

The Cameroon line, West Africa: a comparison between oceanic and continental alkaline volcanism

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SUMMARY: The Cameroon line is the only known intra-plate alkaline volcanic province which straddles a continental margin. It consists of a chain of Tertiary to Recent volcanoes stretching from the Atlantic island of Pagalu to the interior of the African continent. It therefore provides a unique area in which to study the differences, if any, between the sub-oceanic and sub-continental mantle sources for alkali basalt. Although the Cameroon line does not have a graben structure, its origin is closely linked to that of the nearby Benue rift.

Geochemical and isotopic data show no significant differences between basaltic rocks in the continental and oceanic sectors. However, the more evolved rocks in the two sectors are quite distinct. The continental magmas evolve towards peralkaline rhyolite, whereas those in the oceanic sector evolve towards phonolite. Progressive crustal contamination of the continental magmas accompanied by crystal fractionation is required to explain this distinction.

The striking geochemical similarity between basaltic rocks in the two sectors implies that their parental magmas had very similar mantle sources. This source must lie in the asthenosphere. The old lithosphere mantle beneath Africa will be chemically and isotopically different from the young Atlantic Ocean lithosphere mantle and its involvement would be readily detected in the geochemistry of the basalts. Recent models of intra-plate alkaline magma genesis have stressed the importance of metasomatic enrichment of the mantle in large-ion lithophile elements (LILE) as a precursor to magmatism. Evidence for mantle metasomatism is provided by lithosphere-derived mantle xenoliths. Since the lithosphere is the only place where large domains of enriched mantle are likely to be preserved for long periods, it follows that the Cameroon line magmas could not have had a metasomatically enriched lithosphere source.

The Cameroon line alkali basalts are chemically and isotopically similar to most other intra-plate (e.g. ocean island and continental rift) basalts which may also, therefore, have an asthenosphere rather than a metasomatized lithosphere source. The asthenosphere is also the source for mid-ocean ridge basalts (MORB) and must have a bulk composition depleted in LILE. Isotopic differences between MORB and intra-plate basalts require this source to be heterogeneous. The Cameroon line and most other intra-plate alkali basalts can be generated by small-degree melting (less than 1%) of a LILE-depleted MORB mantle source containing LILE-rich streaks.

Introduction

The Cameroon line (Fig. 1) is a prominent volcanic lineament, 1600 km long, which comprises both oceanic and continental volcanoes. It includes four large central volcanoes in Cameroon and extensive lava plateaux in Cameroon and northern Nigeria, and extends offshore into the islands and seamounts of the Gulf of Guinea. The volcanic rocks are dominantly alkaline, ranging from transitional basalt to nephelinite and alkali rhyolite to phonolite. Chemical analyses of representative volcanic rocks from the Cameroon line are given in Tables 1–3. General reviews of the geology and physical geography of various parts of the region have been written by Gèze (1943, 1953), Hedberg (1968), Mitchell-Thomé (1970), Gouhier *et al.* (1974) and Dunlop (1983).

The occurrence of contemporaneous oceanic and continental alkaline volcanism within a

single province makes the Cameroon line a unique area in which to compare oceanic and continental intra-plate volcanism. The results of a project designed to assess the differences between the sub-oceanic and sub-continental mantle sources for alkaline magmas, the melting processes required to produce them and the effects of crustal contamination upon their subsequent evolution are summarized in this paper.

Age of the Cameroon line

The earliest igneous activity on the Cameroon line is represented by a number of small (up to about 10 km) intrusive ring complexes composed of granite and syenite with less abundant gabbro and occasional remnants of trachyte and rhyolite. These extend along virtually the whole of the continental sector (Fig. 2) and give ages ranging

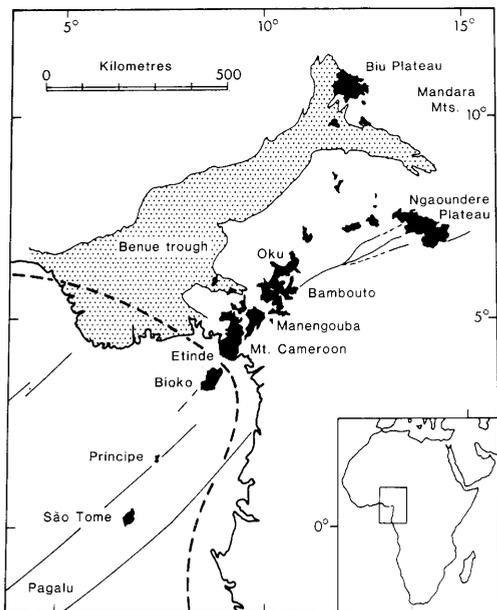


FIG. 1. Geological map of part of W Africa showing the outcrop of the Cameroon line volcanic rocks (black) and the Cretaceous sedimentary rocks of the Benue trough (stippled). The boundary between continental and oceanic crust (bold broken line) is taken from Emery & Uchupi (1984), chart XI. The Ngaoundéré fault system (between Bambouto and the Ngaoundéré Plateau) and oceanic transform faults are from Gazel (1956) and Sibuet & Mascle (1978) respectively.

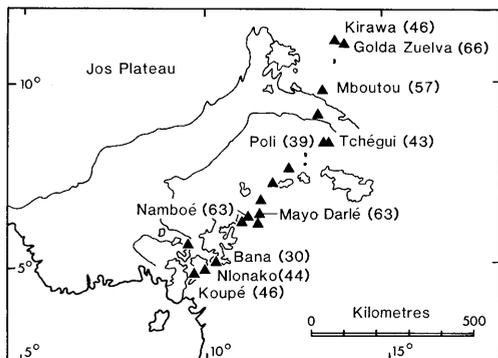


FIG. 2. The distribution of early Tertiary intrusive complexes (▲) with their ages (in Ma) given in parentheses. Data from Cantagrel *et al.* (1978), Lasserre (1978) and Jacquemin *et al.* (1982). The Benue trough and the outcrop of the late Tertiary to Quaternary extrusive rocks of the Cameroon line are shown in outline.

from 66 to 30 Ma (Cantagrel *et al.* 1978; Lasserre 1978; Jacquemin *et al.* 1982).

Seismic profiles across the oceanic sector (Grunau *et al.* 1975) show disturbance of sediments of inferred Upper Cretaceous age. Gorini & Bryan (1976) have suggested that uplift of the ocean floor along the Cameroon line has formed an effective barrier between the Niger Delta and the Douala Basin since the Cretaceous. It seems likely, therefore, that the earliest volcanism in the oceanic and continental sectors was essentially contemporaneous.

The oldest dated extrusive rocks are trachyte lavas and rhyolitic ignimbrite flows associated with a small granophyre and microsyenite ring complex at Kirawa on the Nigeria-Cameroon border (Fig. 2). The trachytes and rhyolites give Rb-Sr whole-rock ages of 51.2 Ma and 45.5 Ma respectively (Okeke 1980; Dunlop 1983). The younger extrusive centres (Fig. 1) give K-Ar ages ranging from 35 Ma (Mandara Mountains) to Recent with the oldest dated oceanic rocks (Principe) at 31 Ma (Dunlop & Fitton 1979; Dunlop 1983; Fitton & Dunlop 1985, and references cited therein). The available K-Ar data on the extrusive centres are summarized in Fig. 3. Signs of geologically recent volcanic activity can be found in most of the volcanic

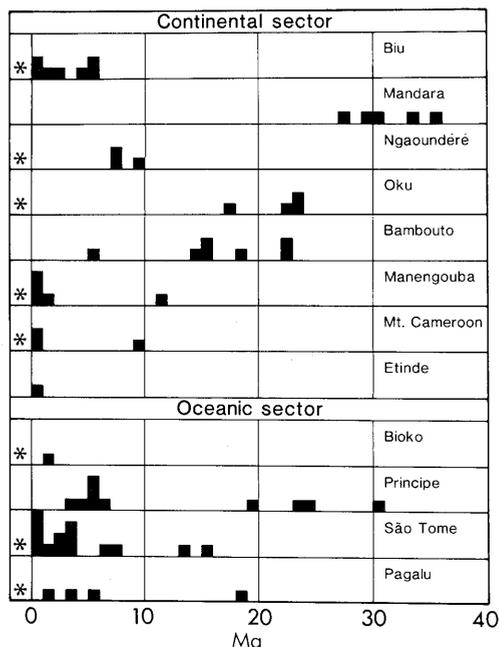


FIG. 3. Summary of available K-Ar age data for the Cameroon line volcanic rocks. Asterisks indicate the presence of morphologically recent cinder cones. (After Fitton & Dunlop 1985.)

TABLE 1. Chemical analyses of representative Cameroon line basic volcanic rocks

	Continental sector							Oceanic sector							
	N31	C147	C174	C89	C112	C65	C5	FP44	P15	P31	P18	ST72	ST96	ST19	AN15
<i>Major elements (wt.%)</i>															
SiO ₂	49.18	43.60	45.47	46.62	43.85	50.14	47.30	44.40	46.77	47.71	37.01	43.80	49.25	44.25	41.67
Al ₂ O ₃	13.69	14.50	15.35	16.86	13.65	14.21	17.20	10.80	12.45	14.07	10.88	14.06	17.43	15.30	10.62
Fe ₂ O ₃ (t)	10.94	12.66	11.55	12.50	12.52	12.40	10.32	13.33	12.66	11.90	14.28	12.46	9.50	12.95	13.76
MgO	9.78	7.42	5.47	5.56	9.21	7.94	4.25	10.80	9.22	8.15	12.24	8.31	4.01	5.63	15.52
CaO	7.94	9.78	9.34	8.50	10.55	7.60	8.87	12.58	9.88	8.00	13.19	10.62	7.08	9.50	9.70
Na ₂ O	3.42	3.32	5.08	3.39	3.47	3.48	5.26	2.48	2.55	3.05	3.10	3.61	3.74	4.76	2.36
K ₂ O	1.57	1.54	2.86	1.37	1.47	1.35	2.09	1.19	0.48	1.67	1.55	1.41	2.72	2.15	0.85
TiO ₂	2.17	3.91	2.58	3.18	3.10	2.24	2.93	3.51	2.38	2.94	4.15	3.17	2.51	3.81	3.12
MnO	0.15	0.16	0.26	0.19	0.19	0.18	0.20	0.19	0.16	0.16	0.21	0.22	0.19	0.21	0.19
P ₂ O ₅	0.47	0.74	0.91	0.86	1.03	0.40	0.92	0.63	0.41	0.67	1.42	1.03	0.99	1.02	0.82
LOI	0.14	1.79	0.93	1.10	0.19	0.08	-0.12	0.00	3.11	1.60	1.62	0.54	2.22	0.04	0.62
Total	99.45	99.42	99.79	100.13	99.21	100.01	99.21	99.90	100.07	99.92	99.95	99.23	99.64	99.61	99.23
<i>Trace elements (ppm)</i>															
Ni	282	88	54	29	148	197	18	215	236	167	193	171	7	43	569
Cr	376	129	116	46	459	275	25	521	351	231	338	381	5	33	588
V	295	503	185	182	232	193	177	294	215	231	307	237	124	234	268
Sc	24	19	15	19	23	23	10	30	24	21	26	18	6	16	25
Cu	44	43	35	22	43	57	30	72	60	42	56	48	9	41	43
Zn	96	90	122	98	96	122	103	107	113	112	108	114	128	133	118
Sr	632	1110	1266	1005	1242	429	1202	823	444	625	1297	1130	1056	1113	779
Rb	35	34	74	22	32	31	51	33	13	38	40	47	62	60	37
Zr	239	250	476	254	190	299	461	336	131	327	337	366	527	472	253
Nb	55	62	114	46	56	47	118	65	25	49	118	92	105	99	57
Ba	527	485	1000	517	874	458	650	562	174	439	987	718	854	728	516
Th	4	0	12	2	1	4	4	7	2	5	10	17	12	11	6
La	37	45	109	46	43	41	97	58	16	43	103	79	89	83	58
Ce	77	87	180	94	89	64	199	121	40	88	203	157	180	163	90
Nd	33	44	65	48	46	33	81	59	22	41	85	66	76	72	43
Y	27	25	34	28	26	29	38	32	23	29	35	33	37	36	30
<i>CIPW norms (Fe₂O₃/(FeO + Fe₂O₃) = 0.3)</i>															
or	9.41	9.42	17.25	8.23	8.82	8.07	12.52	7.08	2.96	10.13	—	8.52	16.59	12.86	5.12
ab	29.38	18.14	10.38	29.18	14.09	29.70	23.30	10.67	22.47	26.43	—	15.05	32.66	14.33	10.74
an	17.62	20.77	10.81	27.26	17.66	19.36	17.37	14.99	21.95	20.31	11.53	18.39	23.51	14.21	16.27
lc	—	—	—	—	—	—	—	—	—	—	7.40	—	—	—	—
ne	—	5.91	18.13	—	8.56	—	11.82	5.70	—	—	14.64	8.76	—	14.35	5.25
di	15.38	19.50	24.68	8.11	23.07	12.95	17.14	35.02	21.11	12.77	24.95	22.92	4.94	21.67	21.87
hy	2.45	—	—	4.37	—	12.51	—	—	14.83	8.66	—	—	4.40	—	—
ol	15.97	11.49	6.82	9.49	14.19	7.12	5.77	12.85	5.65	9.42	19.33	12.58	6.59	7.51	27.00
cs	—	—	—	—	—	—	—	—	—	—	4.63	—	—	—	—
mt	4.48	5.28	4.75	5.14	5.14	5.05	4.22	5.43	5.32	4.92	5.94	5.14	3.96	5.30	5.68
il	4.18	7.67	4.99	6.15	6.00	4.29	5.64	6.74	4.71	5.73	8.12	6.15	4.92	7.33	6.07
ap	1.12	1.82	2.19	2.07	2.48	0.95	2.21	1.51	1.00	1.62	3.46	2.49	2.43	2.44	1.99

All the analyses in Tables 1, 2 and 3 were carried out in Edinburgh using X-ray fluorescence techniques (see Fitton & Dunlop 1985). Samples were dried at 110°C before analysis. Fe₂O₃(t) is total Fe expressed as Fe₂O₃. LOI, loss on ignition at 1100°C.

N31, ol-phyric transitional basalt (5 km NNW of Tila Lake, Biu Plateau, Nigeria); C147, aphyric alkali basalt (hillside to S of Kila, Mandara Mountains, Cameroon); C174, basanite (waterfall on River Vina, 13 km S of Ngaoundéré, Cameroon); C89, plag-(ol-augite)-phyric transitional basalt (4 km SW of Kumbo, E of Mount Oku, Cameroon); C112, ol-augite-phyric basanite (2 km WSW of Lekwé Leloé, near summit of Bambouto, Cameroon); C65, plag-(ol-augite)-phyric transitional basalt (S wall of Lac de la Femme, Eboga caldera (Manengouba), Cameroon); C5, plag-augite-(ol)-phyric tephrite (summit of Mount Cameroon) (see Fitton *et al.* (1983) for more analyses of Mount Cameroon lavas); FP44, ankaramite (coast W of Pico Sta Isabel, Bioko (see Piper & Richardson 1972)); P15, olivine tholeiite from blocks in hyaloclastite breccia (River Cambungo, SE side of Principe); P31, ol-augite-phyric transitional basalt (N-S dyke, Praia de Sta. Rita, N coast of Principe); P18, olivine nephelinite (1 km SE of S. Antonio, Principe) (total includes 0.30% F); ST72, ol-hbl-d-augite-plag-phyric basanite (Maria Luisa, NW side of São Tomé); ST96, hbl-d-phyric hawaiite (NW-SE dyke, Praia Capitango, S coast of São Tomé); ST19, aphyric basanite (recent lava flow from Morro Corregado, N coast of São Tomé); AN15, picritic basalt (N side of Pagalu) (see Piper & Richardson 1972).

TABLE 2. Chemical analyses of representative Cameroon line intermediate and evolved volcanic rocks

	Continental sector							Oceanic sector							
	C144	C155	C85	C87	C104	C116	C63	P11	P12	P16	ST10	ST57	ST73	ST84	ST110
<i>Major elements (wt.%)</i>															
SiO ₂	74.08	55.75	72.70	71.13	68.56	58.19	53.99	54.90	53.98	56.28	57.39	66.69	48.54	61.66	53.64
Al ₂ O ₃	11.40	22.05	10.74	12.56	13.08	19.50	17.03	22.32	19.54	21.47	21.70	18.05	17.42	18.63	18.83
Fe ₂ O ₃ (t)	3.49	2.29	5.69	3.31	5.74	4.39	8.95	3.13	4.80	2.38	2.20	2.16	9.93	3.41	6.41
MgO	0.17	0.09	0.04	0.04	0.05	0.80	3.30	0.30	1.42	0.07	0.11	0.08	3.51	0.16	1.76
CaO	0.09	1.08	0.12	0.02	0.44	2.88	5.39	2.00	5.11	2.00	1.24	0.58	8.16	0.94	4.96
Na ₂ O	5.26	10.93	4.72	5.68	5.87	7.77	5.29	9.53	5.39	8.62	9.43	6.29	5.38	7.93	7.28
K ₂ O	4.05	5.54	4.57	5.26	4.81	4.26	3.02	6.02	4.41	4.86	6.02	4.37	2.45	5.46	3.65
TiO ₂	0.12	0.35	0.37	0.68	0.35	0.91	1.98	0.48	1.08	0.16	0.24	0.30	2.65	0.26	1.63
MnO	0.10	0.19	0.09	0.03	0.13	0.21	0.17	0.12	0.13	0.18	0.15	0.11	0.22	0.24	0.16
P ₂ O ₅	0.01	0.08	0.00	0.09	0.01	0.20	0.65	0.09	0.24	0.02	0.04	0.07	0.93	0.03	0.44
LOI	0.46	1.09	0.22	0.66	0.60	0.84	0.22	0.85	4.08	3.75	1.19	1.15	0.34	0.99	0.55
Total	99.24	99.43	99.26	99.46	99.64	99.93	99.98	99.75	100.17	99.80	99.72	99.86	99.53	99.70	99.31
<i>Trace elements (ppm)</i>															
Ni	1	2	1	1	3	2	24	3	11	2	2	2	19	1	6
Cr	3	5	4	5	3	3	22	5	14	4	4	5	23	4	5
V	7	14	0	0	0	2	91	45	56	5	15	2	173	0	79
Sc	0	0	0	2	0	0	9	0	2	0	0	0	9	0	4
Cu	0	0	0	0	0	0	14	0	4	0	0	0	21	0	7
Zn	307	131	141	261	300	75	106	79	87	185	114	79	120	174	117
Sr	4	466	4	15	6	1549	856	959	1115	11	110	231	1281	19	1128
Rb	402	212	172	207	133	101	87	133	180	314	223	124	64	200	114
Zr	1938	1291	1405	2071	1372	558	643	431	719	1745	965	515	599	1414	751
Nb	576	167	255	379	251	152	107	79	119	136	82	141	133	264	116
Ba	9	406	4	80	31	2501	701	1087	758	11	154	1361	870	70	845
Th	81	31	38	39	30	6	13	13	26	76	34	20	18	39	17
La	41	60	20	325	249	116	76	65	68	98	76	148	105	199	79
Ce	153	105	18	659	398	183	148	93	111	110	95	254	197	338	148
Nd	29	27	19	267	188	73	58	23	36	17	18	78	79	88	52
Y	78	22	26	165	112	49	32	15	20	19	13	38	38	43	27
<i>CIPW norms (Fe₂O₃/(FeO + Fe₂O₃) = 0.3)</i>															
Q	29.92	—	29.11	23.18	16.18	—	—	—	—	—	—	11.58	—	—	—
C	—	—	—	—	—	—	—	—	—	—	—	2.12	—	—	—
or	24.29	33.33	27.40	31.57	28.79	25.47	18.02	36.08	27.19	29.97	36.19	26.20	14.72	32.76	21.97
ab	36.62	25.76	30.18	35.83	41.10	45.74	43.21	21.47	35.23	37.73	29.02	54.00	26.48	50.12	34.35
an	—	—	—	—	—	5.83	13.90	0.34	16.79	5.74	—	2.45	16.40	—	8.03
ne	—	31.61	—	—	—	11.27	1.05	32.65	6.73	20.78	27.26	—	10.66	8.86	15.33
ac	2.85	1.88	4.64	2.70	4.69	—	—	—	—	—	1.58	—	—	1.47	—
ns	1.24	1.86	1.17	2.29	0.91	—	—	—	—	—	—	—	—	—	—
di	0.35	4.31	0.51	—	1.92	6.13	7.18	6.28	6.60	3.98	5.26	—	15.08	3.98	11.57
wo	—	—	—	—	—	—	—	0.78	—	—	—	—	—	—	—
hy	4.46	—	6.26	3.04	5.72	—	—	—	—	—	—	2.03	—	—	—
ol	—	0.38	—	—	—	1.56	7.65	—	2.69	0.43	0.02	—	5.24	1.59	1.91
mt	—	—	—	—	—	1.79	3.64	1.28	2.02	1.00	0.11	0.88	4.07	0.66	2.63
il	0.23	0.68	0.71	1.31	0.68	1.74	3.80	0.92	2.13	0.31	0.46	0.58	5.10	0.51	3.14
ap	0.03	0.20	0.01	0.21	0.01	0.48	1.55	0.21	0.60	0.04	0.10	0.17	2.25	0.07	1.07

LOI, loss on ignition.

C144, rhyolite (Aiguille Mchirgui, 2 km N of Mogode, Mandara Mountains, Cameroon); C155, phonolite (plug near Beka, 20 km N of Ngaoundéré, Cameroon); C85, rhyolite (flow) (Sabga Pass, 16 km ENE of Bamenda; SW of Mount Oku, Cameroon); C87, rhyolitic welded tuff (6 km SW of Jakiri; SE of Mount Oku, Cameroon); C104, trachyte (Pinyin, N slopes of Bambouto, Cameroon); C116, trachyphonolite (Bambouto summit, Cameroon); C63, benmoreite (NW rim of Eboga caldera (Manengouba), Cameroon); P11, phonolite (large block in River Fria, SE side of Principe); P12, tristanite (Os Dois Irmãos (plug), SE side of Principe); P16, trachyphonolite (large block in River Cambungo, SE side of Principe); ST10, phonolite (Cão Grande (plug), S side of São Tomé); ST57, quartz trachyte (NE peak of Ilhéu das Cabras, off NE coast of São Tomé); ST73, hblid-plagiophytic mugearite (Maria Luisa, NW side of São Tomé); ST84, trachyte (S side of Mizambu (plug), E side of São Tomé); ST110, phonolitic tephrite (River Manuel Jorge, 0.5 km NW of Sta. Luisa, E side of São Tomé).

TABLE 3. Chemical analyses of representative nephelinite lavas from Etinde, Cameroon

	C150	C24	C152	C131	C154
<i>Major elements (wt.%)</i>					
SiO ₂	39.60	40.04	39.99	42.65	46.25
Al ₂ O ₃	12.75	14.56	17.54	17.39	19.46
Fe ₂ O ₃ (t)	13.91	12.34	9.69	9.94	6.64
MgO	8.03	5.28	3.74	2.70	1.16
CaO	14.98	14.00	10.87	8.79	6.15
Na ₂ O	2.68	3.47	6.94	6.41	7.80
K ₂ O	1.66	2.61	3.87	4.48	6.22
TiO ₂	4.49	4.06	2.95	2.54	1.06
MnO	0.20	0.30	0.28	0.38	0.36
P ₂ O ₅	0.67	1.18	0.89	0.68	0.16
SrO	0.12	0.28	0.28	0.33	0.78
BaO	0.09	0.12	0.11	0.14	0.38
SO ₃	0.00	0.24	1.94	1.41	0.27
Cl	0.04	0.10	0.37	0.49	0.47
LOI	1.00	1.31	0.79	1.95	3.06
Total	100.21	99.87	100.17	100.18	100.11
<i>Trace elements (ppm)</i>					
Ni	46	12	10	6	2
Cr	126	8	5	6	3
V	456	448	342	416	241
Sc	45	15	4	2	0
Cu	113	64	84	38	5
Zn	93	127	134	194	209
Sr	1054	2401	2395	2825	6553
Rb	89	87	114	349	201
Zr	399	656	626	888	914
Nb	96	223	281	320	295
Ba	788	1067	1006	1254	3424
Th	8	25	21	15	4
La	97	238	225	170	131
Ce	195	489	425	272	168
Nd	84	180	142	86	41
Y	28	47	44	52	37
<i>CIPW norms (Fe₂O₃/(FeO + Fe₂O₃) = 0.3)</i>					
or	—	—	—	18.90	23.20
an	18.36	18.12	13.68	12.61	2.60
lc	7.84	12.40	18.20	6.51	11.83
ne	12.36	15.02	23.79	23.03	34.29
hl	0.07	0.17	0.62	0.83	0.81
th	—	0.44	3.49	2.57	0.50
di	35.37	36.17	25.23	23.75	16.90
wo	—	—	—	—	4.59
ol	7.13	1.38	1.64	1.05	—
cs	2.85	0.41	1.55	—	—
mt	5.72	5.11	3.97	4.12	2.79
il	8.69	7.91	5.69	4.96	2.10
ap	1.62	2.87	2.14	1.66	0.39

LOI, loss on ignition; C150, olivine melanephelinite; C24, nephelinite; C152 and C131, h aüyne nephelinite; C154, leucocratic nosean leucite nephelinite. All the samples were collected from loose blocks in stream beds: C24 from Bonenza, and the others from Batoke.

centres (Fig. 3). Mount Cameroon is still active and last erupted in 1982 (D eruelle *et al.* 1983; Fitton *et al.* 1983). It is clear that volcanic activity on both the oceanic and continental sectors of the Cameroon line has been more or less continuous since the end of the Cretaceous.

Geology and petrology

Continental sector

Intrusive complexes

Relatively little is known about the mineralogy and geochemistry of these complexes. That which is known suggests that they are broadly similar to the ring complexes of the Jos Plateau in Nigeria (Bowden *et al.* 1987). Some of the granite bodies in the Mayo Darl e area contain traces of cassiterite (Chaput *et al.* 1954; Gazel *et al.* 1963). Only the Mbutou complex (Fig. 2) has been described in detail. Parsons *et al.* (1986) have published an account of the mineralogy and crystallization history of this complex and stress the very mildly alkaline nature of its parental magma. Jacquemin *et al.* (1982) have carried out a geochemical and isotopic study of the Golda Zuelva and Mbutou complexes and conclude that the magmas, although initially mantle derived, have interacted with crustal rocks as they crystallized.

The intrusive complexes were not included in the present study because they are only exposed in the continental sector. Furthermore, the effects of crystal accumulation and late-stage alteration processes limit the usefulness of coarse-grained rocks in geochemical studies. The following discussion will therefore be confined entirely to the younger extrusive centres shown on Fig. 1.

Biu Plateau

The small town of Biu in NE Nigeria is built on a plateau composed of basaltic lava flows. These reach a maximum thickness of 250 m (J. W. du Preez, cited by Carter *et al.* 1963) and range in composition from basanite to *hy*-normative transitional basalt. The most recent volcanism in the area produced a large number of cinder cones aligned in a NNW–SSE direction. These usually have well-defined craters which have often been breached by small lava flows. K–Ar ages on the plateau basalts range from 5.3 Ma for the basal lavas to about 0.8 Ma for a flow high in the succession (Grant *et al.* 1972; Fitton & Dunlop 1985).

The lava flows and cinder cones often contain abundant peridotite xenoliths and clinopyroxene and anorthoclase megacrysts. A large crater to

the west of the village of Miringa (14 km NNW of Biu) is particularly noteworthy for its abundant and extensive inclusion suite. These include peridotite xenoliths up to 0.5 m in diameter and large megacrysts of clinopyroxene, garnet, anorthoclase, ilmenite and amphibole (Wright 1970; Frisch & Wright 1971).

Mandara Mountains

The Mandara Mountains lie along the Cameroon–Nigeria border and are composed of granite and gneiss of Pan-African age. In the area around the small settlements of Roumsiki and Mogode the basement rocks have been intruded by numerous plugs of peralkaline trachyte and rhyolite which form spectacular spines. Occasional remnants of alkali basalt lava flows are also found in this area. K–Ar dating (Fitton & Dunlop 1985) gives ages of 30 and 33 Ma for two samples of basalt and 35 Ma for one of the trachyte plugs.

Ngaoundéré Plateau

The town of Ngaoundéré in northern Cameroon lies on the western edge of an extensive basalt plateau. The earliest lavas were erupted into two broad valleys eroded into the Precambrian basement by the headwaters of the Rivers Bini and Vina. Continued volcanism built a plateau of alkali basalt across the area and culminated in the building of a central volcano, Nganha. This volcano is composed of basanite capped by trachyte and phonolite flows which have yielded K–Ar ages of 10–7 Ma (Gouhier *et al.* 1974). Trachyte flows also occur as small outliers over the whole plateau as do trachyte and phonolite plugs. The most recent volcanism is represented by cinder cones aligned in a WNW–ESE direction. These have sometimes produced small lava flows.

All the basic lavas in the area are *ne* normative and range from alkali basalt to basanite. Some of the flows contain peridotite xenoliths and clinopyroxene, anorthoclase and rare zircon megacrysts.

Oku

Mount Oku (3011 m) is a large deeply dissected volcanic massif composed largely of rhyolite and quartz trachyte flows, domes and plugs. Rhyolitic ignimbrite sheets are common in the area and extend as far south as the town of Bamenda (midway between Oku and Bambouto) which sits at the base of a prominent escarpment of welded tuff. The lower part of the volcanic succession is dominated by flows of basalt and hawaiite,

although basic lava flows also occur higher in the succession. K–Ar ages obtained from Oku lavas range from 23 to 17 Ma (Gouhier *et al.* 1974; Fitton & Dunlop 1985). The basic lavas range in composition from slightly *ne* normative to slightly *hy* normative. The volcanic rocks as a whole form a strongly bimodal suite.

Numerous cinder cones (some yielding peridotite xenoliths) and explosion craters on and around Mount Oku provide evidence for recent volcanic activity in the area although very few of the vents produced lava flows. The craters reach 2 km in diameter and many form crater lakes. Two of these (Lake Monoun to the south and Lake Nyos to the NW of Oku) recently released large volumes of carbon dioxide with catastrophic results (SEAN 1985, 1986). The cause has not yet been established.

Bambouto

Unlike Oku, Bambouto still retains some of its original volcanic form despite extensive erosion. The highest peaks (up to 2740 m) lie around a calder rim 10 km in diameter. Tchoua (1972) identified two episodes of caldera collapse.

As with Oku, the lava succession comprises a strongly bimodal basalt–trachyte suite. Basalt lavas dominate the basal part of the succession but are also found on the upper slopes of the volcano. The Bambouto lavas are generally more undersaturated than the Oku suite. All the basic lava samples analysed are slightly *ne* normative and the salic rocks range from quartz trachyte to trachyphonolite. Rhyolites have not been found by the present author although Tchoua (1973) has reported extensive ignimbrite sheets on and around Bambouto.

K–Ar dating of lavas from Bambouto gives ages ranging from 23 to 14 Ma (Gouhier *et al.* 1974; Fitton & Dunlop 1985). Gouhier *et al.* also report an age of 5.8 Ma for a basalt sample collected about 20 km N of the caldera rim. There is no evidence for recent volcanic activity on Bambouto.

Manengouba

Manengouba is a well-preserved central volcano whose summit region is occupied by two concentric calderas. A poorly defined outer caldera (Elengoum; 6 km in diameter) encloses the younger perfectly preserved Eboga caldera (3 km in diameter). The highest point on the mountain (2411 m) lies on the Elengoum rim. This outer rim was breached on its eastern side during the building of the Eboga volcano and allowed large

volumes of lava to flow over the north-eastern flanks of the older volcano.

The oldest Manengouba lava flow so far dated gives a K–Ar age of 1.55 Ma (Gouhier *et al.* 1974). A recent detailed K–Ar and field study of the volcano (C. A. Hirst, unpublished data) has confirmed this date and established a chronology for the evolution of the volcano. The collapse of the Elengoum caldera probably occurred between 0.8 and 0.6 Ma ago. The Eboga lavas give ages ranging from 0.56 to 0.26 Ma, and the collapse of the Eboga caldera is thought to have occurred at about 0.25 Ma. An isolated trachyte spine intruding the eastern rim of the Elengoum caldera gave a K–Ar age of 0.26 Ma. More recent volcanism has produced a SW–NE line of crater lakes and cinder cones across the floor of the Eboga caldera.

The lavas range in composition from *ne-* and *hy-*normative basalts to trachyte, quartz trachyte and rare rhyolite. Intermediate lavas are common. In striking contrast with the lavas of Oku and Bambouto, the Manengouba lavas form a complete compositional continuum.

Mount Cameroon

At 4095 m, Mount Cameroon is by far the highest mountain in West Africa and one of Africa's largest volcanoes. The base of the lava pile is below sea level. Mount Cameroon is the only currently active centre of the Cameroon line and has had five recorded eruptions this century (Fitton *et al.* 1983, and references cited therein). It is a composite volcano built of alkali basalt to basanite lava flows interbedded with small amounts of pyroclastic material. Small cinder cones are abundant in several areas on the flanks of the volcano and on the surrounding lowlands. These cones are often aligned along fissures running SW–NE, parallel to the long axis of Mount Cameroon. The Plain of Tombel, between Mount Cameroon and Manengouba, is likewise dotted with numerous cinder cones and explosion craters, some of which are occupied by lakes. The largest of these, Lake Barombi Mbo near the town of Kumbo, is surrounded by a tuff ring containing abundant peridotite xenoliths. The Mount Cameroon lavas are all *ne* normative and range from almost aphyric to strongly porphyritic (olivine ± plagioclase). Lavas more evolved than hawaiite have not been reported.

The few K–Ar dates available (Hedberg 1968) suggest that all the exposed lavas were erupted in the last million years although there is field evidence that the volcanic edifice was built upon a basement of older lava flows. These are best exposed along the coast to the SE of Limbe (formerly Victoria) but have not yet been dated.

Etinde

Mount Cameroon is not a single volcano but has a large subsidiary peak, Etinde, on its SW flank. Etinde rises from the rain forest a few kilometres from the coast and is densely forested up to its summit (1713 m). Exposures of lava are rare and deeply weathered but the numerous streams contain blocks of fresh volcanic rock unlike anything found on the main volcanic massif. These provide evidence that Etinde is composed entirely of nephelinite lavas. The lavas were first described by Esch (1901).

Attempts to date the Etinde lavas have not yet produced consistent results. A sample of nosean leucite nephelinite gave a K–Ar age of about 0.1 Ma (Fitton & Dunlop 1985). R. M. MacIntyre (unpublished data) has obtained K–Ar ages of 1.1, 0.6 and 6.3 Ma from, respectively, a melane-nphelinite whole-rock sample and nepheline and h a yne separated from two different nephelinite samples. It is possible that the apparent age of the h a yne separate is anomalous and is due to excess radiogenic argon. The lavas appear to form a consanguineous series, and it would therefore be unwise to postulate an eruptive history of over 6 Ma on the basis of this one determination. It seems likely that the lavas are less than a million years old and possibly contemporaneous with some of the Mount Cameroon lavas. More work is needed to establish the relative ages of the two volcanic centres.

The lavas form a compositional continuum from olivine melanephelinite through nephelinite and h a yne–nephelinite to a leucocratic nosean leucite–nephelinite (Table 3). Feldspar is absent throughout the whole series and melilite is observed in a few samples. The observed chemical variation among the lavas can be explained by crystal fractionation dominated by aluminous clinopyroxene. This gave rise to residual liquids close in composition to the nepheline–leucite–diopside cotectic in the system nepheline–sanidine–diopside (Fig. 4). These highly evolved compositions contain very high concentrations of some incompatible elements. The nosean leucite–nephelinite lavas, for example, contain around 7000 ppm Sr which is present as an unusual strontian melilite (Fitton & Hughes 1981).

Oceanic sector

Bioko (formerly Fernando P o)

Despite being only 50 km from the mainland, Bioko is, geologically, the least well known of the Gulf of Guinea islands. From the time the island gained independence from Spain in 1968 until the coup of 1979, the brutal regime of President

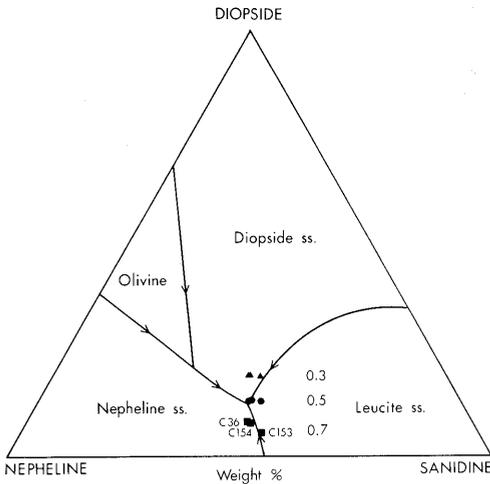


FIG. 4. Normative compositions of three leucocratic nosean leucite-nephelinite samples from Etinde projected into the system nepheline-sanidine-diopside (Platt & Edgar 1972). The norms were calculated using $\text{Fe}_2\text{O}_3/(\text{FeO} + \text{Fe}_2\text{O}_3)$ values of 0.3, 0.5 and 0.7. Chemical analysis of C154 gives a value of 0.659 for this ratio (Fitton & Hughes 1981).

Macias Nguema ensured complete inaccessibility. The current social and economic state of the island has been described by Winchester (1984) who, 4 years after the execution of the President, claimed that Bioko was 'the nastiest place on earth'. Consequently our knowledge of the geology and petrology of the island is based on pre-independence field work and rock collections. The most recent geological expedition to the island was undertaken as part of a palaeomagnetic study of the Cameroon line (Piper & Richardson 1972). The cores they collected were used in the present study. The earlier literature on the geology of Bioko and the other Gulf of Guinea islands has been reviewed by Mitchell-Thomé (1970).

Bioko is a youthful volcanic island with three eruptive centres. Pico Santa Isabel, the highest point (2972 m), forms the northern part of the island and the other two volcanoes, San Carlos and Pico Biao, form the southern part. The lavas range from basanite to *hy*-normative basalt. Unlike the other Gulf of Guinea islands, more evolved rocks have not been reported from Bioko.

A single K-Ar age of 1.1 Ma was obtained by Hedberg (1968) from a basanite sample collected near the town of Santa Isabel on the northern coast. Nearly all the palaeomagnetic sites investigated by Piper & Richardson (1972) are normally magnetized which led these workers to conclude that virtually the whole volcanic edifice

was formed in the Brunhes epoch and is thus younger than 0.7 Ma. Numerous cinder cones occur on all three volcanic centres although there have been no well-documented reports of eruptions. What may have been a small eruption on the east side of the island was reported by natives around the beginning of this century (Hedberg 1968).

Principe

Principe is a small and deeply eroded island surrounded by a broad submarine shelf. The prevailing SE trade winds in the Gulf of Guinea veer SW as they approach the mainland. Therefore the south-western parts of the islands receive the highest rainfall and consequently suffer the most erosion.

The structure, volcanic stratigraphy and petrology of Principe have been described by Fitton & Hughes (1977), and a K-Ar chronology has been established by Dunlop & Fitton (1979) with an additional determination by Hedberg (1968). The oldest exposed rocks are hyaloclastite breccias which crop out in river valleys in the south-eastern part of the island. These represent a submarine phase in the evolution of the island and contain blocks of fresh olivine tholeiite dated at 31 Ma. The oldest sub-aerial lavas (the Older Lava Series) are exposed mostly in the north and range in composition from transitional basalt to hawaiite. They give K-Ar ages ranging from 24 to 19 Ma (Hedberg 1968; Dunlop & Fitton 1979) and are intruded by numerous dykes and small intrusions compositionally similar to the lavas. The Older Lava Series is unconformably overlain by nephelinite and basanite lava flows which have yielded K-Ar ages of 5.6 and 3.5 Ma. This Younger Lava Series is overlain by phonolite lavas in the northern part of the island. The high ground to the south (up to 948 m) is composed of a chaotic assemblage of lava flows and plugs ranging in composition from tristanite to trachyte and phonolite. Four K-Ar determinations on samples collected from these plugs give ages ranging from 7 to 5.3 Ma. There are no signs of recent volcanic activity on Principe.

São Tomé

São Tomé is a roughly conical volcanic island rising from about 3000 m below sea level to a height of 2024 m. The ground rises gently in the drier northern and eastern parts of the island but is much more rugged in the S and W. Phonolite and trachyte spines rising vertically out of the rain forest form a spectacular feature of the southern part of the island where rainfall and erosion rates are highest.

The oldest rocks on the island are conglomerates, sandstones and shales of the Ubabudo Formation (Hedberg 1968). These crop out in stream sections in a small area in the eastern part of the island and are accompanied by oil seepages. They are almost non-fossiliferous and cannot be

dated with certainty, but Hedberg (1968) has suggested a Cretaceous age on the basis of lithological comparisons with Cretaceous sandstones in Gabon and on radiolaria and foraminifera extracted from the shales.

The volcanic succession is dominated by basic

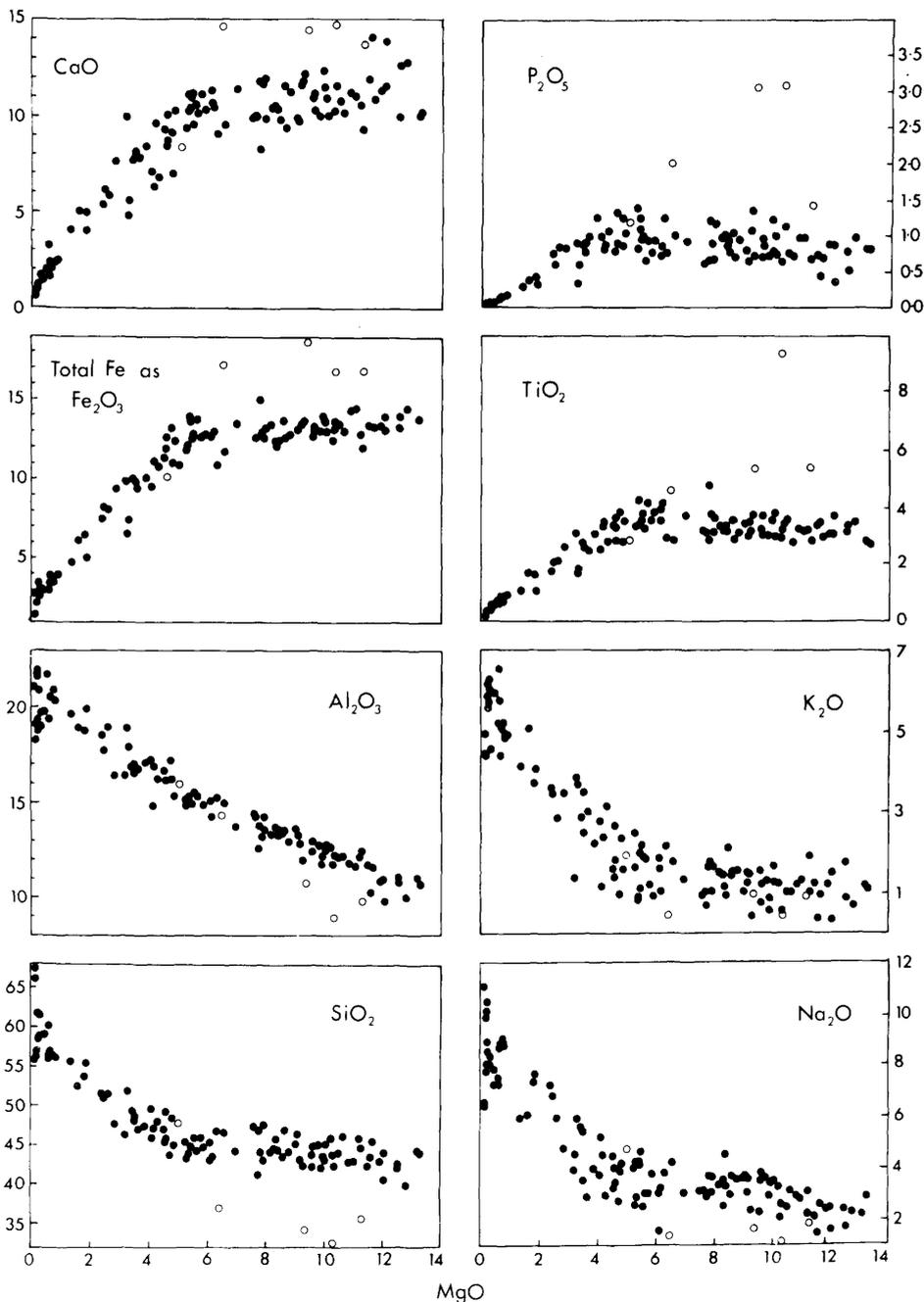


FIG. 5. ● Major-element variation (wt.%) among the volcanic rocks of São Tomé; ○, analyses of cognate xenoliths collected from intermediate and evolved rocks.

lavas, although evolved rocks are common and form part of a continuum from basalt to trachyte and phonolite with no discernible composition gaps (Fig. 5). The chemical variation can be explained by crystal fractionation involving the observed phenocryst phases. Fractionation of the more magnesian magmas ($MgO > 6\%$) was dominated by olivine + augite. The inflection in the variation diagrams marks the onset of plagioclase, hornblende, magnetite and apatite crystallization. Cognate xenoliths composed of these four phases with or without augite are abundant in some of the intermediate and more evolved lavas (open circles in Fig. 5).

The basic lavas range in composition from basanite to *hy*-normative basalt with no clear correlation between degree of undersaturation and age (*cf.* Principe). It may be significant, however, that the rare quartz trachytes are the oldest dated volcanic rocks on São Tomé. Grunau *et al.* (1975) have obtained an age of 15.7 Ma, but they do not state where the sample was collected. The quartz trachyte plug forming the Ilhéu das Cabras off the NE coast gives a K–Ar age of 13 Ma (Fitton & Dunlop 1985). All other dated volcanic rocks from São Tomé are much younger with ages ranging from 7.6 to 0.1 Ma (Hedberg 1968; Fitton & Dunlop 1985). Evolved rocks were erupted over most of this period, although lavas erupted over the last million years appear to have been entirely basaltic. Recent cinder cones are very common over most of the island.

Pagalú (formerly Annobon)

Pagalú is the southernmost and smallest of the Gulf of Guinea islands. After independence from Spain it joined Bioko as part of Equatorial Guinea, thereby becoming virtually inaccessible. A suite of rock samples, however, was collected by Piper & Richardson (1972) and these samples were used in the present study. More recently, a visit to the island by a French research vessel resulted in accounts of the geology of the island and the petrology and geochemistry of its volcanic rocks published by Cornen & Maury (1980) and Liotard *et al.* (1982). The following summary of the geology of Pagalú is based largely on these studies.

The oldest rocks on the island are hyaloclastite breccias which are exposed around the coast. They contain large clinopyroxene megacrysts and are intruded by numerous basaltic dykes. Neither the breccias nor the dykes have been dated. This basal unit is overlain by basic lavas ranging in composition from basanite to *hy*-normative basalt. The lowest flow in this succession has been dated at 18.4 Ma by Piper & Richardson (1972).

The lava pile has been intruded by tristanite (3.9 Ma) and trachyte plugs and by numerous basanite dykes, one of which gave a K–Ar age of 5.4 Ma (Cornen & Maury 1980). The most recent basaltic lava flows have been erupted from a well-preserved crater, now occupied by a lake. These flows contain abundant peridotite xenoliths. A lava sample from one flow gave a K–Ar age of 2.6 Ma (Piper & Richardson 1972).

The geological history and volcanic rocks of Pagalú are broadly similar to those of Principe. The main difference is that quartz trachyte is the most common salic rock type on Pagalú whereas phonolite is the most abundant on Principe. However, of the three quartz trachyte analyses presented by Cornen & Maury (1980), the two with significant amounts of quartz in their norms are also corundum normative (as is the quartz trachyte from São Tomé; Table 2) which suggests that these rocks have suffered some alkali loss. It is not clear, therefore, whether the original magma was oversaturated or undersaturated. The Pagalú basalts are strikingly magnesian, a feature first noted by Tyrrell (1934). Although not unusually rich in olivine phenocrysts, they contain up to 16.5% MgO.

Origin of the Cameroon line

From the above discussion it is clear that there is no evidence of age progression of volcanic activity on the Cameroon line. Magmatism in both oceanic and continental sectors appears to have commenced in the late Cretaceous, and most of the volcanic centres have shown some activity in the course of the last million years or so. The only currently active volcano (Mount Cameroon) lies half-way along the line. The Cameroon line cannot, therefore, be a hot-spot trail as suggested by Van Houten (1983). Over the past 60 Ma, Africa has rotated anticlockwise and drifted northwards by about 12° (Smith *et al.* 1981). If the Cameroon line owes its origin to a deep mantle plume then this must have had the form of a sheet and have been moving so as to keep pace exactly with the lithosphere.

Magmatism on the Cameroon line was accompanied by regional uplift. The average height of the Precambrian basement beneath and around the large volcanic centres of Bambouto and Oku, for example, is about 1200 m above sea level with peaks up to 2000 m. The occurrence of inliers of Cretaceous sandstone on São Tomé suggests that the oceanic basement has been similarly uplifted. This uplift is presumably an isostatic response to low-density (? partially melted) mantle beneath the region.

Freeth (1979) has suggested that the Cameroon line is an extensional feature resulting from membrane stresses generated by the northward movement of the African plate away from the equator. This is consistent with regional uplift and alkaline magmatism but direct field evidence for extension is lacking. Freeth (1979) estimates that about 1 km of extension has taken place across the Cameroon line. No rift faulting or graben structures have been observed along the continental sector, however, and most of the magmatism has been associated with central rather than fissure volcanoes. Furthermore, the intrusive complexes were emplaced at a time when the region should have been under compression (Freeth 1979).

For part of its length the Cameroon line runs almost parallel to a major crustal shear zone, the Ngaoundéré fault (Fig. 1), which was a continuation of the Pernambuco lineament in Brazil before continental separation. Gorini & Bryan (1976) have suggested that the Cameroon line volcanism was caused by reactivation of this ancient lineament. Evidence for this reactivation is provided by the deformation and metamorphism of Cretaceous conglomerates close to the fault zone (Vincent 1968). This explanation is weakened, however, by the lack of evidence that the lineament has exerted any influence over the volcanic activity. For example, the WNW–ESE-trending fissures in the Ngaoundéré Plateau (marked by lines of cinder cones) cut across the fault zone at an angle of about 70° . The oceanic part of the Cameroon line is similarly unaffected by the transform faults it crosses (Fig. 1).

An explanation for the Cameroon line may lie in its relationship with the Benue trough (Fitton 1980, 1983). The two features are so remarkably similar in shape and size (Fig. 1) that they can be superimposed exactly by rotating one with respect to the other about a pole in Sudan. Fitton (1980) suggested that this is not a fortuitous coincidence but results from a displacement of the African lithosphere relative to the asthenosphere. In this model the Y-shaped asthenosphere hot zone which would have underlain the Benue trough during its extensional phase in the Cretaceous became displaced relative to the lithosphere so that it now lies beneath Cameroon and the Gulf of Guinea. The Cameroon line and Benue trough are therefore regarded as complementary features. The former is a line of volcanoes with rift valley affinities but lacking rift faulting. The latter is a rift valley containing relatively few volcanic rocks. Magmas originally destined for the Benue rift reached the surface as the Cameroon line instead. The sequence of events postulated in this model is summarized in Fig. 6.

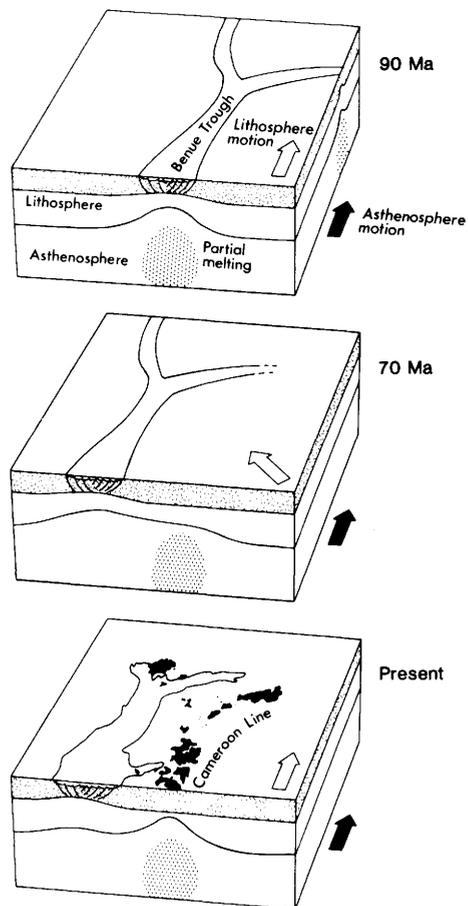


FIG. 6. Postulated sequence of events leading to the development of the Cameroon line. The block diagrams represent segments of crust and upper mantle measuring 1000 km square by 200 km deep. (From Fitton 1983.)

Whatever the cause of the Cameroon line volcanism, there can be little doubt that its oceanic and continental components share a common origin. They need not, however, have similar mantle sources. The geochemistry of the Cameroon line basic volcanic rocks has been useful in constraining the composition of the mantle sources for oceanic and continental alkali basalts. These constraints will be discussed in the next section.

Mantle sources and generation of the basic magmas

Intra-plate basaltic rocks, both oceanic and continental, are strikingly enriched in large-ion lithophile elements (LILE) by comparison with mid-ocean ridge basalts (MORB). The large

degree of asthenosphere melting inferred to take place in a relatively shallow melting zone beneath mid-ocean ridges allows the composition of the MORB source to be closely constrained. This source must be depleted in LILE (e.g. K, Rb, Ba and light-rare-earth elements) in comparison with the bulk Earth. Isotope ratios of Sr and Nd show that the MORB source must have had low Rb/Sr and Nd/Sm ratios (and by implication have been depleted in other LILE) for considerable periods of time. Isotope ratios in most intra-plate basic volcanic rocks also imply a long-term LILE-depleted source, although not quite as depleted as the MORB source. This observation has posed a major problem in the understanding of intra-plate volcanism. How can such LILE-rich magmas be generated from a LILE-depleted source?

One solution involves very small degrees of partial melting (less than 1%) in order to concentrate LILE (which are incompatible in mantle phases) into the melt. Such small melt fractions have been considered impossible to extract and simple melting models have been rejected as a consequence. Complex melting processes such as zone refining (Harris 1957) and wall-rock reaction (Green & Ringwood 1967) have been proposed to account for the concentration of K and other incompatible elements into a melt. More recent attempts to account for the high LILE concentrations in intra-plate volcanic rocks have invoked metasomatic enrichment of the mantle source shortly before it is melted. The current popularity of mantle metasomatism has led to a proliferation of *ad hoc* hypotheses involving variable degrees of enrichment through the agency of CO₂- and H₂O-rich fluids originating in deeper parts of the mantle. Such hypotheses have been applied to both oceanic (Clague & Frey 1982; Wright 1984) and continental (Lloyd & Bailey 1975; Frey *et al.* 1978; Menzies & Murthy 1980; Bailey 1982, 1987) intra-plate magmas.

Direct evidence for the existence of enriched mantle is provided by the occurrence of metasomatized peridotite xenoliths in alkali basalt and kimberlite. These xenoliths almost certainly originate within the continental lithosphere which is the only part of the Earth where large domains of enriched mantle are likely to be preserved for very long periods. The oceanic lithosphere is young and at least as depleted in LILE as the asthenosphere, which accounts for the scarcity of metasomatized mantle xenoliths in oceanic volcanic rocks. The asthenosphere will be well stirred by convection and therefore homogeneous on the scale of individual convection cells. Where the asthenosphere does become locally enriched by metasomatic processes (as it must do above subduction zones, for example), convection will carry the enriched material away from the site of enrichment and mix it with normal depleted asthenosphere. If metasomatically enriched lithosphere mantle plays an important role in the generation of magmas then we would expect significant geochemical differences between oceanic and continental intra-plate basic volcanic rocks. If such differences do not exist then an asthenosphere source is implied and the case for mantle metasomatism as a necessary precursor to intra-plate magmatism will be weakened.

In an attempt to detect systematic differences between oceanic and continental intra-plate volcanism Fitton & Dunlop (1985) analysed a large number of volcanic rocks from the oceanic and continental sectors of the Cameroon line for major and trace elements and Sr isotope ratios. The compositional range of the oceanic and continental sector basic (MgO > 4%) volcanic rocks is shown in Fig. 7. The highly undersaturated Etinde nephelinites are not included in this diagram. Average incompatible-element concentrations in basic rocks from the main volcanic centres are shown in Fig. 8. Despite minor

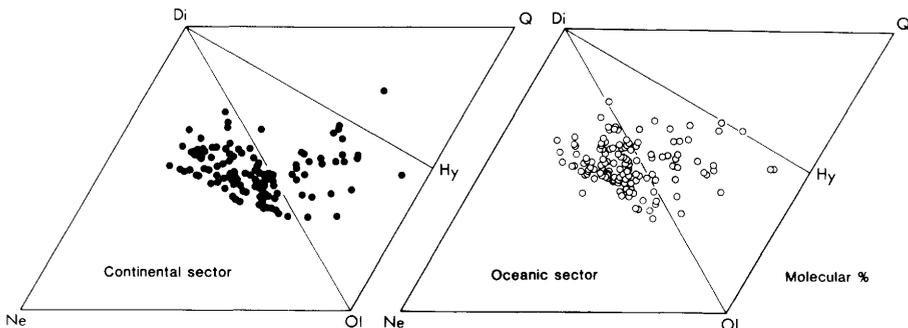


FIG. 7. Normative composition of Cameroon line basic volcanic rocks (MgO > 4%) (Etinde nephelinite data not included). Fe₂O₃/(FeO + Fe₂O₃) is normalized to 0.3 for the norm calculations.

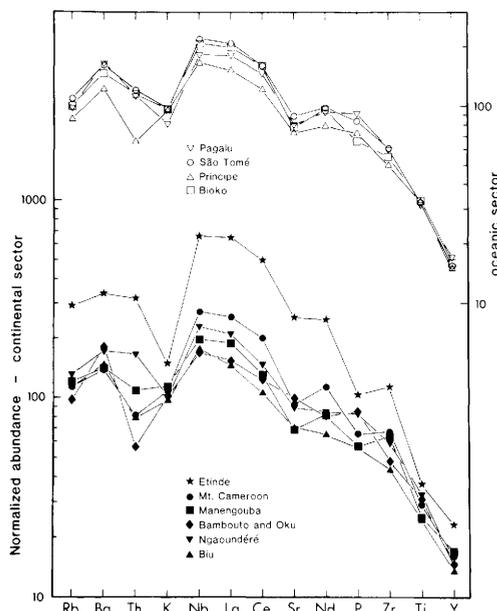


FIG. 8. Average incompatible-element abundances in basic volcanic rocks ($MgO > 4\%$) from the main volcanic centres of the Cameroon line. Number of analyses: Pagalu, 16; São Tomé, 77; Principe, 28; Bioko, 23; Etinde, 15; Mount Cameroon, 24; Manengouba, 15; Bambouto and Oku, 22; Ngaoundéré, 27; Biu, 35. Concentrations have been normalized to chondrite and primitive mantle abundances (Sun 1980). The spread in Th data is due to analytical uncertainty.

chemical variation between centres it is clear that there are no systematic differences between the oceanic and continental suites. The basic volcanic rocks in each of the centres could have been generated by melting very similar mantle sources. Smaller degrees of melting followed by extensive crystal fractionation can account for the higher incompatible-element concentrations in the Etinde nephelinite lavas.

The remarkable similarity between the oceanic and continental basic volcanic rocks is illustrated in Fig. 9 in which the average concentrations of incompatible elements in all the samples (except the Etinde nephelinites) from the two sectors are compared. The two suites of basic rocks are also isotopically indistinguishable (Fitton & Dunlop 1985). These data imply that the oceanic and continental basalts had very similar mantle sources. Since it is most unlikely that the ancient lithosphere mantle beneath the continental sector is chemically and isotopically similar to the young lithosphere mantle beneath the oceanic sector, it follows that lithospheric mantle was not the source of the Cameroon line basalts. The lack of any consistent migration of volcanism with time

rules out a source below the 670 km discontinuity. If the Cameroon line were the product of a deep-mantle plume it would require that lower-mantle convection has kept pace exactly, in both velocity and direction, with the movement of the African plate over the past 65 Ma. Therefore the convecting upper mantle is the only plausible mantle source.

The convecting upper mantle (of which the asthenosphere is the upper part) is also the source of MORB. Fitton & Dunlop (1985) proposed that the LILE-rich Cameroon line basalts could be derived from the LILE-depleted MORB source by small-degree (about 0.2%) melting using bulk partition coefficients consistent with experimentally determined values for mantle silicate phases. They suggested that the convecting upper mantle, although LILE depleted in its bulk composition, contains LILE-rich streaks. The presence of an ancient enriched component in the asthenosphere is necessary to account for the small but consistent isotopic differences between MORB and intra-plate basalts. Such heterogeneities could be produced by the migration of magma through the asthenosphere (*cf.* Hawkesworth *et al.* 1984). They may also be produced by the return of ocean crust (including ocean islands) to the asthenosphere during subduction. Whatever their origin, heterogeneities in the asthenosphere will be drawn out into streaks by convection (Hoffman & McKenzie 1985). Partial melts from these streaks will be selectively incorporated into small-degree melts, whereas larger-degree melting will tend to homogenize the source and result in magmas (MORB) reflecting the bulk composition of the asthenosphere.

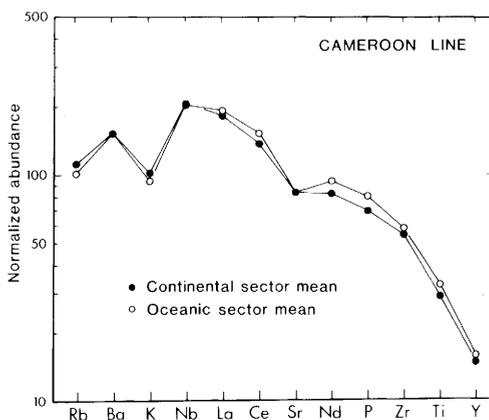


FIG. 9. Normalized abundance patterns (Sun 1980) for incompatible-element concentrations in average continental (134 analyses) and oceanic (144 analyses) basic volcanic rocks from the Cameroon line. Data from the Etinde nephelinites are not included in the averages. (After Fitton & Dunlop 1985).

McKenzie (1984, 1985) has recently derived a set of equations describing the compaction of partially molten rock and concluded that melt will begin to flow at very small degrees of melting (less than 0.5%) provided that its viscosity is low. The degrees of melting required to produce LILE-rich intra-plate basalts from the MORB source should no longer be regarded as impossibly small.

If the Cameroon line basic magmas were generated by small-degree melting of the MORB source then the same must also be true of other intra-plate basic magmas. This has been discussed by Fitton & James (1986) who compared incompatible-element concentrations in a wide range of intra-plate volcanic rocks with calculated concentrations in liquids generated by variable degrees of equilibrium melting of the MORB source. Partition coefficients required by the Cameroon line data were used in the calculations. The comparisons are encouraging and suggest that ocean island and rift valley magmas may share a common asthenosphere source. Metasomatic enrichment of the source is not necessary.

Origin of the salic rocks

The extent to which contamination of magmas by crustal rocks influences their composition and evolution has been debated for many years (see, for example, Moorbath *et al.* 1984). Magmas in continental provinces may receive contributions

from the asthenosphere, the mantle part of the lithosphere and the crust. Radiogenic isotope studies can constrain the possible magma sources in some cases but cannot always distinguish unambiguously between contributions from continental crust and enriched lithosphere mantle.

In the case of the Cameroon line we can be confident that the basic magmas have not interacted to any significant extent with either the mantle or crustal parts of the lithosphere. This provides a unique opportunity of studying the effects of environment on the subsequent evolution of two essentially identical sets of basalt magmas during storage in the oceanic and continental lithosphere respectively.

The compositional range of the Cameroon line volcanic rock samples collected from all the main centres is illustrated on an alkali-silica diagram in Fig. 10. Despite the similarity of the oceanic and continental basalts, the intermediate and evolved rocks from the two sectors show a clear divergence. With very few exceptions, the oceanic magmas evolve to phonolite and the continental magmas to rhyolite. The two samples of oceanic quartz trachyte (Fig. 10) were both collected from the Ilhéu das Cabras off São Tomé; the two continental phonolite samples are from plugs on the Ngaoundéré Plateau. Apart from these four samples, the separation of oceanic and continental salic rocks is complete. This is further illustrated in Fig. 11 in which the CIPW norms of the salic rocks have been projected into the residua system (quartz-nepheline-kalsilite).

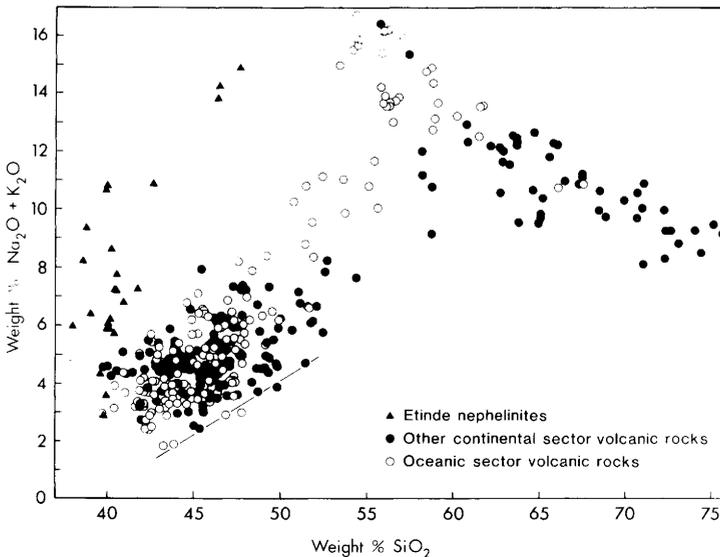


FIG. 10. Alkali-silica diagram for the Cameroon line volcanic rocks. The line separating Hawaiian tholeiitic and alkaline rocks (Macdonald & Katsura 1964) is shown for comparison. (From Fitton 1983.)

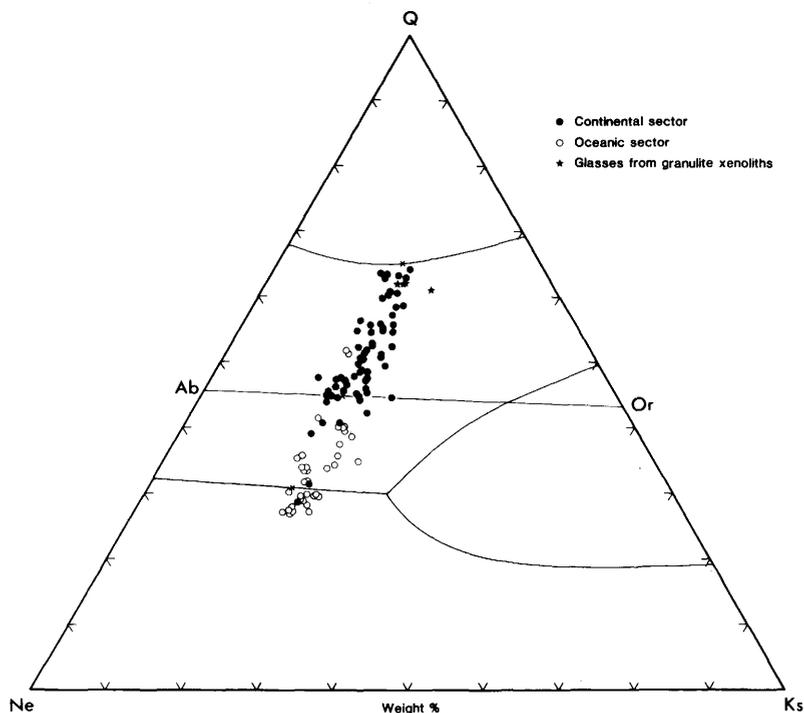


FIG. 11. Normative compositions of evolved (more than 80% normative salic components) volcanic rocks from the Cameroon line projected into the residua system. Additional data from Manengouba (supplied by C. A. Hirst) are included. Phase boundaries ($P(\text{H}_2\text{O}) = 5 \text{ kb}$) are taken from Hamilton & MacKenzie (1965).

The oceanic rocks occupy the thermal valley between the trachyte and phonolite minima, whereas the continental rocks plot in the corresponding valley down to the granite minimum.

This separation into undersaturated oceanic and oversaturated continental salic rocks must be related to the environment in which the magmas evolved. The basaltic parental magmas in both sectors of the Cameroon line are transitional to strongly-alkaline in character (Figs 7 and 10). Low-pressure crystal fractionation alone should therefore have produced salic rocks ranging from phonolite to trachyte, with only small amounts of oversaturated magma (as in the oceanic sector). Progressive crustal contamination of the fractionating magmas provides the simplest explanation for the dominance of oversaturated salic rocks in the continental sector.

Evidence for crustal contamination is provided by the occurrence of partly digested granulite xenoliths in a basalt lava flow from Bambouto, near to the village of Babadjou. The xenoliths are partially melted and contain patches of fresh glass in a residual mineral assemblage of quartz, plagioclase, orthopyroxene and minor-alkali feldspar. Electron microprobe analyses of the glass

show it to have a composition close to the minimum on the quartz-alkali feldspar cotectic (Fig. 11). Contamination of the continental magmas with crustal rocks or their partial melts could allow the magmas to cross the low-pressure thermal divide and evolve towards rhyolite.

Dunlop's (1983) strontium isotope study of the Cameroon line volcanic rocks provides further evidence for crustal contamination. Although the oceanic and continental basalts are generally indistinguishable in their isotope ratios, the distribution of $^{87}\text{Sr}/^{86}\text{Sr}$ in the continental basalt samples is skewed slightly towards higher values (Fitton & Dunlop 1985). The contaminated basalt from Bambouto has the highest initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio (0.70403) of all the continental basic lavas analysed. It is in the evolved rocks, however, that crustal contamination can be most clearly demonstrated. The trachytes and rhyolites from Bambouto, Oku and the Mandara Mountains have initial $^{87}\text{Sr}/^{86}\text{Sr}$ values ranging up to 0.715. For comparison, two of the granulite xenoliths from Bambouto have age-corrected $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of 0.71021 and 0.72088 (Dunlop 1983).

A possible alternative to crustal contamination as a means of crossing the low-pressure thermal

divide is provided by fractional crystallization of hornblende-bearing assemblages. Cawthorn *et al.* (1973), for example, suggested that removal of silica-poor *ne*-normative hornblende from under-saturated basic magma could lead to the evolution of silica-oversaturated residual liquids. This mechanism cannot apply, however, to the Cameroon line magmas since hornblende is seldom found as a phenocryst phase in the continental volcanic rocks. Curiously, though, hornblende phenocrysts are abundant in the intermediate lavas and even in some of the basalts in the oceanic sector.

The abundance of hornblende in the oceanic lavas and its scarcity in the continental rocks is puzzling since the basic magmas in the two sectors were compositionally identical (Fig. 7). The most important factor controlling amphibole stability in basaltic magma is the partial pressure of water vapour (Allen & Boettcher 1978). It is possible that magma stored beneath the islands of the oceanic sector had easier access to water than did the continental magma reservoirs, and that seepage of small amounts of sea-water into the oceanic magma reservoirs stimulated the crystallization of hornblende.

The middle and late stages in the evolution of magmas on São Tomé are dominated by fractionation of hornblende-bearing assemblages (Fig. 5). This is clearly demonstrated by the abundance of apparently cognate hornblende-rich cumulate

blocks. The operation of this process in the oceanic magmas, but not in those of the continent, is well illustrated in a plot of Y against Zr (Fig. 12). Y is a compatible element in amphiboles (Pearce and Norry 1979), and so hornblende removal would cause the concentration of Y to fall. Zr, which is incompatible with all the observed phenocryst phases, is a useful index of fractionation. Figure 11 shows that the Y concentration rises with fractionation in the continental lavas, where hornblende is seldom present, but falls in the intermediate and evolved oceanic sector rocks.

The hornblende-bearing cumulate blocks from São Tomé are relatively rich in Y. Fractional crystallization involving the removal of this cumulate material from the basic to intermediate magmas (*cf.* Fig. 5) would increase the Zr/Y ratios in the residual liquids along a line such as that on Fig. 12. This line was constructed by applying the Rayleigh fractionation equation to the average oceanic-sector basic lava composition (from Fitton & Dunlop 1985). Bulk partition coefficients for Zr and Y were taken as the ratio of their average concentrations in the cumulate blocks to the average in the oceanic-sector basic lavas. The degrees of fractionation indicated by the numbers on this line depend heavily upon the concentrations of Zr and Y in the crystal extract and serve only to show that the array of oceanic-sector data points could be generated by crystal

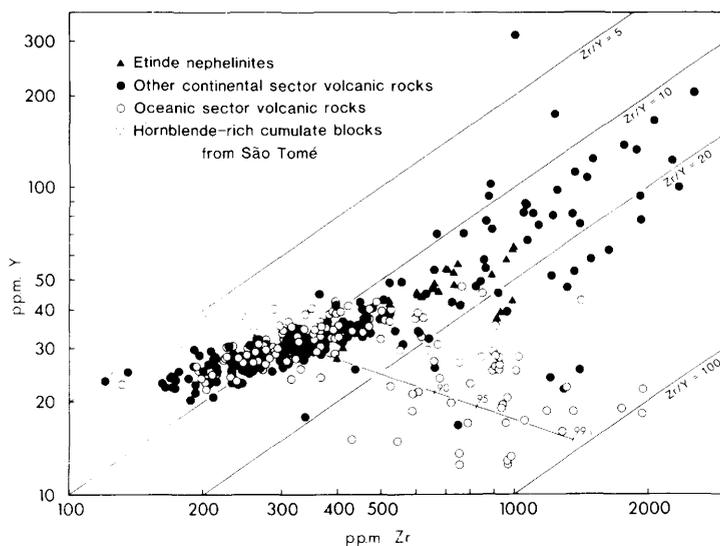


FIG. 12. Y and Zr contents of the Cameroon line volcanic rocks and cognate xenoliths. The fractional-crystallization path has been calculated by subtracting the average xenolith composition from the average oceanic-sector basic volcanic rock composition. The percentages indicate the amount of crystallization.

fractionation of hornblende-bearing assemblages.

There is no evidence that removal of hornblende has caused the oceanic-sector magmas to become less undersaturated. Virtually all these magmas have evolved towards the phonolite minimum (Figs 10 and 11) which shows clearly that crystal fractionation involving hornblende is *not* an efficient mechanism for crossing the low-pressure thermal divide.

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