11 The Kapamba lamproites of the Luangwa Valley, eastern Zambia

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ABSTRACT

The Kapamba lamproites, of the Luangwa Valley in eastern Zambia, comprise 14 pipe-like bodies and an associated suite of dikes. They form a NW-SE trending group approximately 25 km long. The Kapamba bodies cut upper Karoo sediments and may have an age of 220 Ma. The pipes range in area up to 45 ha. They are craters predominantly infilled with pyroclastic, often well bedded, lapilli tuffs composed of variable proportions of juvenile lapilli and xenocrysts. Some of the tuffs are intruded by younger magmatic material that appears to have formed extrusive, ponded lava lakes. The magmatic rocks and juvenile lapilli contain diopside, leucite (or its pseudomorphs), olivine, titaniferous phlogopite, titanian potassic richterite, sanidine, spinel (titanomagnetite and chromite), perovskite and glass. Some of the bodies contain mantle-derived minerals including diamond. The Kapamba bodies, therefore, represent another province of diamond-bearing lamproite. There is a suggestion that the pipes become relatively more evolved from the north-west to the south-east of the province. The olivine lamproites to the north-west produced most of the diamonds, while the leucite lamproites to the south-east produced rare diamonds. The diamonds include abundant yellow and brown stones and the tetrahexahedroid crystal form predominates. The Kapamba diamonds are similar to other lamproitic (and kimberlitic) populations. The Kapamba province occurs off-craton in the Irumide (ca. 1300 Ma) tectonic belt and, together with three kimberlite provinces, is associated with the Luangwa graben.

Keywords: lamproite, leucite, Luangwa Valley, Zambia.

11.1 INTRODUCTION

A group of pipes and dikes are located along the Kapamba River, a tributary of the Luangwa River, approximately 150 km west-north-west of Chipata in eastern Zambia (Fig. 11.1). The four largest pipes (P1 to P4) in the Kapamba group were discovered in 1961 and 1962 by Chartered Exploration Limited during prospecting of the southern Luangwa Valley. Intrusions P5 to P14 and four dike systems were found later by De Beers Prospecting (Zambia, formerly Rhodesian Areas) Limited. Prospecting continued intermittently until 1972. The occurrences were initially referred to as kimberlites (Dawson 1970, 1980). Bulk sampling of the bodies revealed very low diamond grades. The more northerly bodies (P1, P2, P7, P8) contained a few diamonds with the majority of stones being found at P2. Rare diamonds were also recovered from P3, P4, P9 and P10.

11.2 GEOLOGY

11.2.1 General geology

The Kapamba bodies occur in the Irumide tectonic belt or Kibaran province. The Irumide metamorphism and tectonism is dated at 1355 Ma but subsequent metamorphic overprinting has occurred in some areas (Cahen *et al* 1984). The intrusions lie just within (Fig. 11.1) the downfaulted, NE-trending Luangwa graben. The major faulting is post-Karoo or early Jurassic to Lower

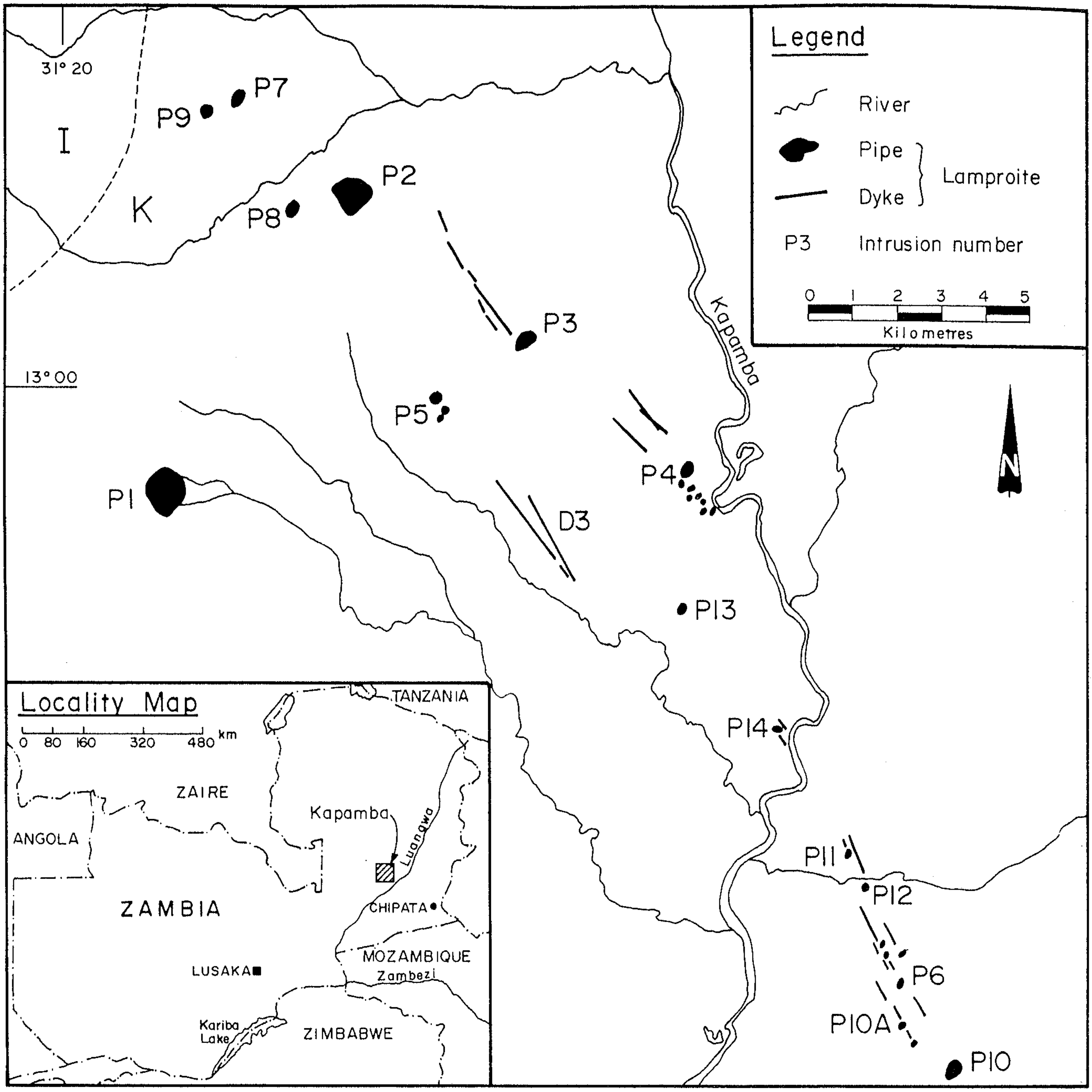

Fig. 11.1 Location of the Kapamba lamproites. The dotted line in the north-west part of the main map is a major graben fault and boundary between the Irumide basement rocks (I) and the Karoo (K).

Cretaceous with some rejuvenation in Tertiary to recent times (Bailey 1961, 1974; Vail 1968; Scholz *et al* 1976). Previous authors (e.g. Holmes 1965; Scholz *et al* 1976; Chapman & Pollack 1977) have related the Luangwa graben to the East African rifting although unequivocal evidence for such a relationship is difficult to establish. The Kapamba lamproites show a NW–SE alignment, a trend which differs notably from that of the main NNE–SSW-trending fault of the Luangwa graben (Fig. 11.1).

11.2.2 Other igneous rocks

It is interesting to note the occurrence of three other provinces of post-Karoo intrusions spatially associated with the Luangwa graben. Two of the provinces (Panela and North Luangwa) occur within 150 km of Kapamba (E and NE respectively) while the Isoka group were found some 300 km away in the northern part of the Luangwa Valley (Hawkes 1974; Thieme & Johnson 1975; Hawthorne 1975). It is important to note, however, that all the North Luangwa bodies occur within the graben but far (50 km to NE and SW) from the main faults, the Isoka province straddles the graben boundary, while the Panela province falls just outside the graben. None of these bodies have been studied in detail but they are all considered to be true kimberlites by the authors. Early Proterozoic to Late Palaeozoic alkaline and carbonatitic intrusions also occur in the Luangwa Valley (Bailey 1961; Snelling *et al* 1964).

11.2.3 Kapamba province

The Kapamba occurrences comprise eleven single pipes and three groups of small, pipe-like bodies (referred to as P1 to P14) which together with a suite of dikes (Fig. 11.1) occupy a NW–SE trending zone approximately 25 km long. This direction is common amongst post-Karoo faults in the mid-Luangwa Valley. Most of the pipes, particularly P1, P2, P3, and the P4 and P5 groups, form distinct, positive, topographic features; in contrast, P7 is marked by an oval shaped depression. Most of the pipes are approximately circular in shape and range in size up to approximately 45 hectares (P1). The country rock sediments are often silicified along the contacts and sometimes form a prominent rim to the pipe. The geology of the pipe is variable and often complex in detail.

11.2.4 Age

The Kapamba intrusions cut through subhorizontal, Upper Karoo sediments (mudstones, siltstones, sandstones and grits) which lie immediately to the east of, and overlie, the Precambrian metasedimentary basement rocks. On the basis of stratigraphic relationships, the intrusions are younger than about 250 Ma in age. Radiometric age dating of samples from the intrusions has been attempted, but the lack of suitable samples has resulted in poorly constrained data. Model age calculations based on Rb-Sr mica analyses by Craig B. Smith (unpubl. data) suggest an emplacement age of 220 Ma (sample 195/53/PK12/1). $^{40}Ar/^{39}Ar$ age determinations have also been attempted on mica by D. Phillips (unpubl. data). Stepheating results suggest a maximum plateau age of about 255 Ma and a minimum age of about 160 Ma with excess argon being present. Considering this information as a whole, an age of about 220 Ma is plausible but further information is required for verification.

11.2.5 Volcanoclastic rocks

All the pipes are composed predominantly of volcanoclastic material. The clasts include dark coloured, juvenile lapilli as well as abundant grains and some rounded pebbles of quartz. Bombs are not common. Xenoliths include angular metamorphic basement rocks. The rocks are mostly lapilli tuffs which are broadly similar, but differ with respect to (i) fragment or grain size, (ii) variation in proportion of the dark coloured juvenile lapilli (from approximately 70 to <1 modal %) to the leucocratic xenolithic material, (iii) the absence or nature of the bedding, (iv) variation in olivine content, (v) the degree of alteration, primarily of the juvenile lapilli, and (vi) megascopic and macroscopic colour (mainly red, blue, green). Other finer-grained rocks, which are thought to be volcanoclastic in origin, are classified as crystal tuffs or coarse (ash) tuffs. Some fine (ash) tuffs may also be present. It is difficult to distinguish tuffs which contain no discernible igneous material (xenocrystic tuffs) from the country rocks because the main constituents of the former are derived from the latter and both may be bedded.

Structures within the tuffs include graded bedding, suggestive of pyroclastic flow and/or fallout deposits, as well as cross-bedding, which

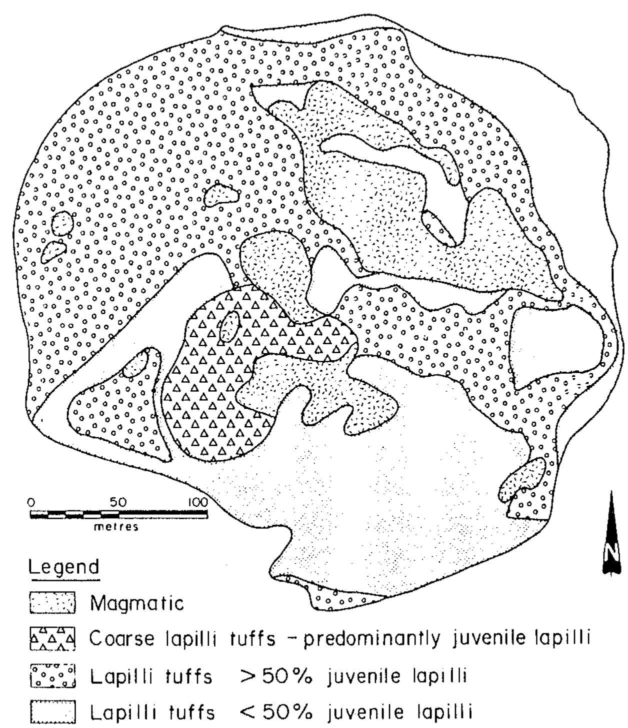

Fig. 11.2 Simplified geological map of Kapamba P10 lamproite.

is typical of pyroclastic surge deposits. Such features, however, could also result from the reworking of pyroclastic material. It is not possible to distinguish between these without further, detailed volcanological studies. Inward dipping bedded tuffs are well displayed at P10. If the bedding is of pyroclastic origin, those deposits may have formed by numerous pulses of pyroclastic activity. Some rocks in which the lapilli show signs of coalescing and contain only minor xenocrystic material are interpreted as being slightly welded tuffs. These tuffs may occur close to a vent and be lava spatter type deposits.

11.2.6 Magmatic rocks

Magmatic rocks (with some autolithic breccias) occur in some of the pipes (P4, P6, P10, P11 and P12). Outcrops of a significant size (up to 200 m, Fig. 11.2) occur and some variation in the nature of the rocks may suggest multiple intrusion.

11.2.7 Pipe geology

The different rock types may display a concentric distribution, albeit sometimes imperfect (Fig. 11.2). The marginal tuffs commonly contain abundant accidental or xenocrystic material (quartz) whereas the more central, later tuffs contain much less or only rare accidental material (e.g. P3, P7). At P10 (and possibly some of the other pipes) the magmatic rocks occur towards the centre of the pipe (Fig. 11.2). Concentric structures are also displayed by variations in colour which often appear unrelated to any primary feature. The marginal rocks are generally red in colour while the more central rocks are typically blue/green. Steeply dipping tuffs and near vertical bedding in the country rock along the margin of some of the pipes may be indicative of crater collapse. The geology of these pipes, particularly P10 (Fig. 11.2), is similar to that of many of the lamproites from the West Kimberley, Western Australia (e.g. Ellendale 4 and 9, Atkinson *et al* 1984 and Seltrust 2, unpublished work by the authors). Similar geological features also occur in the lamproitic intrusions at Prairie Creek, Arkansas (Scott Smith & Skinner 1984a), Smoky Butte, Montana (Bergman, 1987) and south-east Spain (unpubl. data) but are also typical of many small, alkaline volcanic intrusions.

11.2.8 Dikes

The Kapamba dikes are generally highly altered but can be followed using changes in the vegetation as well as the occasional exposure. The country rocks adjacent to the dikes may be indurated up to 1–2 m from the contact. This effect is most strongly developed in the shales. The relationship between the dikes and the pipes could not be determined.

11.3 PETROGRAPHY

11.3.1 Volcanoclastic rocks

The clasts (typically less than 2 cm) comprise juvenile lapilli, single grains of quartz and feldspar and a few country rock xenoliths. The juvenile lapilli are often altered. They have porphyritic textures with olivines set in a fine-grained or glassy groundmass. The olivines comprise abundant microphenocrysts (with many less than 0.25 mm but ranging up to 0.5 mm in size) which are euhedral but typically form complex, multiple growth aggregates. A few larger euhedral grains also occur and are referred to as phenocrysts. Larger grains (up to 8 mm) which may be anhedral and rounded are termed macrocrysts (Clement *et al* 1984). They may show some euhedralism resulting in serrate crystal margins. A few olivine macrocrysts are usually present but their abundance varies. Olivine macrocrysts may range from abundant to rare in different pipes (e.g. P2 and P10 respectively) or within some intrusions (e.g. P1).

The nature of the groundmass of the lapilli varies between pipes and even within a single thin section. Lapilli in some pipes (e.g. P1, P5) contain numerous altered laths of phlogopite (typically 0.1 mm) while in other pipes (e.g. P10) they contain phenocrysts of leucite and clinopyroxene. The groundmass of some lapilli may contain only fine-grained clinopyroxene and leucite (e.g. P7, P8). Fine-grained phlogopite and some opaque grains may also be present. The base to most lapilli is generally glassy and in some cases vesicular. Most of the lapilli are best termed glassy, olivine lamproite although some contain abundant phlogopite, leucite or clinopyroxene (Table 11.1).

The angular grains of quartz and lesser amounts of feldspar (up to 10 mm) are xenocrysts

TABLE 11.1 Modal analyses of magmatic and pyroclastic rocks from the Kapamba lamproites.

Sample type	Magmatic									Lapilli tuffs single lapilli				
	seg.	unif.	unif.	unif.	unif.	unif.	seg.	unif.	unif.					
Sample no.	P4/11	P6/1	P6/2	P6/3	P10/2	P10/3	P10A/6	P10B/1	P12/1	P1/13	P2/18	P5/1	P7/2	P10/6
Olivine	25	14	12	17	18	18	14	7	14	26	61	41	34	18
Phlogopite	27	12	8	7	1	13	16	32	28	14		14		
Clinopyroxene	19	26	31	24	26	43	36	35	30		<1		2	24
Leucite	11	36	37	39	31	17	12	19	10	2	3		9	40
Sanidine	11	2	4	4	3	6	13	6	11					
Amphibole	3	2	<1	<1	<1		5							
Glass/base	1		2		16	<1				58	30	45	54	18
Oxides	<1	6	4	8	4	3	2	1	6		4			
Perovskite		1	<1	1	<1	<1								
Carbonate	1						<1		<1					
Xenocrysts	1						2		<1		1		1	

Notes:
unif., uniform; seg., segregationary texture in groundmass. Results may not be accurate for magmatic rocks with fine-grained and/or segregationary textures and for lapilli in the tuffs.

derived by the fragmentation or disaggregation of the country rock sediments. The occasional xenocryst also occurs within the juvenile lapilli. The fine inter-lapilli matrix is generally composed of indiscernible, clay-like material which probably comprises finely comminuted volcanic constituents.

Other less common types of crater-facies rocks are also present. Some (e.g. P10) are composed predominantly of fine-grained quartz and feldspar with some isotropic material which appears to be volcanic glass. They are often well bedded and are classified as crystal or coarse (ash) tuffs. Other rocks (e.g. P10, Fig. 11.2), composed almost exclusively of juvenile lapilli, are slightly welded tuffs.

The glassy and vesicular nature of the juvenile lapilli suggests that most of the volcanoclastic rocks are pyroclastic lapilli tuffs. Fisher and Schmincke (1984, p. 89) suggest that glassy constituents would not survive epiclastic processes.

11.3.2 Magmatic rocks

The magmatic rocks (Table 11.1) which occur within the pipe-like intrusions have macrocrystic or porphyritic textures, with olivine set in a finer grained groundmass. The olivine is similar to that described from the juvenile lapilli but is usually less altered. Most of the olivines, termed macrocrysts, are considered to be xenocrysts because of the occurrence of polycrystalline grains (i.e. microxenoliths), common anhedral shapes and strong undulose extinction. Some subhedral shapes and serrate margins are thought to be imposed morphology resulting from corrosion or reaction. The small (<1 mm) euhedral grains, which often form complex multiple growth aggregates, are considered to be microphenocrysts. Larger (>1 mm) subhedral to euhedral grains (termed 'phenocrysts') are difficult to interpret as they may represent either true phenocrysts, or xenocrysts with imposed morphology. Undulose extinction cannot easily be used to discriminate between them as weak undulose extinction is noted in some of the microphenocrysts.

Two main varieties of groundmass occur. The first type (e.g. P4/11, Table 11.1) is fine-grained (typically < 0.2 mm) and usually displays a distinct segregationary texture with phlogopite-rich and clinopyroxene-rich areas. The phlogopite occurs as tiny plates often concentrated about olivine crystals. These areas have an orange-brown colour. The clinopyroxene occurs as a felt of small needles imparting an overall grey colour. These rocks, which are commonly glassy, also contain leucite and amphibole (both usually most abundant in the clinopyroxene-rich areas), spinel and apatite.

The second type of groundmass is medium-grained with uniform textures and is composed of leucite, clinopyroxene, phlogopite, amphibole, sanidine, opaque minerals, perovskite, apatite and glass. Leucite occurs as euhedral, equant grains which are commonly near isotropic or, in other instances, have been replaced by a polycrystalline mosaic of sanidine. Secondary alteration is evident in some partly turbid, brown coloured grains.

Leucite typically occurs as groundmass grains (<0.2 mm) but may also occur as phenocrysts up to 0.5 mm in size (e.g. P10/6). Phlogopite occurs as small laths (0.3 mm) or, more commonly, as small (<0.2 mm), interstitial grains often poikilitically enclosing other groundmass minerals. It is pleochroic from a pale to darker distinctive orange-brown colour. Clinopyroxene is present as slender or stubby laths which may be optically zoned. It occurs as phenocrysts (0.5–0.7 mm), microphenocrysts and groundmass (<0.15 mm) grains. The amphibole forms small (<0.15 mm) interstitial grains which display the distinctive pink to yellow pleochroism typical of titanian potassic richterite. Sanidine is a late-stage interstitial mineral. Spinel occurs as small (<0.1 mm) euhedral to anhedral grains. Euhedral to anhedral pinkish grains of perovskite are similar in size to the spinel. Rare autolithic breccias (P10B) have cognate inclusions of earlier crystallized magmatic material in a magmatic host. A few xenocrysts of quartz and/or feldspar are present.

These magmatic rocks are classified as lamproites (Scott Smith & Skinner 1984a, b; Mitchell 1985). It can be seen from Table 11.1 that there is considerable modal variation. The sample from P4 should be termed a clinopyroxene-olivine-phlogopite lamproite. The rocks from P6 are clinopyroxene-leucite lamproites. Those from P10 are clinopyroxene-leucite or clinopyroxene lamproites. The samples examined from P10B and P12 can be classed as phlogopite-leucite lamproites. It is important to note that all the phlogopite in the magmatic rocks occurs as goundmass and that no true phenocrysts were observed. Mitchell (1985) suggests that such rocks should be termed madupitic (as opposed to phlogopite) lamproite. The magmatic rocks which occur within the pipes have relatively fine-grained or glassy groundmasses. This suggests that they cooled relatively quickly and are probably extrusive.

The dike rocks are altered but primary constituents include olivine, clinopyroxene, leucite, phlogopite and sanidine. Olivine macrocrysts are rare. The leucite often displays a glomeroporphyritic texture while the sanidine, although interstitial, may form subhedral rectangular plates.

11.4 MINERAL CHEMISTRY

This part of the investigation was confined to the magmatic rocks. No samples from the lapilli tuffs could be included because of alteration and/or fine grain size which, for some magmatic samples, also precluded the analysis of a full suite of minerals. Representative mineral analyses are given in Tables 11.2 and 11.3.

TABLE 11.2 Selected mineral analyses from the Kapamba lamproites.

Mineral	Phlogopite						Amphibole		Feldspar	Pseud	Perovskite	
Sample no.	P12/1	D3/2	P6/2	P6/4	P6/4	P10/3	P4/10	P10A/6	P12/1	P10/2	P6/4	P10/2
Anal no.	84-366	84-347	85-108	84-376	84-377	84-393	81-750	85-096	84-358	85-075	84-386	85-079
SiO_2	38.47	38.57	38.15	41.47	42.11	36.51	52.44	50.81	64.62	64.02	nd	nd
TiO_2	9.09	8.35	7.83	7.44	7.01	6.46	4.08	4.70	0.11	nd	57.10	56.97
Al_2O_3	10.71	9.59	9.09	7.39	6.35	4.88	1.06	1.62	17.17	18.14	nd	0.07
Cr_2O_3	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	0.02	nd
FeO	11.30	12.12	9.46	10.73	11.40	33.12	4.78	8.02	1.50	0.31	0.42	0.74
MnO	0.10	0.07	0.07	0.06	0.05	0.28	0.09	0.13	0.01	nd	nd	nd
NiO	0.07	0.03	0.04	0.06	0.04	0.02	0.06	0.02	0.01	nd	0.02	nd
MgO	16.05	16.48	18.85	17.74	17.48	4.98	20.32	17.19	nd	nd	nd	0.02
CaO	0.09	0.02	0.09	0.03	0.03	0.09	6.78	5.77	0.17	nd	32.67	38.28
Na_2O	0.56	0.45	0.83	0.74	1.06	0.52	5.33	5.74	2.68	0.07	2.01	0.77
K_2O	8.88	9.26	8.73	9.34	9.52	9.18	2.91	3.02	12.69	16.44	0.05	0.09
BaO	1.25	1.04	2.80	1.56	1.20	0.39	0.03	0.06	0.33	0.16	0.22	0.13
SrO	nd	nd	nd	nd	nd	nd	nd	0.06	nd	nd	3.88	0.97
ZrO	nd	nd	nd	nd	nd	nd	0.03	nd	nd	nd	0.05	0.02
P_2O_5	nd	nd	nd	nd	nd	nd	0.09	0.10	0.12	nd	0.04	0.06
F	1.25	1.78	3.89	2.51	2.73	0.38	1.57	0.87	nd	nd	0.51	0.10
Total	97.29	97.01	98.19	98.04	97.83	96.65	98.91	97.74	99.41	99.14	96.78	98.18
Mg/(Mg+Fe)	0.717	0.708	0.780	0.747	0.732	0.211	0.883	0.793				

Notes:
Pseud — leucite pseudomorphs. nd, not detected. Totals have been adjusted for O=F. 84-366 — centre of 0.4 mm equant grain, defocused beam; 84-347 — 0.31 mm centre of subhedral lath; 85-108 — centre of 0.39 mm poikilitic grain; 84-376 — centre of 0.07 mm equant, euhedral plate; 84-377 — 15 μm from edge of 84-376; 84-393 — 10 μm from edge of groundmass grain with darker rim of tetraferriphlogopite; 81-750 — small interstitial grain; 85-096 — centre of 0.14 mm subhedral grain, defocussed beam; 84-358 — subhedral, interstitial groundmass grain; 85-075 — 0.1 mm clear grain; 84-386 — 20 μm grain; 85-079 — edge of subhedral grain in groundmass pool.

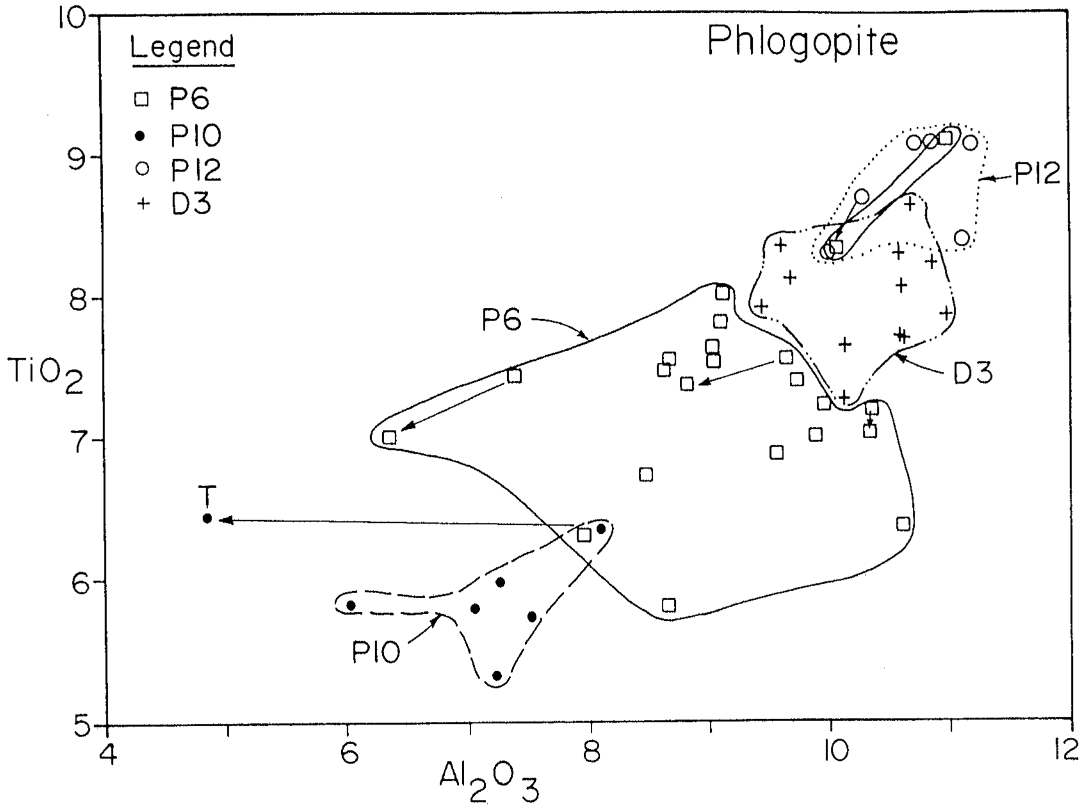

Fig. 11.3 TiO_2 versus Al_2O_3 (wt%) for phlogopites from some of the Kapamba intrusions. (Arrows indicate zoning; T = tetraferriphlogopite.)

11.4.1 Phlogopite

The groundmass or madupitic micas at Kapamba (Table 11.2, Figs 11.3 and 11.4) are similar in composition to phlogopites from other lamproites (e.g. Jaques *et al* 1984; Scott Smith & Skinner 1984a; Mitchell 1985; Wagner & Velde 1986), being titaniferous (5–9 wt% TiO_2) and relatively poor in alumina (4–11.5 wt% Al_2O_3). The Na_2O contents (0.3–1.3 wt%) are also similar to those from other lamproites. Significant amounts of fluorine (1–5 wt%) were detected but high values have been reported elsewhere (Carmichael 1967; Scott Smith & Skinner 1984a). The Kapamba phlogopites have no detectable Cr_2O_3 which is consistent with their late-stage crystallization. FeO contents (7–14 wt%), particularly for P12 and D3, fall at the higher end of the lamproite range. BaO (0.9–3.2 wt%) contents of the Kapamba phlogopites are higher than many lamproites but those in groundmass or madupitic micas from some other lamproites range from 1.6 to 2 wt% (Leucite Hills, south-east Spain) and from phenocrystal mica (up to 3 wt%) at Smoky Butte, Montana (Mitchell 1985; Wagner and Velde 1986).

Micas in the magmatic pipe rocks are small and poikilitic making analysis of zoning difficult but the data obtained for the pipe rocks suggest zoning to lower TiO_2 and Al_2O_3 (e.g. analyses 84-376 and 84-377, Table 11.2, Fig. 11.3). Zoning data obtained for coarser grained mica from D3 is not consistent and is not plotted in Figs 11.3 and 11.4. The phlogopite differs in composition in each of the intrusions analysed (Figs 11.3 and 11.4). This variation may indicate the relative degree of evolution of each intrusion. Using the characteristic lamproitic trends of depletion of Al_2O_3, which is also reflected in the zoning at Kapamba, the data (Fig. 11.3) suggests that P12 is the least evolved while P10 is the most evolved of the pipes examined (P12 $<$ P6 $<$ P10). The micas from D3 are different in mode of occurrence but are most similar in composition to P12.

11.4.2 Olivine

The Mg/(Mg+Fe) atomic ratios of all the olivines analysed (Table 11.3) fall in the range 0.822–0.934 and are similar to those from other lamproites, although these compositions are not

TABLE 11.3 Selected mineral analyses from the Kapamba lamproites.

Mineral	Olivine								Clinopyroxene						
Sample no.	P4/10	P4/10	P4/10	P4/10	P6/2	P6/2	P6/2	P6/2	P4/10	P6/1	P6/1	P10/2	P10/2	P12/1	P2/C
Anal no.	81-576	81-578	81-589	81-590	81-617	81-618	85-285	85-286	81-753	84-441	84-442	85-149	85-150	84-424	106-A
SiO_2	40.61	40.32	40.67	40.26	40.45	40.08	40.29	39.97	52.45	51.85	52.16	53.04	51.88	53.10	54.21
TiO_2	nd	0.01	nd	0.02	nd	0.01	nd	0.02	0.68	1.17	1.48	1.08	1.81	0.90	0.19
Al_2O_3	0.02	0.03	0.02	0.02	0.04	0.03	0.02	0.02	1.12	1.27	0.59	0.85	0.78	1.00	2.42
Cr_2O_3	nd	nd	0.03	nd	nd	0.01	0.01	nd	0.93	0.92	nd	0.42	nd	0.65	1.60
FeO	7.76	8.88	8.70	9.00	7.62	10.31	10.36	11.31	2.43	3.51	4.58	3.50	5.05	3.36	2.60
MnO	0.12	0.16	0.14	0.15	0.10	0.15	0.11	0.19	0.05	0.06	0.08	0.07	0.09	0.08	0.11
NiO	0.39	0.31	0.38	0.34	0.37	0.26	0.37	0.26	0.04	0.03	nd	0.02	nd	0.04	
MgO	51.71	50.27	50.39	50.02	51.13	48.68	48.94	47.81	17.33	16.52	16.06	16.77	15.56	17.08	17.38
CaO	0.04	0.30	0.16	0.31	0.02	0.26	0.15	0.36	24.10	23.40	23.61	24.10	24.17	23.53	19.21
Na_2O									0.48	0.44	0.61	0.30	0.37	0.33	1.45
K_2O									nd	0.01	0.01	nd	0.02	nd	
Total	100.65	100.28	100.49	100.12	99.73	99.79	100.25	99.94	99.61	99.18	99.18	100.15	99.73	100.07	99.16
Mg/(Mg+Fe)	0.922	0.910	0.912	0.908	0.923	0.894	0.894	0.883	0.927	0.893	0.862	0.895	0.846	0.901	0.923

Notes:
81-576 — centre of 3.5 mm polycrystalline xenocryst or microxenolith; 81-578 — edge of 81-576; 81-589 — 0.47 mm centre of euhedral phenocryst; 81-590 — edge of 81-589; 81-617 — centre of 2 mm macrocryst; 81-618 — edge of 81-617; 85-285 — centre of 0.31 mm subhedral microphenocryst; 85-286 — edge of 85-285; 81-753 — centre of 0.7 mm phenocryst; 84-441 — centre of 0.22 mm euhedral microphenocryst; 84-442 — edge of 84-441; 85-149 — centre of 0.3 mm euhedral microphenocryst; 85-150 — edge of 85-149; 84-424 — 0.7 mm subhedral groundmass grain; 106-A — 0.3 mm concentrate grain; 84-485 — 0.05 mm subhedral grain, FeO = 36.44, Fe_2O_3 = 28.53, Total = 100.08; 84-249 — 0.05 mm subhedral grain, FeO = 26.60, Fe_2O_3 = 13.74, Total = 98.99; 85-250 — edge of 85-249, FeO = 38.09, Fe_2O_3 = 38.09, total = 98.05; 85-208 – 0.08 mm subhedral grain, FeO = 47.69, Fe_2O_3 = 21.10, Total = 97.56; 84-487 – 0.1 mm euhedral grain, FeO = 40.84, Fe_2O_3 = 20.67, Total = 94.89; 106-B – 0.3 mm concentrate grain, FeO = 14.76, Fe_2O_3 = 10.69, Total = 100.56; 85-261 – centre subhedral grain; 106-C – 0.15 mm concentrated grain. nd = not detected.

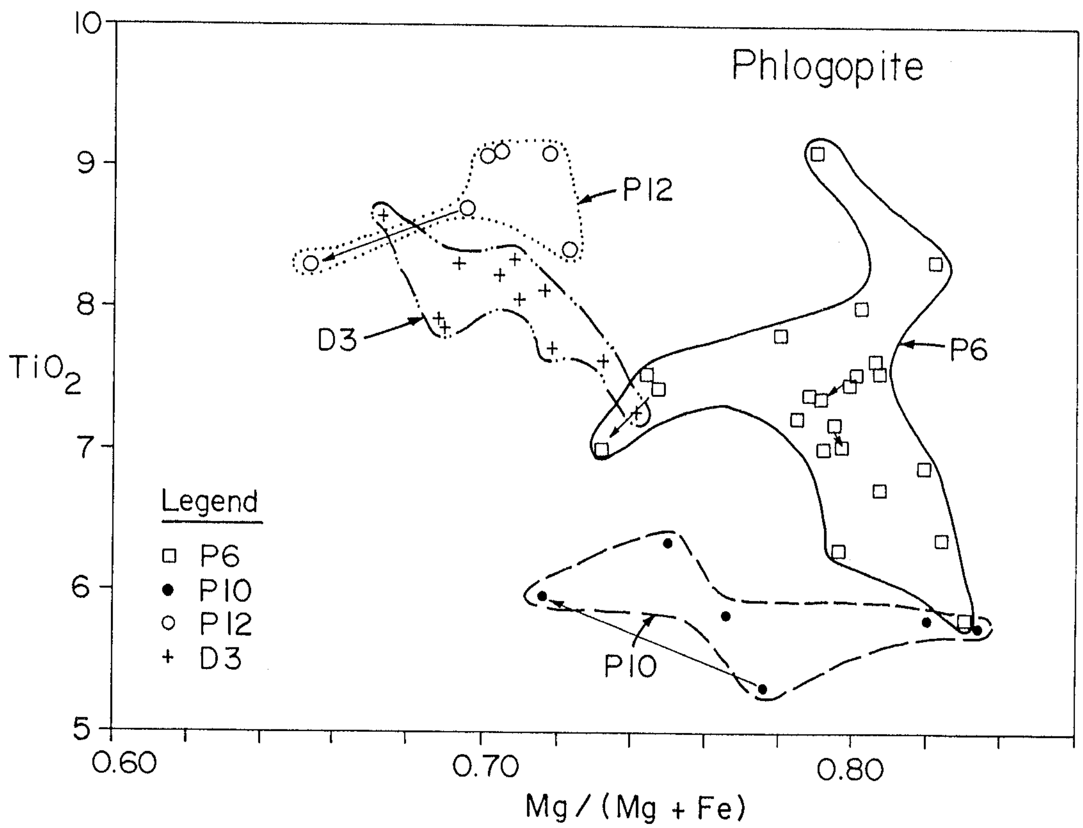

Fig. 11.4 TiO_2 wt% versus Mg/(Mg+Fe) atomic ratios for phlogopites from some of the Kapamba intrusions. (Arrows indicate zoning.)

Spinel						Ilmenite	Garnet
P6/1 84-485	P6/3 85-249	P6/3 85-250	P6/4 85-208	P10/2 84-487	P2/16 106-B	P6/2 85-261	P2/20 106-C
nd	nd	nd	nd	0.17	nd	nd	42.24
16.08	3.90	16.30	23.96	22.32	2.18	49.74	0.08
0.64	5.13	0.27	0.85	1.01	12.88	0.02	20.97
11.47	43.29	10.90	0.11	0.10	46.06	0.15	3.61
62.11	38.97	62.45	66.68	59.43	24.37	43.34	7.87
0.69	0.58	0.70	1.40	9.08	0.17	1.02	0.34
	0.06	0.16	0.11			0.05	
5.78	5.60	4.39	2.26	0.25	13.82	2.37	20.01
0.15	0.08	0.17	0.08	0.10	nd	0.17	5.30
0.30				0.36			
97.22	97.61	95.34	95.45	92.82	99.48	96.86	100.42

distinctive. The compositions of the microphenocryst cores are different from each of the intrusions analysed. Those from P4 have Mg/(Mg+Fe) ratios from 0.900 to 0.915, and those from P6 are 0.885 to 0.905; the few data from P10 fall in the range 0.880–0.890 (Fig. 11.5). This would suggest that, of the three intrusions examined, P4 is the least evolved and P10 the most evolved (P4 < P6 < P10). The cores of most of the xenocrysts (macrocrysts) fall in the range 0.905 to 0.925, although some lower values were encountered. The compositions of the cores of the olivines termed 'phenocrysts' (see Petrography) overlap the data from both the xenocrysts and microphenocrysts, indicating that both true phenocrysts and xenocrysts are represented in this population.

The rims (generally <100 μm) of all the olivines from P4 (0.905–0.915) and P10 (0.880–0.890) show very restricted compositions and may represent late-stage overgrowths or equilibration. Two of the four samples analysed from P6 contain olivines which also have rims with restricted compositions (0.885–0.895) while, in the other two samples, the rims display a wider range in composition. The compositions of the rims suggest that P4 may be the least evolved of these intrusions (P4 < P10 and P6). The olivines are zoned with respect to minor elements with decreasing NiO (0.56–0.16 wt%) and increasing CaO (0.09–0.46 wt%) from core to rim.

11.4.3 Clinopyroxene

Many of the groundmass clinopyroxene laths are too small or too narrow to analyse. The larger (typically > 0.05 mm) grains are usually optically zoned. Those analysed are diopsides (Table 11.3, Fig. 11.6) with variable TiO_2 (0.4–2.7 wt%), Al_2O_3 (0.4–2.2 wt%), Cr_2O_3 (0–1.1 wt%), FeO (2.4–6 wt%), MgO (17.7–14.9 wt%) and Na_2O (0.2–0.9 wt%) contents. The zoning is not consistent, even within one sample, but generally the rims have higher TiO_2, FeO and, to a lesser extent, Na_2O contents.

The high TiO_2 contents of these diopsides are similar to those from other lamproites. The higher Fe values, mostly representing small groundmass grains or rims, from P6 and P10, fall outside the main lamproite field (Fig. 11.6). Jaques *et al* (1984) note that some clinopyroxenes from the Western Australia lamproites are zoned to high FeO and Venturelli *et al* (1984) note clinopyroxenes from south-east Spain show slight Fe-enrichment and give some Fe-rich compositions. In these instances, however, the CaO contents are much lower than those at Kapamba. Many of the Kapamba clinopyroxenes also have higher Al_2O_3 contents (up to 2.2 wt%) than other lamproites (typically <1 wt%, Barton 1979; Kuehner *et al* 1981; Venturelli *et al* 1984; Mitchell 1985) but they are not as high as those found in other potassic rocks (e.g. up to 7 wt% Al_2O_3 from Vico, Italy and Toro-Ankole, Uganda, Barton 1979). This feature is illustrated in terms of atomic Al in Fig. 11.7 (after Mitchell 1985). Clinopyroxenes from many other potassic rocks plot off Fig. 11.7 with higher Al values.

The compositions of the clinopyroxenes from each of the intrusions are somewhat different (Fig. 11.6). Those from P4 show a range in Ca values compared with the other intrusions while those from P6 and P10 have variable Fe values relating to late-stage Fe-enrichment in rims or in small groundmass grains. These data may suggest that P4 may be the least evolved of these intrusions (P4 < P12 < P10+P6).

Rare chrome diopsides occur in heavy mineral concentrates from some pipes. A representative analysis is given in Table 11.3.

11.4.4 Amphibole

The amphiboles are titanian, potassic richterites (Table 11.2; 4–6.3 wt% TiO_2, 4.8–9.5 wt% FeO, 2.7–3.2 wt% K_2O) and are similar to those from other lamproites (Scott Smith & Skinner 1984a; Jaques *et al* 1984; Venturelli *et al* 1984; Mitchell

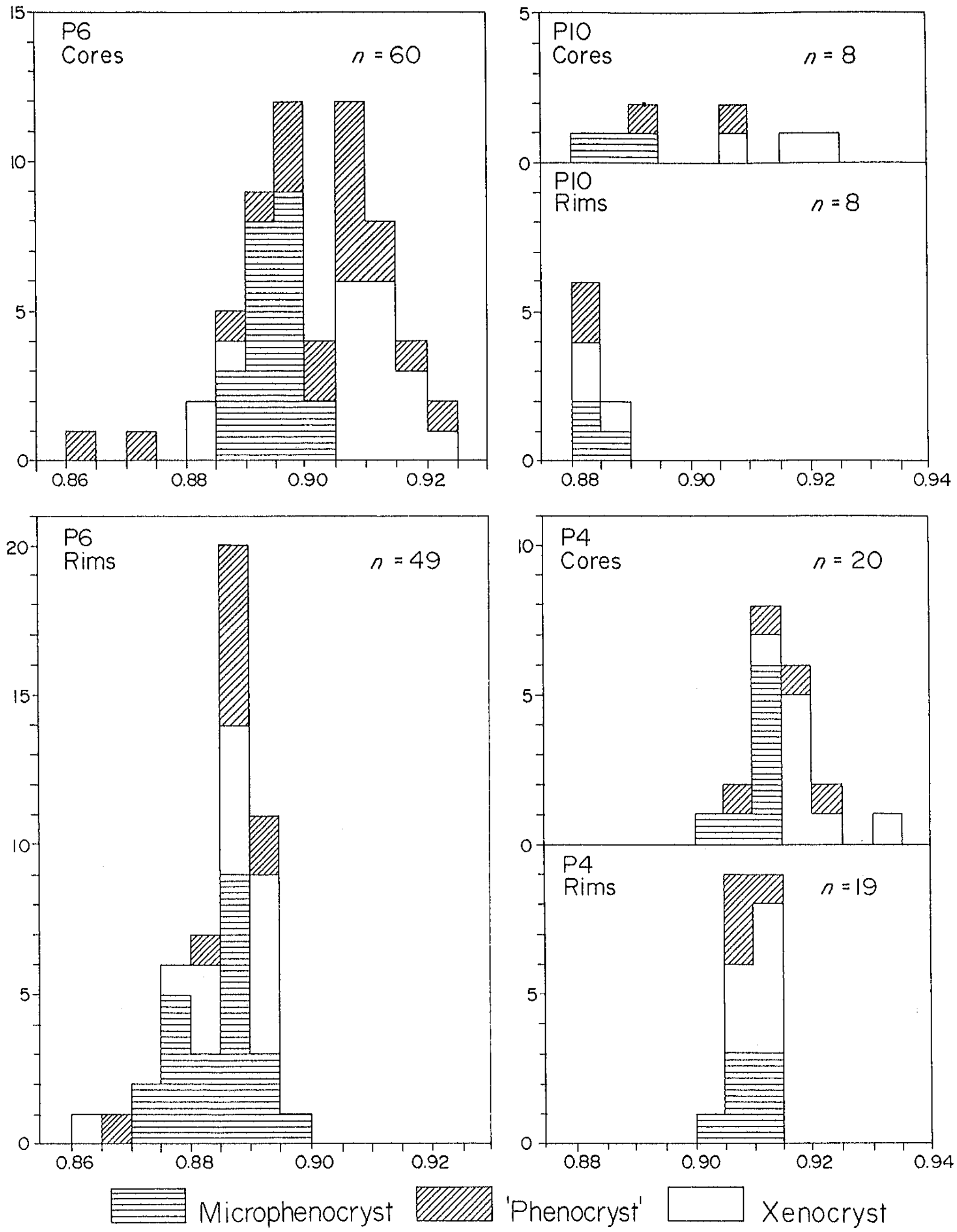

Fig. 11.5 Frequency distribution diagrams of Mg/(Mg+Fe) atomic ratios for olivines from some of the Kapamba lamproites.

1985; Wagner & Velde 1986) while amphiboles similar in composition from other rocks are rare. With relatively high Na_2O contents (5.2-6 wt%) and high Na/(Na+K) ratios (0.73-0.75) the Kapamba amphiboles are most similar to those from south-east Spain. Ba contents are low or below detection. Fluorine contents are moderate (0.9-1.6 wt%). The Kapamba amphiboles, however, have higher Al_2O_3 contents (0.5-2, average 1.5 wt%) than those typical of lamproites (mostly < 1 wt%) and in this respect compare with the rare occurrences of similar amphiboles in other ultrapotassic rocks (e.g. Shiprock minette, New Mexico, Wagner & Velde 1986; New South Wales, Cundari 1973).

11.4.5 Spinel

Most of the groundmass spinels are titanomagnetites, but some grains have chromite cores (Table 11.3, Fig. 11.8). Evolutionary trends are increasing TiO_2, FeO, MnO and decreasing Cr_2O_3, Al_2O_3 and MgO. The Kapamba spinels are

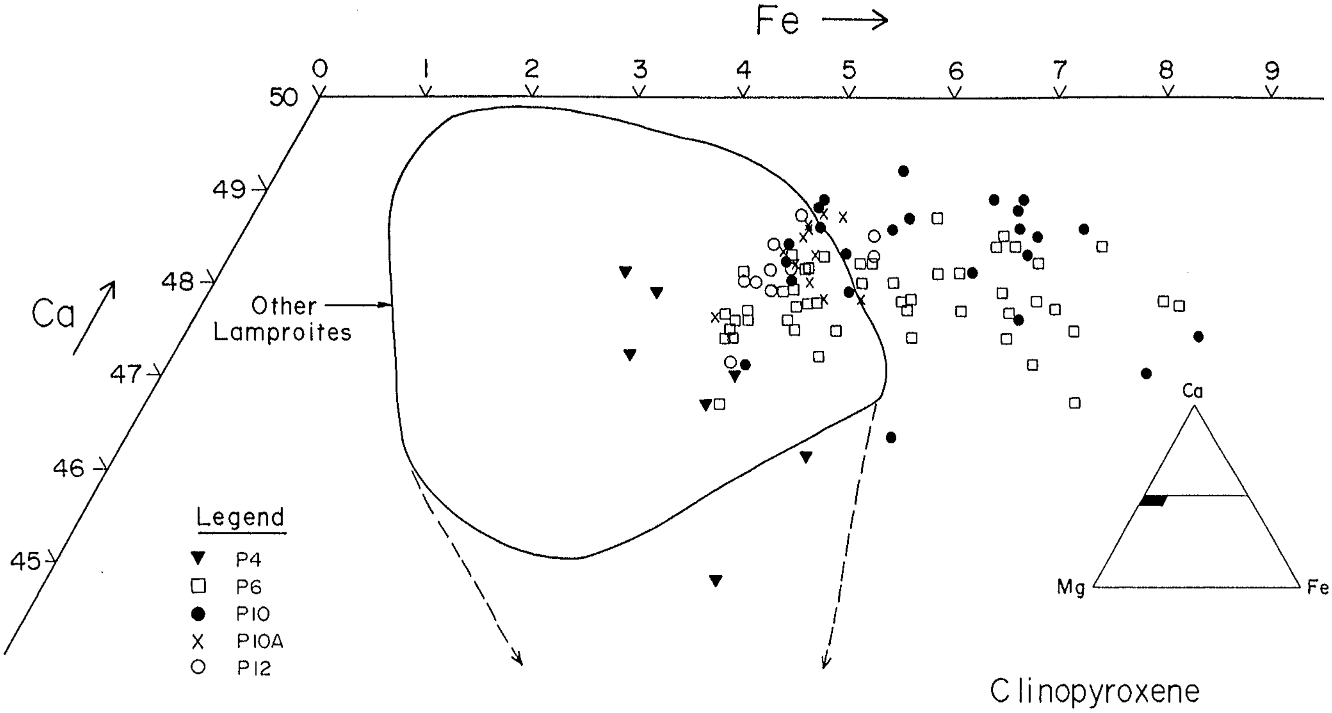

Fig. 11.6 Ca-Mg-Fe atomic proportions plot of clinopyroxenes from some of the Kapamba lamproites. (Main lamproite field after Mitchell 1985; dotted line includes few additional data from Jaques *et al* 1984 and Venturelli *et al* 1984.)

generally similar to spinels in other lamproites, when present (Jaques *et al* 1984; Scott Smith & Skinner 1984a; Mitchell 1985). All the spinels analysed from P10/2 are rich in MnO (up to 10 wt%) but the totals are low (94–95) and the reason for this is not understood. Magnesian chromites (Table 11.3) occur in the heavy mineral concentrates of some pipes.

11.4.6 Leucite

No fresh leucite was encountered and all the grains analysed are composed of sanidine (analysis 85-075, Table 11.2) and/or sodium aluminosilicates. The latter are often browny, turbid and probably altered, and are difficult to analyse because they decompose under the electron beam. The analyses obtained suggest that this material is mostly analcite (e.g. 52.04 SiO_2, 23.58 Al_2O_3, 11.89 Na_2O, all wt%) but a few analyses (e.g. 46.75 SiO_2, 25.77 Al_2O_3, 15.53 Na_2O, all wt%) are similar to nepheline. These minerals may occur within one grain showing a similarity to pseudoleucite and it is considered that all these minerals pseudomorph primary leucite. Analcite occurs at Moon Canyon, Utah and Smoky Butte which have also been interpreted as replacing leucite (Mitchell 1985).

11.4.7 Feldspar

Primary sanidine occurs both as interstitial groundmass and as coarser grains in pool-like segregations. It typically has significant FeO (up to 2.1 wt%), Na_2O (up to 2.7 wt%) and BaO (0.3–1.4 wt%) contents (Table 11.2). The sanidine pseudomorphing leucite contains negligible Na_2O and BaO, while potassium feldspar xenocrysts contain no FeO. Although the Na_2O contents of the Kapamba sanidines may be slightly higher than found in many other lamproites, they are otherwise similar (e.g. Mitchell 1985; Wagner & Velde 1986), particularly to those from south-east Spain (Venturelli *et al* 1984).

11.4.8 Perovskite

Data for perovskites from lamproites are limited (Carmichael 1967), but the Kapamba perovskites are similar in terms of their relatively high Na_2O and SrO contents and their relatively low FeO contents (Table 11.2) compared with the perovskites from Prairie Creek (Scott Smith & Skinner 1984a). The perovskites from P6 and P10 are compositionally different, particularly with respect to Na_2O and SrO.

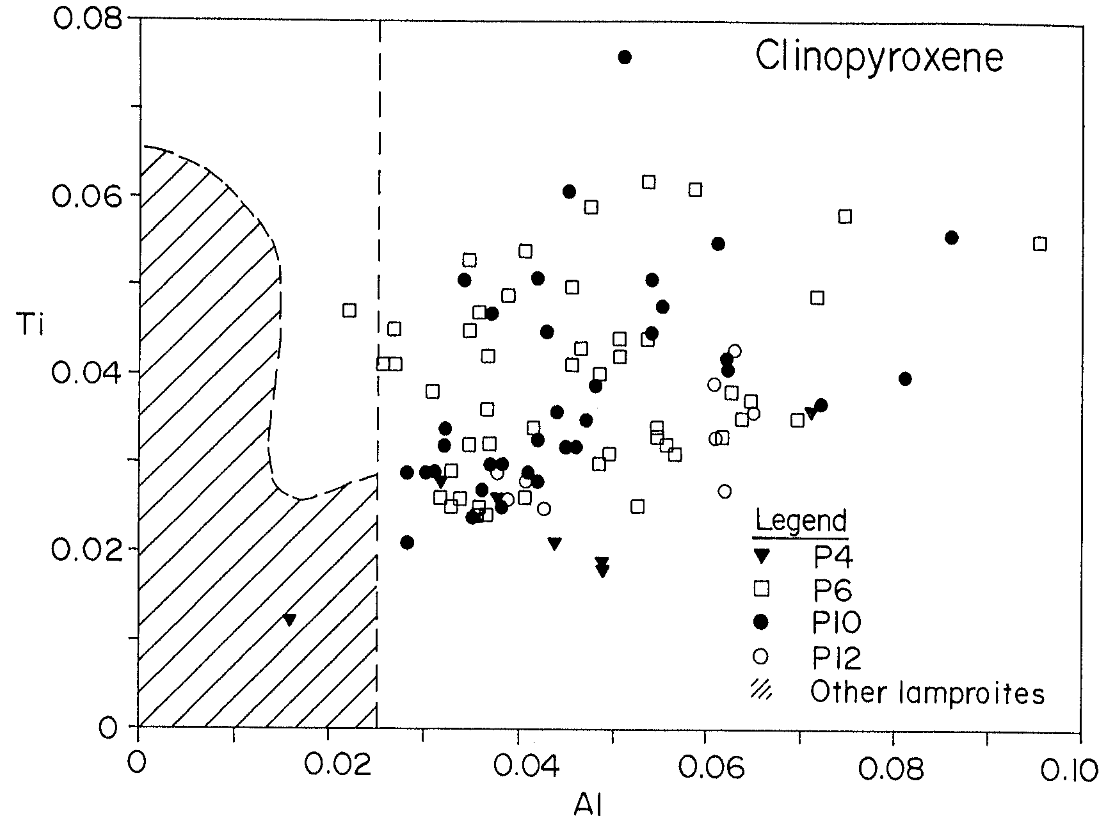

Fig. 11.7 Compositions (Ti versus Al) of clinopyroxenes from some of the Kapamba intrusions compared with other lamproite data (after Mitchell 1985).

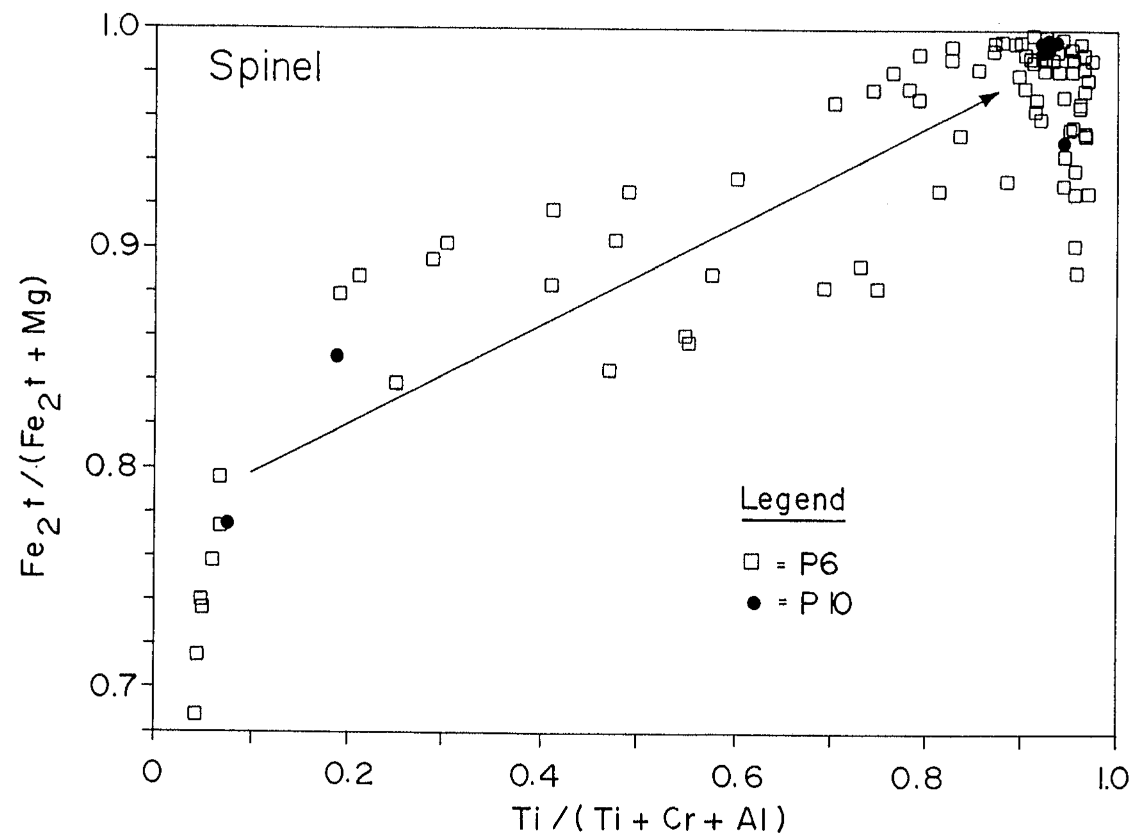

Fig. 11.8 Compositions of spinels from the Kapamba lamproites. (Arrow indicates general direction of zoning.)

11.4.9 Ilmenite

Ilmenite (Table 11.3) rarely occurs as anhedral to cubic grains (<0.1 mm) and it is not clear whether it is primary or xenocrystal. Ilmenite is not common in lamproites, but some with similar compositions to the Kapamba grains have been noted from south-east Spain (Mitchell 1985).

11.4.10 Glass

Fresh glass is not common at Kapamba but where analysed (sample P10/2) its composition resembles that of the phlogopite, except for lower MgO (10 wt%) and higher FeO (20–22 wt%) contents.

11.4.11 Garnet

Garnets were recovered from heavy mineral concentrates of some pipes (e.g. P1). The majority of the garnets (Table 11.3) have peridotitic compositions with over 75% being chrome pyropes (cluster Group 9 of Dawson & Stephens 1975, 1976). Smaller numbers of garnets falling into groups 1 and 3 (titanian pyropes, calcic pyrope-almandines respectively of Dawson and Stephens 1975, 1976) are also present. Almandine-rich garnets can be abundant.

11.5 WHOLE-ROCK GEOCHEMISTRY

The most notable feature of these whole-rock data (Table 11.4) are the variable alkali contents. Some samples have high K_2O contents (>5 wt%) and high K_2O/Na_2O ratios (3–5) while other samples have high Na_2O contents (>4 wt%) and low K_2O/Na_2O ratios (0.3–0.4). Primary mineralogy, in particular the lack of sodic minerals, cannot account for this variation. Microscopically (Table 11.1) and macroscopically visible, although minor, contamination by xenolithic material was noted for most of the samples analysed. The xenocrysts observed in thin section include quartz and feldspar. The few feldspar xenocrysts analysed are potassium feldspar with up to 2.2 wt% Na_2O. Assimilation of this material cannot account for the variation in alkali content, although other xenolithic material may have been completely digested into the magma. Also, of the samples analysed, those with significant modal amounts of xenocrystic material (samples P4/11 and P10A/6, Table 11.1) do not have the highest Na_2O contents. A feature which might explain this variation is the secondary replacement of leucite by analcite (and/or nepheline) but a systematic study of leucite or its pseudomorphs was not undertaken. Sodium aluminosilicates only replace leucite in some samples and their distribution may be patchy even within one thin section. The replacement does not appear to be related to primary, late-stage, magmatic enrichment. It seems more likely, therefore, that the leucite alteration, and hence high Na_2O values, have resulted from some other postconsolidation process, such as the interaction with groundwater, which Gupta and Fyfe (1975) have shown can readily occur in leucite.

The observed xenolithic contamination and possibility of secondary alteration shows that the whole-rock analyses may not be particularly representative of the magma. The relatively high SiO_2 and K_2O contents of samples P4/11 and P10/6 may reflect the visible contamination but it is not obviously reflected in the Al_2O_3 contents. Except for the high Na_2O values, most aspects of the Kapamba whole-rock compositions (Table 11.4), however, are similar to those from other lamproites (see review by Bergman, 1987) although the Ba and Zr contents are comparatively low. The data are most similar to the non-diamondiferous lamproites of Bergman (1987) or the (leucite) lamproites (as opposed to olivine lamproites) of Jaques *et al* (1984) although the TiO_2 contents are lower.

11.6 DIAMONDS

Approximately 150 diamonds from P2 and a small number of stones from P1, P3, P4, P8, P9 and P10 were examined briefly by D.N. Robinson (pers. comm.) and his unpublished results are given here. Most of the diamonds are smaller than 0.1 carat but larger examples, up to 0.5 carat (P2), were found. Yellow and brown stones occur in approximately equal proportions at P2, while brown stones are more common at the other pipes. The yellow colour is particularly deep in some cases and may verge on orange. The tetrahexahedroid (i.e. 'rounded dodecahedron') crystal form predominates but rare octahedra also occur. Tetrahexahedroid surfaces are often unusually smooth. Lamination lines, however, are developed on most of the larger, brown diamonds. Only black inclusions, probably of sulphide in most instances, were observed.

The Kapamba diamonds resemble other lamproitic diamond populations (Hall & Smith 1984) in the strong predominance of the tetrahexahedroid form and the lack of microrelief on tetrahexahedroid surfaces. These characteristics are not in any way different from those of kimberlitic diamond populations.

The depth of the yellow colour in some of the Kapamba diamonds suggests the possible presence of single, substitutional nitrogen. If this is the case, it would imply (cf. Evans & Qi 1982) either that some of the diamonds are not much older than the lamproite or that they are formed from particularly cool mantle. A xenocrystic origin, for at least some of the diamonds, is favoured by the presence of lamination lines which have been shown to reflect plastic deformation (Urusovskaya & Orlov 1964).

11.7 DISCUSSION

The geology and petrography of the Kapamba pipes are typical of lamproites. Many features of the mineral chemistry and whole-rock geochemistry are characteristic, or even diagnostic, of lamproites. A few features, however, are different but are not extreme enough to preclude their classification as lamproites. Such features include, for example, the high Al_2O_3 contents of the clinopyroxenes and amphiboles. Minerals from Kapamba, that have compositions which can be distinguished from other lamproites, begin to approach but seldom attain, the compositions of those found in other potassic rocks (e.g. south-west Uganda and leucitites of western Italy and New South Wales, e.g. Barton 1979; Mitchell 1985). The similarity with south-west Uganda is interesting because the Kapamba province occurs within, what is considered by some to be, a SW-extension of the East African Rift. The potassic rocks of south-west Uganda, however, differ from Kapamba and other lamproites, notably in the presence of nepheline, kalsilite and melilite, absence of amphibole and relative abundance of titanomagnetite.

Most of the compositional differences from other lamproites occur in late-stage minerals (phlogopite, amphibole and later crystallizing clinopyroxene). Xenocrysts, including quartz and potassium feldspar, were observed to be in reaction with the host magma. Late-stage contamination from the partial or total digestion of xenolithic material (processes similar to those discussed by Scott Smith *et al* 1983) could explain features such as the inconsistent zoning and high (relative to other lamproites) alumina contents of the clinopyroxenes. Some rocks may also have been affected by secondary processes, such as later interaction with groundwater, as suggested from whole-rock and mineral chemistry data. The differences in mineral compositions from other lamproites, however, may equally reflect a Kapamba 'signature', because each lamproite province is characterized by certain minerals having different compositions (Mitchell 1985).

Modal analyses (Table 11.1) show that there is considerable variation in the proportions of minerals present in rocks from Kapamba. Bearing in mind the problems of comparing glassy juvenile lapilli from the pyroclastic rocks with the more crystalline magmatic rocks, an overview of the mineralogy suggests that, in general, the pipe rocks become more evolved from the north-west

TABLE 11.4 Whole-rock compositions of some magmatic samples from the Kapamba lamproites. See Table 11.1 for modal analyses.

Sample no.	P4/11	P6/1	P6/3	P10/2	P10/3	P10A/6
SiO_2	51.12	46.06	47.07	45.51	42.63	50.70
TiO_2	1.71	2.58	2.59	2.83	2.44	2.20
Al_2O_3	7.69	9.51	9.18	9.31	7.72	8.70
Fe_2O_3	7.34	10.14	10.01	9.94	10.94	8.43
MnO	0.08	0.13	0.12	0.26	0.15	0.13
MgO	12.09	11.62	10.54	10.40	13.59	8.62
CaO	5.27	8.25	7.95	9.19	9.47	7.49
Na_2O	1.71	4.23	4.49	0.63	4.25	2.33
K_2O	5.69	1.76	1.90	4.98	1.39	5.60
P_2O_5	0.62	0.72	0.41	1.26	0.83	0.85
H_2O+	1.9	3.8	3.9	4.3	3.7	3.5
CO_2	2.4	0.4	0.7	0.4	0.3	0.3
Total	97.62	99.20	98.86	99.01	97.41	98.85
Sc	12	14	13	15	16	13
V	84	144	144	151	150	124
Cr	672	417	405	521	574	350
Co	55	59	59	64	72	49
Ni	739	389	366	427	609	409
Cu	51	84	75	91	111	89
Zn	74	103	97	119	120	96
Rb	103	35	42	220	122	70
Sr	873	1473	1420	1398	1577	1520
Y	15	20	18	20	23	17
Zr	348	424	424	448	339	433
Nb	93	116	115	123	93	107
Ba	2173	3152	2734	2740	2658	4897

to the south-east of the province. The abundance of olivine is not affected by the degree of crystallization of the groundmass and, therefore, olivine is the best petrographic indicator. Pipes with abundant olivine (>30 modal %; P2, P3, P5 and P7) occur in the north-west of the province while those with less olivine (<20 modal %; P6, P10 and P12) occur in the south-east of the province (Table 11.1, Fig. 11.1). P4 in the centre of the province and P1 to the west are intermediate. The abundance of leucite is affected by the degree of crystallinity, but it is most abundant in the rocks from the south-east of the province (up to 40 modal %; P12, P6 and P10). It should be noted that there are significant modal variations within some intrusions.

The mineral compositions also vary between each of the intrusions examined. Using the mineral chemistry, where possible, to assess the relative degree of evolution of the intrusions (phlogopite — P12 < P6 < P10; olivine cores — P4<P6<P10; olivine rims — P4 < P6+P10; clinopyroxene — P4 < P12 < P10 + P6), the data also suggests that the pipes become more evolved to the south-east of the province (P4 < P12 < P6 < P10). The relationship of the dikes to the pipes is not understood.

In general, therefore, the intrusions from the north-west of the province are olivine lamproites which are similar to olivine lamproites elsewhere (Prairie Creek, Ellendale 4 and 9, West Kimberley, Argyle; Scott Smith & Skinner 1984a,b; Atkinson *et al* 1984; Jaques *et al* 1984). The bodies from the south-east of the province are leucite lamproites (±clinopyroxene and phlogopite). Samples suitable for detailed geochemical investigations were only found from the centre and south-east of the province. Compared with other known lamproites, the data obtained for these rocks are, in some instances, most similar to south-east Spain and the Leucite Hills. Both of the latter provinces also comprise relatively olivine-poor and leucite-rich lamproites.

The vast majority of the diamonds were recovered from olivine lamproites while the other bodies produced rare or no diamonds. The Kapamba province is, therefore, comparable to the West Kimberley, Australia occurrences, particularly the Ellendale field, which also include olivine lamproites which may be diamondiferous, together with leucite lamproites which yielded rare or no diamonds (Atkinson *et al* 1984; Jaques *et al* 1984).

Having established the Kapamba bodies as a province of diamond-bearing lamproites, it is interesting to consider their tectonic setting. The so-called classical model, as summarized by Dawson (1980), is that most (diamondiferous) kimberlites occur in old cratonic areas. The Kapamba bodies which occur in the 1355 Ma Irumide tectonic belt obviously do not follow this model. They, therefore, show similarities to other off-craton lamproites including those of the West and East Kimberley (Western Australia) and Arkansas (Atkinson *et al* 1984; Scott Smith & Skinner 1984a; Skinner *et al* 1985; Bergman, 1987). The occurrence of the Kapamba intrusions, as well as the three provinces of kimberlites, close to or within the Luangwa graben is noteworthy. Given the possible ages of the Kapamba lamproites and the main graben faulting, it is possible, but far from proven, that processes associated with the faulting and the magmatism may be related. If so, this association would be unusual for both lamproites and kimberlites, particularly if related to the East African rifting. Other varieties of alkaline magmatism, however, are commonly associated with continental rifting (Bailey 1974). The near surface emplacement of the Luangwa Valley lamproites and kimberlites, however, is not controlled by the major faults.

11.8 CONCLUSIONS

The Kapamba province comprises a suite of pipes and dikes. The geology and petrography of the Kapamba pipes suggests that they are craters predominantly infilled with pyroclastic lapilli tuffs. Some are intruded by younger, magmatic lamproite which may have formed extrusive, ponded lava lakes. The Kapamba bodies are composed of leucite, clinopyroxene, olivine, titaniferous phlogopite, titanian potassic richterite, spinel, perovskite, apatite, sanidine and glass. They are classified as lamproites according to their geology, petrography, mineral chemistry and whole-rock geochemistry. Some of the intrusions yielded mantle-derived garnets, spinels and diamonds. The Kapamba bodies, therefore, represent another province of diamond-bearing lamproite.

Petrography and mineral chemistry within the Kapamba province suggests that, in general, the pipes become relatively more evolved from north-west to south-east. The olivine lamproites from the north-west produced most of the diamonds. These bodies are similar to other (diamondiferous) olivine lamproites (Prairie Creek, Ellendale and Argyle). Certain aspects of samples from the south-east of the Kapamba province (leucite lamproites) are comparable with barren, (leucite) lamproites such as those from south-east Spain or Leucite Hills, Wyoming. The occurrence of diamondiferous olivine lamproites together with leucite lamproites yielding only rare diamonds is similar to the West Kimberley province. The Kapamba diamonds (D.N. Robinson, pers. comm.) are similar to other lamproitic (and kimberlitic) diamond populations. At least some of the diamonds are thought to be xenocrystic.

The Kapamba intrusions are younger than about 250 Ma. Only poorly constrained radiometric age data could be obtained (C.B. Smith & D. Phillips, pers comm.) but an age of about 220 Ma may be plausible. The Kapamba province occurs off-craton in the Irumide (1355 Ma) tectonic belt, a setting similar to that of some other lamproites but different from that typical of kimberlites. The Kapamba lamproites, as well as some kimberlites, occur within or very close to the Luangwa graben. These intrusions could be associated with rifting, but such a relationship would be unusual. The near surface emplacement of the lamproites and kimberlites is not controlled by the major graben faults.

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