Chemical Geology, 97 (1992) 199–218
Elsevier Science Publishers B.V., Amsterdam

Geochemistry and petrogenesis of Upper Cretaceous basaltic rocks from southern Malagasy

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(Received June 13, 1989; revised and accepted October 23, 1991)

ABSTRACT

Dostal, J., Dupuy, C., Nicollet, C. and Cantagrel, J.M., 1992. Geochemistry and petrogenesis of Upper Cretaceous basaltic rocks from southern Malagasy. Chem. Geol., 97: 199–218.

Upper Cretaceous basaltic rocks of southern Malagasy consist of tholeiitic flows and dykes and alkali basaltic dykes. Continental tholeiites include relatively primitive lava types with Mg-numbers ~0.6–0.5 and voluminous evolved basaltic lavas and dykes with Mg-number ~0.4. The evolved tholeiites have compositions similar to the high-Ti (or enriched) basalts from the Karoo and Paraná flood basalt provinces. They are characterized by a high content of strongly and moderately incompatible trace elements and a relative depletion of Nb and Sr. The evolved tholeiites were probably derived from subcontinental lithosphere. The primitive tholeiites resemble enriched MORB and are not directly related to the evolved types. Alkali basalts (1–15% normative nepheline) possess compositional characteristics of oceanic island basalts and show some geochemical similarities to the primitive tholeiites. Both rock-types were probably derived from a mantle plume. The upwelling plume induced lithospheric melting which produced the parental magmas of the evolved tholeiites. The Mesozoic volcanism of Malagasy displays many similarities to other flood basalt provinces including its large volume, rock composition and emplacement near plate margins.

1. Introduction

The Mesozoic evolution of Malagasy is important for understanding and reconstructing the post-Gondwanaland history and for refining plate-tectonic models for the Indian Ocean. After the break-up of Gondwanaland during the period of 130–80 Ma ago, Malagasy was probably a part of the Indian plate. Sea-floor spreading took place between the African and Indian plates. About 80 Ma ago, the spreading axis moved into a new position between India and Malagasy leading to the separation of Malagasy from India. The rifting stopped ~64 Ma ago and shifted northwards triggering the opening of the Arabian Sea (Courtillot et al., 1986). The new rift was located between India and the Seychelles Islands, northeast of Malagasy.

The voluminous Mesozoic flood basalt volcanism in Malagasy has been related to the Marion–Prince Edward hot spot currently located southeast of South Africa (R.A. Duncan, 1981; Hartnady and Le Roex, 1985). The predicted path of this hot spot is subparallel to the lineaments produced by the St. Helena and Tristan da Cunha hot spots (Atlantic Ocean) and coincides fairly well with the Malagasy Ridge. Continental flood basalts which are widespread in Malagasy constitute a significant part of the Mesozoic stratigraphic column. However, very little is known about their age, composition and origin. The purpose of this paper is to geochemically characterize the

0009-2541/92/$05.00

continental basalts of southern Malagasy and put constraints on their genesis. In addition, we present K/Ar data on the basalts. The ages of the Malagasy basalts are important for the refinement of the hot spot model and the reconstruction of the motions of the major tectonic plates. The upper Cretaceous volcanism of Malagasy provides an important temporal connection between major flood basalt events — Lower Jurassic Karoo and Lower Cretaceous Etendeka of southern Africa and Cretaceous/Tertiary Deccan of India — which are associated with the gradual break-up of Gondwanaland.

2. Geological and petrographic notes

Upper Cretaceous basaltic volcanism is widespread around the coast of Malagasy. In the southwestern part of the island (Fig. 1), the basaltic rocks occur predominantly as lava flows although locally some sills are present. Scarce rhyolitic dykes cross-cut the basaltic sequence. The lava flows, which mainly lie on top of Jurassic and Cretaceous sediments, crop out over an area > 200 km long and 2–10 km wide. Drill holes suggest an extension of the volcanic field farther westward toward the coast as well as to the south and probably even to a submarine plateau. Along the southeastern coast, the volcanic belt extends for a distance of > 500 km (Fig. 1). In the southeastern and southernmost part of the island, including the volcanic/subvolcanic complex of Androy, the volcanics were directly emplaced on the Precambrian basement (Nicollet, 1984). There, the volcanic sequence, composed of intercalated basaltic and subordinate rhyolitic and dacitic lava flows, is ~ 500 m thick although locally the thickness reaches 3000 m in the Androy massif (Battistini, 1959). Several gabbro-syenite intrusive complexes, which appear to be associated with the Cretaceous volcanism, occur or may be inferred from the gravity anomalies in southern Malagasy (Rechenmann, 1982).

The basaltic rocks also form numerous dykes (> 500) with a general N60°W orientation. They crop out over a large area (150 × 40 km) spreading between the Androy massif and the lava flows on the southwestern side of the island. These dykes, which occur predominantly in clusters, are tens of meters to several km long with thicknesses varying between 1 m and a few tens of meters. Tholeiitic basalts occur as both lava flows and dykes while rocks of alkali composition are found only as dykes. Field observations suggest that the tholeiitic and alkali basalts are approximately contemporaneous.

The tholeiitic basalts show textures ranging from fine-grained sparsely porphyritic to aphyric. They have a common mineral assemblage composed of olivine, clinopyroxene, plagioclase and opaques, Clinopyroxene (augite) and plagioclase are the dominant phenocryst and groundmass phases. Some samples contain minor amounts of pigeonite in the groundmass. A significant amount of Fe–Ti-oxide is present in the differentiated basalts.

The alkali basalts are fine- to medium-grained rocks with a doleritic texture. The main microphenocryst phase is clinopyroxene (titaniferous augite) with subordinate olivine and Fe–Ti-oxide and occasionally plagioclase. In the most silica-undersaturated rocks, nepheline and clinopyroxene are the predominant groundmass phases with minor quantities of Fe–Ti-oxide and intergranular biotite and amphibole and accessory apatite.

The rocks are usually fresh although some samples display signs of alteration. The more altered samples contain altered glass and olivine replaced by secondary iddingsite. However, the petrographic evidence of alteration does not appear to correlate with the chemical composition. In several samples, we suspect that the low K and Rb values are the result of rainwater leaching as has been demonstrated for the Kohala basalts by Feigenson et al. (1983). The abundances of other elements do not appear to be affected by the secondary processes. However, it is still necessary to use care

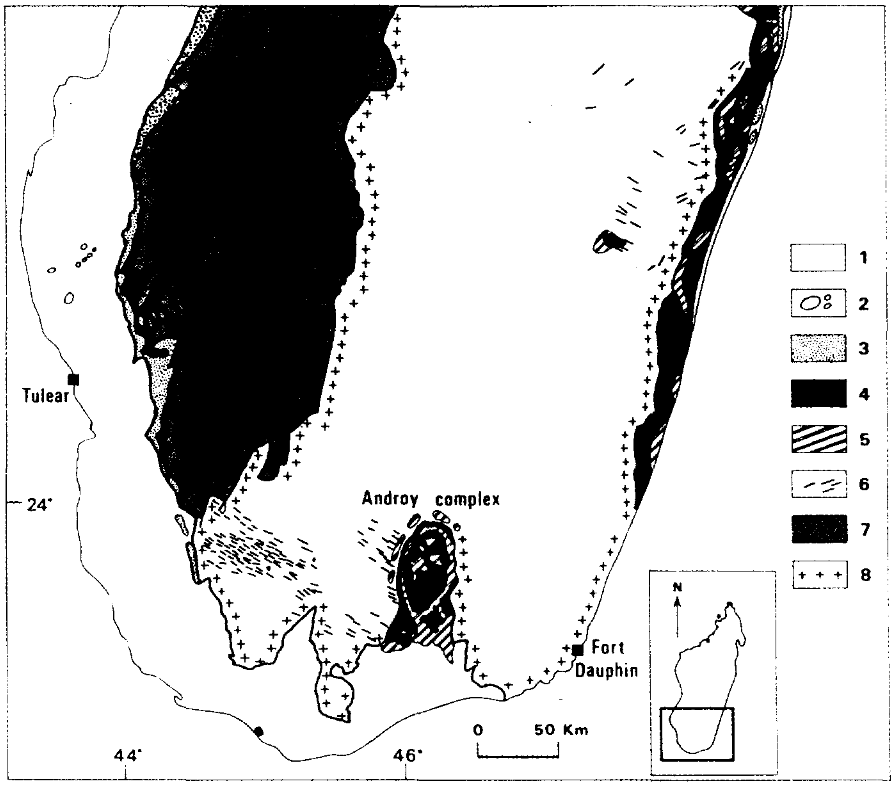


Fig. 1. Generalized geological map of southern Malagasy (modified after Nicollet, 1984). *1* = Eocene marine sediments; *2* = Miocene basalts; *3* = Maastrichtian sediments; *4*, *5* and *6* = Upper Cretaceous basalts, rhyolites and dykes, respectively; *7* = Upper Carboniferous to Cretaceous sediments; *8* = Precambrian basement.

in interpreting the distribution of some of the large-ion-lithophile elements (LILE).

3. Sampling and analytical notes

The samples of basaltic lavas and dykes were collected from accessible terrain throughout the southern one-third of the island. The major elements as well as trace elements Ba, Rb, Sr, Zr, Nb, Y, Ni, Cr, V, Cu and Zn were determined by X-ray fluorescence, whereas Li was analyzed by atomic absorption. From the set, thirty rocks were selected for instrumental neutron activation determinations of rare-earth elements (REE), Co, Th, Hf and Sc. The precision and accuracy of the data have been reported by Dostal et al. (1986) and are generally better than ±10%. The major- and trace-element analyses of representative basaltic lavas and dykes are given in Table 1. A complete set of the whole-rock analyses and locations of the samples can be obtained on request.

4. Age determination

Twenty-five samples were dated by the K–Ar method according to the analytical technique of Cantagrel and Baubron (1983) and the representative results are reported in Table 2. The ages obtained for the tholeiites and alkali basalts overlap. Most of the values are between 65 and 79 Ma, while the rest of the data give mainly significantly younger ages, which are probably the result of Ar loss during alteration. This is particularly true for the samples of the western volcanic belt, where relatively young ages (~30 Ma) were obtained on tholeiitic flows intercalated with Maastrichtian and

TABLE 1

Chemical compositions of representative tholeiitic basalts from southern Malagasy

Rock-type	Evolved tholeiites					
Formation	Western belt					
Sample No. (wt.%)	*1*	*3*	*33*	*34*	*35*	*39*
SiO_2	52.06	50.06	49.70	48.54	49.66	51.45
TiO_2	2.86	3.02	3.18	3.22	3.20	1.50
Al_2O_3	13.80	13.32	13.26	13.35	13.34	13.46
Fe_2O_3	13.22	14.68	14.50	15.11	15.32	13.89
MnO	0.19	0.20	0.16	0.17	0.18	0.17
MgO	4.36	4.00	4.19	4.80	4.92	4.73
CaO	8.31	8.34	8.50	8.47	8.72	9.11
Na_2O	2.51	2.47	2.52	2.54	2.52	2.63
K_2O	1.33	1.50	0.88	0.72	0.90	0.86
P_2O_5	0.47	1.24	0.55	0.52	0.42	0.16
H_2O^+	0.98	0.92	1.49	0.97	0.21	1.21
H_2O^-	0.26	0.44	1.16	0.71	0.09	0.55
Total	100.35	100.19	100.09	99.12	99.48	99.72
Mg-numbers	0.43	0.38	0.39	0.41	0.42	0.43
Li (ppm)	6	5	7	8	7	11
Rb	46	48	32	26	30	31
Sr	387	490	407	411	389	296
Ba	389	552	378	420	306	363
V	350	228	347	360	374	342
Cr	32	35	28	40	43	6
Co	42	37	46	48	45	47
Ni	27	31	38	39	33	29
Sc	26.2	24.8	24.6	23.9	27.6	28.9
Cu	41	36	28	30	42	77
Zn	134	144	136	138	140	114
Y	36	38	35	33	32	29
Zr	259	278	284	258	240	167
Nb	19	19	19	21	17	9
Hf	6.6	6.7	6.9	6.2	6.1	4.3
Th	6.7	5.9	5.6	4.5	5.2	6.7
La	31.5	39.9	31.1	27.3	27.1	23.6
Ce	70.4	92	70.8	62.1	62.6	51.1
Nd	43.3	51.2	42.1	36.7	37.8	28.6
Sm	8.85	11.6	8.87	8.16	8.25	5.78
Eu	2.45	3.81	2.73	2.56	2.48	1.51
Tb	1.15	1.46	1.15	1.10	1.17	0.82
Yb	3.00	3.13	2.95	2.80	2.86	2.90
Lu	0.47	0.46	0.45	0.42	0.44	0.43

						Primitive tholeiites		
Southwestern dyke swarm						Eastern belt		
46	*47*	*49*	*50*	*54*	*55*	*58*	*59*	*60*
52.45	49.40	52.22	49.49	49.90	50.86	50.01	48.84	49.33
2.41	3.41	2.26	3.33	3.36	2.94	2.17	3.60	0.79
12.88	13.50	13.00	13.47	13.53	12.94	13.73	12.75	20.40
14.61	15.00	14.31	15.26	14.59	14.43	13.31	13.27	8.91
0.21	0.20	0.22	0.19	0.20	0.20	0.18	0.15	0.12
4.15	4.48	4.44	4.65	4.52	4.36	6.23	6.41	5.80
7.94	8.17	7.86	8.22	8.14	8.35	10.45	10.00	11.18
2.72	2.54	2.82	2.54	2.61	2.63	2.34	2.31	2.73
1.07	1.36	1.40	1.20	1.54	1.10	0.28	0.49	0.20
0.32	0.89	0.31	0.52	0.89	0.39	0.26	0.32	0.09
0.46	0.16	0.82	0.32	0.31	0.51	0.41	0.87	0.40
0.26	0.02	0.56	0.19	0.02	0.45	0.29	0.06	0.10
99.48	99.13	100.22	99.38	99.61	99.16	99.66	99.07	100.05
0.39	0.40	0.41	0.41	0.41	0.40	0.51	0.51	0.60
5	6	12	10	9	9	–	–	–
64	56	42	37	44	64	7	6	1
276	458	255	402	455	286	252	546	376
365	588	461	383	584	335	73	171	52
390	300	361	355	300	358	329	452	131
9	30	10	25	30	21	153	65	49
45	43	45	46	44	46	–	–	–
13	38	17	31	39	30	97	113	95
31.6	23.3	–	24.9	23.1	28.8	35.8	–	19.1
42	28	88	31	25	77	158	135	48
131	143	124	152	151	147	100	101	63
36	34	38	34	37	38	35	24	9
244	272	232	276	267	260	145	207	51
22	19	20	18	20	19	9	15	5
6.2	6.4	–	6.7	6.4	6.5	3.6	–	1.1
7.4	5.8	–	5.2	5.6	4.9	0.8	–	0.35
29.9	51.5	–	29.1	34.1	26.9	9.70	–	3.80
65.7	78.7	–	66.9	77.7	60.6	24.4	–	9.10
28.1	48.2	–	41.2	45.7	34.4	17.00	–	6.80
7.75	9.72	–	8.75	9.62	8.47	5.48	–	1.74
2.16	3.13	–	2.67	3.09	2.41	1.84	–	0.97
1.08	1.15	–	1.21	1.18	1.17	1.25	–	0.39
3.61	3.41	–	2.98	2.96	3.62	3.30	–	0.85
0.52	0.49	–	0.45	0.47	0.53	0.52	–	0.13

TABLE 1 *(continued)*

Formation	Southwestern dyke swarm						
Rock type	Alkali basalts						
Sample No.	*2*	*8*	*10*	*11*	*12*	*16*	*18*
SiO_2 (%)	46.86	46.65	48.02	48.06	47.06	46.00	45.88
TiO_2	1.48	1.81	1.86	1.98	1.96	1.01	1.62
Al_2O_3	15.27	15.31	16.72	17.19	17.00	13.52	16.93
Fe_2O_3	10.77	12.42	12.90	13.65	13.40	10.30	12.53
MnO	0.16	0.18	0.18	0.18	0.18	0.17	0.19
MgO	9.55	9.06	5.86	5.52	5.50	12.72	7.19
CaO	11.73	9.72	8.10	7.88	7.67	12.03	9.60
Na_2O	2.57	2.98	3.77	4.09	3.88	2.02	3.00
K_2O	0.52	0.88	1.01	1.20	1.10	0.32	0.55
P_2O_5	0.23	0.44	0.47	0.47	0.47	0.21	0.31
H_2O^+	1.03	0.28	0.42	0.21	1.46	1.43	2.20
H_2O^-	0.18	0.04	0.25	0.14	0.40	0.06	0.32
Total	100.35	99.77	100.08	100.57	100.08	99.79	100.32
Mg-numbers	0.67	0.62	0.48	0.48	0.48	0.73	0.56
Li (ppm)	5	9	9	11	9	7	8
Rb	12	22	21	30	22	5	10
Sr	335	316	624	657	631	268	604
Ba	457	500	784	781	847	1,229	602
V	300	218	190	173	177	252	210
Cr	500	450	70	41	20	735	118
Co	49	49	41	41	41	57	45
Ni	184	122	54	36	20	294	73
Sc	45.3	32.9	26.0	–	22.5	39.0	28.0
Cu	113	56	33	36	35	90	15
Zn	73	91	92	95	93	63	87
Y	27	34	32	33	33	24	24
Zr	123	194	161	178	176	74	139
Nb	16	24	28	35	34	14	23
Hf	2.9	4.4	3.8	–	3.9	1.6	3.1
Th	1.3	2.1	3.4	–	3.8	1.0	2.1
La	14.6	19.8	36.6	–	39.2	11.1	20.8
Ce	31.8	44.1	74.1	–	80.1	23.8	44.3
Nd	20.3	27.3	42.8	–	40.7	12.7	23.7
Sm	4.63	6.26	7.21	–	7.48	3.20	4.84
Eu	1.55	1.95	2.22	–	2.31	1.10	1.58
Tb	0.99	1.22	1.18	–	1.22	0.67	0.84
Yb	2.80	3.52	3.23	–	3.10	2.28	2.27
Lu	0.46	0.58	0.52	–	0.51	0.37	0.37

Mg-number = $Mg/(Mg+Fe^{2+})$ with Fe^{3+}/Fe^{2+} assumed to be 0.15. – = no data.

					Southwestern dyke swarm			
					Basanites			
23	*31*	*43*	*44*	*45*	*7*	*20*	*24*	*52*
45.40	46.25	45.68	44.47	45.27	44.06	43.22	45.40	44.04
1.91	1.52	1.43	1.39	2.26	1.77	1.64	1.58	1.04
16.70	14.85	15.38	16.47	16.05	14.51	15.47	16.07	14.60
12.94	11.61	11.89	11.14	13.75	11.38	10.64	11.69	10.43
0.20	0.19	0.18	0.18	0.20	0.14	0.18	0.19	0.18
6.55	10.57	10.32	8.88	7.67	10.13	9.00	9.35	11.13
9.22	11.39	11.38	11.94	8.85	10.77	9.85	11.42	13.43
3.57	2.69	2.22	2.40	3.68	4.11	3.08	2.90	2.50
0.63	0.60	0.27	0.70	0.71	1.26	1.63	0.63	0.62
0.41	0.38	0.17	0.21	0.38	0.88	0.46	0.42	0.40
1.81	0.61	0.32	2.26	0.30	0.93	4.38	0.68	1.05
0.48	0.09	0.12	0.27	0.09	0.18	0.47	0.10	0.18
99.82	100.75	99.36	100.31	99.21	100.12	100.02	100.43	99.30
0.53	0.67	0.66	0.64	0.56	0.67	0.65	0.64	0.70
9	6	6	13	8	8	13	7	7
16	16	5	13	17	30	40	15	17
577	505	371	367	375	1,185	860	556	592
537	448	146	511	372	2,230	1,105	460	573
210	263	297	264	212	200	193	250	268
39	408	368	130	160	390	320	214	400
46	53	56	48	52	52	44	52	52
54	194	146	109	113	204	168	168	225
25.9	37.6	37.9	36.3	25.7	23.2	25.7	–	36.2
40	82	69	71	56	71	64	90	114
82	77	84	76	90	91	58	81	79
30	24	22	26	30	26	27	28	24
161	138	105	112	183	211	172	144	119
37	32	11	23	30	62	68	32	30
3.6	3.2	2.5	2.5	4.3	4.0	3.2	–	2.3
3.5	3.6	0.90	1.3	1.8	10.5	4.1	–	4.8
33.4	37.6	10.0	12.0	17.8	89.6	35.6	–	39.3
68.5	75.9	23.1	27.0	40.1	170	68.3	–	77.8
33.5	35.3	13.0	12.4	22.2	78.7	33.4	–	36.1
6.36	6.26	3.83	4.06	6.24	11.5	5.83	–	6.32
2.07	1.95	1.35	1.42	2.15	3.30	1.88	–	1.86
1.25	0.90	0.77	0.95	1.39	0.95	0.98	–	0.69
2.98	2.45	2.32	2.66	2.97	1.87	2.71	–	2.23
0.49	0.40	0.35	0.39	0.43	0.30	0.43	–	0.37

TABLE 2

K–Ar results for representative samples

Sample	K (wt%)	^{40}Ar ($ng\,g^{-1}$)	Ar (at%)	t (Ma)
Western belt:				
1 tholeiite	1.12	3.61	39	46 ± 0.9
3 tholeiite	1.29	4.73	27	52 ± 1
34 tholeiite	0.649	2.45	36	54 ± 1
39 tholeiite	0.713	1.54	82	31 ± 2
35 tholeiite	0.763	1.96	82	37 ± 2
Southwestern dyke swarm:				
8 alkali basalt	0.753	3.63	33	68 ± 1
11 alkali basalt	0.889	4.29	21	68 ± 1
24 basanite	0.579	2.68	35	66 ± 1
10 alkali basalt	0.839	4.10	27	69 ± 1
45 alkali basalt	0.625	3.32	28	75 ± 2
52 basanite	0.532	2.67	80	71 ± 3
49 tholeiite	1.10	5.01	18	65 ± 1
54 tholeiite	1.34	6.12	11	65 ± 1
Androy massif:				
E*12* tholeiite	1.26	6.02	23	68 ± 1
E*16* tholeiite	1.52	7.86	24	72 ± 1
Eastern belt:				
58 tholeiite	0.240	1.10	56	65 ± 1
59 tholeiite	0.419	2.34	36	79 ± 2
60 tholeiite	0.160	0.85	72	75 ± 2

Ages were calculated using the 1977 constants.

Campanian (66–72 and 72–83 Ma ago, respectively; Haq and van Eysinga, 1987) sediments (Besairie and Collignon, 1972). Since these rocks do not appear to be more altered than similar lavas from the eastern belt, the young ages could also be related to recent faulting or fluid circulation and/or thermal partial resetting of the K–Ar system associated with Upper Miocene to Quaternary alkali volcanism. The basaltic sequence in western Malagasy rests upon Turonian (> 89 Ma ago) sediments, whereas along the eastern coast, marine Maastrichtian sediments are transgressive on top of the basaltic lava (Besairie and Collignon, 1972; Boast and Nairn, 1982). We believe that the true age of the Late Mesozoic volcanism in southern Malagasy is within the range of 65 to 83 Ma.

The Malagasy volcanism is not related to the older Karoo volcanism and also postdates the opening of the Mozambique channel which appeared during the separation of Malagasy from Africa (Norton and Sclater, 1979; Segoufin and Patriat, 1981). The timing of the emplacement of the basaltic rocks in southern Malagasy is close to but slightly older than the eruption of the Deccan Traps in India, where volcanic activity peaked around 66 Ma ago (Courtillot et al., 1986, 1988; R.A. Duncan and Pyle, 1988), the peak of igneous activities at the Seychelles Islands dated at 63 Ma (Dickin et al., 1986) and the initiation of the opening of the Arabian Sea (Courtillot et al., 1986). The volcanic activity appears to be closely related to the continental rifting and lithospheric thinning which must have been accompanied by mantle upwelling (White and McKenzie, 1989).

5. Geochemistry

5.1. Major elements

The basaltic rocks from southern Malagasy (Table 1) show a large range of chemical composition. On the basis of normative composition (assuming $Fe^{3+}/Fe^{2+} = 0.15$), three types of basalts may be recognized:

(1) quartz-normative tholeiites (1–7% norm. quartz) associated with a subordinate amount of olivine-normative tholeiites;

(2) alkali basalts (< 5% norm. nepheline);

(3) basanites (5–15% norm. nepheline).

The two latter types are limited to the southernmost part of the island where they appear as dykes, whereas the tholeiites are from the lava flows on both sides of the island and from the dyke swarm in southern Malagasy (Fig. 1).

The tholeiites can be further subdivided into two groups. The lavas from the western part of the island and the dykes are strongly fractionated with the Mg-number [$Mg/(Mg+Fe^{2+})$ assuming $Fe^{3+}/Fe^{2+} = 0.15$] clustering around 0.40 whereas the tholeiitic basalts from the eastern side are more primitive with the Mg-number between 0.60 and 0.51 (Fig. 2). There is no obvious compositional difference between tholeiites from the lavas of the western belt and the dykes and thus all these evolved basalts are considered together. The evolved tholeiites have a composition similar to that of continental flood basalts from Karoo, South Africa (e.g., Erlank, 1984; Marsh, 1987; Cox, 1988a) and Paraná, Brazil, provinces (e.g., Mantovani et al., 1985; Piccirillo et al., 1989). They usually have low MgO (4–5%) but high $Fe_2O_3^T$ (13–16.5%) and variable TiO_2 and P_2O_5 contents. The latter two oxides separate the rocks into two relatively homogeneous groups. The first group, which encompasses a large majority of the tholeiite samples, has higher contents of TiO_2 (>2%; average 3.03%) and P_2O_5 (>0.3%; average 0.59%) than the second group which includes only two samples (average 1.5% TiO_2 and 0.17% P_2O_5). Similar high- and low-Ti groups of tholeiitic basalts have been reported from the Karoo and Paraná flood basalt provinces (Bellieni et al., 1984; A.R. Duncan et al., 1984; Cox, 1988a). Like in Karoo and Paraná (e.g., Erlank et al., 1988), the two groups of evolved tholeiites do not show any obvious differences in mineralogy, petrography or field occurrence. In the high-Ti tholeiites, SiO_2 (on volatile-free basis) ranges from 49% to 52.5%. The major-element variations which are relatively small include a rough positive correlation of Al_2O_3 and CaO with Mg-number accompanied by a low and relatively constant Al_2O_3/CaO ratio (Fig. 2). The primitive tholeiites from eastern Malagasy are lower in K and some strongly incompatible trace ele-

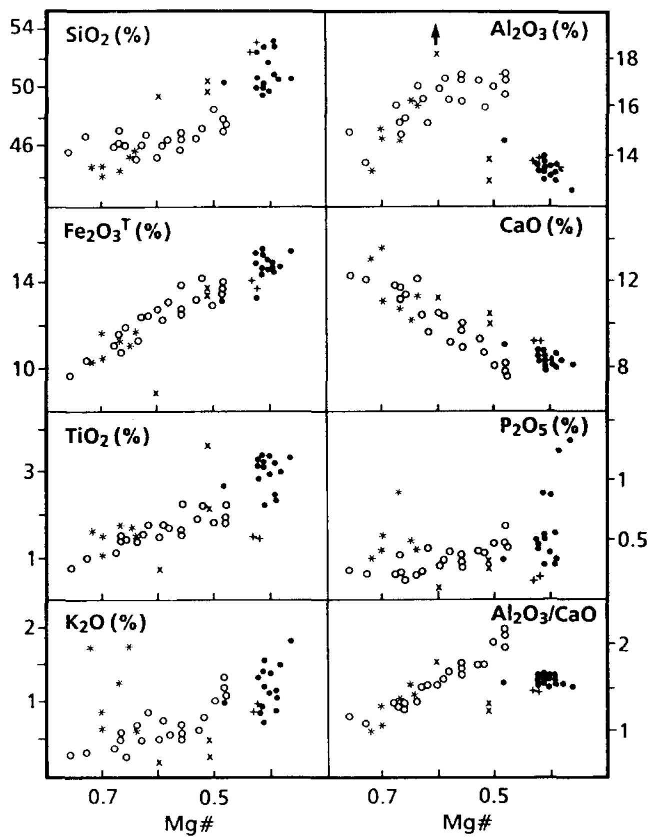


Fig. 2. Plots of SiO_2, Al_2O_3, $Fe_2O_3^T$, CaO, TiO_2, P_2O_5 and K_2O contents and Al_2O_3/CaO ratio against Mg-number [Mg/($Mg+Fe^{2+}$) assuming $Fe^{3+}/Fe^{2+}=0.15$] for the basaltic rocks of Malagasy: evolved tholeiites — high-Ti rocks (●), low-Ti rocks (+); primitive tholeiites (×); alkali basalts (○); basanites (∗).

ments such as Rb and light REE (LREE) than the evolved basalts (Fig. 3). Among the primitive tholeiites, sample 60 probably has accumulated plagioclase as indicated by its high Al_2O_3 content and a higher quantity of plagioclase phenocrysts ($\sim 20\%$).

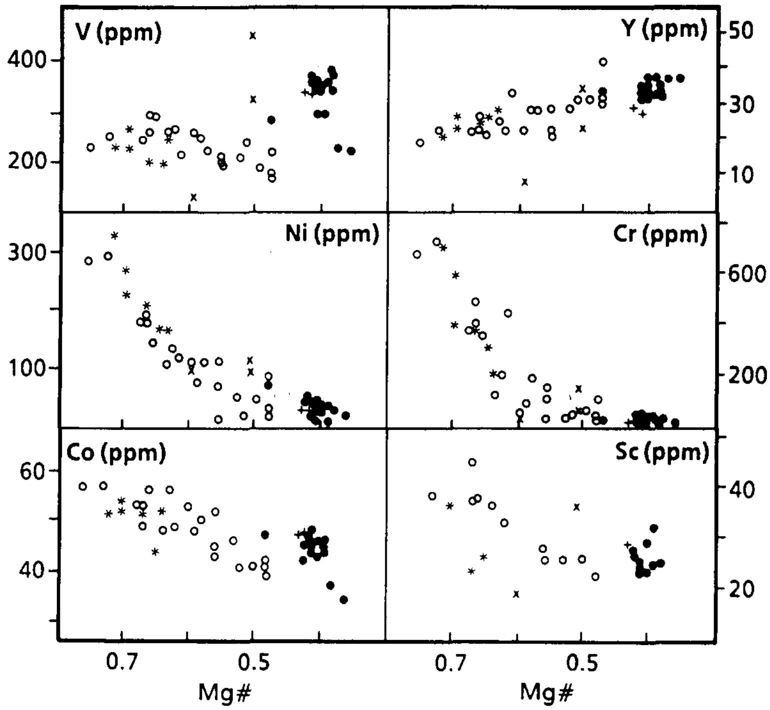
The silica-undersaturated dyke rocks display a range of Mg-number decreasing from 0.76 to 0.48 accompanied by a distinct increase of Fe and Ti with differentiation (Fig. 2). They are poorer in SiO_2 than the tholeiites and have higher Al_2O_3/CaO ratios for a given Mg-number than those of the high-Ti tholeiites (Fig. 2). Most major elements in the silica-undersaturated rocks inversely correlate with the Mg-number, whereas CaO shows a positive correlation (Fig. 2). The Al_2O_3/CaO ratio increases with increasing differentiation (Fig. 2), giving values close to that of chondrites ($Al_2O_3/CaO = 1.1$) in the most primitive undersaturated rocks. The increase of this ratio is the result of pyroxene fractionation. The basanites are higher in K and P and lower in Si than the alkali basalts (Fig. 2; Table 1).

5.2. *Trace elements*

5.2.1. Transition elements. The low abundances of the compatible transition elements, Cr and Ni, in the evolved tholeiites (Fig. 3; Table 1) confirm that the rocks underwent extensive fractional crystallization involving ferromagnesian minerals. The primitive tholeiites have higher abundances of these elements but they are still significantly lower than those expected for primary mantle melts. The concentrations of Cr, Ni, Sc and Co in the silica-undersaturated dyke rocks decrease with differentiation, reflecting the crystallization of olivine and pyroxenes. In agreement with partition coefficients for clinopyroxene and olivine (Wedepohl, 1985), the decrease of Co accompanying the steeper decrease of Ni relative to the other transition elements down to Mg-number ≈ 0.60 (Fig. 3) in the silica-undersaturated dyke samples attests to the influence of olivine, in addition to clinopyroxene, in the first stage of differentiation.

The concentrations of V are higher in the tholeiites than in the undersaturated dyke rocks (Fig. 3; Table 1). In the latter rocks, V slightly increases with differentiation up to Mg-number ≈ 0.65 and then starts to decrease, indicating the crystallization of Fe–Ti-oxides during more advanced stages of the differentiation. Ti/V ratios in the tholeiites and silica-undersaturated dyke rocks are high (Table 3), usually >40; similar values are found in intraplate basalts (Shervais, 1982). The exceptions are the low-Ti basalts with a Ti/V ratio ~ 25, which is still within the range typical of mid-ocean ridge basalts (MORB) and continental tholeiites.

5.2.2. Rare-earth elements. The chondrite-normalized REE distribution of several representative basalts is shown in Fig. 4. The evolved high-Ti tholeiites have patterns enriched in LREE with $(La/Yb)_n$ ratios (n: chondrite-normalized) between 5 and 10 and La_n abundances ranging from 80 to 160. The evolved low-Ti tholeiite has a similar REE pattern with $(La/Yb)_n \approx 5.4$ but slightly lower LREE abundances ($La_n \sim 70$). The REE patterns and abundances of the evolved high- and low-Ti tholeiites (Fig. 4a) are comparable to those of equivalent rock-types from Paraná (Mantovani et al., 1985; Marques et al., 1988) and Karoo (Erlank et al., 1988). The tholeiitic basalts of the Deccan Traps have typically lower REE abundances and less fractionated patterns (Fig. 4b and Mahoney et al., 1982; Dupuy and Dostal, 1984; Lightfoot and Hawkesworth, 1988), although some basalts from the Bombay area (Dupuy and Dostal, 1984) have similar REE distributions. In fact, compared to the main basaltic types of the Deccan Traps (Cox and Hawkesworth, 1984, 1985; Mahoney et al., 1985; Lightfoot and Hawkesworth, 1988), the evolved tholeiites, particularly high-Ti ones, have higher abundances not only of REE but also of many other incompatible ele-



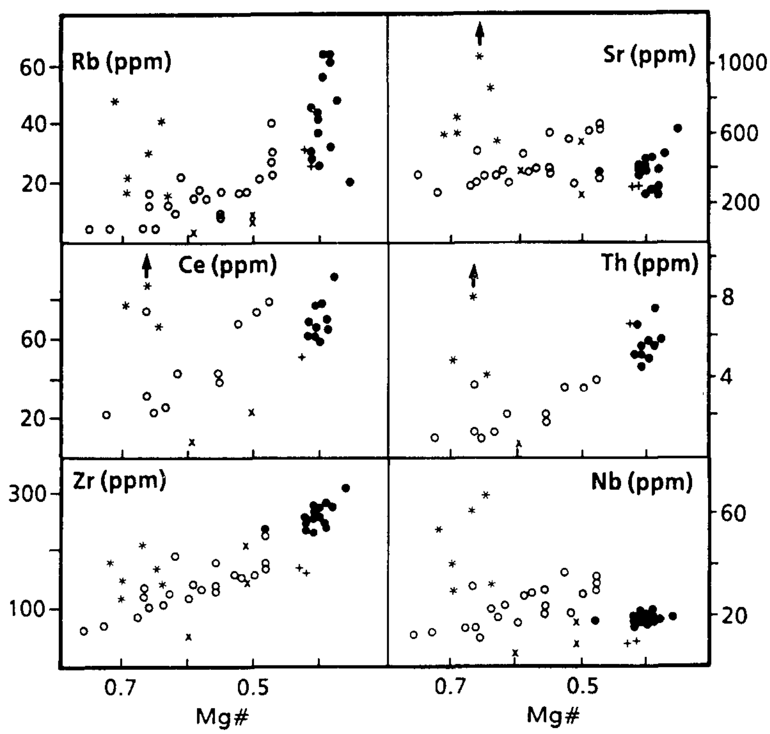


Fig. 3. Plots of selected trace elements against Mg-number [Mg/(Mg+Fe^{2+}) assuming $Fe^{3+}/Fe^{2+}=0.15$)] for the basaltic rocks of Malagasy: evolved tholeiites — high-Ti types (●), low-Ti rocks (+); primitive tholeiites (×); alkali basalts (○); basanites (∗).

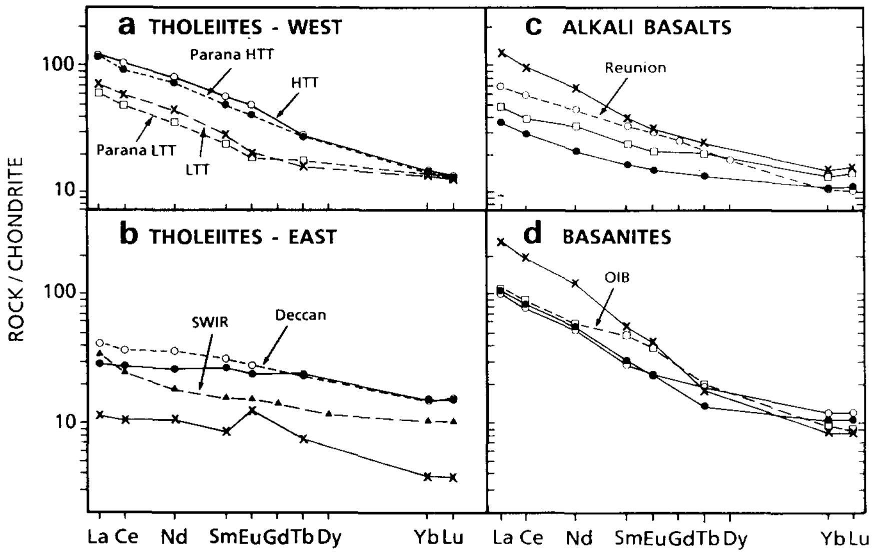


Fig. 4. Chondrite-normalized REE patterns for Malagasy tholeiites (a and b), alkali basalts (c) and basanites (d). Normalizing values after Masuda et al. (1973).
a. High-Ti (*HTT*) (○ = sample *3*) and low-Ti (*LTT*) (× = sample *39*) evolved tholeiites. The averages of HTT (●) and LTT (□) from the Paraná Basin (Mantovani et al., 1985) are plotted for comparison.
b. Primitive tholeiites from eastern Malagasy (● = sample *58*; × = sample *60*) together with the average of the Deccan Trap basalts of the Ambenali Formation (Lightfoot and Hawkesworth, 1988) (○) and the average of E-type MORB from the Southwest Indian Ridge (*SWIR*) (Le Roex et al., 1989) (▲).
c. Alkali basalts (□ = sample *2*; × = sample *12*; ● = sample *16*) together with oceanic island basalts (○) from Réunion Island (Newsom et al., 1986).
d. Basanites (× = sample *7*; ○ = sample *20*; ● = sample *52*) and average of oceanic island basalts (□) of Sun and McDonough (1989).

ments and have higher Zr/Ti, Ba/Nb and Nb/Y ratios (Fig. 5).

Among the primitive tholeiites, sample *60* has the lowest REE abundances and its pattern is characterized by a positive Eu anomaly (Fig. 4b) which, along with its high Al_2O_3 content, confirms that the rock is plagioclase cumulate. On the other hand, sample *58* (Fig. 4b) has a relatively flat REE pattern with $(La/Yb)_n \approx 2$. The pattern somewhat resembles those of the Deccan Trap basalts (Mahoney et al., 1982; Dupuy and Dostal, 1984; Lightfoot and Hawkesworth, 1988) and enriched MORB including basalts from the Southwest Indian Ridge (Le Roex et al., 1983, 1985, 1989).

The REE patterns of the silica-undersaturated rocks are like those of oceanic island basalts (OIB) (Fig. 4c and d). The $(La/Yb)_n$ ratio usually ranges between 2.9 and 10.3 in the alkali basalts and between 9 and 32 in the basanites. The La abundances are increasing from the alkali basalts with values of 30–120×chondrites to the basanites with 100–280×chondritic concentrations. The highest LREE concentrations are encountered in basanite *7* (Table 1) which has 15% normative nepheline. The most silica-undersaturated dyke rocks display a crossing of their REE patterns (Fig. 4d), which may be interpreted to be the result of a variable degree of partial melting from a common source leaving a residue with a variable garnet/clinopyroxene ratio (Kay and Gast, 1973).

5.2.3. Incompatible elements (Ba, Sr, Rb, Nb, Th, Zr, Hf and Y). The abundances of Nb, Ba and Sr increase with the degree of undersaturation from tholeiites through alkali basalts to basanites (Fig. 3; Table 1). The trend is accompanied by a decrease of the Zr/Nb and Y/

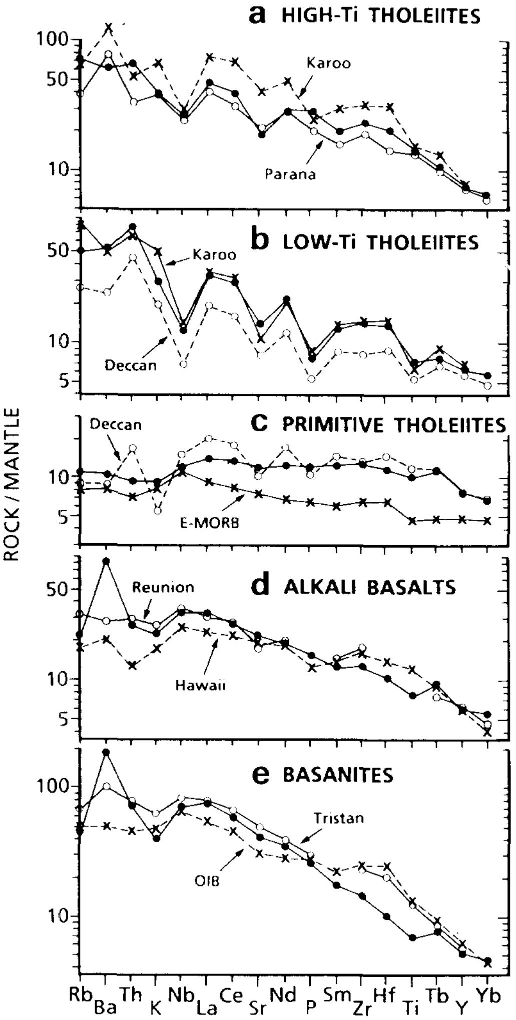


Fig. 5. Plots of incompatible-element abundances normalized to primitive mantle (Sun and McDonough, 1989) for selected basalts from southern Malagasy. Also plotted for comparison are the data for similar rocks. Normalizing values (in ppm): Rb 0.635, Ba 6.989, Th 0.085, K 250, Nb 0.713, La 0.687, Ce 1.775, Sr 21.1, Nd 1.354, P 95, Sm 0.444, Zr 11.2, Hf 0.309, Ti 1300, Tb 0.108, Y 4.55, Yb 0.493.
a. High-Ti tholeiites: ● = average of high-Ti tholeiites (> 2% TiO_2) from southern Malagasy; × = average of Karoo basalts from N. Lebombo (Sabie River basalts; A.R. Duncan et al., 1984; Marsh, 1987); ○ = average of high-Ti tholeiites from northern Paraná (Marques et al., 1988).
b. Low-Ti basalts: ● = low-Ti tholeiite sample *39* from southwestern Malagasy; × = average of Karoo basalts from Etendeka (Tafelberg basalts; A.R. Duncan et al., 1984; Marsh, 1987); ○ = average of the Deccan Traps basalts from the Bushe Formation (Lightfoot and Hawkesworth, 1988).
c. Primitive tholeiites: ● = primitive tholeiite sample *58* from eastern Malagasy; × = average of E-type MORB of Sun and McDonough (1989); ○ = average of the Deccan Traps basalts from the Ambenali Formation (Lightfoot and Hawkesworth, 1988).
d. Alkali basalts: ● = average of alkali basalts from southern Malagasy; also plotted for comparison are data on oceanic island basalts from Réunion (○) (Newsom et al., 1986) and Hawaii (×) (USGS standard rock BHVO-*1*; Gladney and Roelandts, 1988).
e. Basanites: ● = average of basanites from southern Malagasy; ○ = average of oceanic island basalts (moderately to mildy nepheline normative) from Tristan da Cunha (Weaver et al., 1987); × = average of oceanic island basalts of Sun and McDonough (1989).

Nb ratios (Table 3). On the other hand, Th and Rb, like K are enriched in the evolved tholeiites which have contents of these elements close to those of the basanites (Figs. 2 and 3). When compared with the other analyzed basalts, the evolved tholeiites have a distinctly higher Th/La ratio and lower Nb/La, Nb/Th and K/Rb ratios (Table 3).

The variations of Zr, Hf and Y in the alkali basalts and evolved tholeiites are mainly controlled by differentiation as these elements increase within each suite with the decrease of the Mg-number (Fig. 3; Table 1). The evolved high-Ti tholeiites have the highest contents of Zr and Hf (Fig. 3; Table 1). Like in Paraná and Karoo, they also have higher Zr/Y (average 7.3) and Ti/Y (average 508) ratios than the low-Ti basalts (5.8 and 320, respectively). The basanites have generally higher Zr (Fig. 3) and lower Hf than alkali basalts with equivalent Mg-number and thus possess a higher Zr/Hf ratio (~ 53) than the alkali basalts and tholeiites (Table 3). Such an increase of Zr/Hf with the degree of silica-undersaturation has been observed in Hawaiian basalts in the Pacific (Clague and Frey, 1982) and seems to be a common feature among intraplate basalts (Weaver et al., 1987; Dautria et al., 1988; Dostal et al., 1988). It appears that Zr is more incompatible than Hf during the genesis of the strongly undersaturated basalts in the upper mantle. The Hf/Sm and Zr/Sm ratios have near mantle values in the primitive tholeiitic basalts (Table 3) and decrease with the degree of silica-undersaturation towards the basanites. This suggests either an enrichment of LREE relative to high-field strength elements (HFSE) in the respective upper-mantle sources or element fractionation related to small but different degrees of melting.

6. Discussion

6.1. Evolved tholeiites

The differences between the high- and low-Ti tholeiites are also reflected in several incompatible elements. The high-Ti tholeiites have higher contents of P, Zr, Hf, Nb and LREE than the low-Ti rocks of similar Mg-number (Fig. 3; Table 1). The same two groups (high-Ti or enriched and low-Ti or normal groups) have been documented in the Paraná and Karoo provinces. In the three provinces, equivalent rock-groups display close compositional similarities including mantle-normalized distribution patterns of incompatible trace elements (Fig. 5a and b). These patterns are hump-backed, marked by negative anomalies of Nb and Sr characteristic of continental tholeiites (Thompson et al., 1982). In addition, the profile of the low-Ti tholeiite displays a relative depletion in P and Ti. The pattern mimics those of basalts from the Bushe Formation of the Deccan Traps and low-Ti Karoo basalts of Tafelberg, Etendeka (Fig. 5b) which have rather high initial Sr-isotope ratios (>0.715, Cox and Hawkesworth, 1985; Beane et al., 1988 and ~ 0.708–0.713, Erlank et al., 1984, respectively).

Several detailed studies on Paraná and Karoo (e.g., Bellieni et al., 1984; Erlank et al., 1984) demonstrated that the compositional trends within each tholeiitic group can be modelled by low-pressure fractional crystallization of pyroxenes, plagioclase, olivine and Fe–Ti-oxides, which in the case of some low-Ti basalts was probably accompanied by crustal contamination. The observed variations of the evolved tholeiites from Malagasy are consistent with such a mechanism. However, fractional crystallization, even accompanied by crustal contamination or periodically replenished, tapped and fractionated magma chambers (O'Hara, 1977; Cox, 1988b), cannot explain the differences between the high- and low-Ti tholeiites. In Paraná and Karoo, the two groups have been related to distinct parental magmas which have been either generated by varying degrees of partial melting of a homogeneous mantle source (Piccirillo et al., 1989: garnet peridotite — 5% melting for high-Ti and 20% for low-Ti basalts) or derived from a heterogeneous mantle source. The recent isotope and trace-element data favor the heterogeneous source for the two groups (A.R. Duncan et al., 1988; Erlank et al., 1988; Peate et al., 1988; Piccirillo et al., 1989). The same origin can be invoked for the two groups of evolved tholeiites from Malagasy; the melts with their contrasting incompatible-element distribution patterns (Fig. 5) and element ratios (Table 3) cannot be readily derived from the same source by variable degrees of melting. The relative depletion of Nb, Sr and, in low-Ti tholeiite, even of P and Ti (Fig. 5b) while alkali metals, Ba and Th are enriched, suggests the role of a crustal component in the genesis of these rocks. However, available data cannot determine whether the crustal signature is due to crustal contamination of magma or due to the presence of some crustal components in the source.

6.2. Primitive tholeiites and silica-undersaturated dyke rocks

The primitive tholeiites from the eastern side of southern Malagasy seem to be devoid of any crustal component. The ratio of Th/La which is frequently considered to be a sensitive indicator of crustal contamination is low in these rocks, close to the mantle value (Table 3). The mantle-normalized incompatible-element pattern of the primitive tholeiite *58* (Fig. 5c) is

TABLE 3

Average incompatible-element ratios in basaltic rock types of Malagasy

	K/Rb	Ti/V	Th/La	Nb/La	Nb/Th	Zr/Sm	Hf/Sm	Zr/Hf	Zr/Nb	Y/Nb	Hf/Lu	Ba/Nb	Ba/La	$\left(\frac{La}{Sm}\right)_n$	$\left(\frac{La}{Yb}\right)_n$	SiO_2 (wt%)	Mg-number
THOLEIITES:																	
Western Malagasy																	
High-Ti:																	
n	16	16	10	10	10	10	10	10	15	15	10	15	10	10	10	15	15
x	270	56	0.18	0.61	3.5	30	0.73	41	14	1.9	14	24	13	2.2	7.0	51.1	0.41
s	142	15	0.04	0.12	0.5	2.5	0.07	1	1.4	0.1	1.1	8	2	0.4	1.5	1.2	0.03
Low-Ti:																	
n	2	2	1	1	1	1	1	1	2	2	1	2	1	1	1	2	2
x	276	27	0.28	0.38	1.3	29	0.74	39	17	3.0	10	39	15	2.5	5.4	52.8	0.43
s	57	0.4							1.7	0.3		2				0.4	0.01
Eastern Malagasy:																	
n	3	3	2	2	2	2	2	2	3	3	2	3	2	2	2	3	3
x	898	42	0.09	1.1	13	28	0.64	43	13	2.4	7.7	9.5	11	1.2	2.5	50.0	0.54
s	690	6	0.01	0.3	2	2	0.02	4	3	1.3	1.1	1.2	4	0.2	0.7	0.5	0.05
ALKALI BASALTS:																	
n	21	21	11	11	11	11	11	11	21	21	11	21	11	11	11	21	21
x	405	45	0.10	1.2	12	26	0.60	44	6.1	1.3	7.3	23	31	2.5	5.5	46.7	0.60
s	89	15	0.01	0.3	3	3	0.07	1	1.4	0.3	1.4	16	28	0.7	2.6	0.8	0.08
BASANITES:																	
n	6	6	3	3	3	3	3	3	6	6	3	6	3	3	3	6	6
x	336	42	0.12	1.1	9.6	22	0.42	53	3.6	0.59	9.0	21	24	4.1	17.5	44.9	0.68
s	26	11	0.00	0.7	6.0	6	0.11	1	0.7	0.22	3.8	8	8	0.6	12.6	0.5	0.03
PRIMITIVE MANTLE*[1]	394	15*[3]	0.12	1.0	8.4	25	0.70	36	16	6.4	4.2	9.8	10	1.0	1.0		
N-type MORB*[1]	1,071	20–50*[4]	0.05	0.93	19	28	0.78	36	32	12	4.5	2.7	2.5	0.58	0.55		
E-type MORB*[1]	417		0.16	1.3	14	28	0.78	36	8.8	2.7	5.7	6.9	9.0	1.5	1.8		
OIB*[1]	387	>45	0.11	1.3	12	28	0.78	36	5.8	0.6	26	7.3	9.5	2.3	11.5		
BCC*[2]	285	50	0.22	0.69	3.1	42	0.86	33	9.1	1.9	10	23	16	2.8	4.9		
UCC*[2]	252	50	0.36	0.83	2.3	42	1.29	33	7.6	0.9	18	22	18	4.1	9.1		

n = number of samples; *x* = average; *s* = standard deviation; BCC = bulk continental crust; UCC = upper continental crust; Mg = Mg/(Mg + Fe^{2+}) with Fe^{3+}/Fe^{2+} assumed to be 0.15. SiO_2 (wt%) — LOI-free; n = chondrite-normalized.
*[1]Sun and McDonough (1989); *[2]Taylor and McLennan (1985); *[3]Jochum et al. (1986); *[4]Shervais (1982).

relatively flat and smooth and resembles those of E-type MORB including basalts from the southern and Indian Oceans (Le Roex et al., 1983, 1985, 1989). The tholeiite also shows some similarities to the basalts of the Ambenali Formation from the Deccan Traps (Fig. 5c) although there are some important compositional differences which are also depicted by the incompatible-element patterns. The primitive tholeiites are not directly related to the evolved basaltic types. Simple models of partial melting or fractional crystallization of a common parent cannot explain the differences in the trace-element characteristics including the shape of the mantle-normalized patterns and the incompatible-element ratios between these two basaltic types (Table 3). The primitive and evolved tholeiites were derived from different sources. However, the primitive tholeiites have ratios of several incompatible elements similar to those of the undersaturated dyke rocks (Table 3), suggesting that they were both derived from sources which share many geochemical features.

The mantle-normalized incompatible-element patterns of the silica-undersaturated dyke rocks resemble those of OIB (Fig. 5d and e). Compared to typical OIB, they are slightly enriched in Ba and depleted in Zr, Hf and Ti. The absolute abundances of incompatible elements in the alkali basalts are more similar to enriched oceanic island tholeiites than to typical alkali basalts. The alkali basalts and basanites were probably generated from a single source by variable degrees of melting. However, compared to the alkali basalts, the basanites have generally higher contents of strongly incompatible elements and higher La/Yb, but lower Zr/Nb ratios (Table 3), suggesting a decrease of the degree of partial melting towards the most undersaturated rocks. In addition, the crossing of REE patterns in the basanites suggests that the decrease of the degree of partial melting was accompanied by an increase of garnet in the upper-mantle residue. This source had to be enriched in LILE relative to primordial mantle unless the degree of partial melting was $< 1\%$ for the basanites.

Close similarities of the primitive tholeiites to enriched MORB and of the undersaturated dyke rocks to the OIB suggest that the basaltic rocks from Malagasy were derived from an upwelling plume of hot mantle material. However, the abundances of Ba relative to the other strongly incompatible elements in the silica-undersaturated dyke rocks are elevated (Fig. 5d and e). The ratios such as Ba/Nb and Ba/La which are usually not significantly affected by partial melting are also high in the undersaturated rocks (Table 3). Similar Ba enrichment encountered in OIB from Gough Island, South Atlantic Ocean (Weaver et al., 1987) has been interpreted as evidence for the presence of ancient subducted oceanic crust containing a small amount of Ba-rich pelagic sediments in the mantle source region.

6.3. *Geodynamic implications*

Cox (1988a) and Erlank et al. (1988) have delineated two major geochemical subprovinces in Karoo which can be extended into the Paraná basin in Brazil. The northern subprovince is composed of high-Ti basalts with subordinate amounts of low-Ti tholeiites while the southern subprovince contains only low-Ti tholeiites. The contact zone between two subprovinces in a reconstruction of southwestern Gondwanaland during the Mesozoic has been inferred to run E–W across southern Africa and South America for almost 3000 km. The magmatic activity in the two subprovinces lasted for at least 60 Ma as most Karoo basalts in the Lebombo and Lesotho areas are ~ 190 Ma old whereas the age of the Etendeka and Paraná tholeiites is ~ 130 Ma (Erlank, 1984). The lateral extent and longevities of the boundary points to a large-scale heterogeneity of the mantle. Before the break-up and dispersion of Gondwanaland, Malagasy was located along the present southeastern edge of southern Africa. Since the evolved tholeiites closely resemble tholeiitic sequences from the high-Ti subprovinces of Karoo and Paraná, it may be speculated that the large-scale mantle heterogeneity within the Gondwanaland lithospheric mantle extended up to Malagasy. Sweeney and Watkeys (in Erlank et al., 1988) argued that the subprovinces are related to the different lithospheric domains. The high-Ti basalts usually lie on top of the Archean basement which has been stable since ~2.5 Ga whereas the low-Ti basalts overlie post-Archean crust. However, the southwestern part of Malagasy is underlain by the Proterozoic crust (Besairie and Collignon, 1972; Boast and Nairn, 1982), suggesting that the enriched subcontinental lithospheric mantle may also be present under a younger post-Archean crust. The rising mantle plume, from which the primitive tholeiites and the undersaturated rocks were probably derived, could have induced lithospheric melting (White and McKenzie, 1989) which produced the parent magmas of the evolved tholeiites.

7. Conclusions

The Upper Cretaceous basaltic rocks of southern Malagasy include evolved tholeiites with Mg-number ≈ 0.4, relatively primitive tholeiites with Mg-number ranging from 0.6 to 0.5 and silica-undersaturated basaltic dyke rocks. The evolved tholeiites have a composition typical of high-Ti basalts from the Karoo and Paraná flood basalt provinces. They were derived from subcontinental lithosphere which could have been an extension of an enriched domain within the Gondwanaland lithospheric mantle that stretched from South America to southern Africa. The occurrence of rare low-Ti tholeiites in the high-Ti basalt sequence points to smaller-scale heterogeneities within the enriched lithosphere domain.

The primitive tholeiitic basalts are compositionally similar to E-type MORB and are not directly related to the evolved tholeiites. Alkali basalts and basanites resemble OIB. Their distribution of heavy REE is consistent with garnet as a residual phase. The primitive tholeiites share some geochemical features, particularly several element ratios (Table 3), with the silica-undersaturated dyke rocks. Such a relationship may be the result of derivation from a similar source by varying degrees of melting. In the absence of isotope data it is difficult to constrain the composition of such a source. However, it may be speculated that the mantle source contained a recycled oceanic crust component to account for the enrichment of Ba in the silica-undersaturated dyke rocks.

The volcanism is probably related to an upwelling plume of hot mantle material. McDonough et al. (1985) have postulated that for OIB and continental intraplate basalts, the composition of the tholeiites more directly reflects the nature of the plume component than that of alkali basalts. Thus, the primitive tholeiites such as sample *58* may provide evidence of a hot spot trace in southern Malagasy. The rising of the plume triggered extensive melting within the subcontinental lithosphere (White and McKenzie, 1989) under Malagasy which generated the voluminous evolved tholeiites.

The Mesozoic basaltic volcanism of Malagasy displays many similarities to other flood basalt provinces, including large volumes, rock compositions and emplacement near plate margins. The timing of the Malagasy magmatic activities is close to but probably slightly before the emplacement of the Deccan continental flood basalts in western India, where volcanic eruptions probably coincided with the Cretaceous–Tertiary boundary (Courtillot et al., 1986, 1988; R.A. Duncan and Pyle, 1988), frequently dated at 66.4 Ma (Berggren et al., 1985). Compared to the Deccan Traps which were erupted over a short interval (<2 Ma; Courtillot et al., 1988; R.A. Duncan and Pyle, 1988), the Malagasy volcanism was of noticeably longer duration, suggesting the slow migration of the Malagasy plate over the Marion–Prince Edward hot spot.

Acknowledgements

The study was supported by the Natural Sciences and Engineering Research Council of Canada (operating grant A3782) and Centre Géologique et Géophysique of Montpellier. We thank Dr. A.W. Hofmann for helpful comments.

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