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# Temporal evolution of a Polynesian hotspot: New evidence from Raivavae (Austral islands, South Pacific ocean)

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*Key-words.* – Alkali basalts, Basanites, Tholeiites, Unspiked K/Ar ages, HIMU, Hotspot volcanism, Australs, French Polynesia, South Pacific.

*Abstract.* – A new geological mapping of Raivavae island, Austral-Cook linear chain, French Polynesia, combined with 10 unspiked K/Ar ages measured on its lavas, shows that it was built during two successive volcanic phases : 10.6-7.4 Ma (dykes crosscutting Rairua submarine breccias and younger subaerial Rairua flows) and 6.4-5.4 Ma (Anatonu shield volcano and associated trachytic and phonolitic domes and plugs). Geochemical data from the present study and a previous one [Lassiter *et al.*, 2003] demonstrate that the Rairua alkali basalts, picrobasalts and basanites are more enriched in incompatible elements (especially Th and Nb) than the predominantly tholeiitic Anatonu basalts. The isotopic signature of Rairua lavas displays a strong HIMU flavour, while that of Anatonu lavas is more subdued and intermediate between DMM, HIMU and EM end-members. Rairua mafic lavas show obvious petrologic and geochemical similarities with those of the neighbouring island of Tubuai. Both could result from the partial melting of a predominantly HIMU secondary plume, which formed successively Mangaia (19.4-18.4 Ma), the old lavas of Rurutu (12.7-12.1 Ma), Tubuai (10.0-8.8 Ma) and Rairua volcano. The geochemical signature of the younger Anatonu lavas is ascribed to the partial melting, within the same plume, of a distinct filament of more subdued composition. Alternatively, the proportion of pyroxenites with a HIMU character was lower as partial melting degrees increased, generating the Anatonu tholeiites.

## Évolution temporelle d'un point chaud polynésien : l'exemple de Raivavae (îles Australes, Pacifique Sud)

*Mots-clés.* – Basaltes alcalins, Basanites, Tholéiites, Ages K/Ar sans traceur, HIMU, Point chaud (volcanisme de), Australes, Polynésie française, Pacifique Sud.

*Résumé.* – La cartographie géologique de l'île de Raivavae (archipel des Australes, Polynésie française), associée à 10 datations nouvelles de ses laves par la méthode K/Ar sans traceur, permet de montrer qu'elle s'est édifiée en deux épisodes distincts : 10,6-7,4 Ma (dykes recoupant les brèches sous-marines de Rairua et coulées subaériennes de Rairua plus récentes) et 6,4-5,4 Ma (volcan bouclier d'Anatonu et protrusions trachytiques et phonolitiques associées). Les données géochimiques nouvelles, associées aux résultats antérieurs [Lassiter *et al.*, 2003], montrent que les basaltes alcalins, les picrobasaltes et les basanites de Rairua sont enrichis en éléments incompatibles (et notamment en Th et Nb) par rapport aux basaltes principalement tholéiitiques d'Anatonu. La signature isotopique des laves de Rairua est de type HIMU franc, alors que celle des laves d'Anatonu, plus atténuée, est intermédiaire entre les pôles mantelliques DMM, HIMU et EM. Les laves de Rairua sont très similaires, du point de vue pétrologique et géochimique, à celles de l'île voisine de Tubuai. Toutes deux sont susceptibles de résulter de la fusion partielle du sommet d'un panache secondaire à forte dominante HIMU, qui aurait construit successivement Mangaia (19,4-18,4 Ma), l'édifice ancien de Rurutu (12,7-12,1 Ma), Tubuai (10,0-8,8 Ma) et le volcan de Rairua. L'origine de la signature atténuée des laves tholéiitiques plus récentes d'Anatonu peut être attribuée à la fusion partielle, au sein du même panache secondaire, d'un autre filament à signature géochimique différente ; elle est également susceptible de provenir de la « dilution » de l'empreinte HIMU liée à des pyroxénites mantelliques lors de l'augmentation des degrés de fusion partielle.

## INTRODUCTION

Plume-related basalts are characterized by considerable isotopic heterogeneity [Hart, 1988], which mimics that of their mantle sources, because isotopic ratios are not affected by partial melting processes. In Polynesia, Vidal *et al.* [1984] were the first to describe lavas with very high  $^{206}\text{Pb}/^{204}\text{Pb}$  ratios from Tubuai island (Austral-Cook archipelago; fig. 1a). Other occurrences were soon after documented from the same linear chain in Rurutu, Rimatara and Mangaia [Palacz and Saunders, 1986] and much later in Raivavae [Lassiter *et al.*, 2003]. Hart [1984] and Zindler and Hart [1986] labelled this new end-member HIMU (for high  $\mu$ ,  $\mu$  referring to the  $^{238}\text{U}/^{204}\text{Pb}$  ratio). These findings triggered a number of geochemical studies on Austral-Cook lavas (no less than 32 papers from 1984 to 2011), with the common aim of understanding the meaning of their unusual features for the heterogeneity of the Earth's mantle and the origin of mantle plumes.

Comparatively, little attention was paid to the spatial and temporal distribution of the HIMU signature in the Austral-Cook chain. The most extreme, with  $^{206}\text{Pb}/^{204}\text{Pb}$  ratios ranging from 21.4 to 22.0, is found in the oldest island of the group, Mangaia (19.4-18.4 Ma; fig. 1b). From that point, it becomes more attenuated with time and towards the southeastern Macdonald hotspot [Chauvel *et al.*, 1992, 1997], as shown by the slightly less radiogenic lead isotopic compositions of lavas from Rimatara, the old volcanic event in Rurutu (12.7-12.1 Ma), and Tubuai (10.0-8.8 Ma). Quaternary lavas from Rurutu (1.1-1.0 Ma) and the potentially active Arago seamount [Bonneville *et al.*, 2002, 2006] display a more subdued HIMU signature. In the case of Rurutu, the origin of the latter has been attributed to the contamination of young magmas by the overlying oceanic lithosphere, which acquired its HIMU signature through interaction with the older plume-derived melts [Chauvel *et al.*, 1997].

Raivavae lies SE of Rurutu and Tubuai (fig. 1a), and was dated at around 7-6 Ma [Duncan and McDougall, 1976; Diraison, 1991]. Lassiter *et al.* [2003] identified two groups of lavas, one of them displaying a strong HIMU flavor ( $^{206}\text{Pb}/^{204}\text{Pb} = 20.7-21.3$ ), and the second one a more subdued signature with an increased contribution of the Depleted MORB Mantle (DMM) end-member. Although they provided the GPS positions of their samples, Lassiter *et al.* [2003] did not investigate their geological positions nor their ages. The geological mapping of Raivavae [Guille *et al.*, 2011], conducted by BRGM, CEA, CNRS and the University of Brest, allowed us to collect a new set of samples which have been dated by the unspiked  $^{40}\text{K}$ - $^{40}\text{Ar}$  method and analyzed for major and trace elements as well as Sr, Nd, Pb and Hf isotopes [Maury *et al.*, 2011]. In the present manuscript, we show that the edification of Raivavae resulted from two distinct volcanic pulses, and we discuss the implications of this result on the spatial and temporal evolution of the Austral-Cook hotspot lavas.

## THE AUSTRAL-COOK VOLCANIC CHAIN AND RAIVAVAE ISLAND

### Geological setting and previous interpretations

The Austral-Cook volcanic chain (fig. 1a) extends over 2,300 km from Macdonald seamount to Aitutaki, and like

the smaller Society and Pitcairn-Gambier-Mururoa linear chains it follows an ESE-WNW direction (N65°W) with an active seamount at the SE end. The alignment is related to the 11 cm/yr motion of the Pacific plate over the hotspot [Duncan and McDougall, 1976]. However, the Austral-Cook chain displays many unusual features besides its uncommon length and the presence of several old volcanic islands. In particular, the relationship between age and distance to Macdonald (fig. 1b) is abnormal. In this plot the single linear trend normally observed for hotspot chains, e.g. the Society [Guillou *et al.*, 2005], Pitcairn-Gambier [Guillou *et al.*, 1994] and Hawaii-Emperor [Clague and Dalrymple, 1987] does not occur. Indeed, young volcanic activity (< 2 Ma) is observed not only at the SE end of the chain but also in its central part (Rurutu-Arago) and even at its NW end (Rarotonga and Aitutaki islands, Cook islands; fig. 1b).

The Austral-Cook chain is cut in the middle by a major lithospheric discontinuity, the Austral fracture zone (AFZ; fig. 1a) which represents a fossil transform fault linked to the Pacific-Farallon oceanic ridge (the ancestor of the presently active East Pacific Rise or EPR). The area was the site of older hotspot activity, as evidenced by the 34-22 Ma-old Ngatemato and Taukina seamount chains located southeast of the Austral islands [McNutt *et al.*, 1997] and some other 58 to 20 Ma-old seamounts [Clouard *et al.*, 2003; Clouard and Bonneville, 2004; Adam *et al.*, 2005; Bonneville *et al.*, 2006]. The K/Ar ages of the youngest of these seamounts (2.6 Ma for ZEP2-12, west of Rimatara and 8.8 Ma for ZEP2-19, south of Raivavae, fig. 1a) coincide with the activity of the Arago and Macdonald hotspots, respectively (fig. 1b).

The origin of the Austral-Cook chain is still a matter of debate. Turner and Jarrard [1982] interpreted the chain as a hot line, but random activity along that is not observed (fig. 1b). Duncan and McDougall [1976] and Chauvel *et al.* [1997] proposed that it corresponds to a single linear chain built up by three coeval hotspots located below Macdonald, Arago-Rurutu and SW of Rarotonga, respectively. This hypothesis accounts for most of the spatial and temporal patterns shown in figure 1, but not for the almost contemporaneous volcanic activities in Raivavae and Rapa (fig. 1b). Other authors [Diament and Baudry, 1987; Bonneville *et al.*, 2002, 2006; Adam and Bonneville, 2008] distinguished two separate volcanic lines: a northern one from Aitutaki to Président Thiers Bank located 200 km SE of Raivavae, and a southern one including Rarotonga, Mangaia, Rapa, Marotiri and the Macdonald seamount (fig. 1a). This hypothesis does not explain the ca. 1,500 km gap occurring between Mangaia and Rapa, nor the unusual HIMU signature shared by old lavas from Mangaia, Rurutu, Tubuai and Raivavae. Finally, a last set of authors [McNutt *et al.*, 1997; Koppers *et al.*, 2003; Lassiter *et al.* 2003] considered that volcanic activity in the area was mainly controlled by the regional tectonic regime [Hieronymus and Bercovici, 2000; Davies and Bunge, 2006] rather than by plume/hotspot activity.

### Geology of Raivavae

Raivavae (147°40'W and 23°52'S) is a 16 km<sup>2</sup> volcanic island that culminates at 438 m at Mt Hiro and is rimmed by a typical barrier reef delineating a 2 to 3 km wide lagoon (fig. 2). Unlike most of the other Austral islands, Raivavae

has an elongated shape trending WSW-ENE, parallel to the nearby AFZ which lies only ca. 35 km to the NW (fig. 1a). It represents the emerged remnants of a much larger volcano, which has collapsed generating two large submarine debris avalanches [Clouard and Bonneville, 2004]. A 50 km long submarine ridge prolongates the island towards the southwest [Adam and Bonneville, 2008]. Gravimetric data [Clouard *et al.*, 2000] document positive residual isostatic anomalies centered on this ridge, which might correspond to a set of mafic intrusions or to the summit of a magma reservoir. The salient geomorphological feature of Raivavae is

a continuous 200 to 400 m high north-facing cliff, connecting Mt Maunanui, Mt Araua, Mt Hiro, Mt Mouatapu and Mt Turivao (fig. 2). The simplified geological map of Raivavae (fig. 2), drawn from the results of a mapping program conducted from 2005 to 2010 [Guille *et al.*, 2011; Maury *et al.*, 2011], shows two distinct (and superimposed) volcanoes: (i) the old Rairua volcano, which was not identified before the present study; and (ii) the younger, although dissected, Anatonu shield volcano built of thick lava flow piles that are crosscut and/or overlain by trachytic and phonolitic domes and plugs.

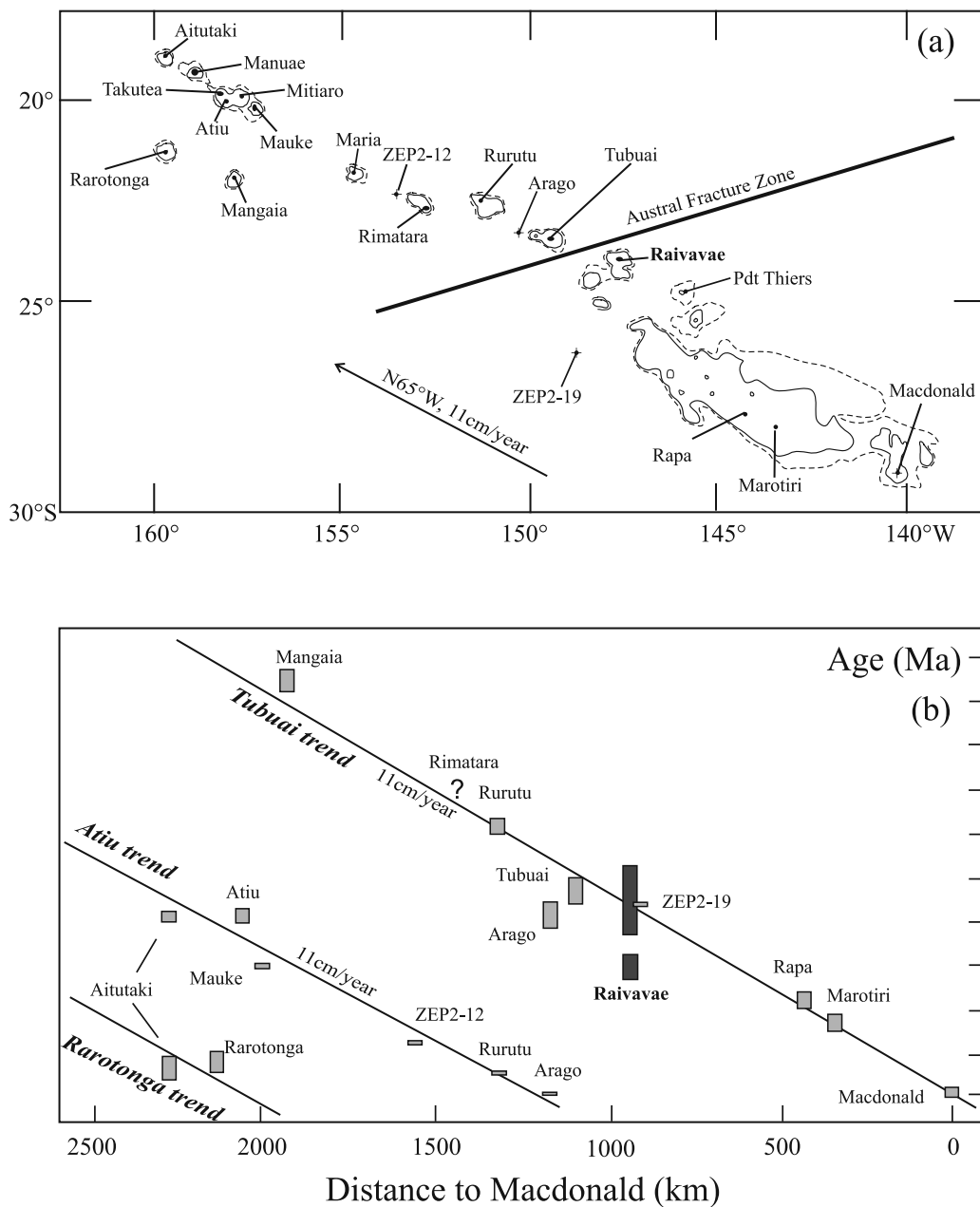


FIG. 1. – The Austral-Cook linear chain, modified after Chauvel *et al.* [1997]. (a) Location and bathymetry. The dotted lines correspond to the 2,400 fathoms (i.e. 4,400 m) isobath, and the full lines to the 2,000 fathoms (3,650 m) isobaths. Crosses indicate the locations of young seamounts, and the arrow corresponds to the Pacific plate motion. Pdt Thiers: Président Thiers Bank. (b) Plot of age versus distance to the Macdonald seamount. Diagram (b) is adapted from Diraison [1991] and includes data from Dalrymple *et al.* [1975], Duncan and McDougall [1976], Turner and Jarrard [1982], Matsuda *et al.* [1984], Diraison [1991], Maury *et al.* [1994, 2000], McNutt *et al.* [1997], Bonneville *et al.* [2002, 2006] and this work (Raivavae). When available, ages measured on groundmass or on separated phases have been preferred to those measured on bulk rocks [Laughlin *et al.*, 1994].

The oldest exposed unit of Rairua volcano is a more than 200 m thick pile of submarine autoclastic basaltic breccias, usually highly altered. They are crosscut by a dense network of picrobasaltic, basanitic and alkali basaltic dykes, most of which trend WSW-ENE, parallel to the AFZ and the elongation of the island. The Rairua breccias outcrop along the axis of the island, and are overlain by either (i) the Rairua basanitic flows which form the Mt Tavaetu plateau in the western part of Raivavae; (ii) Miocene bioclastic limestones (ms in fig. 2) near Anatonu village; and more often (iii) the younger plateaus of Anatonu volcano, e.g. below Mt Maunanui, Mt Taraia and Mt Matotea.

The Anatonu shield volcano is mostly composed of an up to 400 m thick pile of basaltic and intermediate lava flows. These form six gently dipping plateaus, which are, by order of decreasing size, those culminating at Mts Hiro, Taraia, Matotea, Mouatapu, Taamora, and Tapioi, respectively (fig. 2). The variations of the dips and the thickness of the Mt Hiro plateau pile along the northern cliff of the island suggest that these flows were emplaced into a paleodepression dug into the Rairua breccias. Other witnesses of Anatonu volcanic activity are six phonolitic and trachytic domes and plugs, which crosscut either the Rairua breccias (Mt Rareterepa trachytic plug) or the Anatonu shield plateaus (e.g. Mt Turivao). They define a WSW-ENE trend, from Mt Rareterepa to the two phonolitic plugs of Motu Hatuatua at the eastern edge of the island. This trend is parallel to the AFZ and to the main direction of the dykes crosscutting the Rairua breccias. All these features suggest that a deep fracture parallel to the nearby AFZ may have channeled the ascending Raivavae magmas and closely controlled their emplacement.

## Geochemical results

### K/Ar ages

Unspiked K/Ar whole-rock ages (table I) were obtained on 10 samples at the Laboratoire des Sciences du Climat et de l'Environnement (LSCE) in Gif-sur-Yvette. The samples were crushed, sieved to 0.250-0.125 mm size fraction and ultrasonically washed in acetic acid. Potassium and argon were measured on the microcrystalline groundmass, after removal of olivine, clinopyroxene and feldspar phenocrysts using heavy liquids and magnetic separations. This process removes at least some potential sources of systematic analytical error due to the presence of excess  $^{40}\text{Ar}$  in olivine and feldspar phenocrysts [Laughlin *et al.*, 1994]. The K content of the separated groundmass was measured by ICP-AES in Brest using the same method as described below for major elements while Ar analyses were performed in Gif-sur-Yvette using the procedures detailed in Yurtmen *et al.* [2002] and Guillou *et al.* [2004]. Ages were calculated using the constants recommended by Steiger and Jäger [1977].

Previous K/Ar ages measured on whole rocks from the Anatonu volcano [Duncan and McDougall, 1976; Diraison, 1991] cluster around 6.4 Ma (ages shown in italics in fig. 2). They are in agreement with the new unspiked groundmass ages on Anatonu lava flows and domes (table I and fig. 2), most of which range from  $6.42 \pm 0.14$  Ma to  $5.92 \pm 0.13$  Ma. The only younger Anatonu unit is the Tapioi plateau, the basal and summital flows of which have been dated to  $5.74 \pm 0.12$  Ma and  $5.36 \pm 0.11$  Ma, respectively.

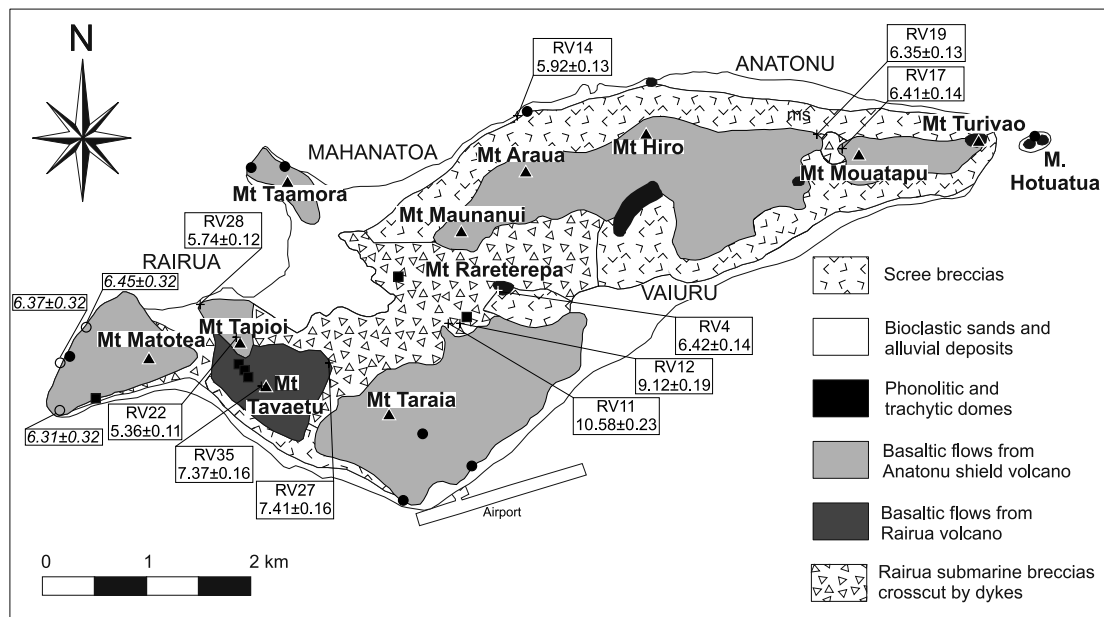


FIG. 2. – Geological sketch map of Raivavae, simplified from Guille *et al.* [2011]. Miocene sediments outcropping beneath the scree breccias near Anatonu are denoted by the letters ms. Filled squares indicate the location of HIMU-type lavas according to GPS data from Lassiter *et al.* [2003], and filled circles indicate the location of the second group lavas according to these authors. K/Ar ages from the present study are shown in bold numbers, and those from previous studies [Duncan and McDougall, 1976; Diraison, 1991] in italics.

TABLE I. – New unspiked K/Ar ages of Raivavae lavas. See text for analytical methods. Petrographic types from figure 3.

Sample Experiment number	Occurrence and type	K (wt. %) $\pm 2\sigma$	Mollen mass (g)	$^{40}\text{Ar}^*\%$	$40\text{Ar}^* (10^{-12} \text{ mol/g}) \pm 1\sigma$	Weighted mean $40\text{Ar}^* (10^{-12} \text{ mol/g}) \pm 1\sigma$	Age (Ma) $\pm 2\sigma$
<b>RV22</b>	Anatonu						
8154	tholeiitic	0,852 $\pm$ 0,009	1,01230	21,828	0,795 $\pm$ 0,004		
8170	flow	« ..... »	1,06068	21,764	0,791 $\pm$ 0,004	0,793 $\pm$ 0,003	5,36 $\pm$ 0,11
<b>RV28</b>	Anatonu						
8155	tholeiitic	0,778 $\pm$ 0,008	1,05550	41,105	0,776 $\pm$ 0,004		
8171	flow	« ..... »	1,00587	33,768	0,774 $\pm$ 0,004	0,775 $\pm$ 0,003	5,74 $\pm$ 0,12
<b>RV14</b>	Anatonu						
7794	picrobasaltic	0,706 $\pm$ 0,007	1,25403	45,535	0,713 $\pm$ 0,004		
7810	dyke	« ..... »	1,29737	63,562	0,739 $\pm$ 0,004	0,726 $\pm$ 0,003	5,92 $\pm$ 0,13
<b>RV19</b>	Anatonu						
7630	benmoreitic	2,067 $\pm$ 0,021	0,97704	42,966	2,284 $\pm$ 0,011		
7646	flow	« ..... »	0,95878	63,387	2,274 $\pm$ 0,011	2,279 $\pm$ 0,008	6,35 $\pm$ 0,13
<b>RV17</b>	Anatonu						
7629	benmoreitic	2,084 $\pm$ 0,021	1,02693	34,332	2,292 $\pm$ 0,012		
7645	flow	« ..... »	1,11624	16,881	2,356 $\pm$ 0,012	2,322 $\pm$ 0,008	6,41 $\pm$ 0,14
<b>RV4</b>	Raraterepa						
7639	trachytic	3,579 $\pm$ 0,036	1,07150	72,436	4,006 $\pm$ 0,020		
7652	dome	« ..... »	1,10875	76,533	3,974 $\pm$ 0,020	3,990 $\pm$ 0,014	6,42 $\pm$ 0,14
<b>RV35</b>	Rairua						
8249	basaltic	0,731 $\pm$ 0,008	1,00763	27,773	0,931 $\pm$ 0,005		
8281	flow	« ..... »	0,50761	29,237	0,942 $\pm$ 0,005	0,937 $\pm$ 0,003	7,37 $\pm$ 0,16
<b>RV27</b>	Rairua						
8153	picrobasaltic	0,755 $\pm$ 0,008	0,50840	18,677	0,971 $\pm$ 0,005		
8169	flow	« ..... »	0,51307	23,612	0,974 $\pm$ 0,005	0,973 $\pm$ 0,004	7,41 $\pm$ 0,16
<b>RV12</b>	Rairua						
7745	alkali basaltic	0,714 $\pm$ 0,007	0,92891	31,721	1,116 $\pm$ 0,006		
7761	dyke	« ..... »	0,97620	31,223	1,149 $\pm$ 0,006	1,132 $\pm$ 0,004	9,12 $\pm$ 0,19
<b>RV11</b>	Rairua						
7628	basaltic	0,697 $\pm$ 0,007	1,11331	15,313	1,278 $\pm$ 0,006		
7644	flow	« ..... »	0,46078	15,198	1,289 $\pm$ 0,007	1,283 $\pm$ 0,005	10,58 $\pm$ 0,23

The ages of the Rairua volcano, not measured before the present study, are clearly older (table I and fig. 2). Two fresh basaltic dykes crosscutting the Rairua breccias yielded unspiked K/Ar ages of  $10.58 \pm 0.23$  Ma and  $9.12 \pm 0.19$  Ma, respectively. The basal and summital flows of the Tavaetu plateau lava pile provided significantly younger unspiked K/Ar ages of  $7.41 \pm 0.16$  Ma and  $7.37 \pm 0.16$  Ma,

respectively. The end of the building of Rairua volcano was therefore separated from the start of that of Anatonu volcano by a ca. 1 m.y. long time gap.

### Major elements and petrographic types

For chemical analysis, all samples were also crushed using an agate mortar. Major and trace element data (table II) were obtained on 30 samples by Inductively Coupled Plasma-Atomic Emission Spectrometry (ICP-AES) at IUEM, Plouzané using the method described by Cotten *et al.* [1995]. The international standards used for calibration were ACE, BEN, JB-2, PM-S and WS-E and the relative standard deviations are  $\pm 1\%$  for  $\text{SiO}_2$ , and  $\pm 2\%$  for other major elements except  $\text{P}_2\text{O}_5$  and  $\text{MnO}$  (absolute precision  $\pm 0.01\%$ ). Major element data for dated samples from table II and additional analyses from Maury *et al.* [2011] have been plotted in a total alkali-silica (TAS) diagram (fig. 3). We included in this diagram previous data [Diraison, 1991; Lassiter *et al.*, 2003] on samples collected from lava flows and dykes (boulders excluded), whose locations are also shown in figure 2. Lavas from the old Rairua volcano are all silica-undersaturated and range from foïdites (analcitites, i.e. analcite-bearing tephrites similar to those described from the nearby island of Tubuai by Caroff *et al.* [1997]), to basanites, alkali basalts and picrobasalts. The younger Anatonu shield volcano is built of plagioclase-phyric tholeiitic basalt flows, with occasional intercalated alkalic

TABLE II. – Major and trace element analyses of dated Raivavae lavas (major oxides in wt.%, trace elements in ppm). Data obtained by ICP-AES, analytical method described in the text.  $\text{Fe}_2\text{O}_3^*$ : total Fe as  $\text{Fe}_2\text{O}_3$ . The chemical analyses of the full set of Raivavae samples shown in figure 3 are given in Maury *et al.* [2011]. Petrographic types from figure 3. bas: basanite; piba: picrobasalt; alba: alkali basalt; thol: tholeiite; ben: benmoreite; tra: trachyte.

Sample Type Location	RV35 bas Rairua	RV27 piba Rairua	RV11 bas Rairua	RV12 alba Rairua	RV14 piba Anatonu	RV28 thol Anatonu	RV22 thol Anatonu	RV17 ben Anatonu	RV19 ben Anatonu	RV4 tra dome
$\text{SiO}_2$	43.70	43.86	44.60	45.50	44.65	46.93	47.80	54.50	55.10	58.90
$\text{TiO}_2$	2.41	2.07	2.52	2.29	1.81	3.15	3.18	1.14	1.26	0.37
$\text{Al}_2\text{O}_3$	15.50	13.53	15.60	13.16	10.32	14.98	14.83	16.00	16.20	17.00
$\text{Fe}_2\text{O}_3$	13.50	12.45	12.70	12.60	13.55	15.38	15.44	11.1	11.37	7.72
MnO	0.21	0.20	0.18	0.18	0.17	0.18	0.20	0.31	0.29	0.20
MgO	6.40	10.45	6.40	10.53	17.26	4.26	4.58	1.49	1.36	0.40
CaO	12.26	12.01	12.40	11.20	9.29	8.88	8.89	4.34	4.28	1.72
$\text{Na}_2\text{O}$	2.62	2.17	2.63	2.45	1.74	3.08	3.66	5.78	6.60	6.95
$\text{K}_2\text{O}$	0.70	0.70	0.77	0.66	0.58	0.86	0.94	2.66	2.54	4.33
$\text{P}_2\text{O}_5$	0.41	0.31	0.34	0.30	0.27	0.41	0.43	0.44	0.48	0.11
LOI	1.70	1.73	1.54	0.98	-0.06	1.08	0.03	2.05	0.30	1.59
Total	99.56	99.47	99.68	99.85	99.58	99.19	99.99	99.81	99.78	99.29
Q	0.00	0.00	0.00	0.00	0.00	0.97	0.00	0.00	0.00	0.00
ne	4.37	3.05	4.15	2.17	0.29	0.00	0.00	0.52	4.25	3.98
hy	0.00	0.00	0.00	0.00	0.00	13.19	6.73	0.00	0.00	0.00
ol	9.77	16.05	11.61	19.42	34.28	0.00	4.14	9.55	8.64	6.19
ac	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Sc	26.0	32.5	30.0	32.5	26.0	25.4	25.1	11.0	12.0	3.5
V	329	282	340	277	212	347	378	2.0	2.5	6.0
Cr	35	461	95	415	720	24	19	3	3.5	8
Ni	60	197	79	240	555	33	27	1.5	1.5	3.5
Co	48	55	48	56	85	43	43	2	3	1
Rb	21.0	9.5	22.0	18.2	12.4	11.6	28.2	65.5	50.5	81.0
Sr	582	457	515	440	415	447	440	677	599	159
Y	29.0	22.6	26.5	22.0	15.7	35.6	35.9	54.5	55.0	57.0
Zr	190	146	173	147	130	193	206	465	450	662
Nb	72.6	56.7	57.5	41.0	28.0	39.1	41.7	108.5	104.0	144.0
Ba	325	265	280	214	146	183	194	1485	1418	2860
La	52.6	40.5	41.0	28.0	20.0	31.0	35.2	89.0	90.0	101.0
Ce	99.9	76.0	80.0	59.0	43.0	62.1	68.3	180.0	180.0	186.0
Nd	43.9	34.1	37.0	28.0	21.0	36.3	41.4	84.0	85.0	83.0
Sm	7.7	6.34	7.1	5.7	4.65	8.19	9.38	15.8	16.4	15.4
Eu	2.3	1.93	2.11	1.84	1.5	2.55	2.83	5.57	5.8	4.47
Gd	6.7	5.35	6.2	5.2	4.1	8.25	8.97	13.55	14.5	12.2
Dy	5.4	4.39	5.2	4.4	3.15	6.51	7.22	10.9	10.9	11.00
Er	3.00	2.10	2.4	1.9	1.35	3.34	3.45	5.00	4.8	5.2
Yb	2.4	1.88	2.08	1.7	1.1	2.64	2.71	4.17	4.12	4.92
Th	4.9	4.83	4.55	3.7	2.65	3.05	3.47	9.7	9.45	13.6

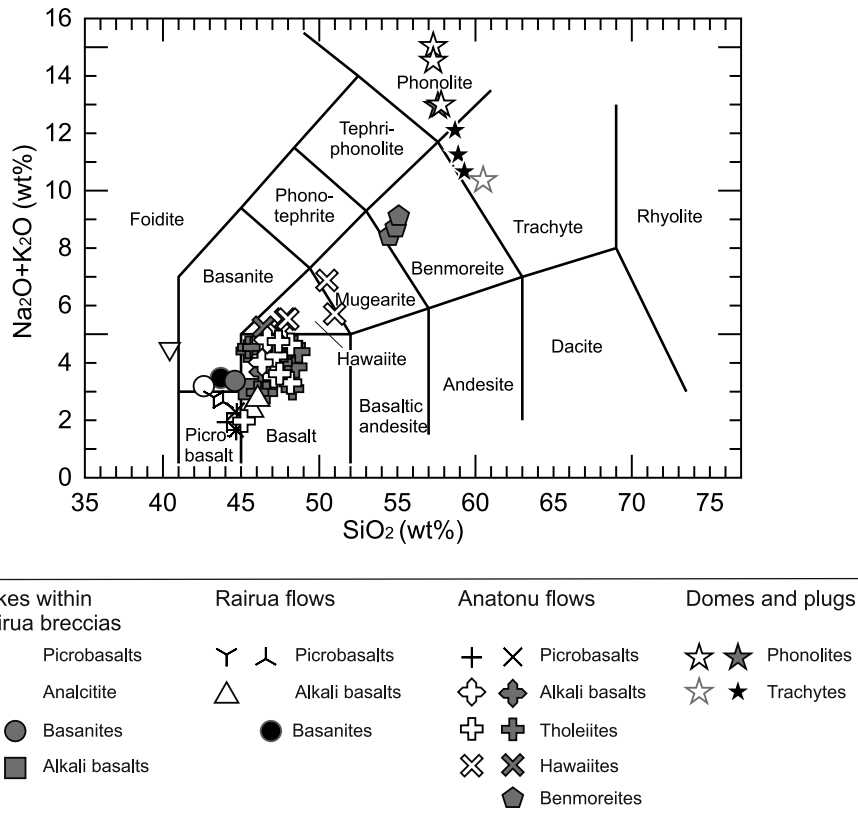


FIG. 3. – Total alkalis-silica (TAS) plot for Raivavae lavas, adapted from Le Bas *et al.* [1986]. Data from this study (table II and Maury *et al.* [2011]) are shown as filled symbols (right column for each group), and data from the literature [Diraison, 1991; Lassiter *et al.*, 2003] as empty symbols (left column). Alternative names: nephelinite = foidite; hawaiiite = trachybasalt; mugearite = basaltic trachyandesite; benmoreite = trachyandesite.

basaltic and intermediate (hawaiitic and benmoreitic) flows. The evolved lava domes and plugs crosscutting these flows are made up of non-peralkaline trachytes and phonolites.

Rairua mafic lavas are highly porphyritic, with up to 40 modal% (in the picrobasalts) calcic clinopyroxene and olivine phenocrysts, the latter containing Cr-spinel inclusions, and minor (< 2 modal%) amounts of calcic plagioclase, all of them set in a microlitic to subdoleritic groundmass containing clinopyroxene, plagioclase, olivine, titanomagnetite, and interstitial alkali feldspar, nepheline and analcite. As a whole, their petrographic and mineralogical features are rather similar to those of Tubuai lavas [Maury *et al.*, 1994; Caroff *et al.*, 1997], although they do not include nephelinites.

Anatonu tholeiitic basalts are less porphyritic, with < 10 modal% phenocrysts among which calcic plagioclase (3-7 modal%), often occurring as large (> 5 mm) crystals, dominates over olivine (1-4 modal%) and occasional calcic clinopyroxene and titanomagnetite phenocrysts. Feldspatoids are absent from their groundmass. Hawaiites, mugearites and benmoreites are subaphyric, with rare phenocrysts of plagioclase, titanomagnetite and apatite. Trachytes and phonolites contain less than 10 modal% phenocrysts, which include anorthoclase, green augite, titanomagnetite, apatite and titanite.

In plots of major element oxides *versus* SiO<sub>2</sub> [Maury *et al.*, 2011], Anatonu lavas define a typical OIB series marked by: a Daly gap at the level of mugearites; an increase of K<sub>2</sub>O/Na<sub>2</sub>O ratios from basalts and hawaiites to benmoreites, trachytes and phonolites; and rather regular decreases of

TiO<sub>2</sub>, total iron, CaO and MgO with SiO<sub>2</sub> while alumina and alkalis contents increase.

### Trace elements and radiogenic isotopes

As for major element data, we have used new trace element data from table II and Maury *et al.* [2011] together with previous data [Diraison, 1991; Lassiter *et al.*, 2003]. In plots of trace elements *versus* Th [Maury *et al.*, 2011], Anatonu lavas display an exponential decrease of the compatible transition elements Sc, Co, Cr and Ni; and a regular increase of most incompatible elements except Sr, Eu and Ba whose contents drop down from benmoreites to trachytes and phonolites.

Primitive mantle-normalized multi-element patterns of mafic Raivavae lavas are shown in figure 4. As a whole, they display the typical characteristics of Polynesian ocean island basalts (OIB), with strong enrichments in the most incompatible rare earth elements (REE) and high field strength elements (HFSE) culminating for Nb, and relative depletions in Rb, Ba and K [Dupuy *et al.*, 1988, 1989; Dostal *et al.*, 1998]. Anatonu basalts display sub-parallel patterns. Mafic samples from the Rairua volcano (dykes crosscutting the breccias and lava flows) tend to be more enriched in Nb, Th, Ce and La and more depleted in K, Ba and Rb than Anatonu ones (fig. 4). The shapes of their multi-element patterns are rather similar to those of Tubuai mafic lavas [Caroff *et al.*, 1997; envelopes shown in fig. 4], although no lava equivalent to the strongly enriched Tubuai nephelinites has been sampled from Rairua volcano. Rairua

dykes and flows and most of Tubuai lavas display Nb/K, Nb/Zr, Th/K and Th/Zr ratios higher than Anatonu mafic rocks (fig. 5).

Numerous Sr, Nd, Pb and Hf isotopic analyses of Raivavae lavas were published by Lassiter *et al.* [2003]. The analyses of twelve new samples, including four Rairua lavas, are given by Maury *et al.* [2011]. They display large variations, especially for Pb. Indeed,  $^{206}\text{Pb}/^{204}\text{Pb}$  ratios range from 19.3 to 21.3 (fig. 6), Rairua mafic samples being more radiogenic than Anatonu lavas.

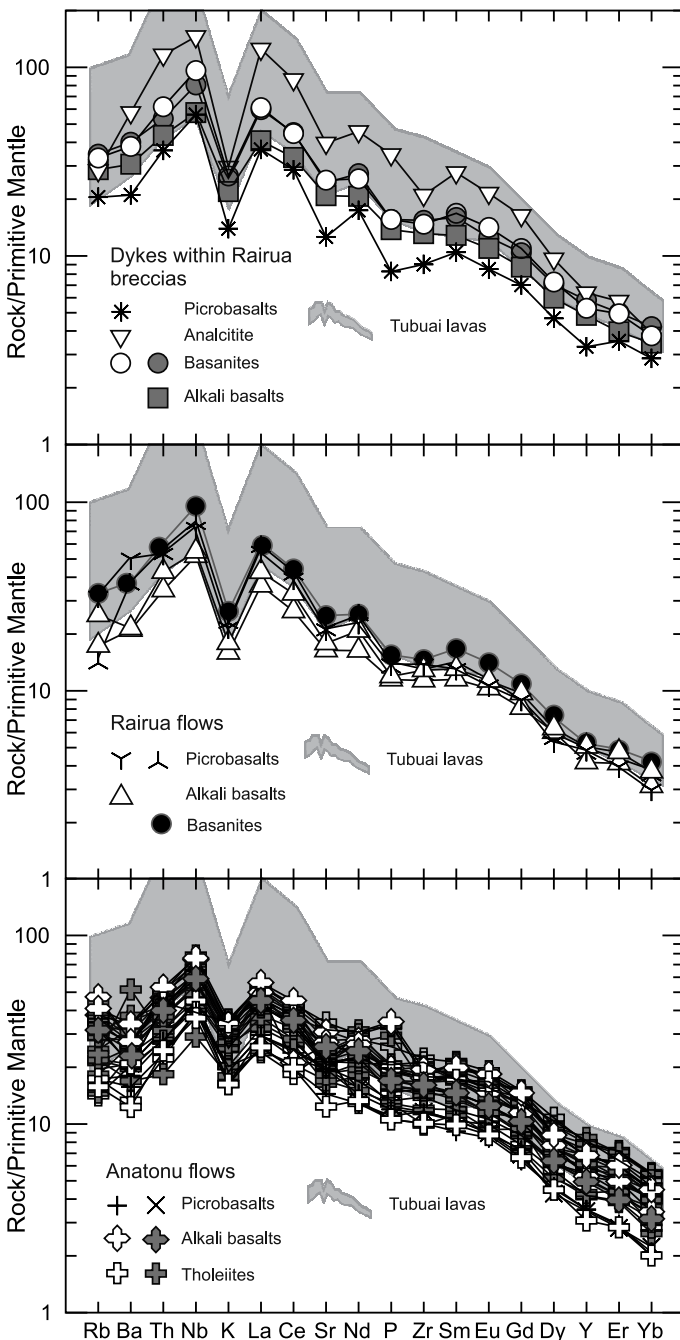


FIG. 4. – Normalised multi-element plots for Rairua and Anatonu mafic lavas. Primitive mantle composition from Sun and McDonough [1989]. Symbols as in figure 3. The outline of Tubuai mafic lavas patterns [Caroff *et al.*, 1997] is shown for comparison (grey field).

## DISCUSSION

### Temporal geochemical evolution of Raivavae lavas

Because of their variable enrichments in incompatible elements (fig. 4), the two groups of Raivavae lavas can be readily distinguished in figure 5, where Rairua mafic lavas display Nb/K ( $> 0.006$ ), Nb/Zr ( $> 0.27$ ), Th/Zr ( $> 0.023$ ) and Th/K ( $> 0.0005$ ) ratios consistently higher than those of Anatonu lavas. The stronger Nb, Th and Light REE (LREE) enrichment of Rairua alkali basalts, picrobasalts, basanites and analcites is typical of Austral HIMU-type silica-undersaturated mafic rocks ranging from alkali basalts to nephelinites [Chauvel *et al.*, 1992, 1997; Caroff *et al.*, 1997; Dostal *et al.*, 1998]. Indeed, the fields of the latter and the locations of Rairua mafic lavas partly overlap in figure 5 plots.

The two groups (Rairua and Anatonu, respectively) distinguished above also show contrasted isotopic (Sr, Nd, Pb, Hf) compositions [Maury *et al.*, 2011]. Especially, the group with a strong HIMU flavour ( $^{206}\text{Pb}/^{204}\text{Pb} > 20.5$ ) defined by Lassiter *et al.* [2003] corresponds to Rairua samples. Indeed, the GPS locations of all the HIMU samples studied by Lassiter *et al.* (filled squares in fig. 2) indicate that they were collected either from dykes crosscutting Rairua breccias or from the Tavaetu plateau [Guille *et al.*, 2011; Maury *et al.*, 2011]. Accordingly, the two dated dykes RV11 and RV12 intruding Rairua breccias as well as the basal (RV27) and summital (RV35) flows of the Tavaetu plateau display similar HIMU-type isotopic signatures, with  $^{206}\text{Pb}/^{204}\text{Pb}$  ratios ranging from 20.58 to 20.90 [Maury *et al.*, 2011; fig. 6], which fit those of Rairua samples from Lassiter *et al.* [2003].

Conversely, Anatonu lavas closely correspond to the second group of samples defined by Lassiter *et al.* [2003], which show a more subdued isotopic signature intermediate between HIMU, DMM and enriched mantle (EM) end-members with  $^{206}\text{Pb}/^{204}\text{Pb}$  20 (fig. 6). Indeed, their GPS locations (filled circles in fig. 2) indicate that they were collected either from the Anatonu shield or from the associated domes and plugs. Accordingly, a similar isotopic signature (with  $^{206}\text{Pb}/^{204}\text{Pb}$  ratios ranging from 19.30 to 19.92) has been identified from newly sampled Anatonu basalts (RV14, RV22, RV28), benmoreites (RV17, RV19), and trachyte RV4 [Maury *et al.*, 2011; fig. 6]. The two geochemical groups were emplaced during distinct periods: 10.6-7.4 Ma for the HIMU group (Rairua) and 6.4-5.4 Ma for the second one (Anatonu). However, we cannot exclude that volcanic activity persisted during the 1 m.y. gap separating the edification of these two volcanoes. Indeed, the island is small although geologically rather complex, and only represents less than 5% of the total volume of the volcanic edifice [Clouard *et al.*, 2000; Clouard and Bonneville, 2004]. The occurrence of submarine Miocene bioclastic limestones and Rairua submarine breccias at an altitude of ca. 200 m in the eastern part of the island suggest tectonic uplift during this time span [Maury *et al.*, 2011].

Rairua lavas show obvious petrologic and geochemical similarities (figs. 4 and 5) with those of the nearby island of Tubuai located 200 km to the NW [Maury *et al.*, 1994; Caroff *et al.*, 1997]. The emplacement of submarine Rairua breccias (10.6-9.1 Ma) and that of the subaerial Tubuai lava flows with similar multi-element patterns (fig. 4) was



almost contemporaneous. Moreover, the time difference (1.4 to 2.6 m.y.) between this first event and the emission of the slightly less enriched (fig. 4) subaerial Rairua lava flows (7.4 Ma) is consistent with the Pacific plate motion velocity (11 cm/yr) and the 200 km distance between the two islands. In addition, the similarities of Tubuai and Rairua mafic lavas (figs. 4-6) do not support the hypothesis that the Austral fracture zone which lies between the two islands (fig. 1a) would separate two contrasted mantle domains controlling the geochemistry of plume-related lavas in contrast with what was suggested by Lassiter *et al.* in 2003.

Anatonu lavas (6.4-5.4 Ma) define a rather typical OIB series ranging from picrobasalts to trachytes and phonolites, with an apparent Daly gap at the level of mugearites (fig. 3). Their trends in major and trace elements [Maury *et al.*, 2011] and the parallelism of the multi-element patterns of mafic lavas (fig. 4), as well as the fact that the K/Ar ages of Anatonu basalts, benmoreites and trachytes overlap (table I), suggest that this evolution was primarily controlled by fractional crystallization. The edification of Anatonu

volcano ended at 5.4 Ma, i.e. only 0.6 m.y. before the subaerial activity started in Rapa island (4.8-4.1 Ma [Diraison, 1991]) located 530 km SE of Raivavae (fig. 1). Taking into account this distance and the 11 cm/year Pacific plate motion, such a short time span appears inconsistent with the hypothesis of derivation of the two islands from a single “conventional” plume with a ca. 200 km-wide melting zone [Farnetani and Hoffmann, 2010].

### Origin of Raivavae HIMU lavas in the framework of the Austral-Cook chain

Although there is still a debate concerning the origin of HIMU end-member, the most widely accepted opinion is that it comes from the ancient (1 or 2 Ga) incorporation into the mantle of residues of subducted oceanic crust that experienced selective extraction of lead with respect to uranium during the dehydration process linked to their subduction [Chauvel *et al.*, 1992; Kogiso *et al.*, 1997; Kawabata *et al.*, 2011]. Most authors envision that these residues were carried down and stored near the mantle-core boundary until their incorporation into ascending mantle plumes [Hofmann, 1988].

The distribution of the HIMU signature within the Austral-Cook chain is still not well understood. Figure 6 shows that the most extreme HIMU imprint is documented at Mangaia [Woodhead, 1996]. The HIMU signature is still very strong in Tubuai, “old” Rurutu [Chauvel *et al.*, 1992, 1997], and some Raivavae lavas [Lassiter *et al.*, 2003] sampled from the old Rairua volcano, but attenuated in “young” Rurutu lavas [Chauvel *et al.*, 1997]. The HIMU flavor is definitively weaker in the other Cook islands, as well as in Rapa, Marotiri, Macdonald [Hémond *et al.*, 1994] and Raivavae lavas sampled from the young Anatonu volcano (fig. 6). A limit drawn at  $^{206}\text{Pb}/^{204}\text{Pb} = 20.5$  as done by Lassiter *et al.* [2003] separates adequately the two sets of Raivavae lavas, but it also separates the set of “old” lavas (19.4-7.4 Ma) with a strong HIMU imprint (Mangaia, “old” Rurutu, Tubuai, Rairua in Raivavae) from that of younger lavas with more subdued signatures (figs. 1b and 6). In Raivavae, the 1 m.y. long gap documented between the emplacement of Rairua lavas and that of the younger Anatonu volcano, is much shorter than that observed in Rurutu [11 m.y.; Guille *et al.*, 1998], and seems inconsistent with the hypothesis of a passage of Raivavae over two different hotspots. Therefore, the whole volcanic history of Raivavae seems related to the fainting stage of a single predominantly HIMU plume which generated successively Mangaia, Rurutu (old) and Tubuai lavas.

Numerical simulations of the evolution of thermal and thermo-chemical plumes [Farnetani *et al.*, 2002; Farnetani and Samuel, 2005; Farnetani and Hofmann, 2009, 2010] suggest that heterogeneous mantle domains occur in the lowermost mantle feeding the plume [Farnetani *et al.*, 2012]. During the ascent of the latter, they are converted into long-lived elongated and narrow filaments within the plume conduit. Such filaments would melt sporadically, and then eventually communicate their specific geochemical fingerprint to basaltic lavas [Farnetani and Hoffmann, 2009]. Distinct filaments could either melt concomitantly such as in Hawaii [Farnetani and Hoffmann, 2010] and the Marquesas [Chauvel *et al.*, 2012], or alternatively during successive volcanic events. The changes that occurred in Raivavae

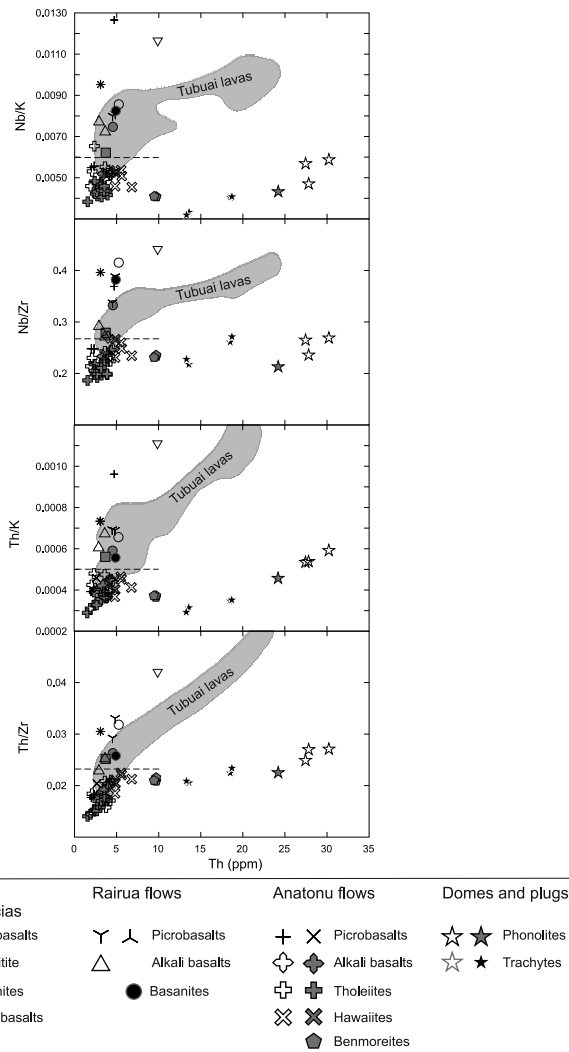


FIG. 5. – Plots of Nb/K, Nb/Zr, Th/K and Th/Zr against Th (ppm), allowing to distinguish mafic Rairua lavas from Anatonu lavas (boundaries shown as dotted lines). The fields for Tubuai mafic lavas [Caroff *et al.*, 1997] are shown for comparison. Symbols as in figure 3.

between 7.4 and 6.4 Ma involved an increase of partial melting degrees illustrated by the evolution of the affinity of the produced lavas from basanitic and alkali basaltic to tholeiitic, and an attenuation of their HIMU signature (fig. 6). This evolution might be ascribed to the contribution of a new filament of different composition (possibly containing a smaller amount of pyroxenites) within a predominantly HIMU plume. Alternatively, the increase of partial melting degrees may have resulted in a “dilution” of the HIMU imprint possibly linked to pyroxenites [Davis *et al.*, 2011], as less fusible peridotitic lithologies would contribute in greater proportions to the melting process. There is no presently identified younger remnant of the activity of this plume, because islands with ages younger than the Anatonu volcano (Rapa, Marotiri) and the active Macdonald seamount have clearly very different Nd and Sr isotopic compositions. This activity could have ended by the building of Président Thiers Bank 200 km SE of Raivavae (fig. 1a), but no fresh volcanic rock has been recovered yet from this submerged edifice [Bonneville *et al.*, 2002, 2006].

### Incidences on hotspot models for the Austral-Cook chain

The unusual concentration of hotspots in Polynesia suggest that they derive from a large-scale structure (“superplume”) storing hot materials [Zhao, 2001] within the lower mantle where it can remain anchored for more than 100 m.y. (type 2 plumes of Courtillot *et al.* [2003]). Analogical [Davaille, 1999; Davaille *et al.*, 2002, 2005] and numerical [Ogawa, 2007] models show that such thermochemical plumes can ascend within the lower mantle in which they form very large oscillating domes of hot materials [Cadio *et al.*, 2011]. A number of much smaller and short-lived plumes (secondary plumes or “plumelets”) can form at the top of the “superplume” and uprise through the asthenospheric upper mantle [Cadio *et al.*, 2011]. Seismic imaging data [Suetsugu *et al.*, 2009] are consistent with the occurrence of such large-scale and secondary plumes beneath the South Pacific. The ascent of secondary plumes or “plumelets” is likely favored by pre-existing discontinuities and zones of weakness in the asthenospheric upper mantle. The track left by a previous plume can certainly create such weaknesses. Numerical models [Sleep, 2008] demonstrate that hot plume materials can be channeled at the top of the asthenosphere. There, they can be guided by the track of an older plume along which the oceanic lithosphere has been thermally eroded. These processes could explain the coeval location in the Austral-Cook chain of several short-lived hotspots along the same track.

The short tracks of Arago (Atiu trend) and Rarotonga (fig. 1b) suggest that the corresponding plumes are young and potentially short-lived “plumelets” [Dickinson, 1998]. Given the existence of a “superplume” in the lower mantle beneath French Polynesia, as evidenced by its very hot character [Zhao, 2001], one may consider that the Macdonald, Arago and Rarotonga “plumelets” emanated from that “superplume”, as short-lived columns of hot mantle materials ascending through the upper mantle [Courtillot *et al.*, 2003; Cadio *et al.*, 2011]. It is thus possible to explain the three coeval present-day hotspots by considering that the oldest plumelet, i.e. the pre-Macdonald HIMU-type one, started to create an ESE-WNW island/seamount chain (the Tubuai trend in fig. 1b) ca. 20 Ma ago (Mangaia), and since then

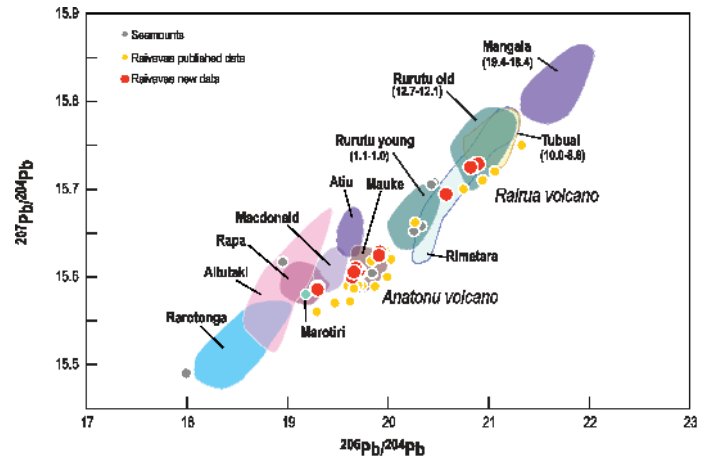


FIG. 6. –  $^{207}\text{Pb}/^{204}\text{Pb}$  versus  $^{206}\text{Pb}/^{204}\text{Pb}$  plot for Raivavae lavas compared to fields defined by lavas from other Austral-Cook islands and seamounts (from the GEOROC data base). Small circles: data from the litterature; large circles: data from Maury *et al.* [2011] on samples described in this paper. Numbers refer to the K/Ar age ranges (Ma) obtained for Mangaia, Rurutu and Tubuai lavas (sources given in figure 1 caption).

left its track in the upper mantle as a weakness zone extending from Mangaia to Raivavae. There, its strong HIMU imprint vanished at 7.4 Ma; between 6.4 and 5.4 Ma, a different source characterized by a more subdued HIMU isotopic signature melted to form Anatonu lavas and then vanished; alternatively, the HIMU imprint linked to pyroxenites was “diluted” as partial melting degrees increased. More or less simultaneously, (i) another plumelet (the present Macdonald plumelet) generated Rapa and then Marotiri lavas, and (ii) the Arago plumelet ascended along the same track beneath Aitutaki and Atiu (fig. 1), creating the Atiu trend. Finally, they were later followed, at less than 4 Ma, by the still poorly known Rarotonga plumelet.

### CONCLUSIONS

The geological mapping of Raivavae, combined with unspiked K/Ar age measurements, shows that it was built during two successive late Miocene and Pliocene volcanic events: 10.6-7.4 Ma (Rairua submarine breccias and younger subaerial flows) and 6.4-5.4 Ma (Anatonu shield volcano and associated trachytic and phonolitic domes and plugs). Geochemical data demonstrate that the Rairua alkali basalts and basanites display a strong HIMU flavour, while the geochemical signature of the predominantly tholeiitic Anatonu lavas is more subdued and intermediate between DMM, HIMU and EM.

Rairua mafic lavas show obvious petrologic and geochemical similarities with those of the neighbouring (and nearly contemporaneous) island of Tubuai, located on the other side of the Austral fracture zone (fig. 1a). Both could result from the partial melting of the top of a predominantly HIMU secondary plume, which formed successively Mangaia, the old edifice of Rurutu, Tubuai and finally Rairua volcano in Raivavae (10.6-7.4 Ma). The geochemical signature of Anatonu lavas (6.4-5.4 Ma), although subdued, is clearly different from that of Rapa lavas which erupted only slightly later (4.8-4.1 Ma) 530 km SE of Raivavae, and

which are similar to those of Marotiri (3.8-3.1 Ma) and the presently active Macdonald seamount.

We propose that the latter group (Rapa and Marotiri lavas) originated from the secondary plume presently located beneath the Macdonald seamount. The building of Raivavae would document the last known activity of a “pre-Macdonald” plume of predominant HIMU composition. The differences between the geochemical signatures of Rairua and Anotonu lavas are attributed to either the contribution to the

latter of a distinct filament of more subdued composition, or alternatively a decrease of the proportion of pyroxenites in the mantle source.

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