

Petrology and age determinations of the ultramafic (lamproitic) rocks from the Yakokut complex, Aldan Shield, Eastern Siberia

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Abstract

Mineralogical, geochemical and isotopic data including K–Ar age determinations are presented for one of the largely unknown, Mesozoic ultramafic (lamproitic) occurrences of the Aldan Shield. Ultramafic, ultrapotassic rocks occur as isolated pipes or as dykes in alkaline complexes as the Yakokut complex. K–Ar dating of phlogopites from different ultramafic dyke rocks of Yakokut give cooling ages of 133.3 ± 1.3 Ma. These rocks generally contain olivine, clinopyroxene, phlogopite and spinel phenocrysts. Olivines are forsterite-rich (Fo_{86-94}) and undeformed, suggesting that they are phenocrysts. The low Ti and Al contents of clinopyroxenes are close to those of lamproites. Phlogopite cores are rich in Al_2O_3 and TiO_2 relative to the rims with two different zonation trends caused by distinct crystallization conditions after emplacement. Spinel is Ti-bearing magnetite and Ti–Al-magnesiochromite with high $Cr/(Cr+Al)$ ratios (> 0.9), indicating their crystallization from a lamproitic melt.

Geochemically, the rocks are ultrabasic – basic with high mg# values, low Ca, Al, Na and strong enrichments of Rb, Ba and K. Their CaO/Al_2O_3 , Zr/Nb or Ba/Sr ratios indicate their lamproitic nature and origin in the subcontinental lithosphere of a depleted mantle source which had undergone metasomatic enrichment. As the rocks all show strong negative Nb anomalies and low Ti, Y and Yb contents, the enrichment is attributed to subduction zone fluids. The Sr–Nd data ($^{87}Sr/^{86}Sr_0$ $0.70573-0.70605 \pm 0.00003$; $\epsilon_{Nd} -10.2 \pm 0.7$) indicate the origin by partial melting of a heterogeneous mantle source with relatively low Rb/Sr ratios and an early enrichment. The evolution model comprises a depletion of ‘basaltic elements’, leading to a harzburgitic source which was enriched by an early *LREE* contribution during the stabilization of the Archaean to Proterozoic Aldan Shield. It is further suggested that the Mesozoic northwesterly directed subduction of the Ochotsk–Chukotsk belt influenced the subcontinental lithosphere underneath the Aldan Shield, leading to the observed subduction-related signature of the Yakokut lamproites.

KEYWORDS: ultramafic rocks, lamproite, geochemistry, isotope data, Aldan Shield, Yakokut complex, Siberia.

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Introduction

THE recent general interest in ultrapotassic rocks has been stimulated by the discovery of diamondiferous lamproites in the West Kimberley region of Western Australia (Jaques *et al.*, 1984, 1986; Atkinson *et al.*, 1984) and from Prairie Creek (Scott-Smith and Skinner, 1984 *a* and *b*). It was further recognized that the parent magmas have been derived from a metasomatized mantle (Vollmer and Norry, 1983; McCulloch *et al.*, 1983; Foley *et al.*, 1987; Mitchell and Bergman, 1991). A widely accepted belief is that volatile-rich, mafic potassic magmas represent some of the earliest melts to be generated when the lithospheric mantle undergoes heating and/or extension (McKenzie, 1989; Thompson *et al.*, 1990; Gibson *et al.*, 1993).

In the Central Aldan, dykes, sills, and pipes of ultrabasic and basic ultrapotassic composition occur which are regarded as lamproites (Bogatikov *et al.*, 1987; Machotkin *et al.*, 1989, 1991). We have studied the mineralogical and geochemical compositions including Sr and Nd isotopes as well as K/Ar ages of the ultramafic dyke rocks from the Yakokut complex to characterize these rocks in more detail. To anticipate the final result, we found out by using the chemical classification scheme of Foley *et al.*, (1987) combined with the criteria of Mitchell and Bergman (1991) that these rocks are lamproitic of which the typical 'intraplate' characteristics are obscured by a subduction zone finger print.

The term lamproite was introduced into petrological literature by Niggli (1923) and perpetuated as a group name by Wade and Prider (1940) in their description of rocks from Western Australia. An early attempt to classify ultrapotassic rocks on a chemical basis was by Sahama (1974) followed later by Barton (1979), Foley *et al.*, (1987), Edgar (1987), Bergman (1987), Rock (1991), Mitchell and Bergman (1991) and Edgar and Vukadinovic (1993). The first classification scheme used by Russian petrologists is that devised by Bogatikov *et al.*, (1985), considering lamproites to be members of the high-magnesium rock group distinguished from komatiites by their potassic nature. Later Bogatikov *et al.*, (1987) proposed a revised version which is quite close to those of Foley *et al.*, (1987) and Mitchell and Bergman (1991).

Foley *et al.*, (1987) define ultrapotassic rocks as those having $K_2O/Na_2O > 2$, $K_2O > 3$ wt.%, and $MgO > 3$ wt.%. They recognize three end-member compositional groups of ultrapotassic rocks, namely lamproites, kamafugites and Roman Province rocks (plagioclucitites) and concluded that rocks of unknown affinity may be placed in the above groups by reference to bivariate plots.

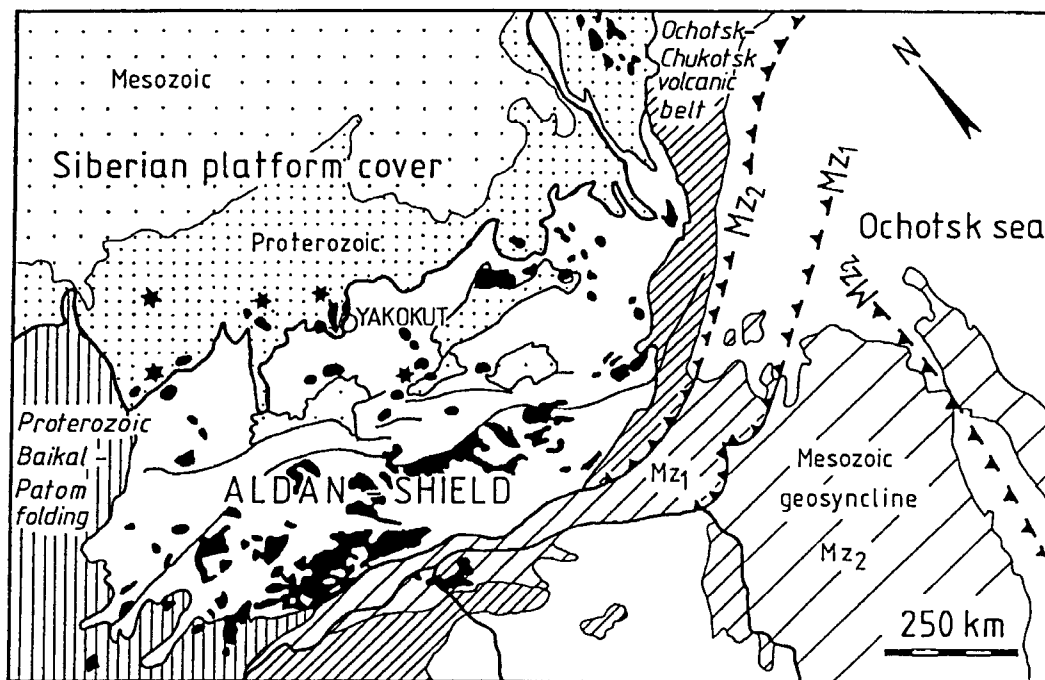
Lamproites as defined by Mitchell and Bergman (1991) refer to a clan of rocks including geochemical and mineralogical criteria. The most striking compositional features are molar $K_2O/Al_2O_3 > 0.8$ and $(K_2O+Na_2O)/Al_2O_3 > 0.7$, typically FeO_T and $CaO < 10$ wt.%, high Ba (> 2000 ppm), TiO_2 (1–7 wt.%), Zr > 500 ppm, Sr > 1000 ppm, La > 200 ppm. Typical primary phases of lamproites are titanian Al_2O_3 -poor phenocrystal phlogopite, titanian groundmass poikilitic tetraferriphlogopite, Ti–K-richterite, forsteritic olivine, Al–Na-poor diopside, Fe-rich leucite or sanidine. Minor but common accessory phases include priderite and wadeite, but also apatite, perovskite, magnesiochromite and others (Bergman and Mitchell, 1991).

Geological setting of the Aldan-Shield

The Aldan-Shield forms the southeastern part of the Siberian platform, and consists mainly of Archaean and Proterozoic rocks (Bilanenko *et al.*, 1984; Perchuk *et al.*, 1985; Fig. 1a). Today, N-S striking Precambrian structures superimposed by WSW-ENE Mesozoic structures are exposed. The Precambrian is subdivided into three lithological units: (1) a granite–greenstone terrane in the west; (2) the Central Aldan zone of mainly Archaean granulites and granites; and (3) to the east the Timpton zone of amphibolites and metabasites intercalated with granulitic rocks. A younger sedimentary cover is of Vendian to Lower Cambrian age (Bilanenko *et al.*, 1984).

The major Mesozoic structures of the Aldan-Shield from south to north are: (1) the Stanovoy zone with Jurassic granitic batholiths; (2) a Mesozoic depression zone with coal deposits; (3) the Aldan uplift zone exposing Cambrian sediments overlain by the Lower Jurassic; and (4) the Aldan Anticline, where the Precambrian basement dips at low angle under the more or less flat lying sedimentary cover of the Siberian platform (Bilanenko *et al.*, 1984; Perchuk *et al.*, 1985; Fig. 1a). The geotectonic complexity can be attributed to plate tectonics of repeated accretion/subduction processes, since the Proterozoic which are subsequently accompanied by magmatic activity. The major events were the accretion of the Aldan Shield to the Siberian platform during the Proterozoic and the accretion of the Burein block to the Aldan Shield in the Mesozoic.

Five distinct periods of alkaline magmatism are recognized in the Aldan Shield: (1) Lower Proterozoic; (2) Upper Proterozoic to Vendian; (3) Devonian; (4) Mesozoic; and (5) Cenozoic (Maksimov and Ugryumov, 1971; Bilibin, 1961). Within the major Mesozoic period potassium-rich rocks such as leucitites, pseudoleucite-syenites and alkaline granites generated, irregularly dispersed throughout the Aldan Shield (Fig. 1b).



a
b

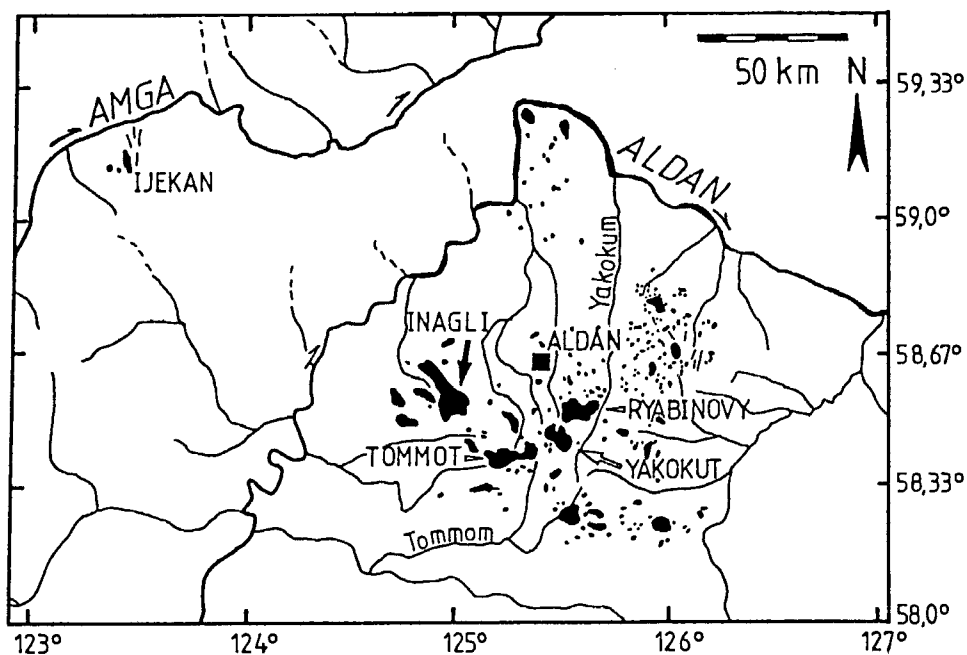


FIG. 1. (a) Generalized geological sketch map of Eastern Siberia. Stars represent lamproite occurrences. (b) Regional dispersion of alkaline rocks in the Central Aldan province (redrawn after Bilibin 1961).

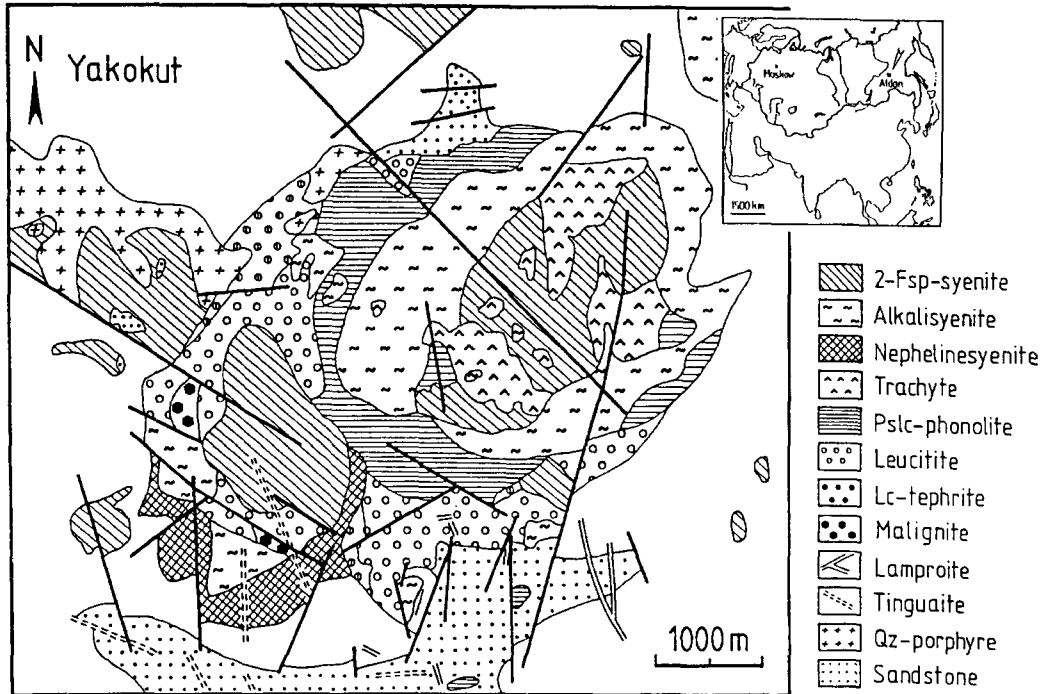


FIG. 2. Sketch map of the Yakokut-complex (redrawn after Maksimov, 1972) with the ultrabasic lamproitic dykes in the southern marginal part. The inset shows the borders of the old Soviet Union and the town of Aldan of which Yakokut is located about 25 km to the southeast.

The Yakokut-complex is one of these Mesozoic alkaline complexes (Machotkin *et al.*, 1989; Kotchetkov and Lasebnik, 1984; Kanucov *et al.*, 1991, Fig. 1*b*). It is made up mainly of alkali-syenitic rocks (Fig. 2) which form two centres of several petrographic varieties of syenite surrounded by leucitites and phonolites (Kotchetkov and Lasebnik, 1984; Maksimov, 1972; see Fig. 2). In the southern part, the Yakokut complex was intruded by several tinguaitic and ultramafic lamproitic dykes, the latter being the object of this study.

Petrography

The dyke rocks from Yakokut are characterized by olivine, clinopyroxene, spinel and apatite with or without phlogopite macrocrysts set in a cryptocrystalline matrix (Fig. 3). The phlogopite-bearing rocks are ultrabasic, whereas the phlogopite-poor ones are of basic compositions. Apatite is frequently found as an euhedral accessory phase. Modal compositions of both rock types are given in Table 1.

Olivine occurs as large (up to 2.5 mm) subhedral macrocrysts comprising up to 28 vol.% of the rocks (Table 1). Their margins are commonly altered with a rim of opaque oxides, mainly magnetite (Fig. 3). Undeformed olivines with euhedral inclusions of chromite showing no undulose extinction do not suggest concentration of olivines by flow differentiation. Single crystals of olivine are characterized by coronas consisting of interlocking laths of phlogopite which appear to have nucleated preferentially upon olivine rather than resulting from a reaction between olivine and liquid.

Abundant colourless or pale-green euhedral up to 2 mm *clinopyroxene* phenocrysts or minute grains are usually untwinned and lack resorption features. Rarely, intensely green pleochroic aegirine-rich patches in colourless pyroxenes are found. The abundance of pyroxene (20–40 vol.%, Table 1) is high compared to lamproites.

Micas are strongly pleochroic from pale yellow-brown to reddish-brown, mostly inclusion-free, although rarely euhedral inclusions of chromite occur. As is usual in lamproites, the micas are

TABLE 1. Modal composition in vol.% of the ultrabasic and basic dyke rocks of the Yakokut complex. Additionally apatite and kalifeldspar occur as phenocrysts and in the groundmass, respectively

Sample	AY 26	AY 27	AY 31	AY 33
olivine	20.8	14.7	25.3	7.1
altered olivine	—	14.9	—	4.1
clinopyroxene	26.2	24.4	21.3	39.9
phlogopite	18.8	21.2	24.6	2.8
oxides	11.4	7.5	5.5	11.8
groundmass	22.8	17.3	23.3	34.3

zoned and have thin rims of tetraferriphlogopite with reddish brown to dark brown reverse pleochroism.

*Spinel*s occur commonly as single-phase euhedral phenocrysts or as euhedral inclusions in olivines and scarcely in phlogopites or clinopyroxenes. Rarely, spinels occur as two-phase crystals consisting of a discrete core (chromite) and mantle (magnetite).

The unresolved matrix of the rocks is mainly colourless to pale-green and Fe-poor phlogopite-like compositions could be detected. Rarely, poikilitic K-feldspar surrounds the macrocrysts as matrix. In the basic varieties, the macrocrysts are set in a matrix of microcrystal K-feldspar, clinopyroxene, magnetite, and titanian phlogopite.

Mineral compositions

The majority of mineral analyses were performed with a Cambridge S4-10 Scanning electron microscope with KEVEX 7000 energy dispersive system. The analytical conditions were 15 kV accelerating voltage, 1 nA beam current, point measuring for 100 seconds dead-time corrected. Some few microprobe analyses were performed on a JEOL 6400 for the minor element contents of minerals.

Olivine. Olivines are Mg-rich with mg-numbers varying between 0.94 and 0.86 (Table 2). Generally, they are unzoned, with occasional large crystals showing normal zoning to slightly Fe-richer rims. All olivines are characterized by uniform NiO contents averaging about 0.20–0.25 wt.%. CaO contents correlate negatively with mg-numbers and CaO is < 0.55 wt.% at mg-numbers > 0.90. Olivines of lower mg-numbers (0.90–0.86) have higher CaO contents varying between 0.60 and 0.80 wt.% (Table 2).

The modal olivine content (up to 25 vol.%, Table 1) in the Yakokut ultramafic rocks is high when compared with most lamproite occurrences where up to 11 vol.% olivine are reported (Mitchell

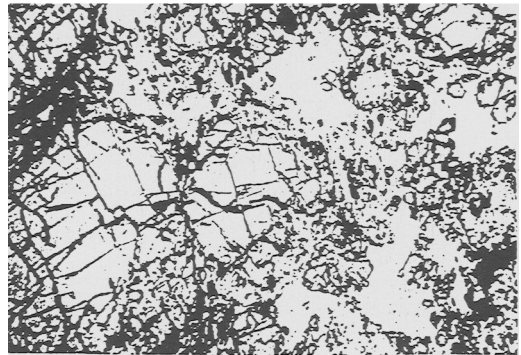
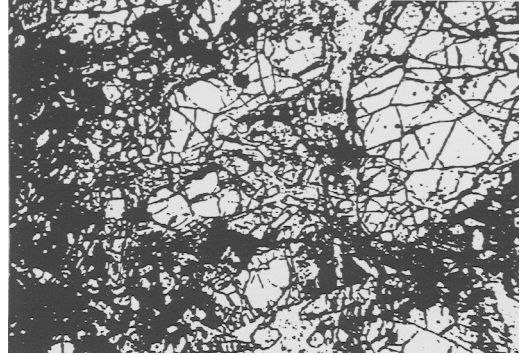


FIG. 3. Characteristic appearance of two different ultrabasic rocks with olivines, spinels, clinopyroxenes, and phlogopites. Note the reversed pleochroism of phlogopite rims in the upper plane polarized light photomicrograph. Fields of view approximately 2.7 mm.

and Bergman, 1991). However, olivine lamproites from Prairie Creek and the West Kimberley region reach maximum amounts of 40 vol.% olivine. Some of these were interpreted as xenocrysts exhibiting undulose extinction and/or kink banding as well as resorption phenomena (Scott-Smith and Skinner, 1984b; Jaques *et al.*, 1986). In the Spanish lamproites Venturelli *et al.*, (1984) distinguished olivine xenocrysts with high NiO contents (0.45–0.7 wt.%) from phenocrysts with only up to 0.32 wt.% NiO.

As olivines from the Yakokut rocks do not show undulose extinction or deformation, the lack of flow alignment textures and the uniform low NiO-contents (≤ 0.3 wt.%, Table 2) lead us to regard all olivines as phenocrysts.

Clinopyroxene. Mitchell and Bergman (1991) have shown that clinopyroxenes of lamproites and other potassic rocks have virtually distinct differences (Fig. 4). Table 3 shows that most of the clinopyroxenes are diopsides ($\text{Ca}_{49}\text{Mg}_{44}\text{Fe}_7$) with low TiO_2

TABLE 2. Representative electron microprobe analyses of olivines of the Yakokut ultrabasic rocks. Formula calculations are based on 4 oxygens; mg# = Mg/(Mg+Fe)

Sample Anal.	AY26 4	AY26 15	AY26 18	AY26 20	AY26 22	AY26 35	AY31 2	AY31 9	AY31 17	AY31 29
SiO ₂	40.30	41.24	40.67	39.97	39.53	40.68	39.36	41.16	40.82	39.40
FeO	12.58	10.44	10.51	10.42	10.51	6.79	13.00	7.73	7.58	9.81
MnO	0.29	0.21	0.25	0.18	0.22	0.13	n.a.	0.16	n.a.	0.19
NiO	0.20	0.24	0.23	0.19	0.14	0.27	0.22	0.23	0.30	0.23
MgO	46.86	48.65	48.28	48.71	48.89	52.68	46.69	49.51	50.69	48.36
CaO	0.67	0.67	0.60	0.59	0.63	0.47	0.67	0.50	0.43	0.72
Total	100.90	101.45	100.54	100.06	99.92	101.02	99.94	99.29	99.82	98.71
Structural formula based on 4 oxygens										
Si	0.995	1.001	0.997	0.986	0.978	0.978	0.984	1.007	0.994	0.984
Fe	0.260	0.212	0.216	0.215	0.217	0.137	0.272	0.158	0.154	0.205
Mn	0.006	0.004	0.005	0.004	0.005	0.003	n.a.	0.003	n.a.	0.004
Ni	0.004	0.005	0.005	0.004	0.003	0.005	0.004	0.005	0.006	0.005
Mg	1.724	1.760	1.765	1.791	1.803	1.888	1.739	1.806	1.840	1.800
Ca	0.018	0.017	0.016	0.016	0.017	0.012	0.018	0.013	0.011	0.019
Total	3.005	2.999	3.003	3.014	3.022	3.022	3.016	2.993	3.006	3.016
mg#	0.869	0.893	0.891	0.893	0.892	0.933	0.865	0.919	0.923	0.898

n.a. - not analysed

(0.2–0.9 wt.%) and Al₂O₃ (0.2–1.1 wt.%). The general lack of extensive zonation results from fast cooling after the beginning of pyroxene crystallization. In terms of Ca, Mg and Fe contents clinopyroxenes have a very similar composition to those of lamproites (e.g. Kapamba, Zambia and Leucite Hills). The positively correlated relatively low Ti contents and variable Al contents are

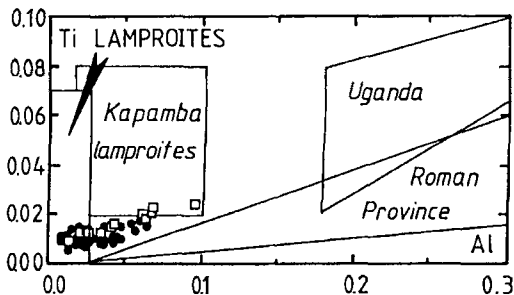


FIG. 4. Ti and Al contents of clinopyroxenes (per formula) from ultrabasic (filled circles) and basic (open squares) dyke rocks from Yakokut. The fields of clinopyroxenes from lamproite occurrences, the Ugandan kamafugites and ultrapotassic rocks of the Roman Province are shown for comparison (after Mitchell and Bergman, 1991).

essentially the same for pyroxenes from ultrabasic (closed circles) and basic (open squares) rocks and are quite similar to Ti–Al contents of lamproites (Fig. 4).

Phlogopite. One of the few minerals that shows distinct zonations and a wide compositional range in ultrapotassic rocks is phlogopite. In particular the contents of TiO₂, Al₂O₃, and FeO_T have been used to assess petrogenetic relationships within and between provinces (Mitchell, 1985).

Phlogopite cores from the Yakokut ultramafic rocks are rich in TiO₂ (2.0–4.0 wt.%), Al₂O₃ (11.5–14.0 wt.%), and BaO (0.7–2.1 wt.%), and poor in FeO_T (10.0–15.0 wt.%). The rims have significantly lower TiO₂ (0.5–1.5 wt.%), Al₂O₃ (11.5–5.0 wt.%), and BaO (0.1–0.3 wt.%) and higher FeO_T (12.5–35.0 wt.%) contents (Fig. 5 and Table 4). We have illustrated this on Fig. 5 by plotting Al₂O₃ vs. TiO₂. For comparison we have also shown the fields of mica from the MARID (mica-amphibole-rutile-ilmenite-diopside) xenolith suite, kimberlites, Leucite Hills, West Kimberley, and Smoky Butte lamproites. Mica in other potassic rocks, those of minettes and Roman Province rocks, have higher Al contents and never develop tetraferriphlogopitic compositions (Mitchell and Bergman, 1991). The overall trend of mica from kimberlites is that from micas relatively poor in Ti,

TABLE 3. Representative electron microprobe (EDX) analyses of clinopyroxenes of the Yakokut ultramafic rocks recalculated to 100%. Formula calculations are based on 6 oxygens with Fe_{tot} calculated as Fe^{2+} ; mg# = $Mg/(Mg+Fe)$. anal. ~ the number of the analyses, not the average of e.g. 7 analyses

Sample Anal.	AY26 7	AY26 9	AY26 12	AY27 1	AY27 3	AY31 9	AY31 16	AY31 21	AY98 2	AY98 14
SiO ₂	54.07	53.36	53.68	53.64	53.11	53.30	53.67	52.67	52.13	52.99
TiO ₂	0.19	0.32	0.30	0.35	0.49	0.25	0.24	0.29	0.56	0.52
Al ₂ O ₃	0.32	0.47	0.57	0.21	0.37	0.44	0.53	0.87	0.86	1.03
Cr ₂ O ₃	0.18	0.34	0.72	0.34	0.15	0.19	0.45	0.28	0.00	0.27
FeO	3.96	4.55	3.60	4.17	6.11	5.03	4.00	5.69	5.99	4.80
MgO	15.98	16.18	16.33	16.25	15.27	16.25	16.08	15.48	14.32	15.52
CaO	25.29	24.79	24.80	25.05	24.50	24.55	25.01	24.72	25.70	24.88
Na ₂ O	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.44	0.00
Total	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00
Structural formula based on 6 oxygens										
Si	1.985	1.965	1.969	1.973	1.968	1.966	1.972	1.951	1.944	1.954
Ti	0.005	0.009	0.008	0.010	0.014	0.007	0.007	0.008	0.016	0.014
Al	0.007	0.010	0.012	0.005	0.008	0.010	0.012	0.019	0.019	0.022
Cr	0.003	0.005	0.010	0.005	0.002	0.003	0.007	0.004	0.000	0.004
Fe	0.122	0.140	0.110	0.128	0.189	0.155	0.123	0.176	0.187	0.148
Mg	0.874	0.888	0.893	0.891	0.844	0.893	0.881	0.855	0.796	0.853
Ca	0.995	0.978	0.975	0.987	0.973	0.970	0.985	0.981	1.027	0.983
Na	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.016	0.000
Total	3.991	3.996	3.977	3.999	3.998	4.003	3.985	3.995	4.003	3.979
mg#	0.878	0.864	0.890	0.874	0.817	0.852	0.878	0.829	0.810	0.852

Al and Fe to types that are relatively rich in these elements, and tetraferriphlogopite is not common, but when present Ti-poor (Mitchell, 1986).

With respect to the Ti and Al contents, the Yakokut phlogopite cores resemble micas of the Leucite Hills lamproites. The Al-depletion and Fe-enrichment trend of the Yakokut phlogopites therefore is not really similar to that of lamproites as Al contents at a given Fe-value are higher than in lamproites.

The observed two divergent crystallization trends are characterized (a) by strong Ti-Al depletion with little increase in Fe and (b) very strong Ti-Al-Fe depletion of the rims towards TiO₂-poor tetraferriphlogopite compositions. They overlap the fields of both kimberlites and lamproites in terms of their TiO₂ and Al₂O₃ contents (Fig. 5). This is mainly because the composition of most of the phlogopites reflects the local post-emplacement crystallization environment rather than the composition of the parental magma. The differences between the two trends are thought to reflect different post-emplacement (redox) conditions of crystallization. Tetraferriphlogopite compositions are restricted to the rims and Si and Al contents are insufficient to fill

all the tetrahedral sites, and consequently Fe³⁺ enters into the tetrahedral position (Farmer and Boettcher, 1981).

The composition of the phlogopite Al-Ti-rich cores of all Yakokut ultramafic rocks is essentially constant, suggesting the derivation from one common parental magma with early Al-rich titanian phlogopites on the liquidus. The incorporation of Ti in phlogopites is controlled essentially by pressure and oxygen fugacity, whereas temperature and composition of the melt have only minor effects (Arima and Edgar, 1981, 1983). Decreasing pressure causes increasing Ti contents in phlogopites (Robert, 1976). Under upper-mantle and lower-crustal conditions, the solubility of TiO₂ in phlogopite is buffered by coexisting mineral phases such as olivine (Tronnes *et al.*, 1985). Oxygen fugacities of at least 2 Δ log units above the iron-wüstite-buffer are responsible for decreasing Ti contents in phlogopites (Foley, 1990). The effect of high oxygen fugacities originated by dissociation of water during ascent is strong enough to compensate the effect of decreasing pressure (Foley, 1990). The Ti-depletion trends of the Yakokut phlogopites are therefore thought to document strong oxidation of the ultramafic melts during ascent.

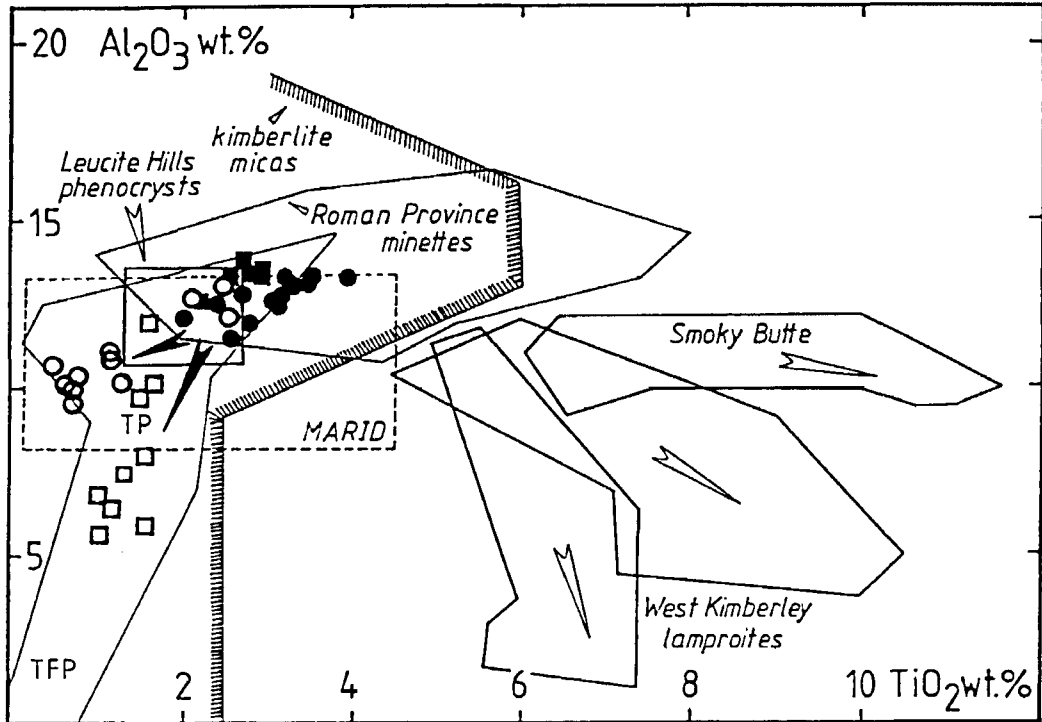


FIG. 5. Al_2O_3 vs TiO_2 for phlogopites of the Yakokut dyke rocks in comparison to phlogopite compositions from West Kimberley, Smoky Butte and Leucite Hills lamproites as well as the fields of micas from kimberlites, MARID, minettes and Roman Province. Cores – filled symbols; rims – open symbols.

Spinel. The Yakokut ultrabasic rocks contain up to 12 vol.% spinels (Table 1) as phenocrysts or as inclusions in olivines and clinopyroxenes. Spinel inclusions generally are Ti-Al magnesiochromites, whereas phenocrysts may be either Ti-Al magnesiochromites or titaniferous magnetites with TiO_2 contents up to 5.5 wt.% (Table 5). Rarely, chromite phenocrysts possess rims of magnetite compositions.

The chromites have high $\text{Cr}/(\text{Cr}+\text{Al})$ ratios (Table 5), essentially varying between 0.80 and 0.99, without significant differences between chromite inclusions and phenocrysts. Such high $\text{Cr}/(\text{Cr}+\text{Al})$ ratios (> 0.90) are typical for lamproite spinels (Mitchel, 1985). In contrast, chromites of other ultrabasic rocks, e.g. lherzolithic rocks, have significantly higher Al_2O_3 contents resulting in distinctly lower $\text{Cr}/(\text{Cr}+\text{Al})$ ratios (Carswell, 1980). Spinels in Roman Province lavas are notably MgO poor (< 1 wt.%) and Al_2O_3 rich (2–5 wt.%) relative to lamproite spinels of similar total Fe and TiO_2 content (Cundari, 1975, 1979; Baldrige *et al.*, 1981). Therefore, the general Ti-Al magnesiochromitic composition and their high $\text{Cr}/(\text{Cr}+\text{Al})$ ratios may

indicate their crystallization from a lamproitic magma.

Apatite. Apatite occurs as a common accessory phase in the ultrabasic and basic dyke rocks of Yakokut. They are fluorapatites with high Sr-contents up to 10.0 wt.% (Table 6). All apatites exhibit strong normal and reversed zoning with respect to Sr and fluorine contents.

Fluorine contents generally vary between 1.3 and 5.4 wt.% with an increasing average of 2.4 wt.% in apatites of ultrabasic to 4.1 wt.% in those of basic rocks (Table 6). The observed fluorine contents are similar to contents reported from apatites of lamproites (Edgar and Charbonneau, 1991) or kamafugites (Edgar *et al.*, 1994). The Sr contents are high and highly variable with decreasing average SrO contents of 5.9 wt.% SrO in ultrabasic to 2.3 wt.% SrO in apatites of basic rocks. Characteristically, SrO greatly exceeds BaO in fluorapatites, with BaO contents reaching only up to 0.3 wt.%. The variation of Sr, Ba, F and Cl in apatites of lamproites is interpreted by Edgar (1989) to be mainly a function of whole-rock chemistry.

TABLE 4. Electron microprobe analyses of phlogopites of the Yakokut ultramafic rocks. Formula calculations are based on 22 oxygens with Fe_{tot} calculated as Fe^{2+} ; mg# = $Mg/(Mg+Fe)$. c – core, m – middle, r – rim

Sample Anal.	AY31 1-c	AY31 1-m	AY31 1-r	AY31 2-c	AY31 2-m	AY31 2-r	AY31 3-c	AY31 3-r	AY33 1	AY33 2
SiO ₂	36.46	37.96	44.55	38.10	39.09	39.50	37.41	39.01	35.73	37.46
TiO _{2.3}	2.87	0.82	0.07	2.64	1.82	0.61	2.24	0.95	4.11	2.57
Al ₂ O ₃	13.48	9.38	11.07	13.12	11.81	9.33	12.10	11.16	11.47	12.99
Cr ₂ O ₃	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.
FeO	10.28	13.87	5.47	10.63	11.34	12.92	10.87	12.03	16.54	14.56
MgO	20.63	21.63	27.61	20.41	22.41	22.55	21.43	21.92	13.74	15.32
CaO	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.	0.24
Na ₂ O	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.
BaO	1.96	0.23	0.05	1.60	1.01	0.27	1.25	0.18	4.94	3.83
K ₂ O	9.08	9.31	8.00	8.85	9.16	8.64	9.52	8.40	7.77	7.11
Total	94.76	93.20	96.82	95.35	96.64	93.82	94.82	93.65	94.30	94.08
Structural formula based on 22 oxygens										
Si	5.458	5.815	6.140	5.627	5.698	5.934	5.592	5.825	5.650	5.761
Ti	0.323	0.094	0.007	0.293	0.200	0.069	0.252	0.107	0.489	0.297
Al	2.378	1.694	1.798	2.284	2.029	1.652	2.132	1.964	2.138	2.355
Cr	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.
Fe	1.287	1.777	0.630	1.313	1.382	1.623	1.359	1.502	2.187	1.873
Mg	4.603	4.939	5.672	4.493	4.869	5.049	4.775	4.878	3.239	3.511
Ca	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.	0.039
Na	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.
Ba	0.115	0.014	0.003	0.093	0.058	0.016	0.073	0.011	0.306	0.231
K	1.734	1.819	1.406	1.667	1.703	1.656	1.815	1.600	1.567	1.394
Total	15.897	16.153	15.657	15.771	15.939	15.999	15.998	15.887	15.576	15.461
mg#	0.782	0.735	0.900	0.774	0.779	0.757	0.778	0.765	0.597	0.652

n.a. - not analysed

Whole-rock geochemistry

As outlined previously, the Yakokut rocks are characterized geochemically using the classification scheme of Foley *et al.* (1987) and the chemical criteria of Mitchell and Bergman (1991). The dyke rocks of Yakokut exhibit a wide range in MgO (12.1–33.1 wt.%) with Fe (FeO+Fe₂O₃) contents (9.5–12.0 wt.%) and mg#-values ranging from 0.72 to 0.87, i.e. they are ultramafic rocks. As we have mentioned above K₂O-contents are > 2.75 wt.% and hence K₂O/Na₂O ratios are variable within these rocks, ranging from 4.7 to 9.5 (Table 7). The rocks may therefore be classified as ultrapotassic. We have plotted our samples in bivariate plots as CaO–Al₂O₃, (K₂O/Al₂O₃)–SiO₂, CaO–MgO, and Al₂O₃–Na₂O (Fig. 6). Most of the rocks have intermediate CaO (4.8–11.4 wt.%), SiO₂ (42.2–47.8 wt.%) and Al₂O₃ (3.2–8.2 wt.%) contents and often plot at the

boundary between lamproites and kamafugites with a slightly higher affinity to lamproites (Fig. 6). The rocks are perpotassic (molar K₂O/Al₂O₃ = 0.7–0.9) and peralkaline (K₂O+Na₂O)/Al₂O₃ = 0.8–1.15) typical features of lamproites with these SiO₂ contents (Fig. 7). In conclusion, they have compositions that seem almost intermediate between lamproites and kamafugites in terms of CaO and Al₂O₃ but have (slightly) higher SiO₂ than kamafugites. In view of their other features and resemblance to the olivine lamproites of West Kimberley and Kapamba lamproites, we classify them as lamproites in the remainder of this study.

The compatible trace elements of the Yakokut ultramafic rocks exhibit large variations in concentration. For instance Cr and Ni contents are in the range 590–2250 and 260–1200 ppm respectively, and show positive correlations with MgO. Scandium

TABLE 5. Representative electron microprobe (WDS) analyses of spinels of the Yakokut lamproitic rocks. $cr\# = Cr/(Cr+Al)$

Sample	AY26 14	AY26 17	AY26 22	AY26 25	AY26 26	AY26 39	AY26 44	AY31 8
TiO ₂	0.65	0.99	0.71	0.84	3.02	2.67	0.52	5.47
Al ₂ O ₃	5.88	4.22	4.16	3.81	0.04	0.06	4.88	0.95
Cr ₂ O ₃	46.41	36.44	44.48	43.44	6.59	1.75	49.38	0.14
FeO _{tot}	39.10	50.19	43.26	45.21	82.29	87.82	33.05	83.87
MnO	1.49	1.13	1.58	1.50	0.85	0.45	0.95	1.35
MgO	4.68	3.45	3.71	1.56	0.17	0.28	7.74	1.04
	98.21	96.42	97.90	96.36	92.96	93.03	96.52	92.82
Recalculated analyses								
Fe ₂ O ₃	16.08	26.30	19.44	18.32	56.12	61.91	14.89	57.70
FeO	24.63	26.51	25.76	28.72	31.78	32.10	19.65	31.93
	99.82	99.05	99.84	98.19	98.57	99.22	98.01	98.59
Mol % end member spinel molecules								
MgAl ₂ O ₄	12.23	9.04	8.79	8.34	0.09	0.14	10.14	2.16
Mg ₂ TiO ₄	1.72	2.71	1.92	0.16	0.44	0.74	1.38	1.90
Mn ₂ TiO ₄	—	—	—	2.19	1.41	0.74	—	2.21
Fe ₂ TiO ₄	—	—	—	—	5.60	5.56	—	9.57
MnCr ₂ O ₄	4.45	3.47	4.80	0.16	—	—	2.84	—
MgCr ₂ O ₄	8.94	4.23	7.22	—	—	—	27.75	—
FeCr ₂ O ₄	51.32	44.61	51.05	63.58	10.16	2.67	38.18	0.21
Fe ₃ O ₄	21.34	35.95	26.22	25.58	82.30	90.14	19.72	83.95
Structural formula based on 32 oxygens								
Ti	0.122	0.195	0.136	0.187	0.668	0.599	0.101	1.258
Al	0.867	0.650	0.624	0.666	0.007	0.011	0.740	0.171
Cr	4.589	3.766	4.474	5.097	0.766	0.206	5.022	0.017
Fe ³⁺	2.726	3.657	3.068	2.046	6.747	7.302	2.370	6.642
Fe ²⁺	5.151	5.796	5.482	7.129	7.817	8.007	4.228	8.169
Mn	0.316	0.250	0.341	0.377	0.212	0.114	0.207	0.350
Mg	1.745	1.344	1.407	0.690	0.075	0.124	2.968	0.474
	15.52	15.66	15.53	16.19	16.29	16.36	15.64	17.08
Cr#	0.841	0.853	0.878	0.884	0.991	0.951	0.872	0.090

is also often present in low (7–23 ppm) abundances, similar to those reported from lamproites and which may be much higher in kamafugitic rocks (25–55 ppm: Gibson *et al.*, 1995).

The Yakokut lamproitic rocks are characterized by high concentrations and variations of incompatible elements, e.g. Ba and Sr are in the range 1100–2900 and 380–3370 ppm, respectively. In order to give an idea of the enrichments of the trace elements, we have plotted normalized multi-element diagrams (Fig. 8a). The patterns of all the studied rocks are generally similar, with all samples having two large troughs, one at Nb and the other one at La–Ce. For the whole element range, the enrichment factor steadily

decreases for more compatible elements with the HFS-elements Ti, Y and Yb showing only low enrichment relative to the primitive mantle (Fig. 8a).

The Yakokut lamproites contain moderate abundances of rare-earth elements (*REE*). For example, La and Lu concentrations reach up to 100 and 5 × chondritic values, respectively (Fig. 8b). The highest content of *HREE* (heavy *REE*; Lu = 0.14 ppm) is present in a basic lamproite (AY33), which has the lowest MgO content. This high *HREE* content may reflect the effect of fractional crystallization in the evolution of this rock. A negative Eu anomaly as observed in the kamafugitic rocks from San Venanzo and Cupaello (Peccerillo *et al.*, 1988) is absent in the

TABLE 6. Representative electron microprobe analyses of apatites of the Yakokut ultrabasic rocks. Formula calculations are based on 24 oxygens (Edgar, 1989). The analyses show core (c) – rim (r) zonations of the apatites. Totals are corrected for F,Cl=O

Sample Anal.	AY26 3-c	AY26 3-r	AY26 16	AY29 31-c	AY29 31-r	AY29 35	AY29 40-c	AY29 40-r	AY29 49	AY33 33-c	AY33 33-r
P ₂ O ₅	38.42	39.37	39.76	38.44	40.55	38.20	39.23	38.59	36.46	39.68	37.69
SiO ₂	0.64	0.44	0.34	0.57	0.06	1.46	1.21	1.49	0.13	2.27	1.98
CaO	51.56	51.62	51.23	51.72	49.77	54.51	55.25	55.53	45.92	52.44	52.61
SrO	4.71	4.91	5.09	2.17	4.60	0.72	0.74	0.78	9.99	1.25	1.20
BaO	0.00	0.00	0.04	0.01	0.10	0.02	0.06	0.03	0.13	0.12	0.10
FeO	0.57	0.42	0.37	0.30	0.32	0.34	0.35	0.31	0.34	0.56	0.48
MgO	n.a.	n.a.	n.a.	0.11	0.01	0.22	0.21	0.21	0.28	0.27	0.22
Na ₂ O	0.16	0.16	0.13	0.09	0.18	0.10	0.06	0.07	0.29	0.06	0.03
K ₂ O	0.27	0.38	0.25	0.07	0.20	0.13	0.16	0.23	0.20	0.21	0.42
F	2.16	2.59	2.36	5.03	3.83	3.91	2.69	3.45	3.24	2.55	3.39
Cl	0.61	0.53	0.49	0.06	0.01	0.09	0.11	0.11	0.00	0.26	0.20
Total	98.29	99.42	99.14	96.48	99.06	99.02	98.99	99.41	95.66	98.66	96.96
Structural formula based on 24 oxygens											
P	5.522	5.590	5.638	5.590	5.779	5.413	5.465	5.393	5.546	5.502	5.398
Si	0.109	0.074	0.057	0.098	0.010	0.244	0.199	0.246	0.023	0.372	0.335
Ca	9.378	9.275	9.194	9.519	8.977	9.775	9.740	9.820	8.840	9.202	9.536
Sr	0.464	0.477	0.494	0.216	0.449	0.070	0.071	0.075	1.041	0.119	0.118
Ba	0.000	0.000	0.003	0.001	0.007	0.001	0.004	0.002	0.009	0.008	0.007
Fe	0.081	0.059	0.052	0.043	0.045	0.048	0.048	0.043	0.051	0.077	0.068
Mg	n.a.	n.a.	n.a.	0.028	0.003	0.055	0.052	0.052	0.075	0.066	0.055
Na	0.053	0.052	0.042	0.030	0.059	0.032	0.019	0.022	0.101	0.019	0.010
K	0.063	0.088	0.058	0.017	0.047	0.030	0.036	0.052	0.050	0.048	0.098
F	1.137	1.232	1.242	2.648	2.016	2.058	1.416	1.816	1.705	1.342	1.784
Cl	0.172	0.158	0.138	0.017	0.003	0.025	0.031	0.031	0.000	0.073	0.056

n.a. - not analysed

Yakokut rocks. Nevertheless, the observed *REE* contents in the Yakokut rocks are lower than in any other ultrapotassic rock (Fig 8c).

The variations and high concentrations of compatible elements such as Cr, Ni, and Sc are comparable with other worldwide lamproite occurrences, especially the olivine-lamproites from West Kimberley, Kapamba and Prairie Creek. The significant relative depletions in high-field-strength elements (Nb and to a lesser extent Ti) is similar to Leucite Hills and the Roman Province. Ba and Sr contents and the resulting high Ba/Sr ratios of *ca.* 2.5 are similar to those reported from the Spanish lamproites and others, but significantly higher than those in Roman Province rocks (0.5–2), kimberlites, lamprophyres (1–1.4) and kamafugites (< 2; Mitchell and Bergman, 1991). Even if the Zr and Nb abundances of the Yakokut rocks are quite lower than those from other ultrapotassic rocks, the resulting Zr/Nb ratios of mainly ± 28 are similar to those reported from lamproites and significantly higher than those of up to 5 in the Toro Ankole kamafugites.

Radiogenic isotopes

Three of the ultrabasic lamproite samples have been analysed by the conventional K–Ar method and for Rb–Sr isotopes. The sample selection depended mostly on the available sample size for mineral separation. The K–Ar and Rb–Sr analyses were carried out in the isotope laboratory of the University in Berne (extended methodical descriptions in Flisch, 1982, 1986; Hurford *et al.*, 1986). Sample ages were calculated using the constants of Steiger and Jäger (1977).

K–Ar ages of the lamproites. Two separate phlogopite concentrates as well as two grain-size fractions of the whole rock were analysed from each of three different ultrabasic samples. Five ages of 133.3 ± 1.3 Ma were determined and one slightly higher age of 135.5 ± 1.3 Ma within the analytical error (Table 8). Whole-rock determinations show a variability between 136.6 ± 1.3 and $144. \pm 1.4$ Ma which are significantly higher than the phlogopite ages (Table 8). A possible explanation is

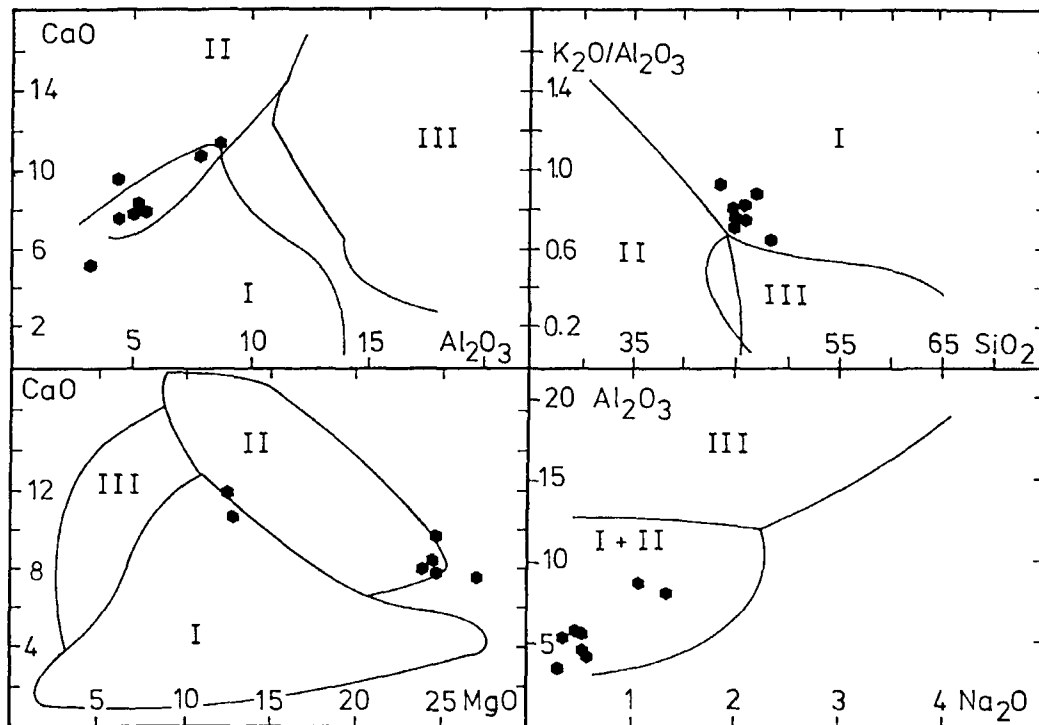


FIG. 6. Variation of CaO–Al₂O₃, CaO–MgO, K₂O/Al₂O₃ (molar)–SiO₂, and Al₂O₃–Na₂O in ultrapotassic rocks (according to Foley *et al.*, 1987), showing the chemical position of the Yakokut ultrabasic rocks as intermediate between Group I (lamproites) and Group II (kamafugites) rocks. The lines drawn serve as approximate group delimiters, but are not intended as strict boundaries (Foley *et al.*, 1987). I — Group I, lamproites; II — Group II, kamafugites; and III — Group III, Roman Province rocks.

incorporated argon in olivines during their crystallization, the so-called excess argon, a factor that is well known for olivines (Flich 1986, and references therein). This has been tested with one olivine concentrate, giving an age of 313.7 ± 3.1 Ma, indicating excess argon in the olivines (Table 8). As these lamproites have up to 28 vol.% olivine, the whole-rock ages are affected by excess argon, and only the phlogopite ages are interpreted to give geologically relevant ages.

Previous K–Ar determinations of Machotkin *et al.*, (1989) on phlogopites of lamproite dykes from Yakokut gave 142 ± 5 and 147 ± 6 Ma which are similar to the whole-rock ages obtained in this study. Nevertheless, the phlogopite ages of 133.3 ± 1.3 Ma are interpreted to be reliable cooling ages of the lamproitic dykes of Yakokut.

Rb–Sr isotopes. Dissolved Rb and Sr samples were mixed with highly enriched ⁸⁴Sr and ⁸⁷Rb spikes respectively. Rb and Sr were separated from

each other with cation exchange columns and the isotopes were measured by the isotope dilution method. Rubidium isotope determinations were carried out on an 'Ion Instruments' mass spectrometer with 35 cm radius and a three filament ion source and strontium analyses on a VG sector mass spectrometer with 26 cm radius with one filament ion source and a five cup system (Hurford *et al.*, 1986). The correspondence of the ⁸⁷Sr/⁸⁶Sr ratio is better than 0.015 %. As standard for strontium analyses NBS 987 was analysed.

Rb–Sr isotopic compositions of phlogopite, whole-rock and olivine were determined of two samples (AY26 and AY31). The results are presented in Table 9 and Fig. 9a. The resulting isochron ages for both samples of 129.5 ± 1.0 Ma (AY26) and of 127.2 ± 1.0 Ma (AY31) are about 4 Ma younger than the corresponding phlogopite K–Ar ages of 133.3 ± 1.3 Ma, respectively (Tables 8 and 9). As only a few measurements could be carried out, the

TABLE 7. Major, minor (wt.%) and selected trace element (ppm) analyses (XRF-data) of the lamproites from the Yakokut complex including some *REE* data (ICP-analyses). mg# = MgO/(MgO+FeO) l.o.i. – loss on ignition. Fe²⁺ was determined by wet chemistry. K₂O/Al₂O₃ and (K+N)/A – (K₂O+Na₂O)/Al₂O₃ are calculated as molar ratios

Sample	AY76	AY31	AY27	AY98	AY26	AY29	AY99	AY33
SiO ₂	42.23	43.37	43.91	44.11	43.57	41.76	45.04	47.85
TiO ₂	0.31	0.45	0.52	0.41	0.52	0.50	0.71	0.78
Al ₂ O ₃	3.23	4.38	5.02	4.19	5.06	5.03	7.63	8.26
Fe ₂ O ₃	4.86	5.40	11.97	5.70	5.31	6.02	11.60	6.92
FeO	4.81	5.49	n.a.	4.44	5.69	4.76	n.a.	4.59
MnO	0.17	0.19	0.19	0.17	0.19	0.18	0.19	0.19
Mgo	33.16	26.13	24.19	23.75	23.63	22.03	12.57	12.14
CaO	4.83	7.10	7.51	9.00	7.72	7.23	10.21	11.38
Na ₂ O	0.29	0.55	0.35	0.53	0.52	0.50	1.30	0.99
K ₂ O	2.75	3.14	3.27	3.12	3.45	3.41	6.14	4.79
P ₂ O ₅	0.31	0.45	0.52	0.34	0.52	0.51	0.70	0.85
l.o.i.	n.a.	2.90	2.36	n.a.	3.01	7.10	n.a.	0.28
Total	96.95	99.55	99.81	95.76	99.19	99.03	96.09	99.02
mg#	0.873	0.826	n.c.	0.842	0.806	0.822	n.c.	0.726
K ₂ O/Al ₂ O ₃	0.92	0.78	0.71	0.81	0.74	0.73	0.87	0.63
(K+N)/A	1.07	0.98	0.82	1.01	0.91	0.90	1.15	0.82
K ₂ O/Na ₂ O	9.48	5.71	9.34	5.89	6.63	6.82	4.72	4.84
Ni	1201	801	726	731	705	700	293	261
Cr	2233	1874	1618	1531	1585	1529	781	591
Co	109	104	103	89	92	98	67	66
Sc	7	12	14	17	16	13	19	23
Ba	1118	2029	2230	1117	2280	1861	2214	2893
Sr	384	830	854	560	840	3373	1488	1214
Rb	83	90	97	105	100	149	196	162
Zr	58	87	88	75	95	201	140	126
Nb	3	3	3	3	3	2	5	5
Pb	13	20	35	13	19	12	8	29
Th	n.d.	2	2	n.d.	3	n.d.	n.d.	1
La		11.63			14.01			20.17
Ce		24.00			28.48			42.68
Pr		2.86			3.47			5.27
Nd		12.52			15.00			22.64
Sm		2.76			3.38			5.14
Eu		0.84			1.01			1.49
Gd		2.28			2.87			4.32
Dy		1.58			1.89			2.83
Ho		0.24			0.31			0.50
Er		0.63			0.84			1.25
Yb		0.57			0.73			1.06
Lu		0.07			0.09			0.14
Y (ICP)		7.41			8.92			12.98

n.a. – not analysed
n.d. – not detected
n.c. – not calculated

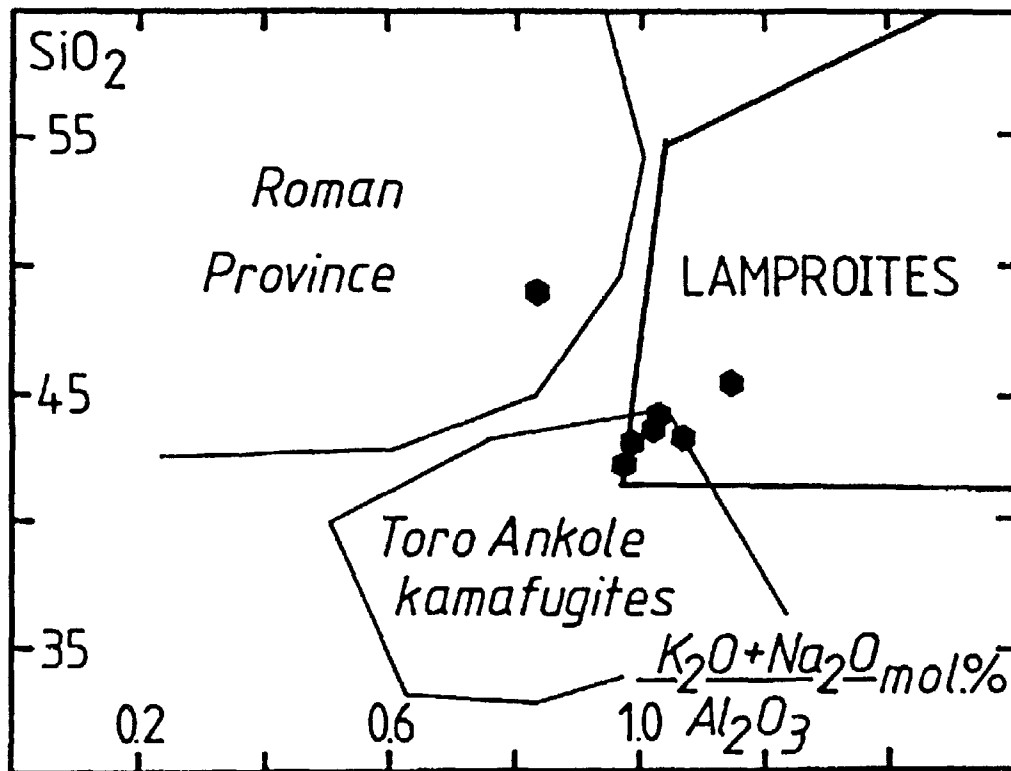


FIG. 7. SiO_2 vs. molar $(\text{K}_2\text{O}+\text{Na}_2\text{O})/\text{Al}_2\text{O}_3$ after Barton (1979) for the distinguish of lamproites, Toro Ankole kamafugites and Roman province rocks.

obtained three point isochrons are of limited value. Nevertheless, as both results are in excellent agreement they give an indication of the reliability of the K–Ar age and accuracy of the $^{87}\text{Sr}/^{86}\text{Sr}$ initial ratios.

The intercepts of both regression lines give $^{87}\text{Sr}/^{86}\text{Sr}$ initial ratios of 0.70572 ± 0.00003 and 0.70575 ± 0.00003 . Both are in excellent agreement with each other (Table 9) and with one value of 0.70572 ± 0.00008 determined by Machotkin (1992). A higher $^{87}\text{Sr}/^{86}\text{Sr}$ initial ratio is shown by the whole-rock value of sample AY98 of 0.70605 ± 0.00001 , which was calculated with the age of 130 Ma obtained by the other samples. To get a $^{87}\text{Sr}/^{86}\text{Sr}$ initial ratio similar to the other determined values of 0.70573, requires an age of 170 Ma for the calculation of the initial Sr ratio. This seems to be unrealistic, because this is much older than the oldest rocks of Yakokut, syenites, showing ages of 160 ± 1.6 Ma (Mues, 1993).

Petrogenetic discussion

The whole set of mineralogical and geochemical data on the ultramafic, ultrapotassic dyke rocks of the Yakokut complex exhibit some almost intermediate features between lamproites and kamafugites. These characteristics include CaO , Al_2O_3 , $\text{K}_2\text{O}/\text{Al}_2\text{O}_3$, and trace element contents. Nevertheless, even if e.g. Ti–Al contents of clinopyroxenes are not as low as in many lamproite occurrences, they are distinctly lower than in kamafugites (Fig. 4). The very high $\text{Cr}/(\text{Cr}+\text{Al})$ ratios of the chromites are as high as typical lamproites and are strong indications for their crystallization from a lamproitic magma.

Additionally, for their alkalinity and $\text{K}_2\text{O}/\text{Al}_2\text{O}_3$ ratios the rocks have too high SiO_2 contents to be classified as kamafugites (Fig. 7). Further lamproitic signatures may include the $\text{CaO}/\text{Al}_2\text{O}_3$ ratios of 1.3–1.6 and lower Sc and high Zr/Nb ratios of 25–30 compared to generally $\text{CaO}/\text{Al}_2\text{O}_3 > 2$ and

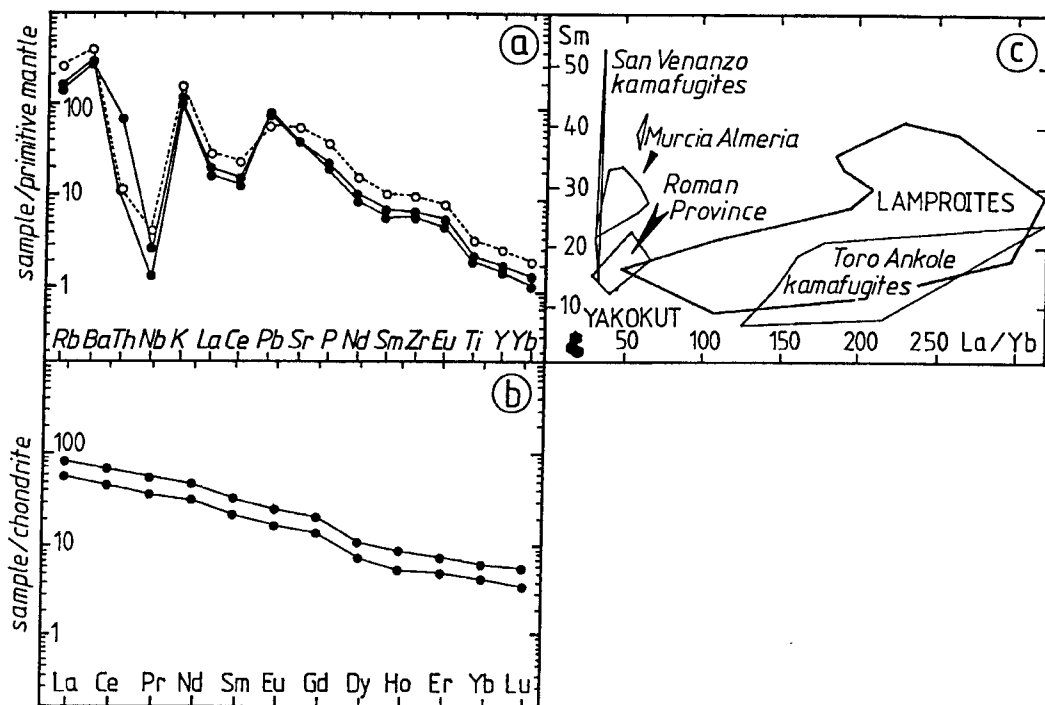


FIG. 8. (a) Patterns of incompatible elements normalized to the primitive mantle values of Sun and McDonough (1989). (b) Chondrite-normalized REE patterns. (c) Sm vs. La/Yb for the Yakokut lamproites and the fields of other potassic rocks for comparison (after Mitchell and Bergman, 1991).

Zr/Nb ratios < 5 in kamafugites (Mitchell and Bergman, 1991). Therefore, we decided as outlined previously to call these rocks lamproites rather than kamafugites, keeping in mind that they are an extreme variety.

It is highly significant in our understanding of the petrogenesis of the Yakokut ultramafic ultrapotassic rocks to estimate the effects of crystal fractionation, accumulation and assimilation. The most magnesian olivines that have been recorded have Fo values of 92–94. The rocks have mg#-numbers of 69.8–83.9, at which these highly magnesian olivines could have crystallized in equilibrium at their liquidus temperature (using a D value for Fe/Mg partitioning between olivine and melt of 0.3–0.33; Roeder and Emslie, 1970). The fractionation and accumulation of olivine seems to have been a dominant influence on whole-rock chemistry of the Yakokut rocks. This is well documented by the chemical trend towards basic compositions with higher SiO₂, Al₂O₃, CaO, Sc and K₂O, Ba, Sr and lower MgO, Cr, Ni and Co. Therefore we shall address the effects of possible olivine accumulation and fractionation in the rocks by concentrating on elemental abundances and ratios

that are unaffected by these processes, such as CaO/Al₂O₃ and La/Yb or the Sr-isotopes.

The ⁸⁷Sr/⁸⁶Sr initial ratios of the Yakokut lamproites (0.7057–0.70605) exhibit a similar variation as reported from the Leucite Hills (0.7057–0.7066; Vollmer *et al.*, 1984), Smoky Butte (0.7083–0.70631) or Sisimiut (0.7045–0.7061; Mitchell and Bergman, 1991). They require the derivation from sources with relatively low Rb/Sr ratios. As lamproites are generally thought to ascend rapidly to the surface, making assimilation unlikely, the isotopic ratios are considered to reflect those of the mantle sources. As the Yakokut rocks are ultramafic the previous assumptions are adopted. The different ⁸⁷Sr/⁸⁶Sr initial ratios of 0.70573 ± 0.00003 and 0.70605 ± 0.00001 imply the derivation of the Yakokut lamproitic magmas from two distinct sources precluding their derivation by (i) fractional crystallization from a common primary magma or (ii) partial melting of a homogeneous mantle source. Bergman (1987) proposed that the Sr isotopic variation within a single province may be caused

TABLE 8. K-Ar data of the ultrabasic lamproitic dyke rocks from Yakokut. Generally m – mesh; only AY26 phlogopite 100–250 μm . phlog – phlogopite, WR – whole rock

Sample	Mineral	Size	% K	Vol ⁴⁰ Ar _{rad} ccSTP/g10 ⁻⁶	% ⁴⁰ Ar	Age (Ma)	
AY 26	WR	35–60m	2.28	12.58	93.59	136.6 \pm 1.3	
		60–90m	2.38	13.31	93.81		
	phlog	60–90m		13.44	93.71	139.1 \pm 1.4	
				39.36	98.91		
		100–250	7.32	39.24	98.69	133.3 \pm 1.3	
AY 27	WR	35–60m	2.20	11.61	94.35		
		60–90m	2.20	11.92	94.95	132.6 \pm 1.3	
	phlog	60–90m		12.12	95.54	136.4 \pm 1.3	
				6.96	37.41	97.92	133.2 \pm 1.3
		35–60m	2.68	15.35	94.77	141.6 \pm 1.4	
AY 31	WR	35–60m	2.39	13.94	94.31	144.1 \pm 1.4	
		60–90m	7.48	40.20	98.44	133.2 \pm 1.3	
	phlog	60–90m		7.64	42.03	98.02	
				41.54	98.87	135.5 \pm 1.3	
	olivine	60–90m		0.102	1.350	83.21	313.7 \pm 3.1

by partial melting of a heterogeneous mantle source with discrete domains of diverse Rb/Sr ratios. To preserve their distinct isotopic characteristics, the lamproitic melts must remain separated during partial melting and/or ascent.

Commonly Nd isotopic compositions of lamproites indicate the derivation from mantle sources that have undergone long-term enrichment in the *LREE*. Nd isotopes in lamproites show low ϵ_{Nd} values varying between -7.4 and -26.0 , suggesting an old source

TABLE 9. Rb-Sr isotopic data of Yakokut lamproitic rocks. The analytical error on the model ages for both samples corresponds to 1.0 Ma. Analytical error is given as $\pm 0.0000xy$ or $\pm 0.000xyz$. WR – whole rock; phlog – phlogopite; oliv – olivine. Generally m – mesh; only AY26 phlogopite 100–250 μm

	Size	Rb ppm	Sr ppm	⁸⁷ Sr/ ⁸⁶ Sr \pm an. error	⁸⁷ Rb/ ⁸⁶ Sr \pm an. error	Age (Ma)
AY 26						
WR		102.66	889.93	0.706333 \pm 18	0.3337 \pm 15	
phlog	60–90m	206.36	116.37	0.715154 \pm 19	5.1344 \pm 83	
phlog	100–250	208.75	125.18	0.714630 \pm 29	4.8280 \pm 307	
				initial 0.70572		129.5
AY 31						
WR		91.45	883.64	0.706249 \pm 22	0.29938 \pm 29	
phlog	60–90m	210.96	116.63	0.715109 \pm 19	5.2370 \pm 478	
phlog	35–60m	204.93	154.54	0.712832 \pm 18	3.8386 \pm 408	
oliv	60–90m	5.13	357.85	0.705821 \pm 47	0.04145 \pm 14	
				initial 0.70575		127.2
AY 98						
WR		111.08	612.82	0.707042 \pm 13	0.52438 \pm 41	
				initial 0.70605		

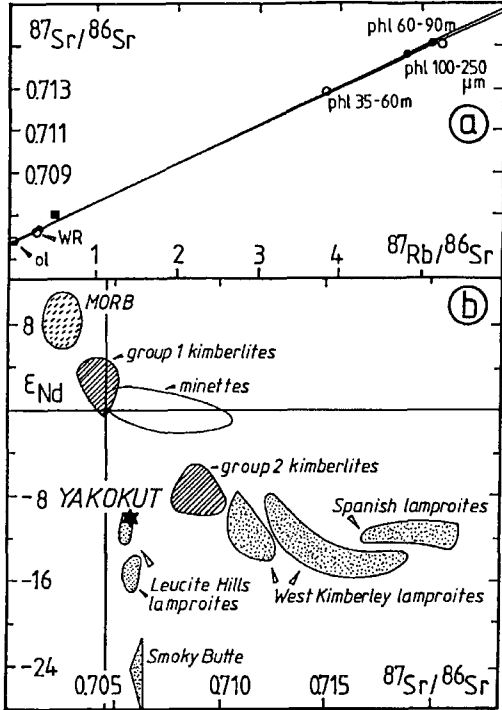


FIG. 9. (a) Diagram for the $^{87}\text{Sr}/^{86}\text{Sr}$ versus $^{87}\text{Rb}/^{86}\text{Sr}$ ratios for the ultrabasic lamproitic dyke rocks from Yakokut. The intercepts of the regression lines as isochrons with the $^{87}\text{Sr}/^{86}\text{Sr}$ -axis give the initial ratios. (b) Diagram of ϵ_{Nd} -values versus $^{87}\text{Sr}/^{86}\text{Sr}$ initial ratios for lamproitic rocks from Yakokut (data from Machotkin, 1992) in comparison with several lamproite provinces (Mitchell and Bergman, 1991) and MORB (Ito *et al.*, 1987).

enriched in *LREE* compared to Bulk Earth with low Sm/Nd ratios (McCulloch *et al.*, 1983; Mitchell and Bergman, 1991; Vollmer *et al.*, 1984; Fraser *et al.*, 1985; Nelson *et al.*, 1986). For the Yakokut rocks only one $^{143}\text{Nd}/^{144}\text{Nd}$ ratio of 0.512101 ± 0.000037 corresponding to an ϵ_{Nd} value of -10.2 ± 0.7 of Machotkin (1992) is available and shown in a ϵ_{Nd} value versus $^{87}\text{Sr}/^{86}\text{Sr}$ initial ratio diagram (Fig. 9b) in comparison with data from several lamproite provinces. The Sr–Nd composition of the Yakokut sample is similar to that of madupitic lamproites from the Leucite Hills (Vollmer *et al.*, 1984), which is explained by the contribution of an old, *LREE* enriched source without increase of the Rb/Sr ratio.

Any attempt to outline the source characteristics of the Yakokut lamproitic rocks of the Aldan Shield must account for the following observations:

(1) The ultramafic, ultrapotassic rocks are characterized by low contents of SiO_2 , Al_2O_3 , CaO, Na_2O and Sc, and exhibit a wide range of Cr, Ni, Ba and Sr abundances.

(2) The rocks have high concentrations of incompatible elements and are moderately enriched in *LREE* relative to *HREE*.

(3) On normalized multi-element plots, the rocks are distinctly depleted in Nb and La–Ce, and to a lesser extent Ti, relative to other incompatible trace elements.

(4) The corresponding magmas are derived from a source with time-integrated Rb/Sr and Sm/Nd ratios that are respectively slightly higher and lower than Bulk Earth. The source has undergone long-term enrichment in *LREE*.

The silica-undersaturated nature, fractionated *REE* patterns and high abundances of the incompatible elements of the Yakokut rocks suggest that either their parental magmas or their precursor melts or metasomatizing fluids were the products of small degrees of partial melting. As the Sr and Nd isotopic ratios are outside the range of oceanic magmas (Fig. 9b), a large contribution to the magmas must have come from the subcontinental lithospheric mantle. In order to generate melts with high abundances of *REE* and large-ion-lithophile (LIL) elements, the mantle source of the Yakokut lamproitic rocks must have been previously enriched (e.g. Vollmer and Norry, 1983). Mantle metasomatism may be caused by (i) volatile-rich, low-viscosity melts that have migrated more or less continuously from the asthenosphere and accumulated in the subcontinental mantle (McKenzie, 1989) or, (ii) melts or fluids derived from a subducting slab (Pearce, 1983).

The characteristic geochemical feature of subduction-related magmas, such as relative depletions in high-field-strength elements, especially Nb, is observed on normalized multi-element plots of the Yakokut rocks (Fig. 8a), suggesting that mantle enrichment was associated with fluids or melts derived from a down-going slab. This metasomatism seems to have been caused by fluids rather than by melts, as fluids better dissolve and transport the LIL elements, whereas the HFS elements and *REE* seem to be best transported in melts (Pearce, 1983). This may explain the relatively low at La–Ce in the multi-element plots.

As outlined above the Yakokut rocks have intermediate to low contents of CaO and Al_2O_3 compared to kamafugites and lamproites, and lower CaO/ Al_2O_3 (1.3–1.6) ratios than generally observed in kamafugites (< 2). *LREE* abundances are moderately enriched and La/Yb ratios are low compared to most kamafugites and lamproites (Fig. 8c). As these abundances cannot readily be explained by crystal fractionation or accumulation, they must

reflect variations in their mantle source compositions. The intermediate to low CaO, Na₂O and Sc and moderately *LREE* abundances suggest that their parental magmas were derived from a mantle source that was depleted in clinopyroxene and in 'basaltic' elements.

Olivine lamproites are widely believed to represent melts derived from a metasomatized mica harzburgite, whereas kamafugites are generally thought to come from a clinopyroxene-rich source. The latter source type is not compatible with the element abundances observed in the Yakokut rocks. It is therefore assumed that these unusual lamproite geochemical features of the Yakokut rocks are due to their distinct tectonic setting compared to most lamproite occurrences.

Generally lamproite magmatism is restricted to continental environments and occurs in the outer parts of old stable cratons as within continental plate magmatism (Mitchell and Bergman, 1991; Wilson, 1989). Most lamproites, with the exception of those from the Leucite Hills and Smoky Butte, show a shallow Nd-Sr trend (Fig. 9b) along with ultrapotassic rocks from the Roman Province and the kamafugites. These sources are characterized by low Sm/Nd but both high and low Rb/Sr. Lamproites of this trend exhibit therefore low Rb/Ba and K/Ti ratios, suggesting that their source has been metasomatized by the introduction of small-volume partial melts. As the Yakokut lamproites do not belong to this trend, this clearly contrasts with the earlier stated influence of subduction-related fluids for the Yakokut source. The Nd isotopic composition requires an early *LREE* enrichment of the subcontinental lithospheric mantle source. In this region, this can be correlated to the consolidation of the Aldan Shield during the Archaean to Proterozoic. The general enrichment of LIL elements is much younger, probably younger than 200 Ma. The widespread large scale potassic alkaline magmatism in Central Aldan, including the lamproitic dykes of Yakokut is related to the collision between the SE part of the Siberian platform, the Aldan Shield, and the Burein block. This is caused by the ongoing northwesterly directed subduction in the Ochotsk-Chukotsk belt farther in the east (Fig. 1a). It is suggested that the subduction slab or detached pieces of the slab reached and influenced the subcontinental lithosphere underneath the Aldan Shield obscuring the characteristic 'normal intraplate' signature as shown by many lamproite occurrences.

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