

## The mineralogy of K-richterite-bearing lamproites

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### Abstract

The mineralogy and chemistry of eleven phlogopite-richterite-bearing lamproites has been studied. It has been found that this mineral assemblage (also found in metasomatically altered ultrabasic xenoliths) is typical of peralkaline lamproites. These lavas are typically potassium- and magnesium-rich, with a positive correlation between these two elements, and consequently between peralkalinity and magnesium content. Titanium contents are variable, but may reach 7% TiO<sub>2</sub>. The peralkaline character was acquired before crystallization, all minerals, including the first magmatic phases, having extremely low Al contents. Loss of alkalis, after partial or complete solidification, can be demonstrated in some specimens. Typical associations include besides phlogopite and richterite, Cr-rich Al-poor spinels, olivine, diopside, Ti-rich oxides (ilmenite, pseudobrookite, priderite). The peralkaline character is emphasized by the presence of roedderite-like phases in two specimens.

An electron microprobe study of lamproite minerals shows a general tetrahedral deficiency in most Al-bearing silicates, the sum Si + Al being systematically lower than its theoretical value. Most elements are regularly partitioned between co-crystallizing phases: e.g., Fe, Mg and Ti between richterite and phlogopite; Na and K between feldspar and richterite; tetrahedral deficiency between phlogopite and richterite.

These observations lead to the conclusion that the peralkaline character of these lamproites imposes constraints on the mineral compositions. However, the mineral compositions are not affected by the whole-rock silica saturation.

Xenocrystic material has been identified and is abundant in lamproites. Isolated relic crystals correspond to most major phases, except for richterite, that crystallize from the liquid, i.e., olivine, clinopyroxene, mica, ilmenite, pseudobrookite and apatite.

A comparison can be made between major authigenic phases from lamproites and equivalent phases from associations interpreted, in ultrabasic nodules, as metasomatic assemblages and presumed to have formed in the mantle under high pressures. Most major element partitioning trends are not affected by pressure but Ti solubility in silicates is considerably reduced by increasing pressure.

The origin of lamproitic liquids cannot be linked to basaltic magmas; basalts are indeed absent from lamproitic provinces. The observed bulk composition of lamproites could result from melting of richterite + phlogopite-bearing ultrabasic associations in the mantle. Calculations indicate that for all specimens considered in the present study the initial proportions could be richterite 30–50% and phlogopite 56–68%. Precipitation of olivine (with or without diopside) could then lead to the observed lamproitic lava compositions.

### Introduction

The term lamproite was coined by Niggli (1923, p. 105) for rocks whose compositions were extreme with respect to their potassium and magnesium (high) and aluminum (low) contents. The term has been used since, to designate rocks with compositions similar to those of lamprophyres, but whose field relations clearly indicate a volcanic situation (Velde, 1969, 1975).

Recent definitions take into account mineralogical compositions. The term is thus restricted by Rock (1984) to

volcanic or subvolcanic rocks with a unique paragenesis that includes the rare minerals, K-richterite and priderite. According to Scott Smith and Skinner (1984), current research shows that "lamproites are characterized by Ti-rich silicates and oxides, while the absence of plagioclase and significant amounts of melilite separates them from other potassium-rich alkaline rocks." Lamproite is used in the present work in Rock's general definition but, as pointed out by Scott Smith and Skinner (1984), the presence of Ti-rich silicates and oxides definitely appears to characterize this group of lavas.

Table 1. List of the samples studied, geographical localization, age, mineral assemblages and references to previous studies

	Localities	Specimen number	Rock type	Age	Mineralogical Assemblage	References
	SPAIN				Ol+Cpx+Phl+Richt+Kspar+Cr+Ilm	Lewis (1882) Osann (1906) Jeremine and Fallot (1929) Hernandez Pacheco (1935) San Miguel de la Camara et al. (1952)
1	Cancarix, Albacete	74A1	Cáncalite	5.7 to 7.6 m.y.	Ac+Ap+"Roed"+Ru+Cc X-Ol ? X-Cr ?	Fuster and Gastesi (1965) Fuster et al., (1967).
2	Calasparra, Murcia	72A7		Bellon et al. (1981) Bellon (1976; 1981)	Ap+Glass+X-Ol ?	Borley (1967) Velde (1969)
3	Jumilla, Murcia	11A8	Jumillite	Nobel et al. (1981)	Lc+Ac+Ap+Glass	Ruiz and Badiola (1980)
4	Barqueros, Murcia	73A8	Fortunite	Montenat et al. (1975)	Psb+Glass X+Phl+X-Al-Sp+X-Psb+X-Ilm+X-Ru	Nixon et al. (1982)
	UNITED STATES					
5	Smoky Butte, Montana	7B8	Lamproite	27 m.y. Marvin et al. (1980)	Ol(alt.)+Cpx+Phl+Richt +Kspar+Anal+Ap+Glass +Psb+Prid+Cr	Matson (1960) Velde (1975)
6	Shiprock dyke, New Mexico	5B6	Minette	27-30 m.y. Naesser (1971) Armstrong (1969)	Ol(alt.)+Cpx+Phl+Richt- Arfv+Kspar+Ap+Glass+ Ilm+Mt+Ac	Williams (1936) Nicholls (1969) Velde (1969) Roden (1981)
7	Moon Canyon, Utah	MC77	Lamproite	36-40 m.y.	Ol(alt.)+Cpx+Phl+Richt+ Kspar+Ap+Glass+Ilm+ Psb+"Roed"+Cr+X-Phl	Crittenden and Kistler (1966) Best et al. (1968)
	FRANCE					
8	Sisco, Corsica	61A3	Minette	15.4-13.5 m.y. Feraud et al. (1977)	Ol(alt.)+Cpx+Phl+Richt- Arfv+Kspar+Ilm+Prid+Cr	Velde (1967)
9	Sisco, Corsica	8A12		Civetta et al. (1978) Bellon (1981)	+Ru+Ap+Sph+Cc	Velde (1968)
	AUSTRALIA					
10	Mount North, West Kimberley	S0136	Wolgidite	17-22 m.y. Jaques et al. (1982)	Ol(alt.)+Cpx+Phl+Richt+ Lc+Ilm+Prid+Ap+Glass	Wade and Prider (1939) Carmichael (1967) Kaplan et al. (1967) Mitchell (1981) Jaques et al. (1984)
	ITALY					
11	Orciatico, Pisa	60A7	Selagite	4-1 m.y. Borsi et al. (1967)	Ol+Cpx+Phl+Kspar+Ilm +Richt+Cr+Ap	Barberi and Innocenti (1967) Stefanini (1934)

**Abbreviations :** Ac, acmite ; Al-Sp, aluminous spinel ; Anal, analcite ; Arfv, arfvedsonite ; Cc, calcite ; Cpx, clinopyroxene (diopside) ; Cr, chromite ; Ilm, ilmenite ; Kspar, potassium feldspar ; Lc, leucite ; Mt, magnetite ; Ol, olivine ; Ol(alt), altered olivine ; Phl, phlogopite ; Psb, pseudobrookite ; Prid, priderite ; Richt, richterite ; "Roed", roederite-like phase ; Ru, rutile ; Sph, sphene. X indicates that the species whose abbreviated name follows is present, but interpreted as a xenocryst.

The phlogopite-K-richterite association is also recorded from nodules found in kimberlites, the MARID suite as defined by Dawson and Smith (1977). The mineralogical associations as well as the composition of the phases present is similar in the nodules and in the lamproites. It seemed warranted to examine the compositions of these phases and their crystallization relationships in lamproites and consider more closely the similarities and differences between the constituent minerals of lamproites and the metasomatic assemblages from mantle rocks that could be the source of lamproites.

Experimental bulk rock analyses were performed using a Philips PW 1400 spectrometer. Ferrous iron was determined using a method outlined by Peck (1964) and H<sub>2</sub>O<sup>+</sup> by a modified Penfield method.

Electron microprobe analyses were carried out on an MS-46 CAMECA instrument with natural or synthetic minerals as standards for olivines, micas, amphiboles and clinopyroxenes. Counting time = 30 s; accelerating voltage = 15 kV; current = 30 nA (20 nA for K and Na in micas and amphiboles). Other mineral compositions were determined with a CAMEBAX automated electron microprobe using either oxides or natural minerals as standards.

Counting time = 30 s; accelerating voltage = 15 kV; current = 30 nA for oxides; 5 nA for feldspars.

#### Chemical composition and peralkaline character of the lamproites

The eleven specimens studied in this paper come from most of the classical localities. Their ages vary from 40 to 4.1 m.y.; they usually occur as pipes, dikes or necks but flows are also found. Table 1 gives the list of all samples studied and Table 2 the corresponding bulk chemical compositions.

Lamproites are magnesium-rich: their CIPW norms show 20% normative enstatite (or diopside) with or without forsterite. They are potassium-rich, and analyses show a positive correlation between potassium and magnesium, a significant difference with usual trends in magmatic series. No correlation is observed between alkalis and iron content. In the specimens studied, K<sub>2</sub>O varies between 7 and 12% (except the Jumilla specimen where the present K<sub>2</sub>O content is only 3.7%), and is largely superior to Na<sub>2</sub>O. Titanium contents are variable, and can reach 7% TiO<sub>2</sub>. This is particularly striking when considered in connec-

Table 2. Chemical composition of the samples studied. All analyses are new but for Nos. 6, 8, 10

	1	2	3	4	5	6	7	8	9	10	11
	74A1	72A7	11A8	73A8	7B8	MC72	5B6	SO136	61A3	8A12	60A7
SiO <sub>2</sub>	55.41	56.61	47.31	58.77	51.41	55.73	53.70	45.82	57.46	56.23	57.81
Al <sub>2</sub> O <sub>3</sub>	8.83	7.96	6.52	11.34	8.36	10.72	11.06	6.86	10.88	12.06	10.94
Fe <sub>2</sub> O <sub>3</sub>	1.17	2.12	3.83	1.31	3.02	3.73	3.57	6.07	1.46	1.91	3.55
FeO	3.40	3.13	2.88	3.49	2.31	0.95	3.08	1.98	2.54	2.91	1.44
MgO	13.59	12.45	17.16	9.22	8.05	6.96	7.86	10.90	7.24	6.90	8.26
CaO	3.97	2.97	8.82	2.71	4.71	3.85	7.38	4.70	2.96	3.48	4.43
Na <sub>2</sub> O	1.13	0.75	1.88	1.08	1.25	1.21	2.31	0.84	1.14	1.03	1.09
K <sub>2</sub> O	8.63	9.21	3.71	8.46	7.87	10.49	6.64	8.82	10.75	10.00	7.94
MnO	0.08	0.08	0.12	0.10	0.06	0.08	0.10	0.10	0.05	0.07	0.07
TiO <sub>2</sub>	1.41	1.61	1.32	1.35	5.46	2.82	1.81	7.34	2.19	1.14	1.37
P <sub>2</sub> O <sub>5</sub>	1.35	1.23	2.00	0.79	2.22	1.13	0.91	1.83	0.74	0.79	0.72
BaO						0.35		1.27			
CO <sub>2</sub>								0.08		0.41	
H <sub>2</sub> O+	0.59	1.86	3.77	0.96	3.58	1.14	1.53	0.75	1.70	1.56	1.18
H <sub>2</sub> O-	0.23	0.65	1.31	0.31	1.74	0.89	0.62	2.40	0.72	1.58	1.06
F	0.36	0.41	0.32	0.30	0.46		0.16		0.38	0.35	0.33
Clppm	226	321	174	87	78				45	319	74
Total	100.15	101.04	100.95	100.19	100.50	100.05	100.73	99.76	100.21	100.42	100.19
O≡F	0.15	0.17	0.13	0.13	0.19		0.07		0.16	0.15	0.14
Total	100.00	100.87	100.82	100.06	100.31		100.66		100.05	100.27	100.05
Na+K Al	1.27	1.41	1.09	0.96	1.26	1.24	0.99	1.59	1.24	1.04	0.95
Q		2.90		2.27	3.14				2.33		4.61
Or	48.15	43.37	21.90	50.04	45.65	58.53	39.25	37.25	59.32	59.10	46.98
Ab			12.42	9.12			17.34			6.36	9.22
An				1.06			0.14				1.45
Ac	3.37	5.57	2.68		8.73	9.02		6.47	4.20	2.08	
Ne			0.23				1.19				
Ns	1.33				0.16				1.13		
Ks	0.80	3.08			0.26	0.96		4.16	1.18		
Di	8.65	5.29	24.07	5.75		4.34	23.80		7.76	7.92	12.33
En	21.20	28.87		20.77	20.13	15.32		27.20	14.69	10.55	14.94
Fs	2.44	2.85		3.92					0.92	1.60	
Po	6.30		22.35				6.03			2.38	
Fa	0.80		0.53							0.40	
Mt		0.28	4.20				5.01			1.73	0.90
Hem						0.61	0.11	4.00			2.93
Ilm	2.68	3.05	2.51	2.57	5.00	2.18	3.44	4.41	4.16	2.17	2.60
Tn					6.39	3.85		1.37			
PF						0.18		6.12			
Ru					0.22			0.88			
Ap	3.19	2.92	4.70	1.88	5.24	2.73	2.15	4.03	1.75	1.73	1.71
Cc								0.20		0.93	

Analysis 6, Best et al., 1968; Analysis 8, Wade and Prider, 1940; Analysis 10, Velde, 1967.

tion with the low iron content of these rocks in general (4.0 to 6.7% FeO + Fe<sub>2</sub>O<sub>3</sub> and 8% in specimen SO136).

The specimens examined are either definitely peralkaline or close to peralkalinity. This character is demonstrated by the presence of normative acmite in the CIPW norm, accompanied by sodium and/or potassium metasilicate, the maximum content of normative anorthite being 1.45% (Table 2). Fluorine (between 0.2 and 0.47%) is positively correlated with potassium content following the relationship discovered by Aoki et al. (1981).

Two lines of evidence indicate the primary origin of the peralkaline character of the lamproitic magma. One ar-

gument is the fact that lamproites often occur with glassy textures, frequent among the Spanish lamproites, present in Smoky Butte and the Leucite Hills, and recently described in Australia (Prider, 1982). The second line of evidence is that the spinels found in olivines of the Spanish lamproites, Sisco, Moon Canyon, the selagites and the Smoky Butte lamproite are uncommonly chrome-rich (51 to 65% Cr<sub>2</sub>O<sub>3</sub>) but always alumina-poor (0.23% to 4% Al<sub>2</sub>O<sub>3</sub>) (Table 3). In comparing data on various olivine-bearing lavas, it appears that the Al<sub>2</sub>O<sub>3</sub> content of Cr-rich spinels is related to the composition of the rocks in which they are found. Using, for example, analyses published by

Table 3. Representative electron microprobe analyses of chrome spinels found as inclusions in the olivines. Fe<sup>3+</sup> calculated assuming 3 cations and 4 oxygens

	1	2	3	4	5	6	7	8	9	10	11
	74A1	74A1	72A7	72A7	11A8	11A8	73A8	6B18	8A12	60A7	60A7
Al <sub>2</sub> O <sub>3</sub>	0.23	0.15	3.04	1.72	2.00	1.16	1.10	0.71	4.16	3.56	1.87
Cr <sub>2</sub> O <sub>3</sub>	51.41	62.64	65.34	61.43	55.78	51.48	31.58	42.78	63.00	55.64	36.22
FeO	37.05	26.40	20.33	29.99	29.34	34.97	46.72	32.84	15.31	30.17	47.95
MgO	4.42	8.00	10.41	4.72	9.80	7.43	5.24	9.54	13.80	5.77	3.30
MnO	0.84	0.82	0.69	0.90	0.79	0.47	0.55	0.61	0.48	0.80	0.86
TiO <sub>2</sub>	4.52	1.16	1.23	1.63	1.03	0.79	10.90	9.50	2.14	1.12	5.70
ZnO	nd	nd	nd	nd	nd	nd	nd	0.32	0.20	nd	nd
Total	98.47	99.17	101.04	100.39	98.74	96.30	96.09	96.30	99.09	97.06	95.90
Fe <sub>2</sub> O <sub>3</sub>	9.77	7.18	3.22	4.54	13.32	16.39	15.79	9.59	2.97	7.73	19.27
FeO	28.26	19.94	17.43	25.90	17.35	20.23	32.51	24.22	12.64	23.21	30.61
Al	0.010	0.006	0.120	0.072	0.081	0.049	0.048	0.030	0.163	0.151	0.082
Fe <sup>3+</sup>	0.267	0.190	0.081	0.121	0.345	0.444	0.436	0.257	0.074	0.209	0.538
Cr	1.476	1.742	1.736	1.720	1.520	1.464	0.916	1.205	1.656	1.580	1.062
Fe <sup>2+</sup>	0.858	0.587	0.490	0.767	0.500	0.609	0.997	0.721	0.351	0.697	0.949
Mg	0.239	0.420	0.522	0.249	0.504	0.399	0.286	0.506	0.684	0.309	0.182
Mn	0.026	0.024	0.020	0.027	0.023	0.014	0.017	0.018	0.014	0.024	0.027
Ti	0.123	0.031	0.031	0.043	0.027	0.021	0.301	0.254	0.053	0.030	0.159
Zn								0.008	0.005		

FeO° = Total iron as FeO  
nd : not sought for.

Eissen (1982) on MORB basalts and Barbot (1983) on spinels from alkali basalts, it can be seen that for a given Cr content, spinels from tholeiitic basalts are richer in Al<sub>2</sub>O<sub>3</sub> than spinels found in alkali basalts. Within the field of alkaline rocks, spinels are richer in aluminum in alumina saturated melilite-bearing compositions than they are in peralkaline compositions, the two fields showing no overlap (Velde and Yoder, 1983; unpublished data). In the lamproites the Cr-rich spinels can be considered as the liquidus phase. From their rather unique composition, they definitely belong to the peralkaline field as defined

Table 4. Electron microprobe analyses of the glass (1 = average of 3 determinations) and olivine (2) included in an apatite crystal from specimen 11A8

	1	2
SiO <sub>2</sub>	55.00	40.00
Al <sub>2</sub> O <sub>3</sub>	12.00	
FeO°	6.00	11.00
MgO	1.30	50.00
CaO	1.60	0.20
Na <sub>2</sub> O	5.00	
K <sub>2</sub> O	12.00	
MnO		0.20
TiO <sub>2</sub>	2.60	
Total	95.50	101.40

FeO° = Total iron as FeO

by Velde and Yoder (pers. com.). The peralkaline character of the lamproites could, in some instances, have been more pronounced than presently evidenced by their bulk composition. Specimen 11A8, from Jumilla, contains apatite phenocrysts with frequent glass inclusions. One of these (0.06 × 0.03 mm) contains besides glass, a euhedral crystal of olivine and a small mica blade, both phases appearing as phenocrysts in the rock. This glass (Table 4) is considerably richer in alkalis than the bulk rock and the (Na + K)/Al of the glass inclusion is 1.72 as opposed to 1.09 for the bulk rock composition. Least squares calculations made to estimate whether the glass composition could be derived from the bulk composition of the rock, taking into account the presence of the olivine and biotite crystals, show that a reasonable fit is obtained only if it is assumed that the bulk composition has lost alkalis (mostly potassium) after inclusion of the glass in the apatite.

From the above one can consider that the lamproite liquids that produced phogopite and K-richterite assemblages had primary peralkaline compositions. This character was acquired before the first liquidus phase (Cr-spinel) crystallized. In at least one specimen (11A8) it can be demonstrated that the composition was altered after the crystallization of the feldspar (see below) and originally had a stronger peralkaline characteristic.

Finally, the silica saturation is not a factor critical to the mineralogical association under consideration since lamproite compositions can vary from undersaturated with respect to silica (with modal feldspathoid) to oversaturated (with modal quartz).

Table 8. Representative electron microprobe analyses of feldspars from the lamproites studied. Total iron = Fe<sub>2</sub>O<sub>3</sub>

	1	2	3	4	5	6	7	8	9	10	11	12	13
	74A1	72A7	11A8	11A8	73A8	7B8	5B6	5B6	MC77	61A3	61A3	60A7	60A7
SiO <sub>2</sub>	63.90	63.91	61.28	63.11	63.81	63.61	63.22	65.02	63.23	64.24	64.63	64.99	64.65
Al <sub>2</sub> O <sub>3</sub>	17.46	17.12	12.85	14.49	18.59	15.76	19.00	18.41	15.86	17.29	18.24	18.57	18.52
Fe <sub>2</sub> O <sub>3</sub>	1.12	1.26	4.95	3.17	0.31	1.93	0.62	0.62	2.91	1.14	0.20	0.51	0.78
MgO	0.02	0.03	0.10	0.20					0.08				
CaO							0.11	0.10					
Na <sub>2</sub> O	0.44	0.63	0.66	0.32	1.39	0.55	3.25	3.53	0.32	0.42	0.56	2.25	2.29
K <sub>2</sub> O	16.15	15.75	16.38	16.20	14.50	15.83	10.98	11.52	15.84	16.21	15.97	13.34	12.98
BaO	0.25	0.39	0.11	0.08	0.64	1.21	1.87		0.50	0.11	0.34	0.06	0.89
TiO <sub>2</sub>	0.27	0.29	0.42	0.22	0.16	0.33		0.26	0.29	0.12	0.07	0.17	0.40
Total	99.61	99.38	96.75	97.79	99.40	99.22	99.05	99.46	99.03	99.53	100.01	99.89	100.51
Si	2.986	2.993	3.012	3.028	2.969	3.013	2.945	2.980	2.994	3.000	2.994	2.982	2.966
Al	0.962	0.945	0.745	0.820	1.020	0.880	1.043	0.995	0.885	0.952	0.996	1.004	1.002
Fe <sup>3+</sup>	0.039	0.044	0.183	0.114	0.011	0.069	0.022	0.021	0.104	0.040	0.007	0.018	0.027
Mg	0.001	0.002	0.007	0.014					0.006				
Ca							0.005	0.005					
Na	0.040	0.057	0.063	0.030	0.125	0.051	0.294	0.314	0.029	0.038	0.050	0.200	0.204
K	0.963	0.941	1.027	0.992	0.861	0.957	0.653	0.674	0.957	0.966	0.944	0.781	0.760
Ba	0.005	0.007	0.002	0.002	0.012	0.023	0.034		0.009	0.002	0.006	0.003	0.016
Ti	0.009	0.010	0.016	0.008	0.006	0.012		0.009	0.010	0.004	0.002	0.006	0.014
	5.005	5.001	5.050	5.008	5.003	5.004	4.996	4.997	4.994	5.002	4.999	4.992	4.988

### Mineralogy

A general order of crystallization can be deduced from the textural relationships of the lamproites studied. In all specimens olivine appears to be the first silicate phase to appear. The Cr-rich spinels are only found as inclusions in olivine. Diopside and phlogopite have probably syncrystallized; amphibole and feldspar have appeared subsequently and are contemporaneous with an acmitic rim on the clinopyroxenes and in some instances with a deeply red colored edge around the phlogopites. Oxides appear to have formed after the diopside and the phlogopitic centers of the micas. The specimen from Moon Canyon shows phlogopite surrounding olivine crystals, and richterite rimming the diopside, thus supporting the order stated. It should also be mentioned that, when present, feldspar often shows a skeletal habit. Modal analyses are given in Table 5.<sup>1</sup>

### Olivine

Olivine is found unaltered only in some Spanish specimens and the Orciatico selagite. In both types, some crystals (usually anhedral, and large in size) represent xenocrysts with typical kink-bands. Phenocrysts are generally euhedral in outline, and contain numerous euhedral spinel inclusions. Compositions of the xenocrysts are similar to that of the cores of the phenocrysts, in major elements (Fo%) or trace elements (Ni for example). Compositions, including xenocrysts (Table 6<sup>1</sup>) vary from 92 to 77% Fo. Nickel contents are consistently high, vary from 0.10 to 0.70% NiO, but usually occur between 0.30 and 0.50%. Such high Ni contents are above the values given for olivine in basalts (between 0.10 and 0.30 NiO, Simkin and Smith, 1970; Anderson and Wright, 1972). They are comparable to values reported by Reid et al. (1975) in olivine from peridotite xenoliths, close to Ni

contents found in olivines from nodules occurring in kimberlites (Nixon and Boyd, 1973; Scott, 1981) and similar also to values recently determined in the Australian lamproites and kimberlites (Jaques et al., 1984).

The NiO content of the olivines is inversely correlated with the fayalite content. In the selagite, the most ferrous olivine (Fo<sub>77</sub>) contains 0.23% NiO. Calcium in the olivines is usually low, particularly in the xenocrysts, but always below 0.20% CaO. There is a positive correlation between CaO and fayalite contents of the olivine.

The manganese contents are consistently high and positively correlated with the fayalite content; MnO is close to 0.30% at Fo<sub>80</sub>.

### Pyroxene

Clinopyroxene (Table 7<sup>1</sup>) in the lamproites studied falls mostly in the endiopside field of the pyroxene quadrilateral. Its abundance varies from 6 to 22% in volume. It most often occurs as unzoned phenocrysts which contain around 1% TiO<sub>2</sub>. The Cr<sub>2</sub>O<sub>3</sub> content varies between 0.1 and 0.3 but, in the Spanish lamproites falls in the 0.4–0.9% range. These values are considerably higher than those commonly found in lavas and are, for the highest ones, consistent with percentages recorded from diopsides from dunite inclusions (e.g., Berger, 1981).

In all samples, save the specimen from Shiprock (5B6), the tetrahedral site of the pyroxene is incompletely filled by Si and Al; the tetrahedral site occupancy varies from 1.94 to 2.00, usually being between 1.97 and 1.98 atoms. The site can be considered as completed by Fe<sup>3+</sup> atoms that are present when the structural formula is calculated on 4 cations and 6 oxygens. This deficiency in Si + Al was noted by Carmichael (1967) for the Leucite Hills diopside and is mentioned by Jaques et al. (1984) for the Australian lamproites. This situation can be related to the low-alumina content of the liquid from which the pyroxenes crystallized. Acmite is present in many samples, as a thin, late rim around the diopside crystals, or small groundmass grains.

<sup>1</sup> To receive copies of Tables 5, 6, 7, 14, 15, and 16, order Document AM-86-292 from the Business Office, Mineralogical Society of America, 1625 I Street, N.W., Suite 414, Washington, D.C. 20006. Please remit \$5.00 in advance for the microfiche.

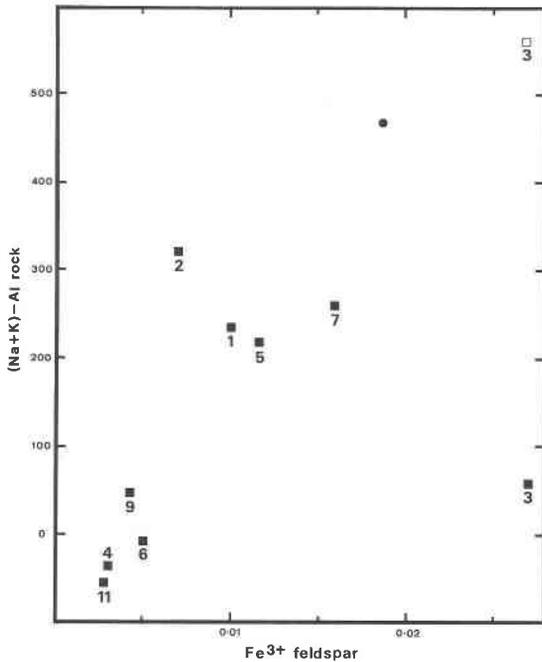


Fig. 1. The joint distribution of  $(\text{Na} + \text{K}) - \text{Al}$  calculated from the bulk rock composition and the  $\text{Fe}^{3+}$  content of feldspar (mole%). Solid squares: data from this study (specimen numbers are listed in Table 1); solid circle: LH9, Carmichael (1967). Open square: Jumilla composition corrected for calculated alkali loss.

In two specimens (Shiprock, 5B6 and Smoky Butte, 7B8) some of the colorless diopside phenocrysts show a green center. In Shiprock the green cores are richer in Al, Fe and Na than the diopside zone. Barton and Bergen (1981) have reported the presence of green clinopyroxenes with a salitic composition in a Leucite Hills wyomingite, and this type of clinopyroxene has frequently been found in alkali-rich, potassic rocks (Barton and Bergen, 1981; Lloyd, 1981). These green centers in diopsides are commonly interpreted as being xenocrysts, coming from the disaggregation of clinopyroxenite nodules, and appear to be a ubiquitous feature of potassic volcanics.

### Feldspar

Feldspar is abundant (19 to 48% in volume) and appears late in the crystallization sequence; it is absent from the Mount North Australian lamproite which carries leucite and from some of the glassy specimens. In all but two cases the feldspar composition is close to that of the potassic end member (Or 95–98). Feldspars from the Shiprock and Barqueros specimens are more sodic (64 to 72, and 87 to 89% Or respectively) (Table 8).

Iron contents are high, usually varying between 1 and 2%  $\text{Fe}_2\text{O}_3$ , but reaching 2.5% at Moon Canyon and 5% in the jumillite. Carmichael (1967) reports values between 3 and 4% in the Leucite Hills lamproites and Best et al. (1968) 3.83% in some of the Moon Canyon specimens. According to Smith (1974) the iron content of feldspars could be related to a rapid growth rate. Though feldspars in lamproites have crystallized rapidly as exemplified by their common skeletal habit, the iron content cannot be exclusively linked to the growth rate, but can definitely be correlated with the peralkalinity of the liquid (Fig. 1). The only exception is the Jumilla feldspar whose abnormal position on the diagram is consistent with the alkali loss already addressed above.

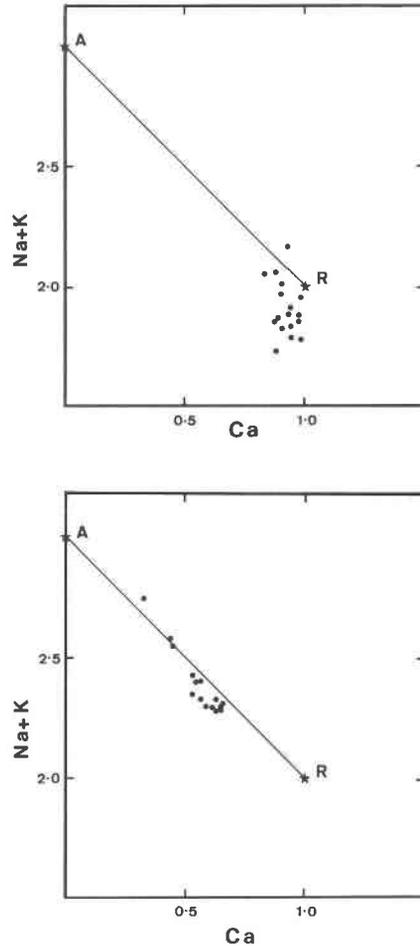


Fig. 2. (a) The distribution of  $\text{Na} + \text{K}$  and  $\text{Ca}$  in richterites from specimen 7B8. (b) The distribution of  $\text{Na} + \text{K}$  and  $\text{Ca}$  in richterites from specimen 5B6.

The feldspars are titanium-rich. Percentages of  $\text{TiO}_2$  in the lamproite feldspars analyzed reach 0.45%; Best et al. (1968) give 0.50% for a Moon Canyon feldspar, values that are considerably higher than the 0.07 to 0.08% recorded by Smith (1974) in his treatise. Feldspars from the Sisco minette, which coexist in the groundmass with abundant minute sphene needles, show the lowest values (0.07%  $\text{TiO}_2$ ). No simple relationship is apparent between the  $\text{TiO}_2$  content of the feldspars and the  $\text{TiO}_2$  content of the liquid from which they crystallized.

Barium varies between 0.1 and 0.3%  $\text{BaO}$ , but can reach 0.62% (73A8) and 1.3% (7B8). Two specimens show large variations in the barium contents of the feldspar: in sample 60A7,  $\text{BaO}$  varies from 0 to 0.89%, and in sample 5B6 from 0.0 to 3.22%  $\text{BaO}$ . In both cases there is an inverse correlation between  $\text{BaO}$  and the sum  $\text{Na}_2\text{O} + \text{K}_2\text{O}$ .

### Amphibole

Amphiboles found in peralkaline lamproites belong to the richterite group and are characterized by the presence of a high proportion of potassium in the A site. They are among the last phases to appear and are found, albeit in small amounts, even in some lamproites that do not have a peralkaline composition (e.g., the selagite). In this case, their presence indicates that towards the

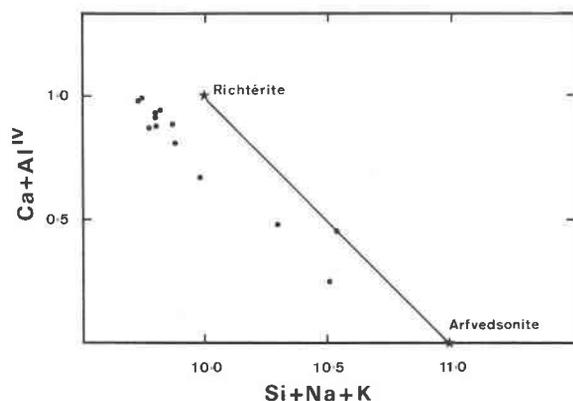


Fig. 3. The distribution of Ca + Al<sup>IV</sup> and Si + Na + K in richterites from specimen 61A3.

end of the crystallization sequence at least the residual liquids were peralkaline. Richterites belong to the calcalkaline amphibole group and their general formula can be written (Na,K)NaCa(Mg,Fe<sup>2+</sup>),Si<sub>8</sub>O<sub>22</sub>(OH)<sub>2</sub>. The richterites analyzed have at most one Fe<sup>2+</sup> atom in the C site, except for the amphiboles found in the Shiprock and Sisco specimens that are richer in iron (varying between Mg<sub>2</sub>Fe<sub>3</sub> and Mg<sub>3</sub>Fe<sub>2</sub>).

Their composition is usually close to the theoretical richterite composition (e.g., 7B8, Fig. 2a), or intermediate (as in Shiprock, Fig. 2b) between richterite and arfvedsonite (Table 9). The structural formulae were computed following the method of Papike et al. (1974) as advocated by Leake (1978). Initially the A site (Na + K) is overfilled and the tetrahedral site is underfilled (Si + Al < 8). In Figure 3 the trend shown by the experimental points is parallel to the theoretical richterite-arfvedsonite line demon-

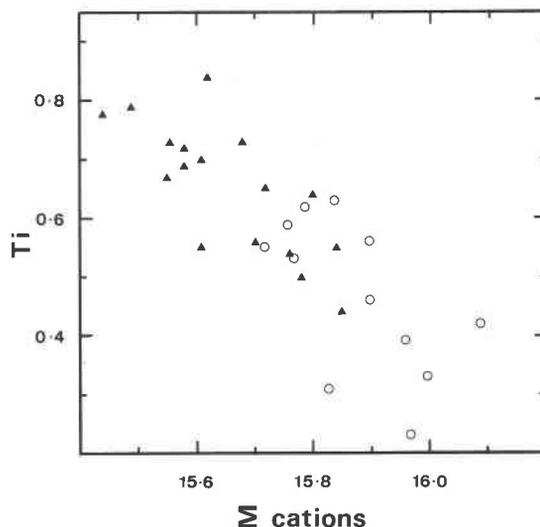


Fig. 4. The distribution of Ti and the sum of cations in richterites from 74A1 (solid triangles) and 72A7 (open circles).

strating this tetrahedral deficiency. The tetrahedral site must then contain another ion which could be Fe<sup>3+</sup>. When calculated Fe<sup>3+</sup> is assigned to the tetrahedral site, the structural formula usually becomes acceptable, the tetrahedral site is filled and there is no excess of ions in the A site. In some analyses however (e.g., in Calasparra) the tetrahedral site cannot be filled by Si, Al and Fe<sup>3+</sup> and the presence of yet another ion should be considered.

The richterites are Ti-rich: the minimal values found are close to 2% TiO<sub>2</sub>, in Sisco for example, but reach 9% in Moon Canyon. In amphiboles with high Ti contents (Cancarix, Calasparra and

Table 9. Representative electron microprobe analyses of K-richterites from the lamproites studied. Structural formulae calculated on the basis of 23 oxygens

	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16
	74A1	74A1	72A7	11A8	11A8	73A8	7B8	5B6	5B6	MC77	MC77	61A3	8A12	SO136	SO136	60A7
SiO <sub>2</sub>	52.89	52.92	51.79	52.31	53.22	54.40	53.26	51.10	48.85	51.12	50.06	51.55	52.02	52.46	51.24	54.50
Al <sub>2</sub> O <sub>3</sub>	0.41	0.55	0.36	1.01	0.32	0.33	0.38	1.75	1.62	0.51	0.25	0.36	0.40	0.33	0.53	0.93
Fe <sup>o</sup>	4.51	13.64	6.62	7.99	9.55	7.08	4.22	13.21	17.85	5.53	8.29	16.78	12.35	2.94	5.45	8.05
MgO	18.65	11.17	17.24	17.09	16.70	18.58	18.93	15.02	12.36	17.06	14.68	10.14	16.25	20.55	18.12	18.13
CaO	6.68	4.44	5.79	3.51	4.05	5.84	6.37	3.87	1.99	6.38	4.30	1.21	6.44	6.58	6.68	8.34
Na <sub>2</sub> O	4.24	4.67	4.75	6.33	6.43	5.01	4.27	6.95	8.26	4.33	4.96	6.66	3.85	3.29	3.64	3.40
K <sub>2</sub> O	3.64	4.12	3.69	2.89	2.94	2.51	4.25	1.62	1.52	4.22	3.94	4.34	3.56	5.87	5.55	1.32
MnO			0.10	0.13	0.14	0.09	0.22	0.17	0.21	0.12	0.11	0.04	0.24	0.04	0.05	0.11
TiO <sub>2</sub>	5.14	7.16	5.66	5.42	4.19	3.81	5.10	3.55	3.23	7.44	8.47	6.54	2.56	5.04	6.34	1.64
Total	96.16	98.67	96.00	96.68	97.54	97.65	97.00	97.25	95.88	96.71	95.06	97.62	97.67	97.10	97.60	96.42
Si	7.632	7.725	7.577	7.588	7.711	7.747	7.633	7.527	7.488	7.432	7.470	7.736	7.657	7.532	7.422	7.833
Al	0.069	0.094	0.062	0.173	0.055	0.055	0.065	0.303	0.293	0.087	0.044	0.063	0.069	0.055	0.090	0.157
Fe <sup>2+</sup>	0.544	1.665	0.809	0.969	1.157	0.843	0.506	1.627	2.288	0.672	1.035	2.106	1.520	0.353	0.661	0.967
Mg	4.011	2.430	3.759	3.695	3.606	3.943	4.043	3.299	2.824	3.697	3.265	2.267	3.565	4.397	3.911	3.884
Ca	1.033	0.694	0.908	0.546	0.629	0.891	0.979	0.611	0.326	0.994	0.688	0.194	1.016	1.012	1.037	1.285
Na	1.185	1.321	1.347	1.780	1.806	1.383	1.187	1.986	2.456	1.221	1.435	1.938	1.099	0.916	1.024	0.948
K	0.669	0.767	0.688	0.535	0.543	0.456	0.777	0.305	0.297	0.783	0.750	0.830	0.669	1.076	1.026	0.242
Mn			0.012	0.016	0.017	0.011	0.022	0.021	0.027	0.015	0.014	0.005	0.030	0.005	0.006	0.013
Ti	0.558	0.786	0.622	0.591	0.457	0.408	0.550	0.394	0.372	0.813	0.951	0.738	0.283	0.544	0.690	0.177
Total	15.701	15.485	15.784	15.892	15.980	15.737	15.767	16.073	16.370	15.726	15.652	15.877	15.908	15.890	15.867	15.506

Fe<sup>o</sup> = Total iron as FeO

Table 10. Representative electron microprobe analyses of micas from the lamproites studied. Structural formulae calculated on the basis of 11 oxygens

	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	
	74A1	74A1	72A7	72A7	11A8	73A8	73A8	7B8	7B8	7B8	5B6	5B6	5B6	5B6	MC77	8A12	8A12	8A12	SO136	SO136	60A7
SiO <sub>2</sub>	41.50	41.42	39.94	40.12	38.68	41.21	42.90	39.74	38.20	38.80	38.60	36.51	38.32	36.73	40.43	40.24	40.99	39.99	39.82	39.42	
Al <sub>2</sub> O <sub>3</sub>	11.00	10.24	12.41	10.28	8.79	12.65	11.92	11.27	10.59	10.43	13.13	10.34	13.57	10.83	12.15	11.77	10.59	10.89	6.93	11.05	
FeO°	5.11	7.63	3.07	6.53	11.49	6.92	3.87	3.88	6.95	7.71	4.79	11.16	5.76	11.82	2.70	4.41	11.78	4.13	10.66	11.03	
MgO	20.79	18.19	24.44	19.69	16.24	22.18	24.60	22.12	18.51	18.21	23.26	15.59	21.79	17.37	25.17	23.73	17.51	23.31	19.63	17.63	
MnO	0.03	0.01	0.01	0.04	0.08	0.06	0.03	0.02	0.03	0.04	0.04	0.12	0.03	0.10	0.03	0.05	0.13	0.01	0.10	0.10	
TiO <sub>2</sub>	6.88	9.51	3.90	8.78	11.19	4.79	1.98	7.04	9.13	9.93	3.14	8.50	3.73	9.43	2.30	3.38	5.11	5.49	7.95	6.25	
Cr <sub>2</sub> O <sub>3</sub>	0.24	0.09	1.33	0.09	0.03	0.07	0.76	0.76	0.15	0.03	1.32	0.51	0.51	0.51	1.07	1.57	0.06	1.19	0.02	0.04	
Na <sub>2</sub> O	0.07	0.14	0.01	0.20	0.46	0.34	0.45	0.10	0.16	0.18	0.60	0.70	0.46	0.50	0.11	0.15	0.36	0.23	0.23	0.55	
K <sub>2</sub> O	9.42	9.33	10.24	9.63	8.30	9.48	9.87	9.63	9.13	8.86	10.74	9.97	10.84	10.56	11.12	10.53	10.38	11.39	11.00	8.96	
BaO	0.84	0.61	0.05	0.51	2.11	0.19	0.08	1.11	2.80	2.62	0.34	0.31	0.33	0.28	0.13	0.13	0.29	0.12	0.55	1.53	
Total	95.89	97.15	95.40	95.87	97.37	97.89	96.48	95.67	95.65	96.81	95.86	93.21	95.34	97.64	95.18	95.97	97.21	96.53	96.87	96.56	
Si	2.949	2.936	2.837	2.882	2.883	2.880	2.997	2.843	2.810	2.818	2.777	2.791	2.779	2.693	2.888	2.870	2.974	2.849	2.929	2.888	
Al	0.921	0.855	1.039	0.871	0.759	1.042	0.982	0.950	0.918	0.893	1.113	0.932	1.160	0.938	1.020	0.990	0.906	0.915	0.601	0.953	
Fe <sup>2+</sup>	0.303	0.452	0.182	0.392	0.704	0.405	0.227	0.233	0.428	0.469	0.285	0.713	0.350	0.725	0.162	0.263	0.715	0.246	0.656	0.676	
Mg	2.203	1.922	2.588	2.108	1.773	2.310	2.562	2.359	2.030	1.971	2.495	1.776	2.355	1.898	2.681	2.523	1.893	2.476	2.152	1.925	
Mn	0.002	0.001	0.003	0.003	0.005	0.004	0.002	0.001	0.002	0.003	0.003	0.008	0.002	0.008	0.002	0.003	0.008	0.001	0.006	0.006	
Ti	0.368	0.507	0.209	0.474	0.616	0.252	0.104	0.379	0.505	0.543	0.170	0.489	0.204	0.520	0.124	0.182	0.279	0.294	0.440	0.339	
Cr	0.014	0.005	0.075	0.005	0.001	0.004	0.042	0.043	0.009	0.002	0.076	0.029	0.029	0.029	0.060	0.089	0.004	0.067	0.007	0.002	
Na	0.010	0.019	0.001	0.028	0.065	0.046	0.062	0.015	0.023	0.025	0.083	0.105	0.065	0.072	0.015	0.021	0.051	0.033	0.033	0.079	
K	0.854	0.843	0.928	0.882	0.776	0.846	0.879	0.879	0.857	0.821	0.986	0.973	1.004	0.988	1.013	0.958	0.961	1.035	1.032	0.837	
Ba	0.023	0.017	0.001	0.014	0.061	0.005	0.003	0.031	0.081	0.075	0.010	0.009	0.010	0.008	0.004	0.004	0.008	0.004	0.016	0.044	
Total	7.647	7.557	7.860	7.659	7.593	7.794	7.860	7.733	7.663	7.620	7.995	7.796	7.958	7.850	7.967	7.903	7.799	7.888	7.866	7.751	

FeO° = Total iron as FeO

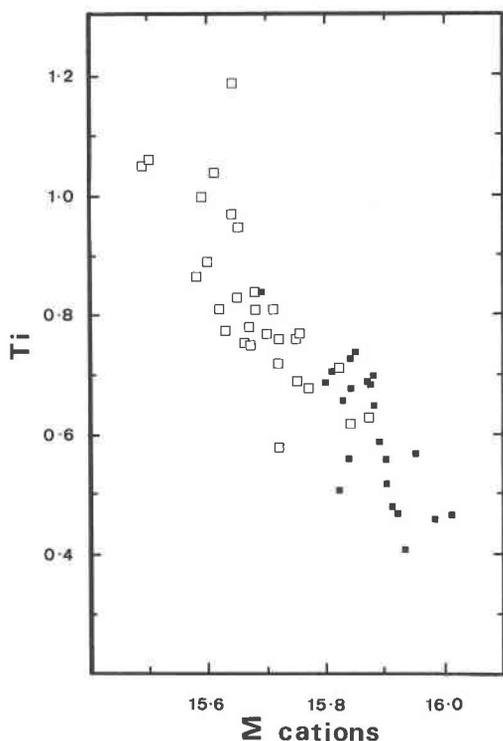


Fig. 5. The distribution of Ti and the sum of cations in richterites from MC77 (open squares) and 11A8 (solid squares).

Moon Canyon) the inclusion of titanium creates a vacancy in the structure as can be shown on a diagram showing the variation of Ti vs. the sum of ions (Figs. 4 and 5), an analogous situation to that described below in trioctahedral micas. Amphiboles with lower Ti-contents from all other specimens display a negative correlation between titanium content and the number of Si + Al ions on the tetrahedral site, but no relationship is found between Ti content and the sum of cations.

Some amphiboles are zoned: richterite from the Cancarix and Australian specimens show a darker color towards the rims of the crystals, and this variation corresponds to an increase in titanium and iron and a decrease in magnesium without any significant variation of the other elements, probably indicating the substitution  $\square + \text{Ti} + \text{Fe}^{2+} = 3\text{Mg}$ .

The main difference between these amphiboles and richterites found in the MARID suite nodules rests with their  $\text{TiO}_2$  content, higher in the lavas than in the high-pressure xenoliths.

### Mica

Trioctahedral micas (representative analyses are listed in Table 10) represent an important constituent of the lamproite mineralogy, from 15 to 30% in volume.

Two kinds of micas are present, some are authigenic crystals whereas others can be interpreted, with various degrees of certainty, as xenocrysts. In two specimens (73A8 and MC77) large independent crystals of phlogopite appear to have been destabilized, with abundant exsolution of needle-like Ti-rich phases. Analyses of these micas show tetrahedral site with 4.00 or close to 4.00 atoms, low  $\text{TiO}_2$  (2 to 3%) and high  $\text{Al}_2\text{O}_3$  (12 to 15%) contents, unusual values for lamproitic micas. These crystals are considered to be xenocrysts. Other large phlogopites may be found

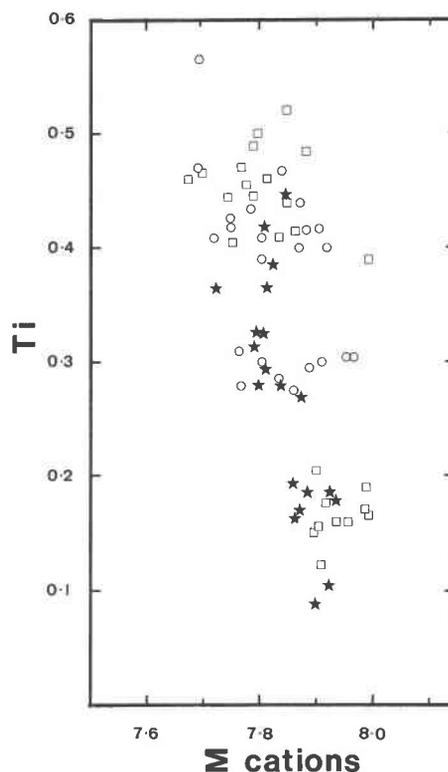


Fig. 6. The distribution of Ti and the sum of cations in phlogopites from 3 specimens. Open squares, 5B6; open circles, SO136; solid stars, 8A12.

either as cores of complex micas, or have served as nuclei for lamproitic micas crystallization. They may (7B8, 5B6) or may not (74A1, 72A7, 7B8, 5B6, 8A12) appear unstable, with Ti-oxide exsolutions. Surrounding lamproitic micas are higher in Ti, Fe and lower in Al than the cores which show  $\text{TiO}_2$  contents around 2%, and significant  $\text{Cr}_2\text{O}_3$  percentages (from 0.30 to 1.0). Such high  $\text{Cr}_2\text{O}_3$  contents have been recorded from peridotites (e.g., Boyd and Nixon, 1978; Delaney et al., 1980; Jones et al., 1982) or from phlogopite centers in some Monument Valley minettes (Jones and Smith, 1983). Typical lamproite micas plot near the phlogopite composition, the  $\text{Fe}/(\text{Fe} + \text{Mg})$  ratio always below 0.30. They are often zoned concentrically, with an increase of both Ti and Fe and a concomitant decrease in Al from center to edge. They show a deficit in the tetrahedral site,  $\text{Si} + \text{Al} + \text{Cr} < 4$  atoms.

Concerning the substitution in the tetrahedral site, several possibilities have been investigated. The presence of  $\text{Fe}^{3+}$  on the tetrahedral site of micas has long been known. Tetraferriphlogopite is found in nature (Cross, 1897) and has been synthesized in the laboratory. Silicon and magnesium can proxy for aluminum on the tetrahedral site (Seifert and Schreyer, 1971; Tateyama et al., 1974).

Robert (1981) showed (using infrared techniques) the presence of tetrahedral Mg in a Smoky Butte lamproite phlogopite. He was also able to demonstrate that this particular mica does not contain any  $\text{Fe}^{3+}$ , and that Ti is exclusively located on the octahedral sites.

*Ti in micas.* Various substitution mechanisms have been proposed to accommodate Ti in phlogopite. If Ti is fixed on the oc-

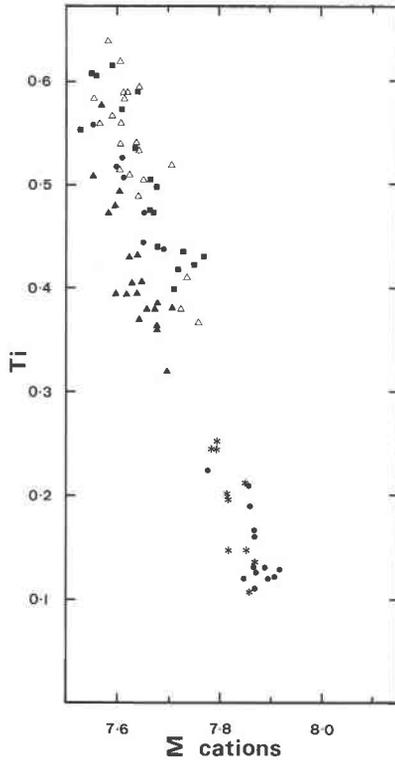


Fig. 7. The distribution of Ti and the sum of cations in phlogopites from 5 specimens. Solid triangles, 74A1; solid circles, 72A7; solid squares, 11A8; asterisks, 73A8; open triangles, 7B8.

tahedral site, it replaces a divalent ion, and hence all proposed mechanisms are coupled substitutions. These substitutions can be limited to the octahedral site or involve either one or two other structural sites. The various mechanisms proposed are as follows:



Substitution (1) was proposed after consideration of the composition of natural phlogopites, and has been recognized by Velde (1979) for micas found in melilite-bearing rocks where it is probably coupled with a substitution involving the incorporation of barium on the interlayer site. When this mechanism is operative a vacancy is created for every 2 Ti atoms entering the structure and the octahedral site occupancy decreases. The existence of substitution (1) is demonstrated in Figures 6 and 7 where Ti is related to the total number of ions. It is of course assumed that both the tetrahedral and interlayer sites are filled and that no  $\text{Fe}^{3+}$  is present in the structure. It appears that for specimens 5B6, 8A12, SO136 the slope of the line defined by the experimental points is close to the expected value, whereas in Figure 7 there are more vacancies than would be expected from substitution (1) alone.

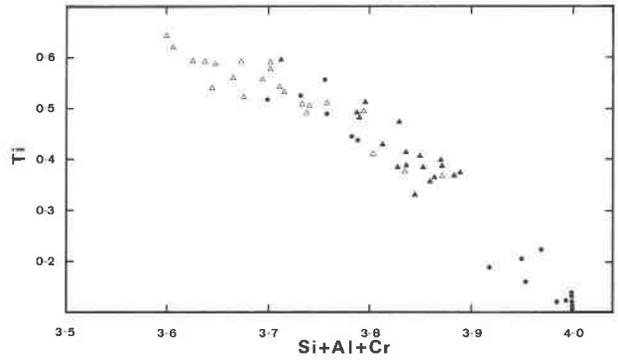


Fig. 8. The distribution of Ti and Si + Al + Cr in phlogopites from 3 specimens. Solid triangles, 74A1; solid circles, 72A7; open triangles, 7B8.

The incorporation of titanium on the octahedral site allows for the introduction of a divalent ion, magnesium (Robert, 1981) on the tetrahedral site (equation 4). In this case, a positive correlation would exist between Ti and Mg (on the tetrahedral site) and hence a negative correlation between Ti and Si + Al + Cr would ensue. Such a correlation is indeed observed for micas of specimens 74A1, 72A7, 11A8, 7B8, 8A12, SO136 (Figs. 8, 9 and 10) with Ti contents above 0.26 atoms per structural formula. Among those micas that do show a negative correlation three different groups can be distinguished: micas from the first group (74A1, 72A7 and 7B8) and the second group (8A12 and 11A8) show similar slopes of the line defined by the experimental points on Figures 8 and 9, but for a given Ti value the site occupancy by Si + Al + Cr of micas from the first group is higher than the occupancy shown for the second group. This difference is mostly due to a difference in Al content, the micas from the first group being richer in Al than micas of the second group. In the third type, which only includes specimen SO136, mica analyses plot along a slope different from that defined by the two other groups. Such a substitution scheme however implies an inverse correlation between Si and Ti that is apparent only in specimen 8A12 (Fig. 11) and, to a lesser extent, specimen 7B8.

There is no correlation between the number of Si + Al + Cr atoms and the Ti content when the number of Ti atoms is below 0.26, indicating a different substitution mechanism. Robert (1976) has proposed substitution mechanism (3) when the Ti content is

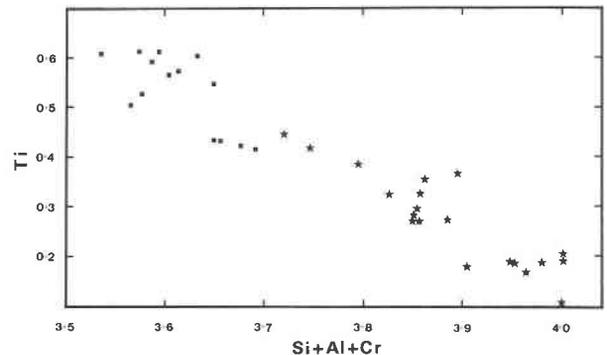


Fig. 9. The distribution of Ti and the sum Si + Al + Cr in phlogopites from specimens 11A8 (solid squares) and 8A12 (solid stars).

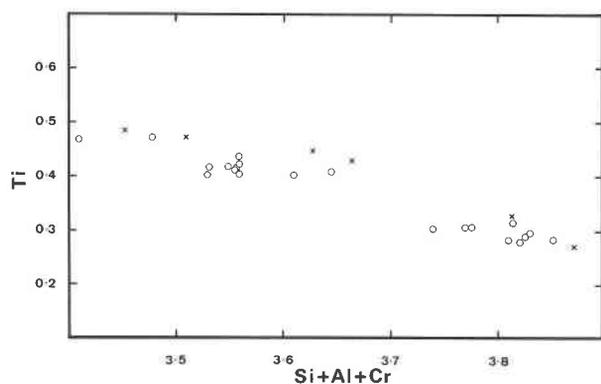


Fig. 10. The distribution of Ti and the sum Si + Al + Cr in phlogopites from specimen SO136 (open circles) and data from Mitchell (1981), crosses.

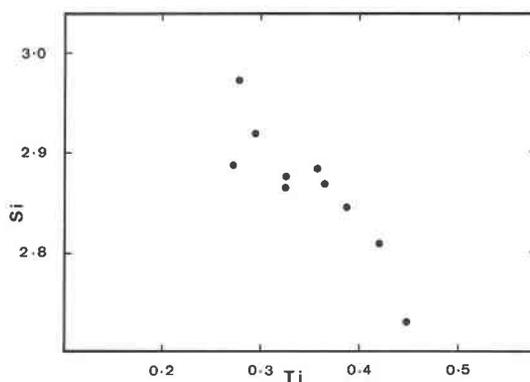


Fig. 11. The distribution of Si and Ti in phlogopites from specimen 8A12, analyses of crystal edges.

below 0.35. This substitution was looked for in analyses corresponding to the centers of micas from specimens 72A7 and 5B6 as well as for micas from 73A8 which show low titanium contents. The possibility of this substitution is confirmed by positive correlations between Ti and Al in all three cases.

In conclusion it seems that the most prominent substitution mechanisms for Ti in the micas studied are substitutions (1) and (4) for Ti-rich compositions. For Ti contents below 0.26 per structural formula the likely mechanisms are (1) and (3).

*Interlayer ions.* Ions in the interlayer site (K + Na + Ba) usually total close to the theoretical value of 1.0 atom per structural

formula, the lowest value found being 0.85. Barium contents are low, usually below 1.0% BaO except for specimens 7B8 and 11A8 where it reaches 3%. This low barium content of micas in rocks where barium is abundant (see for example data for the Leucite Hills volcanics in Carmichael (1967), or as exemplified by the presence of Ba-rich phases in some lamproites such as priderite or barite) is probably due to the alumina-deficiency of the bulk composition. Most barium substitutions involve the mechanism  $Ba + Al = K + Si$ , an unlikely substitution in peralkaline rocks. In the lamproites examined the substitution appears to be of the kind  $Ba + \square = 2K$ . This vacancy-creating substitution could play a role, as does the possible vacancy-creating substitution for Ti. It should be pointed out that micas from specimens 7B8 and 11A8 show a correlation between Ti and Ba, with a different slope in each specimen.

Authigenic micas in lamproites differ from phlogopites from ultrabasic xenoliths by their high  $TiO_2$  contents, but xenocrystic micas found in lamproites are similar, even in their trace element contents, with phlogopites from peridotites.

Table 11. Representative electron microprobe analyses of the roedderite-like phases, from Moon Canyon (analysis 1) and Cancarix (analysis 2). For comparison are listed analyses of roedderite (A, B) and eifelite (C) from Hentschel et al. (1980) and Abraham et al. (1983) respectively

	1	2	A	B	C
$SiO_2$	70.29	69.86	70.60	71.32	69.06
$Al_2O_3$		0.17	0.50	0.36	0.57
FeO <sup>a</sup>	12.60	9.83	5.80	0.49	0.19
MgO	11.44	13.90	15.70	17.86	15.47
MnO	0.23	0.21	0.21	0.31	0.53
$TiO_2$	0.02	0.21	0.07	0.07	0.03
$Na_2O$	0.04	0.25	1.80	5.29	8.07
$K_2O$	4.48	5.14	4.20	4.16	4.16
Total	99.08	99.57	98.88	99.79	98.05
Si	12.310	12.116	12.100	11.818	11.92
Al		0.035	0.101	0.070	0.08
Fe <sup>2+</sup>	1.845	1.426	0.831	0.068 $\alpha$	0.03
Mg	2.987	3.594	4.012	4.412	3.98
Mn	0.034	0.031	0.030	0.044	0.08
Ti	0.003	0.027	0.009	0.009	
Na	0.014	0.084	0.598	1.699	2.70
K	1.001	1.137	0.918	0.879	0.92

FeO<sup>a</sup> = Total iron as FeO

$\alpha$  : Fe 3+

1, Cancarix ; 2, Moon Canyon.

A, B : Hentschel et al., 1980.

C, Abraham et al., 1983.

#### Roedderite-like phases

Both the Cancarix and the Moon Canyon lamproites contain remarkable small prismatic crystals of a pleochroic blue mineral. Electron microprobe analyses of these blue minerals in both specimens have shown that they have similar compositions to minerals of the roedderite-eifelite series (Hentschel et al., 1980; Abraham et al., 1983). Representative analyses from Cancarix, Moon Canyon and of roedderite and eifelite are reproduced in Table 11. The roedderite-like minerals from the lamproites are distinct from those previously described in terrestrial rocks in their composition and in the fact that, for the lamproites, they are part of the mineralogical association, and apparently crystallized from the liquid. They are closely associated with feldspars, amphibole and late crystallizing micas. Their chemical compositions are different from that reported for the Eifel roedderites. The Cancarix mineral (a rare phase in the rock) is a sodium-free and iron-rich relative to the Eifel specimens. The Moon Canyon mineral (quite abundant in the specimen studied) is slightly less iron-rich, more magnesian and sodic than that from Cancarix. The lamproite minerals show a deficit in alkali metals for an  $R^{2+}$  total close to the theoretical 5.0 value. Their general formula could be derived from the roedderite formula  $(Na,K)_2Mg_3Si_{12}O_{30}$  through a substitution of the type  $Fe^{3+} = Fe^{2+} + Na$ . According to this scheme, the entry of  $Fe^{3+}$  in the structure creates one vacant site (an alkali site), the corresponding end-member being

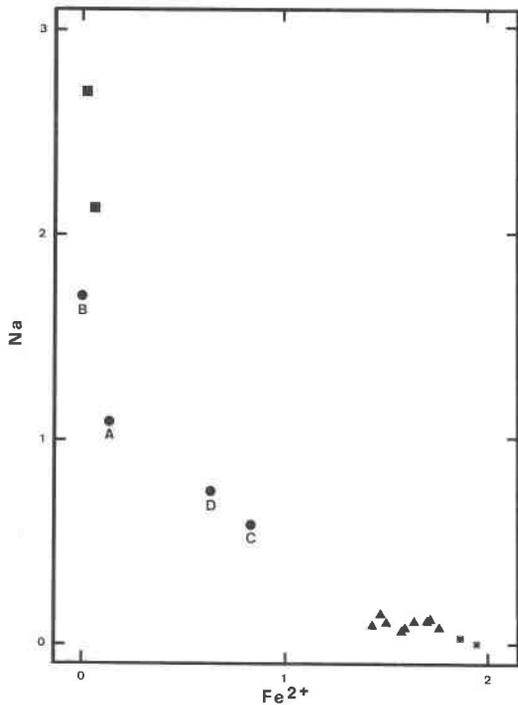


Fig. 12. The distribution of Na and  $\text{Fe}^{2+}$  in roedderite-like phases. Solid triangles, MC77; small solid squares, 74A1; solid squares, eifelites (Abraham et al., 1983); solid circles, roedderites (Hentschel et al., 1980).

$\text{KFe}^{3+}\text{Fe}^{2+}\text{Mg}_3\text{Si}_{12}\text{O}_{30}$ . This substitution has been proposed by Hentschel et al. (1980) and Abraham et al. (1983) for roedderites found in the Bellerberg volcanic field of the Eifel in Germany. The Cancarix and Moon Canyon specimens are close to this theoretical end-member.

These relationships are demonstrated in Figure 12 which represents the variation of the Fe vs. Na. Na is absent, totally replaced by  $\text{Fe}^{3+}$ , in the Cancarix specimen. The iron-free eifelites

Table 12. Representative electron microprobe analyses of priderites from Australian, Sisco and Smoky Butte lamproites

	1	2	3	4	5	6	7
	SO136	SO136	7B8	7B8	7B8	8A12	8A12
$\text{Fe}_2\text{O}_3^\circ$	10.97	9.19	9.53	9.09	7.66	10.75	9.59
MgO	0.92	0.74	1.15	1.02	0.76	0.17	0.03
$\text{TiO}_2$	76.17	72.89	70.43	68.18	66.60	71.46	70.83
$\text{Cr}_2\text{O}_3$	0.03	4.36	0.83	1.85	5.16		0.26
$\text{K}_2\text{O}$	6.82	7.15	2.56	2.73	2.67	2.27	1.37
BaO	5.43	6.30	16.22	16.73	16.93	16.39	17.08
Total	100.34	100.63	99.77	98.69	99.01	99.96	99.16
$\text{Fe}^{3+}$	0.980	0.830	0.901	0.877	0.741	1.013	0.922
Mg	0.163	0.132	0.215	0.195	0.146	0.032	0.006
Ti	6.798	6.580	6.653	6.573	6.442	6.732	6.804
Cr	0.003	0.413	0.082	0.187	0.525		0.026
K	1.032	1.095	0.410	0.446	0.438	0.363	0.223
Ba	0.252	0.296	0.798	0.840	0.853	0.803	0.853
Total	9.227	9.346	9.060	9.118	9.144	8.943	8.834

$\text{Fe}_2\text{O}_3^\circ = \text{Total iron as Fe}_2\text{O}_3$

Table 13. Representative electron microprobe analyses of pseudobrookites from various lamproites studied.  $\text{Fe}^{3+}$  calculated assuming 3 cations and 5 oxygens

	1	2	3	4
	74A1	73A8	7B8	MC77
$\text{Al}_2\text{O}_3$	0.02	0.47	0.30	0.06
$\text{Cr}_2\text{O}_3$	0.82	0.43	0.99	0.05
$\text{FeO}^\circ$	28.97	19.66	19.86	23.46
MgO	8.65	9.39	8.92	7.61
MnO	0.33	0.15	0.14	0.20
$\text{TiO}_2$	60.33	69.02	69.04	66.32
Total	99.12	99.12	99.25	97.70
$\text{Fe}_2\text{O}_3$	26.38	8.57	7.71	11.03
FeO	5.24	11.94	12.92	13.54
Al	0.001	0.020	0.013	0.003
Cr	0.023	0.012	0.028	0.001
$\text{Fe}^{3+}$	0.714	0.232	0.209	0.307
$\text{Fe}^{2+}$	0.157	0.359	0.390	0.419
Mg	0.464	0.504	0.480	0.420
Mn	0.010	0.005	0.004	0.006
Ti	1.631	1.868	1.875	1.844

$\text{FeO}^\circ = \text{Total iron as FeO}$

plot along the horizontal axis, as they only show a  $2\text{Na} = \text{Mg}$  substitution.

The deep blue color of the roedderite-like phase appears to be related to the presence of ferric iron in the structure. The Cancarix mineral, richer in iron (total iron) than the Moon Canyon one, has a deeper blue color. Hentschel et al. and Abraham et al. pointed out that their roedderites C and D, with 5.8 and 4.45% total FeO show a blue color.

Roedderite appears among the low-pressure breakdown products of richterite (Gilbert et al., 1982) and testifies to the strongly peralkaline character of these two lamproites.

### Oxides

Titanium-rich oxides are a general feature of lamproites, as pointed out by Scott Smith and Skinner (1984). In this respect, lamproites differ from peralkaline siliceous rocks that are known for the absence of oxides (Carmichael, 1967). Besides chromites found as inclusions in olivine (considered below) these rocks contain one type of Fe-Ti oxide: ilmenite (11A8), or an association of two oxides: ilmenite + pseudobrookite (MC77 and 73A8); ilmenite + titanomagnetite (5B6); ilmenite + priderite (SO136, Sisco); pseudobrookite + priderite (7B8). These minerals and mineral associations are uncommon and will be described in some detail.

Priderite has been found and analyzed (Table 12) in rocks from Sisco, and Smoky Butte besides the Australian type locality. The original priderite contained 0.32 atoms of Ba and 0.87 atoms of potassium. Carmichael (1967) gave two electron microprobe analyses of priderites from the Leucite Hills and Australia. The priderites from Smoky Butte (where priderite is present in Type 1 lamproite, as noted by Velde (1975) but has also been found in specimen 7B8) are particularly Ba-rich, and the Ba content can reach 0.85 atoms, whereas K is only 0.4 per structural formula calculated on 16 oxygens. These high Ba-values imply, as hypothesized by Post et al. (1982), the existence of a Ba end-member of the priderite series. Another striking variation in the analyzed

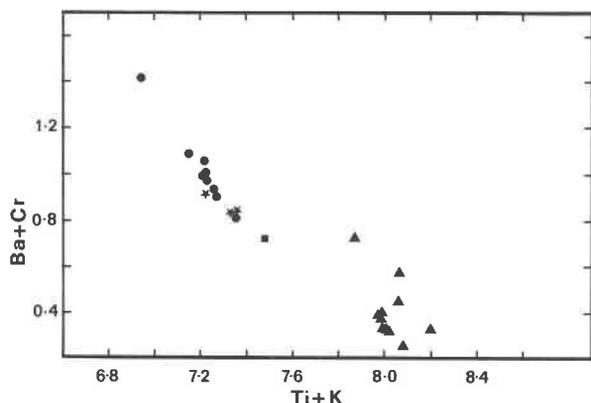
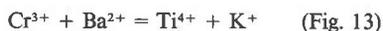


Fig. 13. The distribution of Ba + Cr and Ti + K in priderites. Solid circles: 7B8; solid stars: 8A12; solid triangles: SO136; solid square: LH9, Carmichael (1967).

priderites is their Cr-content: the priderites found in the Australian specimen may contain 4.36%  $\text{Cr}_2\text{O}_3$ , the minimum found being 0.03. The  $\text{Cr}_2\text{O}_3$  content of the Smoky Butte priderite reaches 5.16%. The general formula of priderite (Norris, 1951),  $\text{A}_{2-y}\text{B}_{8-z}\text{O}_{16}$ , is not adequately known, but it seems that the variations observed in the three lamproites mentioned are compatible with a substitution of the type:



This scheme does not however account for the observed variation in the iron content, which could be at least partly in the ferric state and consequently participate in the above equation.

The presence and abundance of *pseudobrookite*, a rare mineral in terrestrial rocks, is peculiarly common to some of the lamproites studied. *Pseudobrookite* usually forms euhedral plates, hexagonal in shape, opaque or barely translucent in transmitted light. When translucent (as in Smoky Butte) crystals can appear either green or purple brown a difference that does not appear to be simply related to compositional variations. In specimen MC77 *pseudobrookite* appears as transparent golden brown crystals, with optical properties very close to those of rutile. In reflected light, these *pseudobrookite* crystals show reddish internal reflections. The composition of these oxides is 40% ferropseudobrookite, 12 to 15% *pseudobrookite* and 42 to 48% *karrooite*. Representative analyses are listed in Table 13. Specimen 73A8 contains *pseudobrookite* crystals, interpreted as xenocrysts which are extremely rich in magnesium (up to 52 mole% of the *karrooite* component). This particular *pseudobrookite* is close in composition to *armalcolite* (Anderson et al., 1970; Haggerty, 1973; El Goresy and Bernhardt, 1974). *Pseudobrookite* has been found, as a primary mineral, in few terrestrial rocks: in meteorite impactites (El Goresy and Chao, 1976), in kimberlites (Haggerty, 1975; Raber and Haggerty, 1979), in ultrabasic nodules (Cameron and Cameron, 1973) and lamproites (Velde, 1975). It has also been found in some basalts (Anderson and Wright, 1972; Van Kooten, 1980; Von Knorring and Cox, 1961), and in meteorites (Fujimaki et al., 1981). It was described by Rice et al. (1971) as a quenched mineral in a dike rock of basaltic composition. According to Haggerty (1976), the presence of primary *pseudobrookite* implies a fast cooling rate that prevents a reaction of the *pseudobrookite* with the liquid, a reaction whose end product would normally be *ilmenite*. It should be pointed out that in no lamproite was any *ilmenite* rim found around the early-formed *pseudobrookite*.

*Ilmenite* is accompanied by *priderite* in specimen SO136 where *priderite* forms abundant small elongated crystals, as well as in some of the Sisco specimens (Velde, 1968) where it is quite rare. *Ilmenite* forms small crystals (5 to 50  $\mu\text{m}$  maximum) without evidence of exsolution or oxidation, except for specimen 74A1. *Ilmenites* appear after olivine, diopside and the centers of the phlogopites crystals. They are usually poor in hematite, containing from 64 to 90% of the *ilmenite* component, 11 to 30% *geikelite* and 1 to 2% *pyrophanite*. Representative analyses are listed in Table 14. Specimen 73A8 contains probable xenocrysts of a magnesian *ilmenite* with a rutile core; the composition of the *ilmenite* is similar, though not as Mg-rich, to that of xenocrystic Mg-*ilmenites* described by McMahon and Haggerty (1984), a composition close to the composition of *ilmenites* found in kimberlites (Dawson, 1980).

*Chromites* (Table 3) found as inclusions in authigenic olivines are similar by their high  $\text{Cr}_2\text{O}_3$ , low  $\text{Al}_2\text{O}_3$ , and high  $\text{Mg}/(\text{Mg} + \text{Fe})$  to *chromites* found by Danchin and Boyd (1976) in *harzburgites*, themselves, as pointed out by these authors, not unlike *chromites* found as inclusions in diamonds. Their low Al is in keeping with the peralkaline chemistry of the magma, their high Cr could be related to the low Al and  $\text{Fe}^{3+}$  contents of the liquid.

Ti-rich oxides are commonly abundant in peralkaline lamproites. This presence and abundance could be ascribed to the high  $\text{TiO}_2$  and high FeO contents of these rocks. When projected in an  $\text{FeO}-\text{Fe}_2\text{O}_3-\text{TiO}_2$  diagram the composition of most *pseudobrookite* (+ *ilmenite*) bearing samples plot close to the *ferropseudobrookite*-*pseudobrookite* line or do so when the amount of iron (and  $\text{TiO}_2$ ) partitioned in the early-crystallized silicates is taken into account.

Most of the oxides do not show any exsolution or oxidation textures and most compositions can be recalculated into ferrous, rather than ferric, end-members, or minerals with high ferrous iron contents. This observation is taken as an indication of the low oxygen contents of the melts, compatible with the low  $\text{Fe}^{3+}/\text{Fe}^{2+}$  ratio of most ferromagnesian silicate phases present in these lavas.

Oxides are known to occur with *phlogopite* and *richterite* in ultrabasic nodules, rutile and *ilmenite* being common, *pseudobrookite* rare.

### Conclusion to the mineralogical description

Lamproites described above can be considered as presenting a magmatic assemblage consisting of Al-poor, Cr-rich spinel, Ni-rich olivine, Ti-phlogopite, Ti-rich K-*richterite*, diopside, Fe-rich feldspar and Ti-rich oxides. Olivine crystals having evident textural (kink-bands) analogies with *peridotitic* olivines are also found and can be interpreted as being inherited from the solid layers of the earth through which the lamproitic magma percolated. Other such inherited minerals are Cr-rich diopsides, Fe-rich clinopyroxenes, Cr-rich *phlogopites*, *pseudobrookites*, Mg-*ilmenites* and large rutile crystals (accidental in some Spanish lamproites). Jones and Smith (1983) have also demonstrated the presence, in the Navajo minettes, of *apatite* and *clinopyroxenes* xenocrysts. The possible presence of various xenocrystic species in the Leucite Hills lavas was also discussed by Kuehner et al. (1981) who, for example, described *pleonaste* in the Deer Butte *wyomingite*.

Hence care should be exercised before discussing either

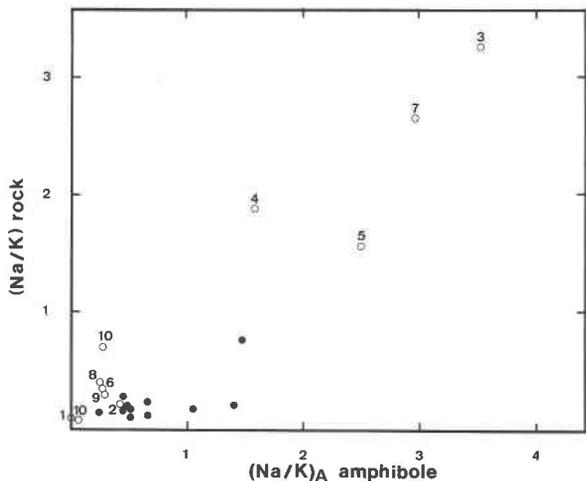


Fig. 14. The distribution of Na/K calculated from the bulk rock composition and the Na/K content of the A site of the richterite. Solid circles: data from this study; open circles: data from the literature: 1, LH9, Carmichael (1967); 2, Type 1, Velde (1975); 3, Velde (1978); 4, S84, Bryan (1970); 5, RH1, Barker and Hodges (1977); 6, Brooks et al. (1978); 7, 121, Nicholls and Carmichael (1969); 8, Sheraton and England (1980); 9, S1, Carmichael (1967); 10, Jaques et al. 1984.

the partitioning of elements between phases or the significance of the trace element geochemistry of these rocks.

**Partitioning of elements between phases**

Some elements are found in certain phases in proportions which can be correlated with the bulk rock composition. The Fe<sup>3+</sup> content of the feldspar is simply correlated with the peralkalinity index of enclosing rock. Micas are more potassium rich in the most potassium rich bulk composition, a relation already described by Edgar and Arima (1983). Richterites contain variable amounts of potassium and there is a relationship between the potassium content of the richterite and the amount of potassium present in the rock expressed in the form of the

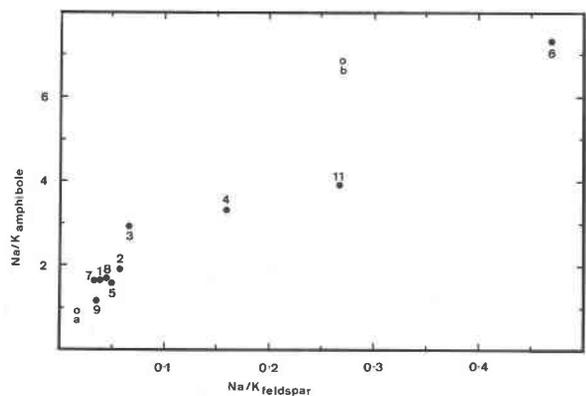


Fig. 15. Partitioning of Na and K between richterites and feldspars. Data from this study: solid circles; data from the literature: open circles (a, LH9, Carmichael, 1967; b, Velde, 1978).

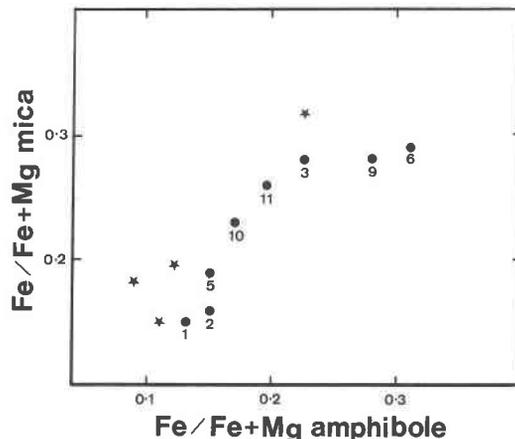


Fig. 16. Partitioning of Fe and Mg between phlogopites and richterites. Data from lamproites: solid circles; data from ultramafic nodules (Aoki, 1974, 1975; and Dawson and Smith, 1977): solid stars.

Na/K ratio. Figure 14 shows this correlation, as deduced from specimens under consideration and examples taken from the literature. Richterites also appear to be more potassic when the rock is more peralkaline, which could be another way of expressing the previous correlation, the peralkalinity in lamproites being directly related to the potassium content.

This being established, how are these elements distributed between co-crystallizing phases? Since richterite and phlogopite coexist in some ultrabasic nodules brought to the surface by kimberlitic magma, these nodules represent physical conditions different from those of lamproitic crystallization, in particular different pressure conditions. These data have been included in Figures 16 to 18. In

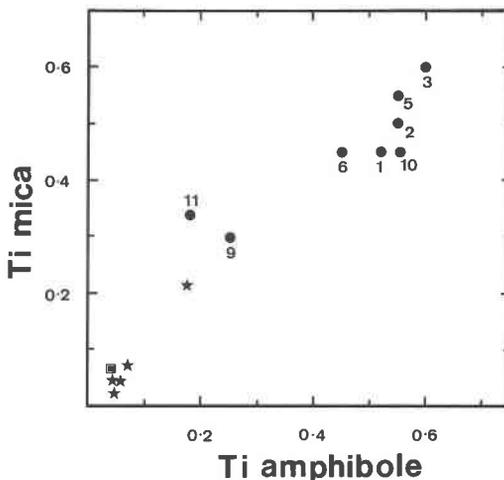


Fig. 17. Partitioning of Ti between micas and richterites. Data from lamproites: solid circles; data from nodules: solid stars (Aoki, 1974, 1975; and Dawson and Smith, 1977) and solid squares (data from Jones et al., 1982).

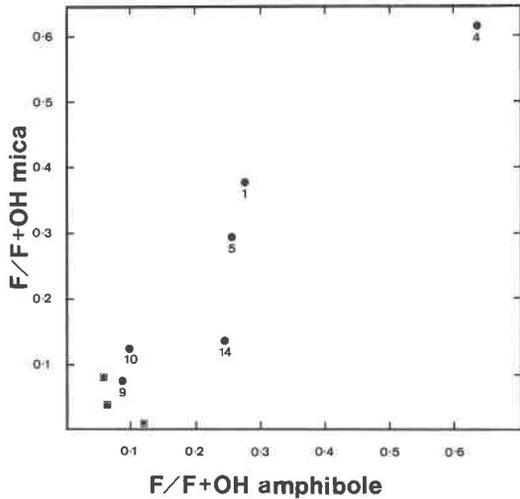


Fig. 18. Partitioning of fluorine between phlogopites and richterites. Data from lamproites: solid circles; data from nodules (Jones et al., 1982): solid squares.

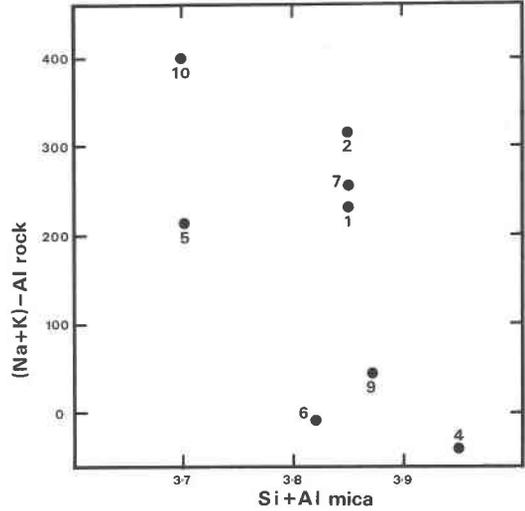


Fig. 20. The distribution of  $(Na + K) - Al$  calculated from the rock bulk compositions and the  $Si + Al$  content of the phlogopites.

this way the influence of pressure on the composition of the minerals under consideration can be evaluated.

Sodium and potassium: there is a direct relationship between the  $Na/K$  ratio of richterite and the coexisting feldspar (Fig. 15).

Iron is equally distributed between amphibole and phlogopite (Fig. 16) in lamproites as well as in ultrabasic nodules found in kimberlites but this is only valid when considering those micas and amphiboles in equilibrium, judged so on the basis of textural arguments (Table 15<sup>1</sup>).

There is a positive correlation (Fig. 17) between  $Ti$  in micas and amphiboles. It should be noted that the localization of titanium in the amphibole structure is still ambiguous, and that the problem, for micas, cannot be considered as unambiguously solved even if  $Ti$  is probably incorporated on the octahedral site in lamproite micas. Lower  $Ti$ -contents appear to characterize micas and amphiboles from ultrabasic nodules.

Fluorine: the maximum content of fluorine found in a

natural mica is 8.5% (Peterson et al., 1982; Valley et al., 1982) and 4.5% in a natural richterite (in a meteorite, Olsen et al., 1973). Synthetic minerals contain up to 9.2%  $F$  for phlogopites (Kohn and Hatch, 1955) and 4.7% in richterite (Kohn and Comeforo, 1955). In lamproite phlogopites the fluorine content is, in most cases, below 1%, but can reach 4% (specimen 73A8). These  $F$  contents are in agreement with what can be calculated from the  $F$  content of the rock and the modal amount of phlogopite, with the exception of specimen 11A8, a rock particularly rich in  $F$ -bearing apatite. Representative analyses of  $F$ -rich micas and amphiboles from lamproites are listed in Table 16.<sup>1</sup>

Figure 18 seems to indicate that  $F$  is equally distributed

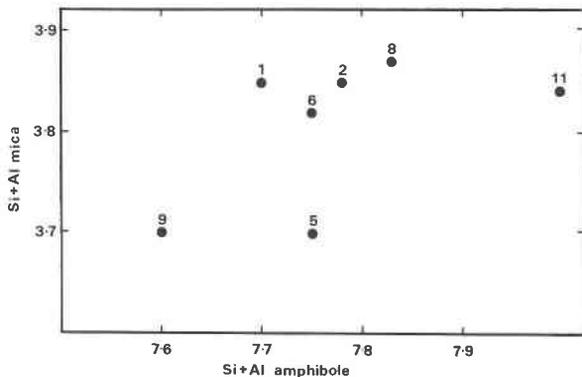


Fig. 19. The distribution of  $Si + Al$  in phlogopites and richterites.

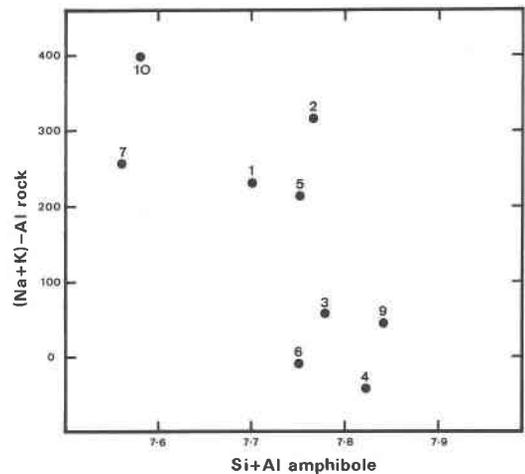


Fig. 21. The distribution of  $(Na + K) - Al$  calculated from the rock bulk compositions and the  $Si + Al$  content of the richterites.

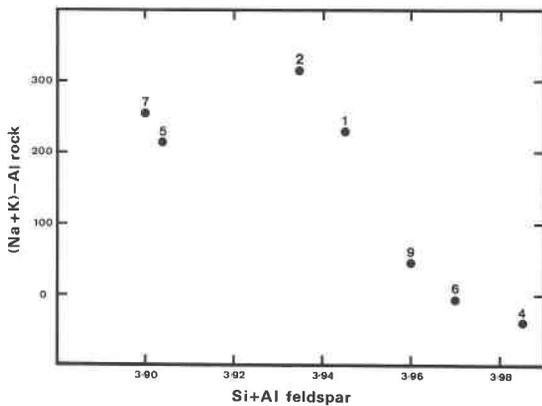


Fig. 22. The distribution of  $(\text{Na} + \text{K}) - \text{Al}$  calculated from the rock bulk compositions and the  $\text{Si} + \text{Al}$  content of the coexisting feldspars.

( $K_D = 1$ ) between the two phases phlogopite and richterite (Table 15). It should be pointed out that the diagram includes values for specimen 73A8 in which the two phases are not considered to be in equilibrium (textural evidence); their  $F$  values follow the same trend as that determined for phases thought to be in equilibrium. This observation could be interpreted as indicating that  $F$  and  $\text{OH}$  can equilibrate after crystallization of the two phases. The  $K_D$  values calculated using the  $F$  determinations published by Jones et al. (1982) for mica and amphibole pairs in metasomatized peridotites are widely different from one another and from those given here. This anomaly cannot be explained at present. Note should be made of the fact that, in the Spanish lamproites, phlogopites show a distinct correlation between  $F$  and  $\text{Ti}$ . This correlation was found in richterites only from specimens 74A1, 73A8 and 7B8.

**Si + Al on the tetrahedral sites:** most of the phases present in the peralkaline lamproites do not contain enough  $\text{Si} + \text{Al}$  to fill the tetrahedral site to its theoretical number of ions. This point was discussed in the preceding section and, in all cases, the nature of the other ion (ions) could not be decided upon with the data at hand.

In spite of the unknown nature of the supplementary ion (or ions) on the tetrahedral site it is interesting to note that the amplitude of this  $\text{Si} + \text{Al}$  deficiency is similar in the coexisting minerals within a given specimen. Figure 19 shows the relationship between the value of  $\text{Si} + \text{Al}$  in the micas and in the coexisting amphibole. There is a positive correlation between the tetrahedral  $\text{Si} + \text{Al}$  content of both phases. Because tetrahedral deficiency appears to be linked to the peralkalinity of the liquid, the  $\text{Si} + \text{Al}$  content of various phases has been plotted against the  $(\text{Na} + \text{K}) - \text{Al}$  value of the corresponding bulk composition (Figs. 20 to 22). The apparent correlation is not valid for the Jumilla specimen, already commented upon.

Barium is preferentially incorporated in micas (Higuchi and Nagasawa, 1969; Wörner et al., 1983) but is also present in coexisting feldspars. The distribution of Ba be-

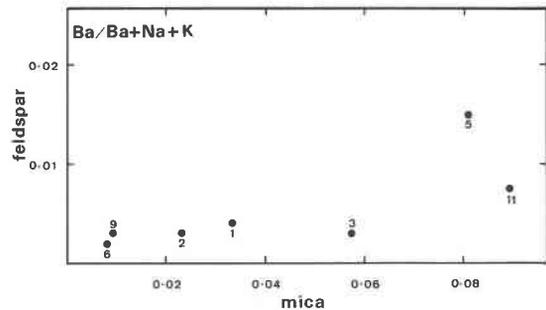


Fig. 23. The distribution of  $\text{Ba}/(\text{Ba} + \text{Na} + \text{K})$  in feldspar and coexisting mica.

tween feldspar and mica in the lamproites is regular as shown on Figure 23. Data from Wörner et al. (1983) indicate that  $\text{BaO}$  contents in phlogopites are twice the  $\text{BaO}$  contents in associated sanidines. Three specimens show  $K_D$ 's for Ba between feldspar and mica that are quite comparable: 0.196 (7B8), 0.192 (74A1) and 0.175 (72A7) but low when compared to values found in other types of volcanic rocks.

Most elements are regularly partitioned between coexisting amphiboles and phlogopites. The bulk composition (either iron, magnesium or titanium contents) exerts some control on the corresponding contents of the minerals in these elements. The silica saturation is however not an important factor in the control of the mineral compositions, whereas aluminum saturation, on the contrary, is a determinant factor particularly on the content of the tetrahedral site.

#### Temperature estimations of cooling history

A number of experimental studies have been made on either phlogopites, richterites or oxides. These data can tentatively be used to estimate temperature of crystallization.

Initially a spinel-olivine geothermometer (Sack, 1982) can be used. The authigenic rim around olivine xenocrysts from the Calasparra lamproite give  $1000^\circ\text{C}$  (72A7) whereas the olivines from the rock give  $800^\circ\text{C}$ . The selagite spinel-olivine associations give values between  $914$  and  $1160^\circ\text{C}$ , whereas a  $1000^\circ\text{C}$  phenocryst temperature has been calculated for the jumillite (11A8). In this specimen an olivine coexisting with glass inside an apatite crystal gave a high crystallization temperature of  $1221^\circ\text{C}$  with the method outlined by Watson (1979).

The maximum thermal stability of the lamproite amphiboles is between  $985$  and  $1030^\circ\text{C}$  (Charles, 1975, 1977) taking into account only the  $\text{Mg-Fe}$  substitution. Richterite is more stable when it is more magnesian and Barton and Hamilton (1978) have shown that a richterite from the Leucite Hills was stable up to  $1075^\circ\text{C}$  at  $\text{H}_2\text{O}$  pressures of 1 kbar. A crystallization temperature, based upon the partitioning of Na and K between amphibole and liquid, has been calculated for two specimens (Smoky Butte and Jumilla) where amphibole and coexisting glass composi-

tions were known, using the relation established by Helz (1979): the values found are 965°C for Smoky Butte and 770°C for Jumilla.

Attempts to estimate temperature of crystallization using the oxide phases were obviously hindered by the absence of magnetite and by possible secondary oxidation. However one sample (specimen 5B6) contains coexisting ilmenite and titanomagnetite which show neither oxidation nor exsolution. These oxides have appeared early in the crystallization sequence, judging from textural relationships, after clinopyroxene. While titanomagnetite appears homogeneous, there is a variation in the  $\text{Fe}_2\text{O}_3$  of the ilmenites which decreases from the center towards the rims of the crystals. Using a program published by Giorso and Carmichael (1981), two sets of temperatures were obtained: for the centers of the ilmenites and the magnetites the calculated temperature is 827°C for a log  $f_{\text{O}_2} = -13.5$  whereas considering the edges of the ilmenites the calculated crystallization temperature is 755°C for log  $f_{\text{O}_2} = -15$ . These results are similar to those deduced from the phase diagrams published by Haggerty (1976).

These temperature estimations for various minerals or mineral pairs allows one to establish the following:

- (1) Xenocrysts with included spinels were not found. Olivine xenocryst rims crystallized at temperatures of 1000–1200°C.
- (2) Groundmass olivine crystallized at 800–900°C.
- (3) Amphiboles formed at temperatures between 960 and 770°C.
- (4) When ilmenite and magnetite are present (Shiprock) temperatures between 800–700°C are indicated.

These values pertaining to a range of lavas with differing compositions are in broad agreement with the liquidus determinations of Carmichael (1967) who found 1165° and 1200°C for two different wyomingites. The data published by Barton and Hamilton (1978) also for wyomingites, in water-saturated conditions, are in general agreement with the crystallization order found here.

*High pressure–low pressure phases.* A comparison should now be made with mineral composition recorded from ultrabasic inclusions. Common characteristics between richterites found in nodules and richterites found in lamproites are the following: their potassium content is similar, as is Na/K, their tetrahedral site is incompletely filled by Si and Al atoms, but  $\text{Fe}^{3+}$  is present in amounts great enough to saturate the tetrahedral site. These richterites can be distinguished from those found in lamproites by a lower  $\text{TiO}_2$  content (less than 0.1 Ti atoms per structural formula). Kushiro and Erlank (1970) have obtained recrystallization at 1100°C and water pressure of 30 kbar of a potassic richterite close in composition to the Australian richterite used as starting material, except for  $\text{TiO}_2$  content: the  $\text{TiO}_2$  content of the starting material was 3%, whereas the artificial richterite contained only 0.5%  $\text{TiO}_2$ . This result implies that the entry of titanium in the richterite structure is favored by low pressure.

Phlogopite, as well as richterite, is an important con-

stituent of some of the nodules found in kimberlites and both minerals have been found as inclusions in diamonds (Meyer and McCallum, 1984). Phlogopite-bearing nodules are more frequent than nodules carrying both phlogopite and richterite. These phlogopites (Dawson and Smith, 1977; Aoki, 1974, 1975) contain less Ti than the phlogopites found in lamproites (less than 0.1 atoms per structural formula), but they also show a tetrahedral site incompletely filled by Si + Al.

The influence of titanium on the stability of micas is not precisely known. Forbes and Flower (1974) envisioned that Ti stabilized the phlogopite; Robert (1981) and Robert and Chapuis (1982) have considered Ti-rich micas as high-pressure phases and this hypothesis is also accepted by Bachinski and Simpson (1984) who consider the Ti-rich (2.0 to 11.3%  $\text{TiO}_2$ ) cores of minette phenocrysts to have crystallized at elevated subcrustal pressures. However, Edgar et al. (1976) established that between 10 and 30 kbar Ti-poor phlogopite is stable relative to Ti-rich phlogopite. Wendlandt and Eggler (1980) have furthermore shown that the Ti content of the phlogopite decreased with pressure for identical compositions, a proposition confirmed by Arima and Edgar (1981). All lamproite micas considered in the present study, excepting xenocrysts, have crystallized at low pressures being found in rocks that have solidified at the surface. Consequently, Ti-rich phlogopites and amphiboles crystallize under pressures corresponding to the surface of the earth. The absence of Ti-rich phlogopites in kimberlite nodules, in assemblages that, for the most part, include Ti-oxides could be interpreted as meaning that Ti-phlogopite is not stable at high pressure.

### Discussion and conclusions

K-Ti-richterites and Ti-phlogopites are the prominent minerals found in peralkaline lamproites. They commonly coexist with Ti-rich Fe-Ti oxides and may be accompanied by rare alkali-rich phases such as priderite, or roederite-like minerals. These associations testify to the peralkaline composition of the liquids from which these minerals precipitated. Coexistent with, but genetically unrelated to this magmatic association, are found a number of phases that have been scavenged by the ascending lamproitic magma. These phases are the same species as those precipitating from the lamproitic magma, i.e., olivine, clinopyroxene, trioctahedral mica, ilmenite, pseudobrookite, rutile and apatite. The rate of dissolution being governed by the minerals stable on or close to the liquidus (Brearley and Scarfe, 1984). The volumetric importance of these relics is not considerable (less than 10%) but, taking into account experiments performed by Scarfe et al. (1980) on basalts, material might have been dissolved into the liquid. Incorporation of Ni-rich olivines at depth could, for example, account for the precipitation of Ni-rich olivines from the magma at the surface; incorporation of Cr-rich phases for the precipitation of Cr-diopside or Cr-rich spinels.

Results of the present study show that peralkalinity of

lamproites is a characteristic of the lamproitic liquid, because of the correlations between some mineral compositions and the alkalinity index, or the precipitation of Al-poor chromite as a liquidus phase. Hence contamination during the ascent of the magma cannot account for the high alkali (potassium) and low Al contents.

Partitioning of elements between phases, excluding xenocrysts, is regular and follows trends in keeping with those that can be deduced from mantle metasomatic assemblages of phlogopite and richterite. A conclusion is that the high K-richterites found in mantle xenoliths are probably derived from the reactions of highly potassic fluids with ultrabasic rocks. The difference between Ti contents of phlogopite and richterite in lamproites (low pressure) and metasomatized peridotites (high pressure) reflects the higher solubility of Ti in these phases at low pressure.

A deep seated origin of these liquids is suggested by two series of observations:

The presence of peridotite xenoliths known to occur in the Leucite Hills (Carmichael, 1967; Barton and Bergen, 1981), in the Australian lamproites (Jaques et al., 1984) and in the Spanish lamproites (Velde, 1969; Nixon et al., 1982).

The discovery of diamonds in a number of Australian lamproite localities (Jaques et al., 1984). This observation alone places the lamproites in a category of rocks akin to kimberlites, at least for the depth at which they originate. Trace element data, and in particular rare earth profiles, when available, indicate for lamproites, as for kimberlites, an extreme enrichment in light rare earth elements (500 to 2000 the chondritic values), and a weaker enrichment in heavy rare earth (Nixon et al., 1982; Mitchell and Hawkesworth, 1984).

The bulk composition of lamproites is so different from common magmatic rocks that their deviation from, for example, basaltic magmas is not permitted. As pointed out by Carmichael (1967) when considering the genesis of the Leucite Hills volcanics, the temptation is to impose on the starting material many of the requirements of the result that is supposed to be achieved. Since Carmichael's study, however, the concept of mantle metasomatism has been widely commented upon (Bailey, 1982). Ultrabasic rocks have been described with a metasomatic assemblage resulting from the reaction of mantle associations with K-rich fluids (Jones et al., 1982; Barton and Hamilton, 1982) that could be precisely the source of ultrapotassic peralkaline rocks.

It is tempting to pursue this reasoning and calculate what percentage of the bulk compositions of lamproites can be explained with the fusion of a richterite-phlogopite-rutile (or ilmenite)-diopside assemblage. Least square calculations performed using the compositions of these phases as listed by Dawson (1980, p. 186) lead to the following conclusions: all the bulk compositions of lamproites listed in Table 2 can be obtained by the fusion of the above kimberlitic nodule assemblage. In all cases, required percentages of phlogopite are between 50 and

70% (weight), but most cluster between 56 and 68%. Required percentages of richterite vary between 30 and 50%, except for Jumilla, whose bulk composition has been altered and will not be considered further, Shiprock and the Australian lamproite. In these two last samples, diopside must be included in the starting material. In all calculations rutile is added, the proportions being above 1% (weight) only for the most titaniferous compositions, Smoky Butte and Australia. Recent experiments (Dickinson and Hess, 1984) have indeed shown that the solubility of rutile ( $K^*$ ) increases dramatically with the  $K_2O/K_2O + Al_2O_3$  of the liquids, from 9.5 wt.% ( $K^* = 0.37$ ) to 41.2 wt.% ( $K^* = 0.90$ ).

For the calculated values to match the observed lamproite compositions, fractionation of olivine, and clinopyroxene, and a certain alkali loss must be called for.

The fact that lamproite major elements compositions can be matched in such a fashion cannot be considered as proof of a particular genetic process. The coincidence simply permits this scheme to be used as a working hypothesis for a more rigorous treatment of the problem that would include data on trace elements combined with the precise determination and evaluation of xenocrystic material that could further constrain the problem.

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