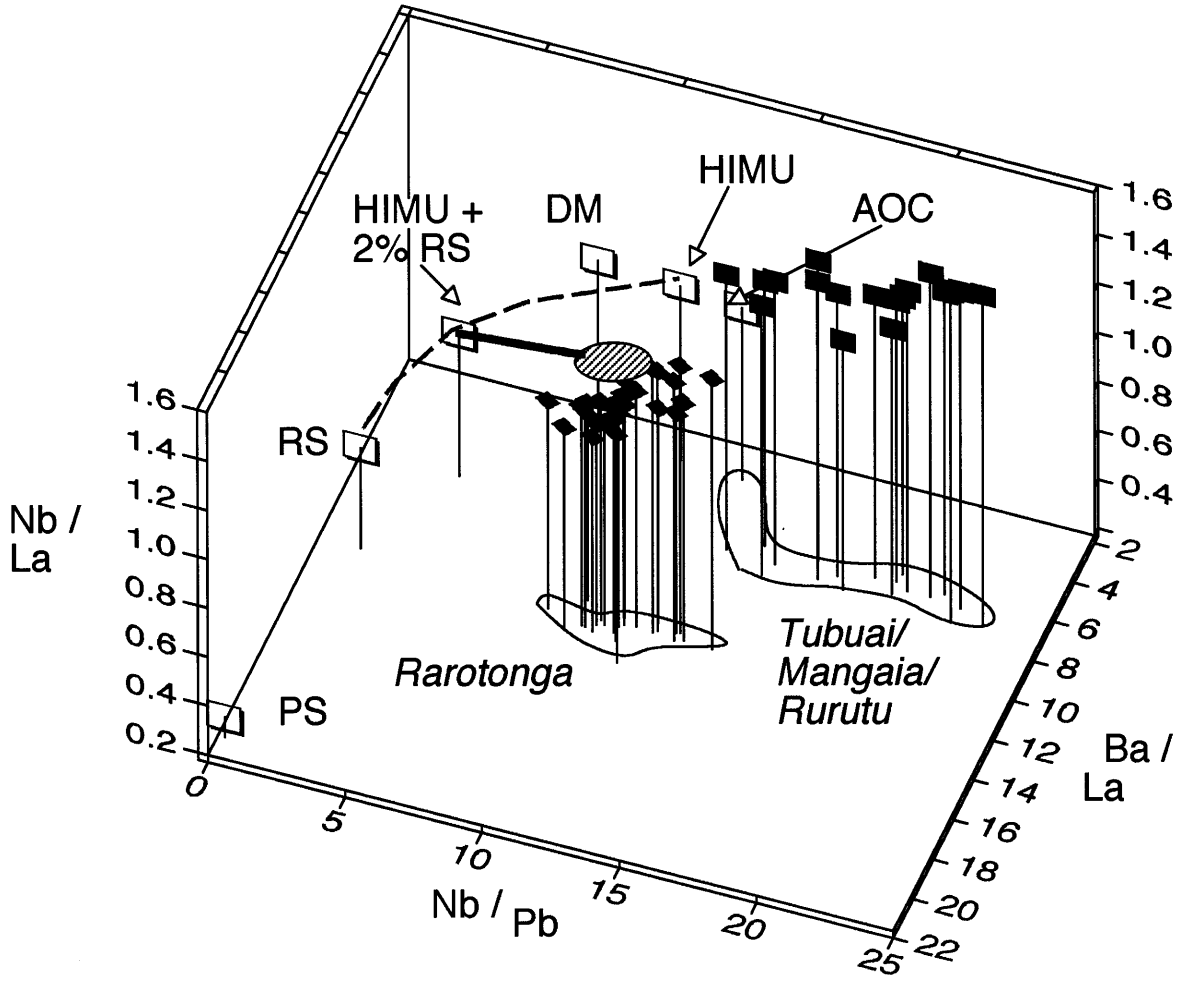

Fig. 5. Three-dimensional scatterplot of Nb/La–Nb/Pb–Ba/La in basalts from French Polynesia. Tie-lines are shown extending from the data points to the base of the plot. □, Mantle end-members: DM, depleted mantle; AOC, altered oceanic crust; PS, pelagic sediments; RS, residual sediment after devolatilization. Model 'HIMU + RS' source is 74% DM, 24% AOC and 2% residual sediments. The dashed line is mixing curve between HIMU and RS, and the thick continuous line is a batch partial melting curve (Shaw, 1978) for the HIMU + RS source. Lined oval area accentuates composition of <1% melt of the HIMU + RS source, and plots just above the EMI data array in this scatterplot.

well as Rb relative to Sr (Dostal & Capedri, 1978; Rudnick & Taylor, 1987). This process has little effect on the relative abundances of Ba, Sr, REE and HFSE (Rudnick & Taylor, 1987). An important mineralogical reaction that releases uranium during metamorphism of subducted sediments may be the transformation of K-rich mica (phengite, high U/Pb) to orthoclase (low U/Pb) (Tilton *et al.*, 1989; Irifune *et al.*, 1994).

Alternatively, the observation that EMI basalts do not lie on a mixing line between HIMU and either PS or RS in isotope–isotope plots may indicate that the Chauvel *et al.* (1992) HIMU model source is not a mixing end-member for EMI basalts. A mixture of 5–10% AOC and 90–95% DM (Table 2), rather than the 25:75 mix used to model HIMU by Chauvel *et al.* (1992), would yield a better candidate for the non-sediment component in EMI basalts (Fig. 8). At first glance, the lower proportion of AOC in EMI-type compared with HIMU-type OIB is puzzling if subducted sediment and its underlying oceanic crust remain together as they are subducted and descend through the mantle. However, given the comparatively lower melting temperature of metasediment (Irifune *et al.*, 1994), partial melts of the sediment may migrate away from its associated oceanic crust within an ascending mantle plume. In this way

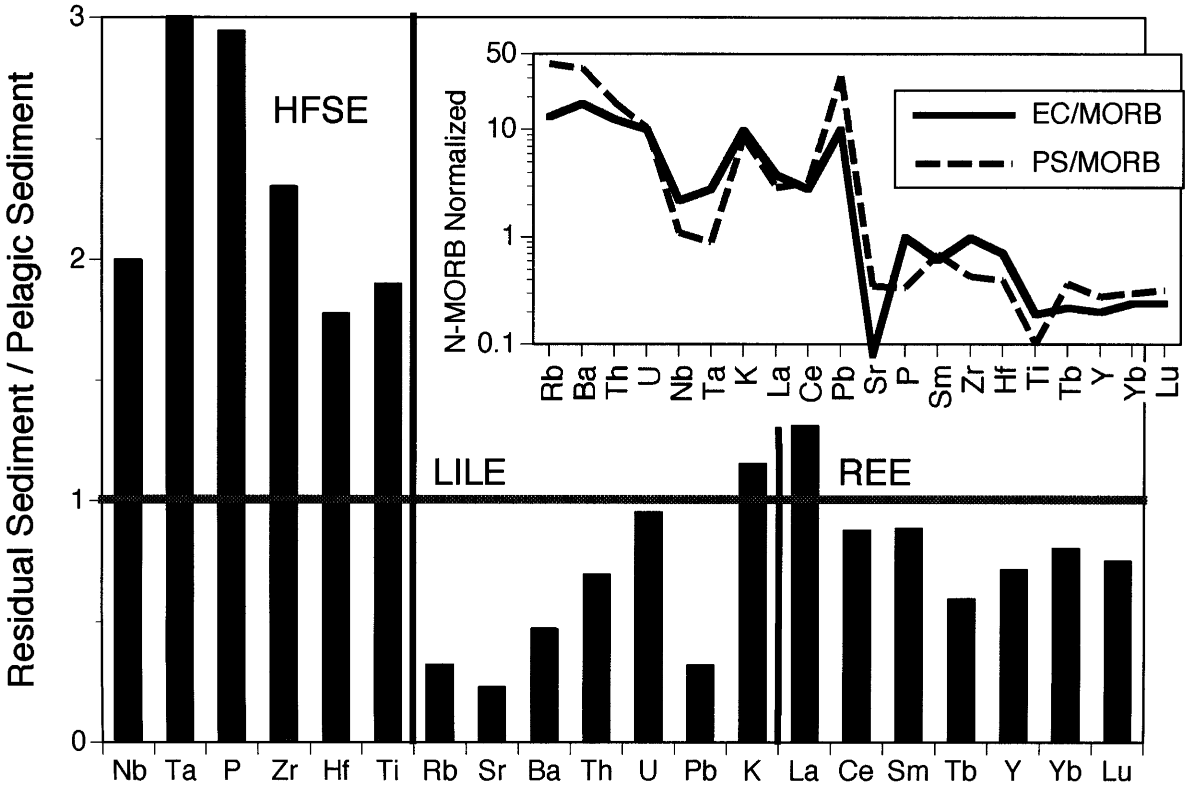

Fig. 6. Bar chart showing relative losses and gains of incompatible elements in the enriched (subduction) component identified in back-arc basalts from the Japan Sea relative to average pelagic sediments (Cousens *et al.*, 1994). HFSE, high field strength elements; LILE, large ion lithophile elements; REE, rare earth elements. Inset: MORB-normalized incompatible element patterns for pelagic sediments (PS) and enriched component (EC) (Cousens *et al.*, 1994). The enriched component was interpreted to be a fluid from devolatilized sediments.

Table 3: Phases and partition coefficients used in batch melting calculations

Phase:	Olivine	Opx	Cpx	Garnet	Sediment
% in AOC:	0	0	50	50	0
% in peridotite:	55	25	10	10	0
% of AOC in mix 25%:	41	19	20	20	0
D (mineral/liquid)					
Ba	0·001	0·005	0·05	0·001	0·2
Th	0·001	0·005	0·05	0·001	0·2
U	0·001	0·005	0·05	0·001	0·2
Nb	0·001	0·005	0·05	0·001	0·2
La	0·001	0·005	0·10	0·01	0·2
Ce	0·001	0·005	0·18	0·02	0·2
Pb	0·001	0·005	0·18	0·02	0·2

Mineral proportions and mineral/liquid *D* (except for sediment) values follow Chauvel *et al.* (1992). *D* values for sediments are bulk *D* values from Irifune *et al.* (1994) at pressures <6 GPa.

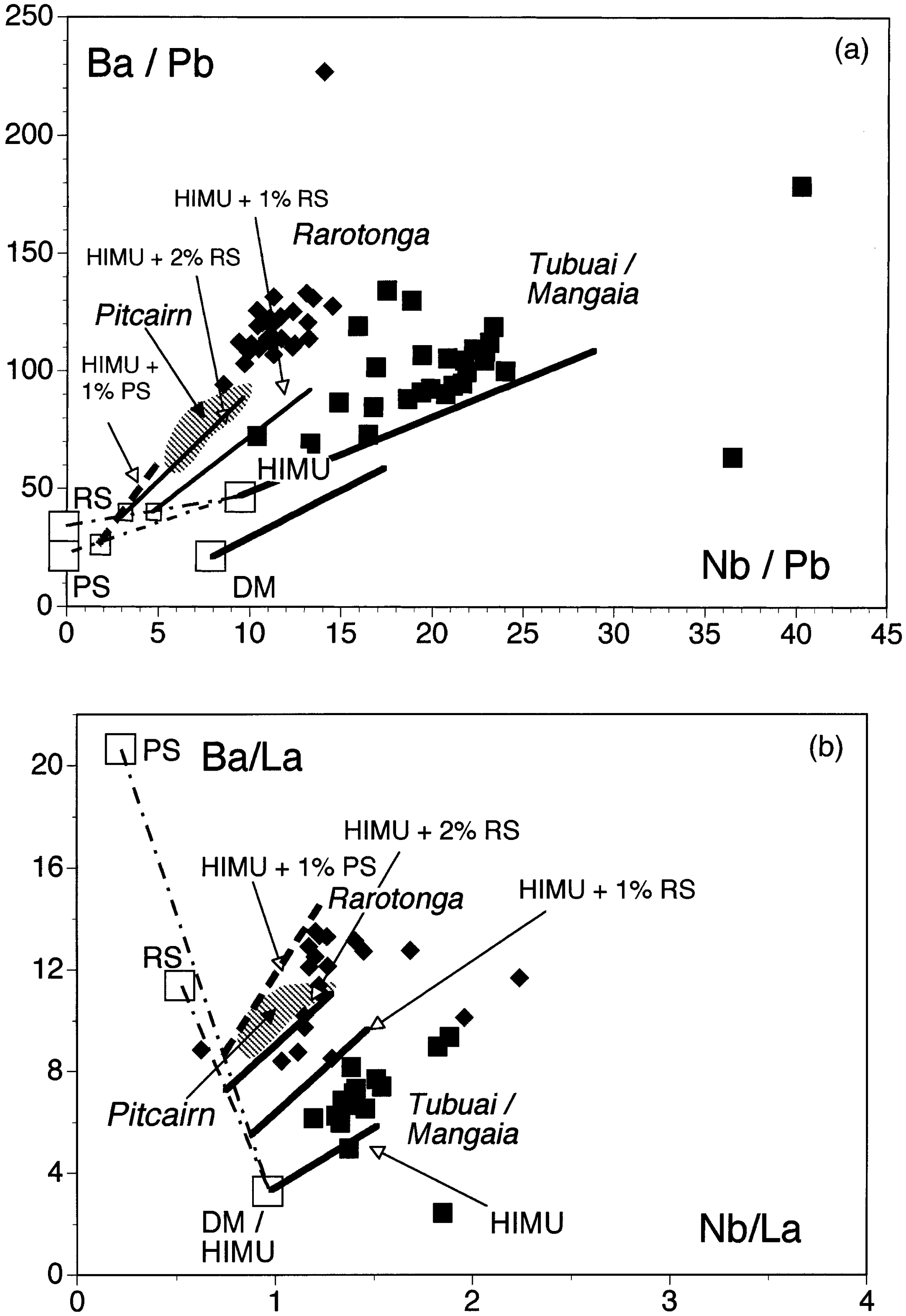

Fig. 7. (a) Ba/Pb vs Nb/Pb and (b) Ba/La vs Nb/La ratios in alkali basalts from Tubuai–Mangaia and Rarotonga in French Polynesia. ☐, Mantle end-member compositions, which are defined in the text and Table 2. Continuous lines are batch melting curves (Shaw, 1978) for pure DM and HIMU, and for mixtures of HIMU and 1% or 2% of RS. Thick dashed line is melting curve for HIMU plus 1% of PS.

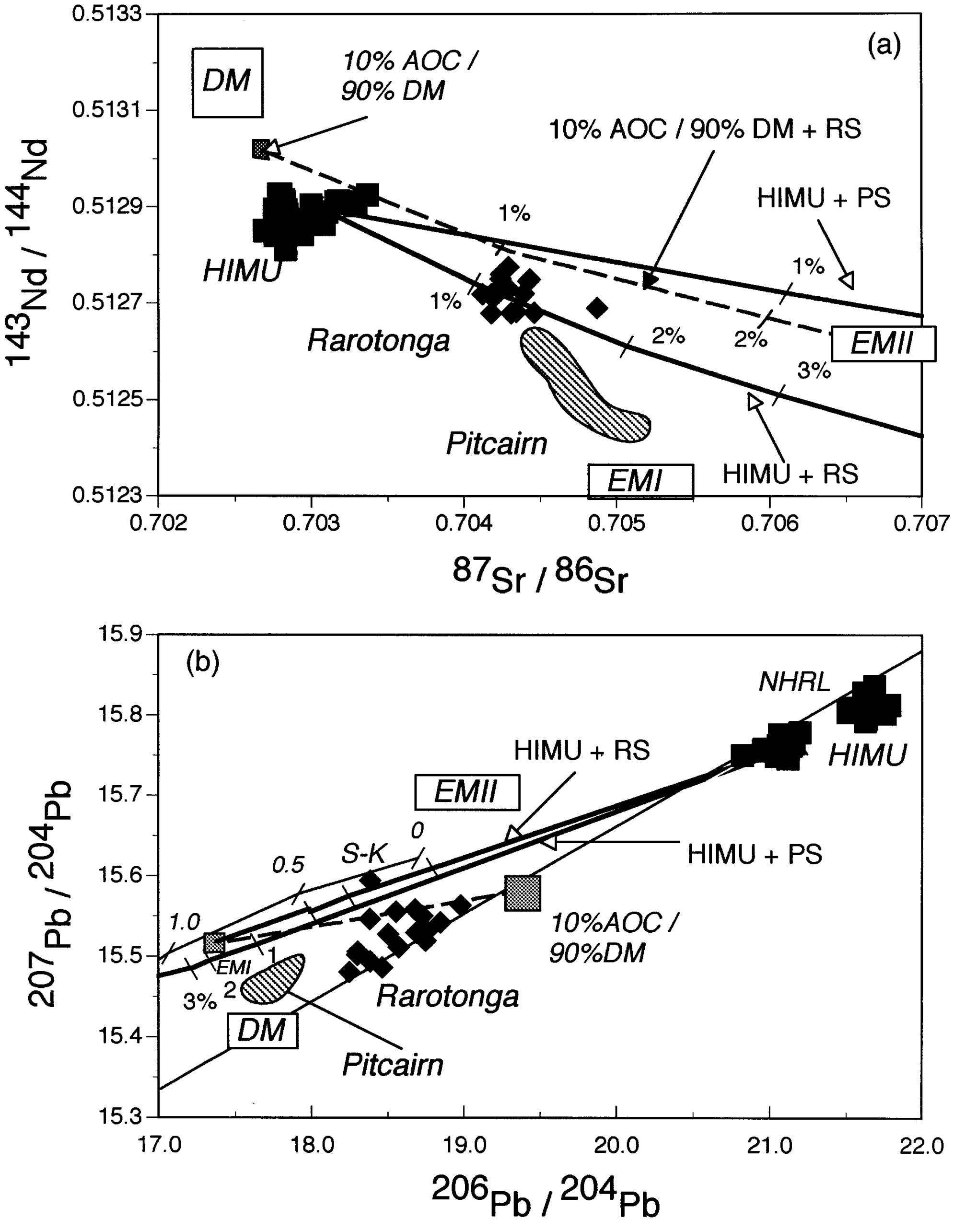

Fig. 8. (a) $^{87}Sr/^{86}Sr$ vs $^{143}Nd/^{144}Nd$ in basaltic rocks from Tubuai (HIMU), Rarotonga and Pitcairn Island (Palacz & Saunders, 1986; Nakamura & Tatsumoto, 1988; Woodhead & McCulloch, 1989; Chauvel *et al.*, 1992). Continuous lines are mixing curves between HIMU and PS, HIMU and RS, whereas dashed line is mixing curve between 10% AOC–90% DM peridotite component and RS. Tick marks indicate 1, 2 and 3% of sediment in the mix. (b) $^{207}Pb/^{204}Pb$ vs $^{206}Pb/^{204}Pb$ in Tubuai (HIMU), Rarotonga and Pitcairn Island basaltic rocks. NHRL, Northern Hemisphere Reference Line (Hart, 1984); S-K, Stacey & Kramers crustal evolution curve, with tick marks every 0·5 Ga (Stacey & Kramers, 1975).

the physical link between crust and overlying sediment may be broken. The incompatible element ratios in partial melts of 10% AOC + 90% DM + RS are very similar to those of HIMU + RS and thus this proposed change in end-member composition remains consistent with incompatible element considerations. In summary, the EMI source requires both a lower proportion of the AOC component relative to HIMU and a metamorphosed, low-μ sedimentary component.

CONCLUSIONS

Basaltic rocks from the Cook–Austral and Pitcairn–Gambier island chains (French Polynesia) exhibit the extreme isotopic and trace element compositions of two proposed mantle components, HIMU and EMI. Both components are considered to be ancient, recycled, subducted crustal materials (oceanic crust and oceanic crust plus subducted sediments, respectively). Whereas the averages of certain incompatible elements are different in HIMU- and EMI-type basalts, ratios of certain key elements such as U/Pb, Th/Pb and Nb/La overlap considerably. In OIB from individual islands, U/Pb and Th/Pb increase from tholeiites towards more undersaturated basalts because during melting U and Th are more incompatible than Pb. Both the overlap in trace element ratios and the significant differences in Pb isotopic compositions in these basalt types can be explained by variable degrees of melting of mantle sources including differing proportions of depleted mantle, altered oceanic crust and devolatilized–metamorphosed pelagic sediments. The EMI-type source apparently includes a smaller oceanic crustal component compared with the HIMU-type source (~10% vs 25%, respectively), as well as a small proportion (<2%) of metasedimentary component. The effect of subduction-related devolatilization and metamorphism on the composition of pelagic sediments is an important factor in modelling of the isotopic evolution of the subducted metasedimentary component: U/Pb and Th/Pb in the sediment are first increased by devolatilization at shallow depths but subsequently decreased by deeper metamorphism, whereas Rb/Sr is decreased and Sm/Nd is largely unaffected by both processes.

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