Canadian Mineralogist
Vol. 26, pp. 23-43 (1988)

ROCK AND MINERAL CHEMISTRY OF THE LINHAISAI MINETTE, CENTRAL KALIMANTAN, INDONESIA, AND THE ORIGIN OF BORNEO DIAMONDS

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ABSTRACT

Exploration for the primary source of diamonds in central Kalimantan led to the discovery of 8 Ma, small, shallow, non-diamondiferous mafic to felsic minette dykes. These dykes intrude Early Tertiary, non-marine, interbedded mudstones and sandstones and Late Cretaceous marine mudstones in the upstream Karamu River area, northern Barito Province, Indonesia. The dykes are vesicular, massive and contact-brecciated magmatic rocks that range from mafic to intermediate compositions. Their mineralogy includes moderately zoned phlogopite (core to rim: Mg# 82→73); pargasitic hornblende with two zones, a euhedral Fe-rich, green, sieve-textured core zone [Mg# 70→47 (c→r)] and a Mg-rich, brown rim zone [Mg# 65→71 (c→r)]; clinopyroxene zoned from augite core (Mg# 80) to diopside (salite variety) rim (Mg# 72), sanidine ($Ab_{58}Or_{37}An_5$), titaniferous magnetite, ilmenite, serpentine pseudomorphs after olivine, apatite, rutile, and zircon. Rock compositions are typical of minettes in terms of major-, minor-, and trace-element abundances. The primitive nature of the Linhaisai minette (high Mg#, elevated Ni, Cr, Co and Sc contents) indicates a mantle derivation. However, the wide range in Mg# (50–73), SiO_2 (47–59 wt.%), and alkali (3.6–8.5 wt.%) contents and systematic chemical variation all indicate a significant amount of differentiation (or contamination); magma mixing is supported by the uniquely zoned amphiboles. The elevated *LIL*-element contents (Ba, Zr, Rb, *REE, etc.*) in the more mafic members of the suite suggest an enriched mantle source. Despite some similarities with kajanite and cocite, the Linhaisai minette is grouped with typical minettes. In addition, the Linhaisai minette bears no lamproite affinities beyond those of typical minettes. We suggest that the alluvial diamonds of central Kalimantan owe their surficial existence to an unidentified lamproite or kimberlite in Borneo or adjacent southeast Asia.

Keywords: lamprophyre, minette, major- and trace-element geochemistry, petrology, mafic magmas, mineral chemistry, geochronology, Kalimantan, Indonesia, diamonds.

SOMMAIRE

Dans le cadre d'une recherche de la source primaire de diamants au Kalimantan central, en Indonésie, nous avons découvert de petites dykes non-diamantifères de minette mafique à felsique, mis en place récemment (8 Ma) à faible profondeur. Ces dykes recoupent des mudstones et des grès interlités non-marins d'âge Tertiaire précoce et des mudstones marins d'âge Crétacé tardif dans la région d'amont de la rivière Karamu, dans le nord de la province de Barito. Les roches sont vésiculaires, massives et bréchifiées, et varient d'une composition mafique à intermédiaire. Elles contiennent: phlogopite à zonation modérée, allant d'une valeur du Mg# [rapport 100Mg/(Mg + Fe)] de 82 à 73 du coeur à la bordure, hornblende pargasitique montrant deux zones, un coeur idiomorphe vert, riche en fer, à texture en tamis (Mg# du coeur 70, de la bordure 47), et une bordure brune plus magnésienne [Mg# de 65 (intérieur) à 71 (bordure)], clinopyroxène zoné à coeur d'augite (Mg# 80) et bordure de diopside au variété de salite (Mg# 72), sanidine ($Ab_{58}Or_{37}An_5$), magnétite titanifère, ilménite, serpentine (en pseudomorphose de l'olivine), apatite, rutile et zircon. Les compositions globales sont typiques de minettes en termes des éléments majeurs, mineurs et en traces. Le caractère primitif de la minette de Linhaisai (Mg# élevé, hautes teneurs en Ni, Cr, Co et Sc) est une indication de sa provenance du manteau. Toutefois, les intervalles en Mg# (50–73) et en concentrations de SiO_2 (47–59%) et des alcalins (3.6–8.5%), ainsi que les variations chimiques systématiques, indiquent une différenciation (ou une contamination) importante. Un mélange de magmas semble indiqué par la zonation particulière de l'amphibole. Les concentrations élevées en éléments incompatibles à large rayon (Ba, Zr, Rb, terres rares, *etc.*) dans les membres relativement mafiques de l'association font penser à un manteau enrichi comme source. Malgré quelques ressemblances à la kajanite et la cocite, on regroupe la minette de Linhaisai avec les minettes typiques. De plus, elle ne montre pas plus d'affinité avec les lamproïtes qu'une minette typique. Les diamants des alluvions du Kalimantan central seraient reliés à un gisement lamproïtique ou kimberlitique non identifié du Borneo ou des régions contiguës du Sud-Est de l'Asie.

(Traduit par la Rédaction)

Mots-clés: lamprophyre, minette, chimisme (éléments majeurs, mineurs) pétrologie, magmas mafiques, chimisme des minéraux, géochronologie, Kalimantan, Indonésie, diamants.

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INTRODUCTION

A minette is a biotite–K-feldspar-rich "shoshonitic" or "calc-alkaline" lamprophyre (Wimmenauer 1973, Velde 1969, Rock 1977, 1984). Minettes generally are regarded as being derived from the mantle (*e.g.*, Kay & Gast 1973, Jahn 1973, Bachinski & Scott 1979, 1980, Jahn *et al.* 1979), although their genesis is the subject of ongoing vigorous debate (*e.g.*, Velde 1971, Němec 1973, 1975, Rock 1980, 1984, 1987).

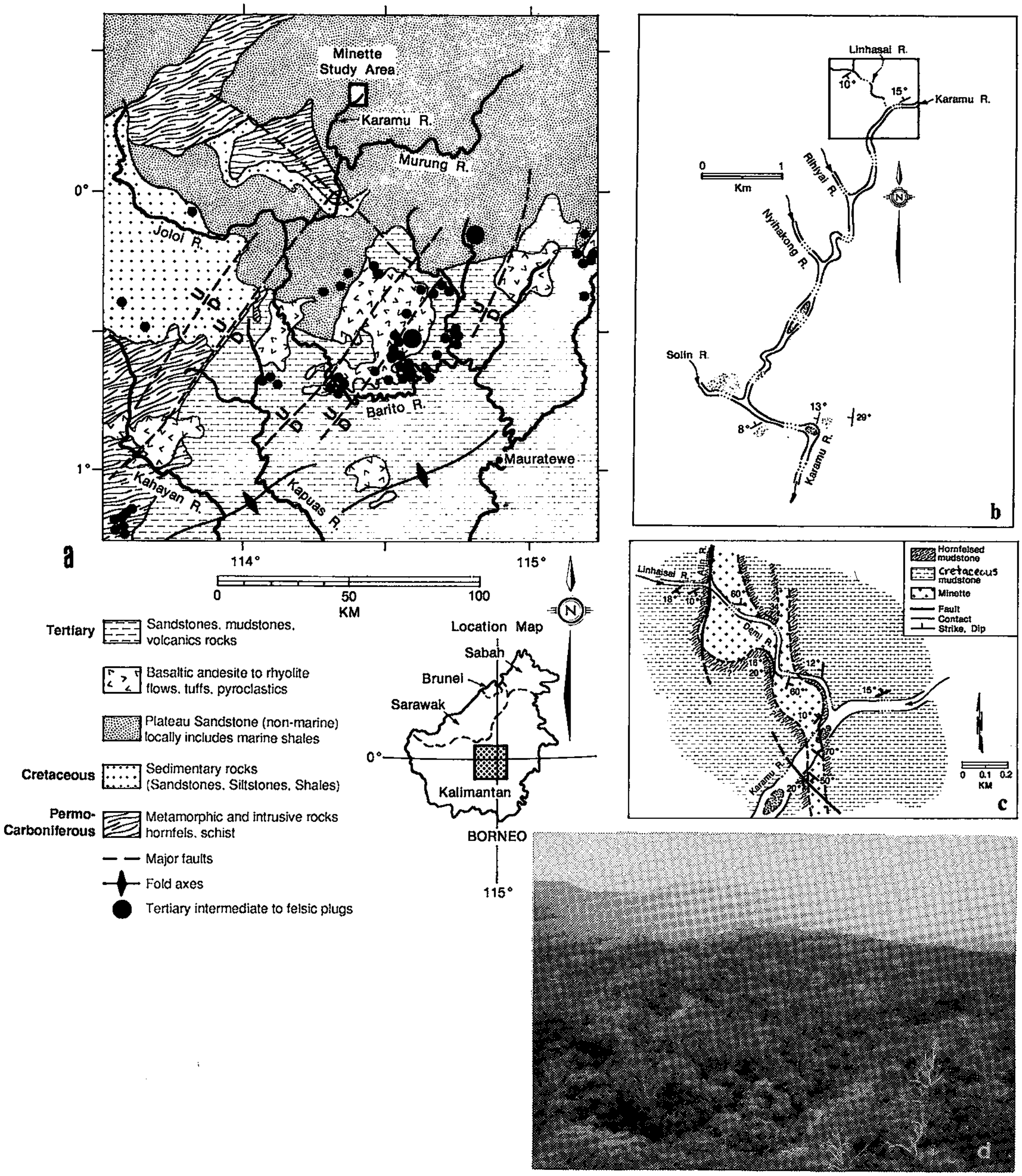

FIG. 1. Geological and location maps showing the Linhaisai minette. (a) Regional geological map (based on van Bemmelen 1949, Zeylmans van Emmichoven 1939, and unpublished Anaconda and C.P. Anekatambang maps). (b) Local location map showing the occurrences of minette subcrop and outcrop in the Linhaisai and Karamu and River areas. (c) Local geological map showing the Linhaisai minette dyke, the best-discovered exposure in the area. (d) Oblique aerial photograph of the obscured Linhaisai River at the LM.

This study contributes petrological and geochemical data on the Linhaisai Minette (LM) suite from Central Kalimantan, Indonesia, which indicate a complex petrogenesis of mantle-derived melts. The 8-Ma LM is among the youngest minettes reported in Rock (1984), second only to those of the Karakorum district of Pakistan (Desio 1979).

Alluvial diamonds have been recovered in Kalimantan (literally translated as "River of Gold and Diamonds") for over 1500 years (Koolhoven 1935). These diamond deposits occur chiefly in three provinces: the Landak region of West Kalimantan (Witkamp 1932), the Barito Province of Central Kalimantan (Hovig 1930) and in the Meratus Range region of southwest Kalimantan (Koolhoven 1935, and references cited therein). The primary source for these diamonds has been an enigma for some time, although most authors have regarded the Pamali breccia of southeast Kalimantan as a kimberlite and the "Borneo-type" source rock for the alluvial diamonds (*e.g.*, Bardet 1977, Orlov 1977). Since the recent reinterpretation of the Pamali breccia as a sedimentary conglomerate with local ophiolite provenance (Bergman *et al.* 1987), the Borneo diamond saga has returned to an unresolved status in the field of diamond genesis.

The LM was discovered by the third author durint a search for the elusive primary source of alluvial diamonds in Central Kalimantan. Minette cobbles were traced upstream to their parent outcrop during routine stream-sediment sampling. This discovery was significant because, as discussed later, minettes are closely related to lamproites, which are now recognized as a well-established, major, primary source of diamonds in Australia and the U.S.A. (*e.g.*, Jaques *et al.* 1984, Scott-Smith & Skinner 1984). The LM occurs within 150 km of the type locality of kajanite, an olivine-leucite-nepheline-phlogopite-richterite-magnetite-chromite rock (from the Kajan River) initially reported by Brouwer (1909) and later discussed by Lacroix (1926) and Wagner (1986). This kajanite possesses some compositional and mineralogical similarities with lamproites (Bergman 1987, Wagner 1986) but is not a lamproite *sensu stricto* because of the presence of nepheline and anomalous mineral compositions. Despite some compositional similarities with lamproites, the LM possesses no more affinity to lamproite than does any other typical minette. This paper discusses the age, mineralogy, geochemistry and petrogenesis of the LM and places it in the context of magmatic activity in Kalimantan, as well as in the context of the setting of minettes from elsewhere in the world.

GEOLOGICAL CONTEXT

Regional geological and tectonic setting

The LM occurs in the upstream Karamu River area in the northern Barito Province of Central Kalimantan at 0°20′N, 114°53′E (Fig. 1). The local elevation is about 200 m but rises rapidly in the high-gradient stream system to a 1200-m drainage divide. This portion of Central Kalimantan forms the southeastern margin of the Sundaland continental craton, locally formed by Permo–Carboniferous to Triassic metamorphic and granitic rocks and younger rocks (*e.g.*, Umbgrove 1938, van Bemmelen 1949, Hamilton 1978). The region is bound to the north by a Tertiary subduction complex of the Baram, Lupar and Rajang Groups (ophiolite, radiolarian cherts, greywackes and argillites in a structurally imbricated complex of southward-dipping thrust blocks). The Lupar chert–ophiolite line, the Rajang Group (mélange) and Late Cretaceous to Eocene, intermediate to acid, calc-alkaline volcanic and plutonic rocks form three parallel belts that have generally been regarded as defining an Early Tertiary southwestward-dipping ocean–continent Early Tertiary convergent margin in Borneo (Katili 1971, 1973, Haile 1968, Hutchison 1973, Hamilton 1978, Williams & Heryanto 1985) which is continuous with the East Asian (China–Indochina) complex of similar age (Taylor & Hayes 1982). The southward-dipping subduction zone is also consistent with the zonation of Cretaceous to Miocene calc-alkaline igneous rocks; tonalites, diorites and granodiorites are most abundant in the northern Schwaner Mountain area and alkaline granites and monzonites are most abundant to the south (Sudradjat 1976, Batan 1976, Hamilton 1978).

The LM occurs just east of the Müller Mountains on the central portion of the Kuching zone of van Bemmelen (1949) and Haile (1974). The Kuching zone records a complex history of late Paleozoic, Mesozoic and Cenozoic sedimentation, folding and magmatism. Widespread Tertiary sedimentation occurred in a shallow marine environment, which was later subjected to the intrusion and extrusion of over 400 km^3 of magmatic rocks in the mid-Tertiary and over 800 km^3 in the latest Tertiary (Rose & Hartono 1978, Haile 1974). These Tertiary deposits have been deformed by several fold-systems whose axial trends range from northerly to east-northeasterly in the area of the LM.

Local setting

The local geology is characterized largely by pre-Cretaceous metamorphic rocks, Cretaceous marine shales and sandstones, and the early to middle Tertiary Plateau Sandstone (*e.g.*, Koolhoven 1935, van Bemmelen 1949). The most extensive country rock is the Oligocene Plateau Sandstone. Cenozoic andesitic to basaltic volcanic rocks (plugs, necks and flows) are widely dispersed immediately south of the LM area (Fig. 1).

The LM locally intruded northeasterly striking, gently dipping, upper Cretaceous, interbedded marine mudstones and sandstones (Horowu Formation?). The Cretaceous strata are exposed only locally, where the more resistant and thick Plateau Sandstone has been eroded. This complex (and unmapped) outcrop pattern is not apparent in Figure 1a, which shows the area covered by the more extensive Plateau Sandstone that overlies the Cretaceous strata. Also not shown in Figure 1a are scattered exposures of the Late Paleozoic metamorphic rocks occurring as windows in the Plateau Sandstone. These rocks are best exposed in low-lying stream valleys where the overlying Cretaceous and Tertiary strata have been removed by erosion.

Where massive, the intrusive LM bodies generally form positive topographic features, although they are locally extensively weathered near contacts and form topographic depressions. The intrusive rocks comprise several dykes that crop out within a 5 × 5 km area; the largest bodies discovered are cut by the Karamu River and its tributary, the Linhaisai River. The macroscopic textures are extremely variable; the rock ranges from medium-grained massive to fine-grained vesicular or brecciated. Well-developed columnar jointing characterizes one section of the Linhaisai River exposure. Most exposed intrusive contacts display discordant breccia-rich zones, although a few contacts are sill-like and possibly concordant in nature. One well-exposed dyke has thermally altered the nearly unconsolidated mudstone wallrock to shale within 0 to 2 meters of the intrusive contacts.

The intrusive rocks are extremely hard where fresh; some moderately weathered zones display a patchy texture reminiscent of the globular structures or ocellar textures characteristic of all lamprophyres (*cf.* Rock 1984, Cooper 1979). Extremely weathered zones are represented by a dark green-brown mica-rich soil. The local emplacement of the minette bodies was structurally controlled by a series of small northerly trending faults. Parts of the dykes display veining by a more felsic pegmatitic material that forms both sharp and gradational contacts with the minette. These vein contacts are characterized by abundant, zoned, coarse-grained amphibole and feldspar.

TABLE 1. K-AR AGE DATA ON PHLOGOPITE SEPARATES* FROM THE LINHAISAI MINETTE

Sample**	K (wt.%)	^{40}Ar rad (x 10" mol/g)	^{40}Ar rad / ^{40}Ar total	Age (Ma)
K-4	6.05	8.26	0.414	7.8±0.3
K-7	6.51	8.76	0.348	7.7±0.2

* The analyses were performed in the geochronology laboratories of the Department of Mineralogy and Geology, Ohio State University, Columbus, by K.A. Foland. ** Each analysis represents an average of two determinations.

Late Cenozoic magmatism in north-central Kalimantan

Van Bemmelen (1949) defined a NE-trending 150-km-wide belt of Neogene–Quaternary volcanic rocks that bisects Borneo. In the upper Barito and Kapuas River region, this belt consists of Neogene volcanic rocks of the (from NE to SW) Nieuwenhuis and Iran Mountains, Upper Kapuas Mountains, Müller Mountains and Madi uplands. The belt is bounded to the southeast by Quaternary volcanic centers at Murai, Beluh, and Bawang Aso. Although detailed geochronological work has not been conducted on the upper Barito River Tertiary volcanic centers and plutonic complexes, the rocks are lithologically and structurally similar to rocks that have been studied to the west and north (Bergman, unpublished data). Williams & Heryanto (1985) described the widespread 16–30-Ma calc-alkaline Sintang intrusive bodies, 200–300 km to the west of the LM (mafic to felsic high-level stocks of diorite, granodiorite and granite, and sills of basalt, andesite and dacite). Kirk (1968) discussed a 16 - 19-Ma calc-alkaline intrusive suite in Sarawak, located more than 150 km to the north of the LM. The widespread Tertiary volcanic rocks to the south of the LM include the following recognized units: the Habaon andesite porphyry, the Masuparia Volcanics (andesite to dacite flows and volcaniclastic rocks), the Kuatan and Lahung Volcanics (basalt flows, tuffs, and volcaniclastic rocks), the Sian Andesite, the Kasale Basalt, Tupuh Volcanics (flow and volcaniclastic rocks) (Metal Mining Agency of Japan 1979, Supriatna & Rustandi 1981). The Masuparia and Tupuh systems possess epithermal-type gold mineralization. Williams & Heryanto (1985) attributed the upper Tertiary to Quaternary magmatism to post-subduction melting due to crustal thickening and increased heat-flow.

Although lamprophyre dykes had not been found previously in the vicinity of the LM in Central Kalimantan, earlier Dutch workers systematically mapped their widespread distribution in western Kalimantan — *e.g.*, the minette, kersantite, spessartite, camptonite and vogesite dykes noted by van Es (1918), Zeylmans van Emmichoven (1939) and Zeylmans van Emmichoven & ter Bruggen (1935).

AGE

Conventional ^{40}K–^{40}Ar methods were used to date phlogopite separates from two unaltered rocks (Table 1). The samples yielded concordant ages of 7.7 and 7.8 (±0.3 Ma). The intrusive age of the LM is limited by field relationships to post-Oligocene (<30 Ma) on the basis of the age of the host sedimentary rocks. An isotopic age of late Miocene is therefore consistent with the geologic cross-cutting relationships. This age is similar to that of the

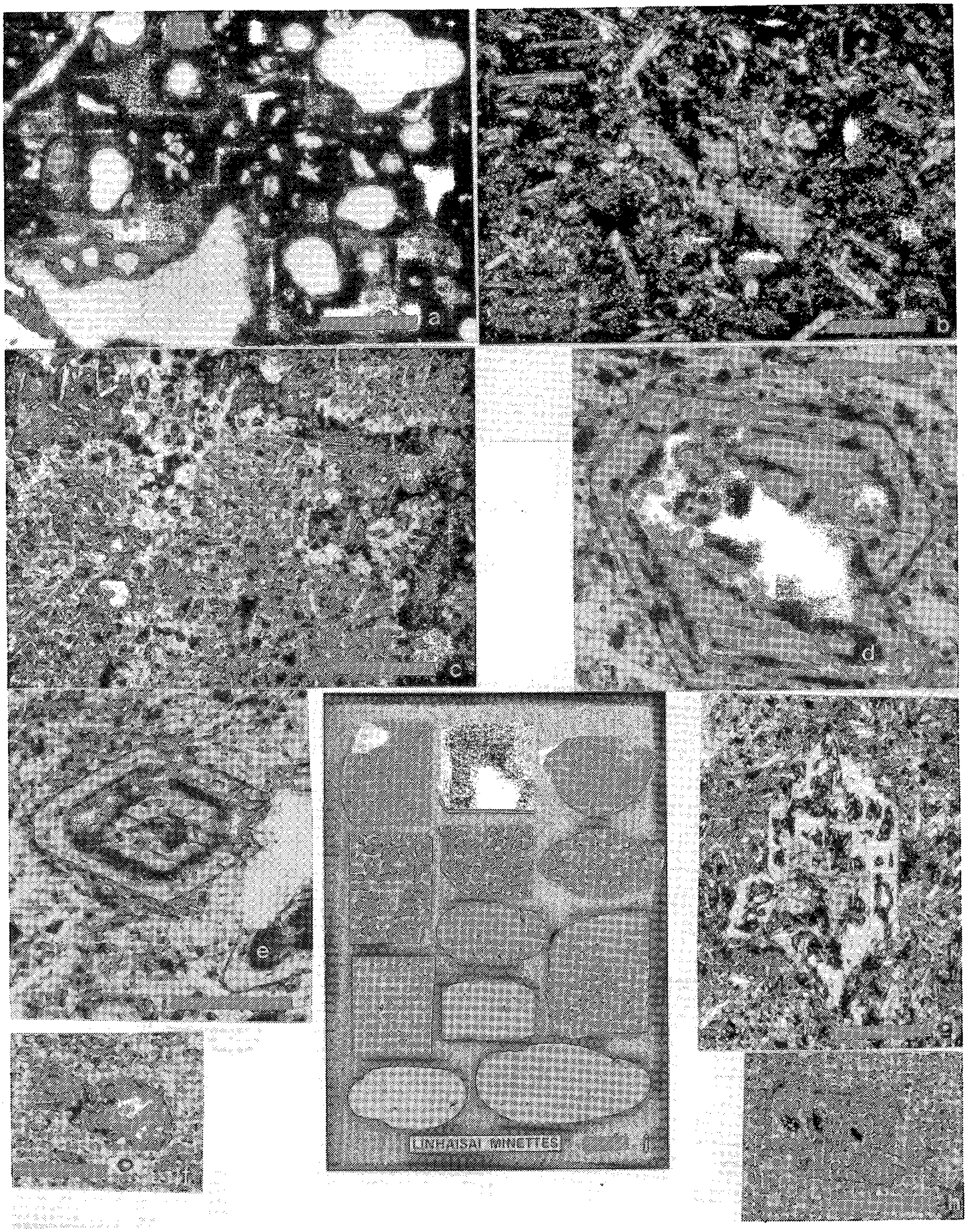

FIG. 2. Photomicrographs and photomacrographs of various Linhaisai minette samples. (a) Photomicrograph of K-1 showing vesicularity, glassy groundmass and quench crystals of augite and olivine. (b) Photomicrograph of K-5 showing abundant phlogopite laths. (C) Photomicrograph of K-2 showing ocellar texture. (d–f) Amphibole phenocrysts in specimen K-4. (g) Olivine phenocryst pseudomorph in K-SR. (h) Olivine phenocryst pseudomorph in K-5-1. (i) Photograph of slabs of the LM showing the varied textures. Scale bars: (a–c, f, g,h) 1 mm, (d,e) 0.5 mm, (i) 5 cm.

TABLE 2a. REPRESENTATIVE ELECTRON-MICROPROBE COMPOSITIONS OF LINHAISAI MINETTE PHASES

											Phlogopite											
	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22
SiO_2	38.5	40.1	38.7	38.9	38.1	38.5	38.7	39.2	37.8	38.6	38.4	38.9	37.5	40.3	38.6	38.5	36.9	39.1	38.1	38.7	39.1	39.2
TiO_2	4.05	4.00	4.07	4.82	4.31	4.06	3.68	5.05	5.24	4.21	4.20	4.10	7.47	3.21	3.63	3.05	7.30	3.93	7.33	3.04	3.56	3.11
Al_2O_3	12.7	12.5	13.9	14.0	13.7	13.8	13.6	13.3	13.3	13.9	14.3	14.5	13.3	13.7	13.7	14.2	14.3	13.4	14.6	14.1	13.7	14.0
Cr_2O_3	.50	.45	.39	.03	.21	.43	.41	.18	.00	.42	.38	.43	.00	.92	.10	2.08	.00	.01	.01	2.03	.15	1.73
FeO^T	11.6	8.80	8.43	9.10	9.20	8.16	7.97	8.87	10.1	8.28	8.70	8.32	10.3	4.96	7.51	4.65	10.5	7.76	9.77	4.80	8.04	4.61
MnO	.13	.00	.02	0.5	.06	.03	.03	.15	.16	.03	.10	.04	.16	.00	.09	.00	.15	.07	.17	.00	.08	.00
MgO	17.4	19.6	19.4	17.5	18.3	19.7	19.9	19.3	18.0	19.2	19.1	19.6	16.7	22.4	21.1	21.9	16.3	20.6	15.7	21.5	20.7	21.9
CaO	.04	.01	.04	.28	.03	.00	.03	.01	.15	.00	.00	.00	.02	.02	.00	.01	.03	.03	.15	.00	.03	.01
K_2O	9.54	9.92	8.91	8.09	8.50	8.78	8.73	8.49	8.64	8.92	9.20	9.55	9.07	9.90	9.40	9.97	8.36	9.19	8.22	9.31	9.10	9.61
Na_2O	.16	.15	.56	1.00	.66	.62	.54	.72	.78	.56	.70	.59	.70	.44	.66	.35	.66	.55	.77	.42	.59	.38
F	.72	.85	1.14	1.80	1.60	.79	.75	2.09	2.16	.57	1.32	.83	1.06	.86	1.76	.77	1.07	.86	1.06	.73	1.51	.92
Cl	.01	.02	.02	.03	.00	.01	.00	.03	.01	.02	.01	.02	.02	.01	.02	.00	.03	.01	.04	.00	.02	.01
Ba	.08	.09	.17	.26	.11	.16	.07	.28	.22	.14	.10	.15	.00	.00	.00	.00	.06	.00	.00	.00	.00	.00
Σ	95.4	96.5	95.7	95.9	94.8	94.9	94.4	97.6	96.5	94.8	96.4	96.9	96.3	96.7	96.6	95.4	95.7	95.5	95.9	94.6	96.5	96.5
Mg#	72.5	79.9	80.3	77.3	77.9	81.0	81.6	79.2	75.7	80.5	79.4	80.7	74.3	88.9	83.4	89.3	73.2	82.4	73.8	88.9	81.9	89.4

Data in wt.%. Mg#: 100*Mg/(Mg+Fe^T) (atomic proportions).

Notes: analyses 1-12 from sample K-4, 13-22 from K-5, (1) rim and (2) core of 150 x 550 µm plate, (3) margin of 75x500 µm grain (adjacent to cpx analyses #1-2 - Table 2c), (4) rim and (5) core of 25 x 300 µm plate, (6) rim and (7) core of 60 x 400 µm plate core of 20 x 200 µm plate, (9) rim and (10) core of 120 x 500 µm plate, (11) rim and (12) core of 20 x 250 µm plate, (13) rim and (14) core of 100 x 500 µm plate, (15) rim and (16) core of 130 x 940 µm plate, (17) rim and (18) core of 100 x 500 µm plate, (19) rim and (20) core of 130 x 700 µm plate, (21) rim and (22) core of 160 x 330 µm plate.

TABLE 2b. REPRESENTATIVE ELECTRON-MICROPROBE COMPOSITIONS OF LINHAISAI MINETTE PHASES (CONT'D)

									Amphibole									
	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18
SiO_2	42.4	40.5	40.0	43.1	41.4	41.4	43.0	43.1	44.7	42.8	42.4	40.8	43.5	42.9	42.3	41.5	42.9	41.5
TiO_2	2.88	3.16	2.63	2.34	3.40	2.40	2.81	2.88	2.07	2.91	3.11	2.13	2.75	2.76	1.29	1.61	2.81	2.45
Al_2O_3	11.4	11.0	13.6	11.7	11.0	12.1	10.0	10.5	12.2	10.3	11.0	12.6	9.87	10.3	11.2	11.4	10.5	11.1
Cr_2O_3	.00	.00	.00	.01	.03	.04	.03	.03	.34	.03	.02	.00	.05	.00	.03	.00	.00	.03
FeO^T	11.6	10.7	16.0	12.4	11.7	14.1	11.1	11.2	9.90	11.1	11.1	15.5	10.7	12.5	15.9	17.6	11.0	14.9
MnO	.08	.13	.20	.14	.15	.19	.10	.14	.13	.12	.11	.22	.12	.10	.25	.28	.10	.21
MgO	14.0	14.9	10.5	12.9	14.2	11.8	14.9	14.8	15.5	14.7	14.2	10.8	14.8	14.0	11.3	10.2	14.7	12.1
CaO	11.4	11.1	10.6	10.3	11.5	11.2	11.2	11.4	10.2	11.4	11.2	11.1	11.4	11.2	11.4	11.4	11.5	11.3
K_2O	1.17	1.42	1.02	1.20	.99	1.26	1.13	1.18	.30	1.10	1.20	1.02	1.16	1.11	1.35	1.36	1.15	1.50
Na_2O	2.16	2.26	2.38	2.46	2.24	1.98	2.25	2.16	2.07	2.16	2.23	2.34	2.37	2.25	1.89	2.00	2.16	2.04
BaO	.02	.00	.00	.00	.00	.00	.00	.02	.03	.00	.00	.00	.01	.00	.00	.04	.03	.04
F	.25	.32	.08	.40	.69	.30	.28	.33	.08	.45	.43	.54	.81	.27	.31	.26	.44	.30
Cl	.03	.00	.06	.01	.02	.03	.04	.00	.03	.02	.02	.08	.03	.01	.05	.06	.01	.04
Σ	97.5	97.5	97.1	97.1	97.3	96.8	96.9	97.8	97.6	97.1	97.0	97.0	97.5	97.4	97.2	97.6	97.3	97.5
Mg#	68.2	70.9	53.6	64.6	68.1	59.5	70.3	69.9	73.4	70.1	69.2	55.2	70.9	66.4	55.6	50.2	70.3	58.8
Σ^*	16.868	16.936	16.890	16.849	17.015	16.878	16.898	16.597	16.910	16.899	16.987	16.987	16.994	16.930	16.917	16.948	16.919	16.967

Data in wt.%.

*per O=23 (including OH=1) Notes: all analyses are from sample K-4. (1) margin and (2) interior parts of brown rim zone, (3) margin and (4) interior parts of green core zone of 500 µm euhedral grain, (5) margin of rim zone, (6) margin and (7) core of green core zone of 500 µm grain aggregate, (8) rim margin and (9) interior core zone of euhedral 150 µm grain, (10) rim margin of 400 µm euhedral grain (11) margin and (12) interior of brown rim zone of 200 µm euhedral grain, (13) margin and (14) interior parts of brown rim zone, (15) margin and (16) interior parts of green core zone of 500 µm euhedral grain, (17) rim and (18) core of 250 µm euhedral grain.

TABLE 2c. REPRESENTATIVE ELECTRON-MICROPROBE COMPOSITIONS OF LINHAISAI MINETTE PHASES (CONT'D)

					Clinopyroxene							
	1	2	3	4	5	6	7	8	9	10	11	12
SiO_2	50.6	51.2	49.8	51.2	51.0	50.4	50.1	51.4	49.7	53.6	49.3	52.6
TiO_2	1.01	1.02	1.16	.87	.95	.93	1.10	.89	1.34	.37	1.55	.50
Al_2O_3	3.62	3.72	4.40	2.94	3.62	3.37	4.03	3.19	3.18	1.20	3.72	1.51
Cr_2O_3	.50	.81	.00	.07	.01	.09	.04	.08	.01	.35	.00	.43
FeO^T	6.87	6.69	8.49	7.39	8.37	7.63	9.25	7.34	6.75	3.87	6.70	4.11
MnO	.18	.13	.22	.14	.29	.12	.16	.11	.13	.09	.15	.09
MgO	14.9	15.3	13.7	15.0	14.5	14.6	13.6	15.8	15.3	18.4	14.9	17.3
CaO	20.9	20.8	21.7	21.2	20.0	21.0	21.1	20.8	22.4	21.2	22.4	21.5
K_2O	.00	.00	.03	.01	.18	.02	.01	.02	.03	.01	.03	.02
Na_2O	.60	.59	.49	.52	.53	.60	.63	.49	.39	.37	.47	.39
Σ	99.1	100.3	100.1	99.3	99.5	98.9	100.1	99.9	99.2	99.5	99.4	98.4
Mg#	79.0	80.0	73.7	78.0	74.9	77.0	72.1	78.7	79.8	89.3	79.5	88.0
Σ^*	4.019	4.019	4.034	4.019	4.018	4.026	4.031	4.021	4.057	4.028	4.061	4.026
Wo	44.3	43.9	45.7	44.2	42.6	44.4	44.5	43.2	45.6	42.5	46.2	44.0
En	44.1	44.9	40.1	43.6	43.0	42.9	40.0	44.8	43.4	51.3	42.8	49.3
Fs	11.7	11.2	14.3	12.3	14.4	12.8	15.5	12.1	11.0	6.2	11.0	6.7

Data in wt.%. Wo, En, Fs: atomic proportions of wollastonite, enstatite & ferrosilite, normalized to 100%.

*per O=6 Notes: analyses 1-8 are from sample K-4, 9-12 from sample K-5. (1) rim and (2) core of 110 µm euhedral grain (adjacent to phlogopite analysis #3 - Table 2a), (3) rim and (4) core of 100 µm euhedral grain (5) rim and (6) core of 30 µm euhedral grain, (7) rim and (8) core of 100 µm euhedral grain, (9) rim and (10) core of 135 µm euhedral grain, (11) rim and (12) core of 350 µm euhedral grain.

TABLE 2d. REPRESENTATIVE ELECTRON-MICROPROBE COMPOSITIONS OF LINHAISAI MINETTE PHASES (CONT'D)

	Alkali Feldspar					
	1	2	3	4	5	6
SiO_2	64.1	63.2	64.9	63.6	64.2	64.5
Al_2O_3	19.2	19.5	19.3	19.9	19.3	18.8
Fe_2O_3	.47	.51	.59	.45	.48	.49
MnO	.02	.01	.07	.03	.06	.04
MgO	.01	.01	.01	.01	.01	.01
CaO	.95	1.08	.80	1.12	.89	.47
BaO	.00	.07	.00	.22	.00	.00
SrO	.16	.31	.07	.61	.11	.19
Na_2O	4.33	4.14	4.53	4.75	4.17	3.82
K_2O	10.1	9.83	10.0	9.06	10.4	11.3
Σ	99.4	98.7	100.3	99.8	99.6	99.7
Ab	37.6	37.0	39.2	41.9	36.2	33.2
Or	57.8	57.7	57.0	52.6	59.5	64.6
An	4.6	5.3	3.8	5.5	4.3	2.3
Σ*	5.026	5.020	5.026	5.028	5.028	5.029

	Fe-Ti Oxides								
	1	2	3	4	5	6	7	8	9
SiO_2	.10	.26	.14	.10	.59	.10	.14	.14	.11
TiO_2	6.99	8.90	7.72	8.28	8.88	9.39	9.11	36.3	30.5
Al_2O_3	.57	1.55	.54	.88	1.96	1.90	1.46	.00	.11
Cr_2O_3	.03	.05	.05	.02	.05	.06	.08	.01	.04
Fe_2O_3	52.9	48.9	50.8	50.2	48.8	48.2	49.0	.00	6.37
FeO	34.6	33.4	34.8	34.6	33.2	35.0	34.4	55.2	54.3
MnO	1.17	1.06	.93	1.04	1.04	1.11	1.11	1.61	.68
MgO	.54	2.61	.82	1.39	2.84	2.13	2.20	.86	1.59
NiO	.01	.02	.09	.07	.00	.05	.02	.00	.06
CaO	.08	.22	.49	.40	.49	.15	.16	.74	.50
Σ	97.0	96.9	96.4	97.0	97.9	98.1	97.7	94.8	94.3
Σ**	23.991	23.978	24.022	24.024	23.950	23.999	23.992	23.590	24.027

Data in wt.%. Ab, Or, An: atomic proportions of albite, orthoclase & anorthite, normalized to 100%.

*per O=8, **per O=32 for spinels and O=36 for ilmenite (cation sum)

Notes: Feldspar: all from K-5 groundmass grains. (1) core of 75 x 100 μm grain, (2) rim and (3) core of 50 x 120 μm grain, (4) rim and (5) core of 100 x 150 μm grain, (6) core of 75 x 100 μm grain.

Fe-Ti oxides: all from K-5 groundmass grains. Titanomagnetites - (1) core of 12 μm grain, (2) core of 6 μm grain, (3) rim and (4) core of 14 μm grain, (5) rim and (6) core of 22 μm grain, (7) core of 25 μm grain, Ilmenite - (8) rim and (9) core of 24 μm grain.

7-10-Ma intrusions in the Sintang area to the west of the LM (P. Williams, written comm. 1987).

PETROGRAPHY

In hand specimen (Fig. 2), the LM is extremely variable in color, ranging from grey-black fine-grained mafic rock (some glassy), through dark green medium-grained varieties, to light brown to grey medium-grained feldspathic types. All varieties possess an abundance of phlogopite phenocrysts (>10 - 30 modal %). Vesicularity is characteristic of all rocks and ranges from ≈1-5% (<1 mm vesicles) to 20-30% (2-4 mm vesicles).

In thin section (Fig. 2), the LM consists of phenocrysts of abundant phlogopite laths (commonly zoned, with a dark rim), amphibole with two "reversely zoned" (internally normally zoned) concentric domains (restricted to the more felsic variants), serpentinized olivine, and (restricted to the more mafic variants) augite in a groundmass of the same mineralogy in addition to a glassy mesostasis and sanidine, apatite, Fe–Ti oxides, pyrite and rare plagioclase (andesine) and quartz. Rocks of intermediate composition display a well-developed ocellar texture (Fig. 2) consisting of 2-10 mm ovoid domains defined by a concentric rim rich in feldspar and a central portion rich in phlogopite grains.

MINERAL CHEMISTRY

Wavelength-dispersion electron-microprobe analyses were performed on major phases from two compositional end members from the Linhaisai exposure: a primitive mafic sample (K-5) and a more evolved felsic variety (K-4). In all, core, rim and intermediate spots (Σ $n \approx 300$) were analyzed on approximately 10 grains each of phlogopite, amphibole, clinopyroxene, oxide phases, and feldspar from each sample. Representative analytical data are presented in Tables 2a–d; analytical methodology is summarized in the Appendix.

Phlogopite

In the most primitive rock, trioctahedral mica phenocrysts are Ti- and Mg-rich and normally zoned; the core with Mg# [atomic 100* Mg/(Mg+Fe)] = 89, 2.0 wt.% Cr_2O_3, 3 wt.% TiO_2, and rim with Mg# = 73, <0.05 wt.% Cr_2O_3 and as much as 8 wt.% TiO_2. The more evolved sample contains mica phenocrysts with a core of Mg# = 82, 0.6 wt.% Cr_2O_3 and 4 wt.% TiO_2 and a rim of Mg# = 73, <0.05 wt.% Cr_2O_3, and 4-5 wt.% TiO_2 (Fig. 3). Groundmass grains and small phenocrysts in both the primitive and evolved types are intermediate in composition but approach the above rim compositions. In general, Al and Ba are most enriched in the grain rim relative to the core (13 wt.% core - 14 wt.% rim Al_2O_3; <0.02 wt.% core - 0.4 rim wt.% BaO). Fluorine contents are variable but are generally highest in the grain rim (0.5-1 wt.% F in core *versus* 1-2.4 wt.% in the rim). Chlorine contents are generally very low (<0.04 wt.% Cl) and are most often below the detection limit of 0.01 wt.%. Structural formulae calculated on the basis of 22 oxygen atoms all show that Si+Al sums from 7.77 to 8.00 (average ≈7.93), indicating slight tetrahedral Fe^{3+} (or Mg, Ti) occupancy. Rim formulae show the largest *Z*-site deficiencies in Si + Al. Al is therefore confined to tetrahedral sites in the LM micas, in contrast to those in the kajanite reported by Wagner (1986).

LM phlogopite compositions are generally in the range found in minette and potassic mafic rocks from a wide variety of locations (*e.g.*, Luhr & Carmichael 1981, Allan & Carmichael 1984, Jones & Smith 1983, 1985, Schulze *et al.* 1986, Bachinski & Simpson 1984, and references cited therein). The LM phlogopite is compositionally also similar to those

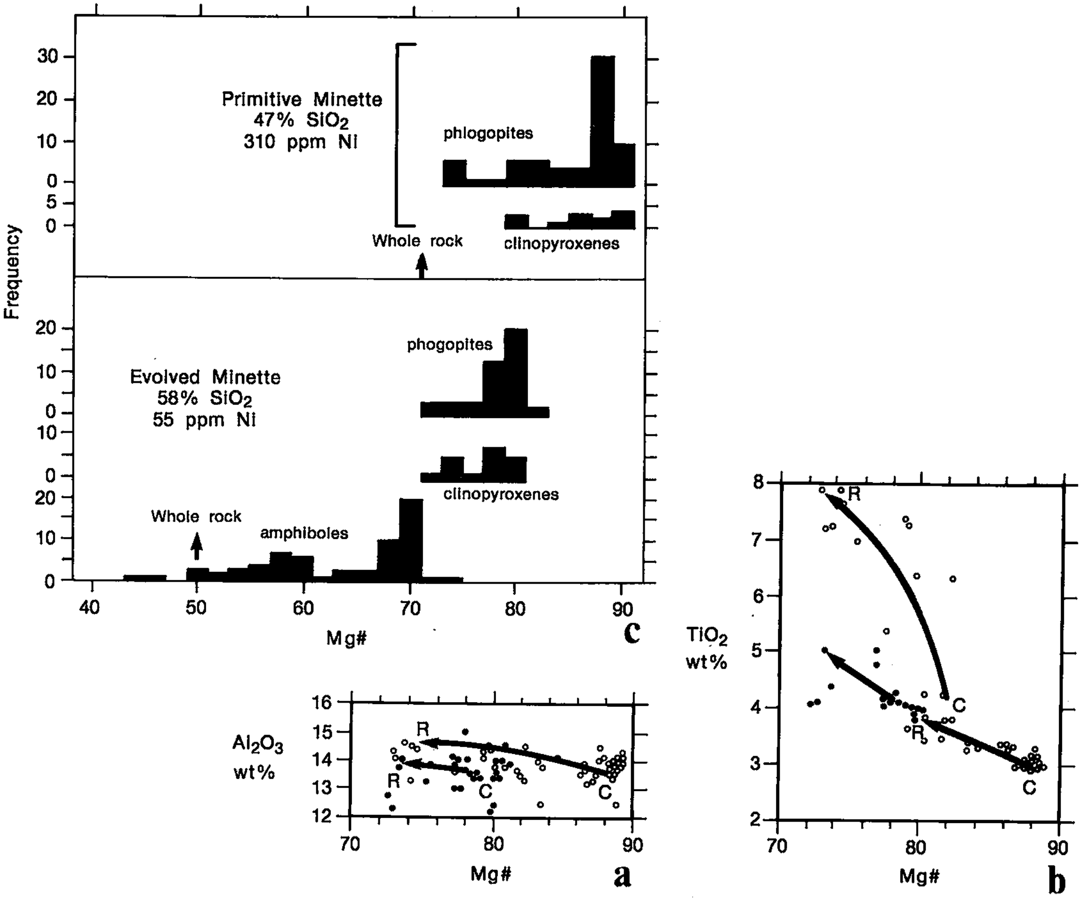

FIG. 3. Plot of LM phlogopite (a) Al_2O_3 *versus* Mg# and (b) TiO_2 *versus* Mg#; open symbols represent the more primitive sample K-5; filled symbols the more evolved K-4; where significant, core (C) to rim (R) zoning trends within individual phenocrysts are schematically shown with the arrows. (c) Histograms of Mg# of phlogopite, clinopyroxene and amphibole comparing the primitive with evolved phases. Whole-rock Mg# values are also shown.

of the nearby kajanite (Wagner 1986). Like the minette phlogopites discussed by Bachinski & Simpson (1984), the LM mica shows no appreciable correlation between Al and Mg#, but it shows a stronger Ti–Mg# negative covariation. In phlogopite of the evolved LM sample K-4, Ba and Ti show a positive correlation like that displayed by phlogopite in minette from Devonshire (Jones & Smith 1985), Arizona (Jones & Smith 1983) and Czechoslovakia (Schulze *et al.* 1986).

Compositional profiles are variable and complex; representative profiles are illustrated in Figure 4a. The following distinct profile patterns are observed: (a) grain interior with uniform Si, Al and gradually decreasing Cr, K, Mg and Si or increasing Fe, Ti, Na and F contents near margin, producing bell-shaped profiles (Fig. 4a-i, ii, iii); (b) homogeneous profiles; and (c) more complex profiles with core characterized by low K, Fe or high Na, F contents and intermediate zones with respective higher or lower contents of the elements and a rim with the reverse patterns (Fig. 4a-iv).

Clinopyroxene

Clinopyroxene grains range from an augite core to an augite–diopside (var. salite) rim (Fig. 5). In the primitive variety, augite grains are normally zoned with a core of Mg# = 90, 1–2 wt.% Al_2O_3, 0.4 wt.% Cr_2O_3, 0.5 wt.% TiO_2 and a rim of Mg# = 80, 3–4 wt.% Al_2O_3, 0.01 wt.% Cr_2O_3 and 1.6 wt.% TiO_2. The more differentiated rock contains augite of similar but more evolved compositions with a core of Mg# = 80, 0.8 wt.% Cr_2O_3, 2–3 wt.% Al_2O_3 and 0.9 wt.% TiO_2 and a grain margin with Mg# = 72, 0.1 wt.% Cr_2O_3, 3–4 wt.% Al_2O_3 and 1.2 wt.% TiO_2 (Fig. 6). The margin is generally

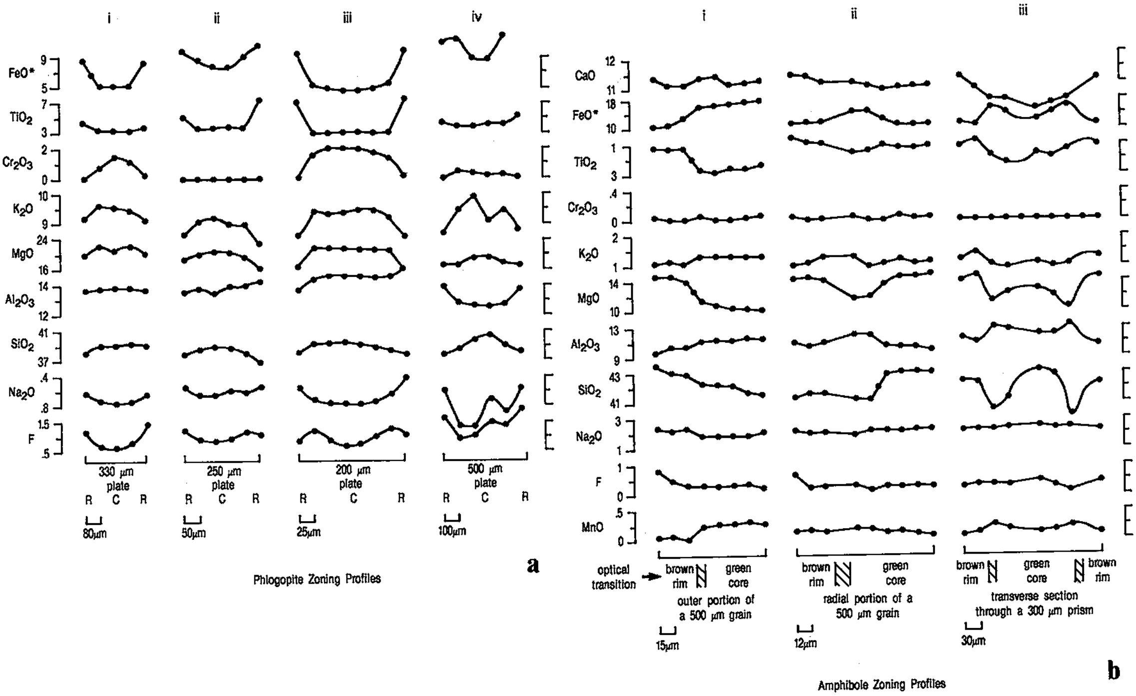

FIG. 4. Electron microprobe step-scan transverses (wavelength-dispersion) across representative (a) phlogopite and (b) amphibole phenocrysts; all phlogopite traverses are longitudinal (parallel to *b* or *c*). All concentrations are in wt.%. All phlogopite profiles are from sample K-5, amphibole profiles from K-4. See text for discussion.

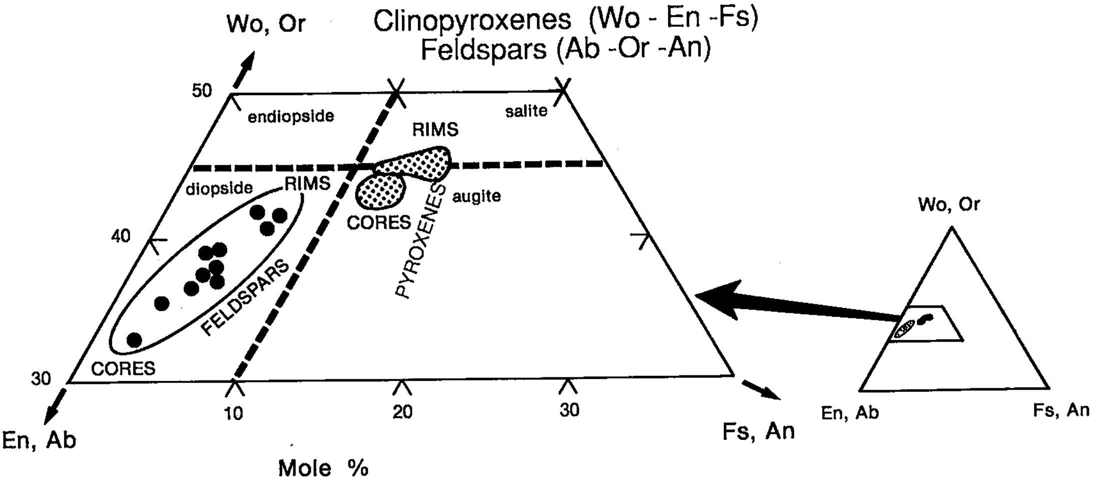

FIG. 5. Combined clinopyroxene (Mg–Fe–Ca) and feldspar triangular (K–Na–Ca) projections (mole %) showing the core and rim compositions for LM clinopyroxenes and schematically for alkali feldspars (plagioclase compositions not shown).

slightly enriched in Na_2O compared to the core (0.3–0.7 wt.% Na_2O). Structural formulae calculated for all analyzed grains show both octahedral and tetrahedral Al; tetrahedrally coordinated Al accounts for 50–80% of the total Al.

With respect to Ca–Mg–Fe, Ti, Al, Cr and Na, LM clinopyroxene compositions are broadly similar to those of minettes from the Navajo volcanic field, Arizona (Jones & Smith 1983) and Devonshire (Jones & Smith 1986) and rocks from the Colima Graben, Mexico (Luhr & Carmichael 1981, Allan & Carmichael 1984).

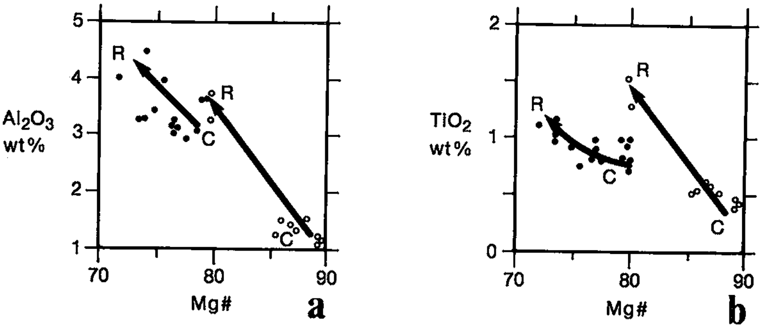

FIG. 6. Plot of clinopyroxene (a) Al_2O_3 *versus* Mg# and (b) TiO_2 *versus* Mg#; symbols and zoning trends are the same as in Figure 3.

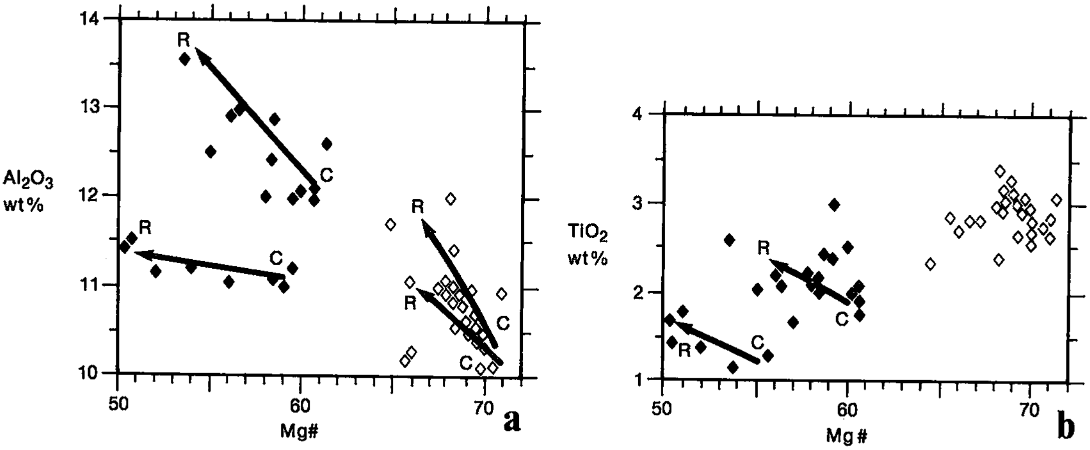

FIG. 7. Plot of amphibole (a) Al_2O_3 *versus* Mg# and (b) TiO_2 *versus* Mg#; all analyses from K-4; open symbols represent analyses of the brownish rim zone, filled symbols the dark greenish core zone. Where significant, core (C) to rim (R) zoning trends are schematically shown.

Amphibole

In the more differentiated rocks, euhedral grains of pargasitic hornblende (terminology of Rock & Leake 1984) are reversely zoned, with a deep-green core and a brown margin, each separated by a mottled black reaction or transition zone (Fig. 2c). Within each zone, "normal" (*i.e.*, decreasing Mg# toward margin) zoning is observed, but the green core is generally much more "evolved" than the rim (Fig. 4b). The core zone displays a compositional range (interior → margin) of 68–50 Mg#, 1.5–3.0 wt.% TiO_2, <0.01 wt.% Cr_2O_3, and 10–13 wt.% Al_2O_3, whereas the rim zones are in the range 71–69 Mg#, 2.7–3.4 wt.% TiO_2, and <0.01–0.05 wt.% Cr_2O_3, 10–11 wt.% Al_2O_3. In both zones (Fig. 7), Na and K contents vary little (ranges 1.9–2.3 wt.% Na_2O, 1-2 wt.% K_2O). Fluorine typically reaches its highest concentrations in grain margins (≈0.8 wt.%). Figure 4b illustrates the core-to-rim patterns of zoning in several representative amphibole phenocrysts. Despite the marked optical discontinuities observed, the following distinct patterns are apparent: (a) grain interiors with uniform or gradually decreasing Al, Fe or increasing F, Na, Si, Mg, Ca contents near grain margins (see Fig. 4b-i); (b) homogeneous profiles, *e.g.*, Cr, Na, and (c) more complex profiles, with cores characterized by uniformly low Al or high Si contents, intermediate zones with higher Fe or lower Mg contents, and rims with lower Fe or higher Mg contents (Fig. 4b-iii). The LM amphibole compositions greatly contrast with those of richterite from the more primitive kajanite (Wagner 1986).

Feldspar

Like the extensively zoned mafic phases, sanidine grains are relatively heterogeneous and within the compositional limits (molar) of $Or_{51-65}Ab_{33-42}An_{2-7}$

TABLE 3. MAJOR- AND TRACE-ELEMENT COMPOSITION OF THE LINHAISAI MINETTE, AND AVERAGE COMPOSITION OF MINETTE, LAMPROITE AND KAJANITE FOR COMPARISON

	K-1	K-2	K-4	K-5	K-5-2	K-5-4	K-SR	S-6-1	S-6-2	S-6-3	ave.*** Linhaisai minette (n=10)	Kajanite& M146		ave.** minette	ave.** lamproite
												A	B		
SiO_2	46.3	51.8	57.7	46.8	49.5	52.1	48.7	51.4	52.0	44.2	53.3	46.0	43.7	52.6	53.3
TiO_2	1.02	1.55	1.36	1.80	1.12	1.59	1.82	1.42	1.43	2.45	1.7	2.20	1.84	1.9	3.0
Al_2O_3	13.3	12.0	15.4	10.8	13.8	11.7	10.5	12.4	12.5	12.0	13.2	12.4	9.27	12.6	9.1
Fe_2O_3	5.97	5.51	3.38	3.87	3.30	3.19	1.71	1.44	2.43	3.66	-	3.54	6.06	-	-
FeO	2.5	2.2	3.1	4.1	3.6	4.3	5.0	5.4	4.5	6.4	7.7#	5.58	4.79	7.6#	6.3#
MgO	8.38	6.67	3.40	10.2	7.70	7.62	9.79	7.58	7.85	9.13	8.6	12.6	12.4	8.0	12.1
MnO	.14	.05	.12	.13	.11	.18	.09	.10	.13	.18	.12	-	.18	.13	.10
CaO	8.87	4.21	5.09	9.05	6.88	5.21	6.33	5.74	4.99	8.48	7.2	8.38	8.88	7.9	5.8
Na_2O	1.35	1.71	3.99	1.74	4.18	2.15	1.20	2.20	2.28	2.75	2.5	1.62	1.72	2.0	1.4
K_2O	2.15	6.30	3.62	3.71	2.54	5.15	5.62	4.57	4.63	2.85	4.6	4.87	4.79	6.0	7.2
P_2O_5	.94	1.18	.58	1.45	.61	1.00	1.27	.77	.76	1.18	1.1	-	1.89	1.2	1.3
$H_2O^+ + H_2O^-$	7.9	4.2	.7	2.1	5.2	2.6	5.0	3.4	4.0	3.9	3.1	3.55	3.08	2.2	2.7
CO_2	.3	.1	.6	.7	.3	.6	.2	.8	.5	.3	0.4	-	-	1.8	2.8
F	.11	.31	.22	.36	.13	.29	.38	.24	.23	.17	.24	-	-	-	-
S	.00	.04	.13	.06	.02	.03	.03	.22	.13	.03	.07	-	-	-	-
Σ	99.4	98.0	99.6	98.2	99.6	98.2	98.3	98.3	99.0	98.5	100	100.9	98.6	100	100
Mg#	65	62	50	71	68	65	93	67	68	63	63	-	-	65	81
Cr	320	370	64	520	300	310	450	280	290	340	300			340	580
Ni	120	470	55	310	260	340	280	290	320	280	273			155	420
V	210	240	200	210	160	210	220	190	190	260	209			200	123
Sc	30	17	14	21	18	18	20	17	17	26	20			21	17
Co	49	36	19	39	55	37	48	52	51	80	47			37	37
Cu		170	80	140		150					135			56	52
Zn		190	130	140		130					150			110	84
Sb		.2	.2	.1		.1					.15				
As		3	1	1		<1					1.5				
Pb		34	38	36		28					34			23	44
Bi		.4	.1	.2		.3					.3				
W		2	<1	<1		2					1				
Ta		<.5	.8	1.5		.8					1.0			2	4.7
Hf		33	16	26		30					26			9	39
Cs		<1.	1.5	2.0		1.0					1.5			.5	1.7
Rb	30	220	110	160	70	230	250	170	170	120	170			115	272
Sr	880	1260	1660	1340	1350	1070	960	970	950	820	1100			1010	1530
Ba	1210	2200	1000	1500	650	4000	1200	1110	1320	790	1500			1345	5120
Y	10	60	<10	40	30	20	30	20	20	30	20			36	27
Zr	230	860	450	760	280	840	970	700	720	450	650		340	350	920
Nb	20	10	20	<10	10	10	20	20	10	40	15		37	83	95
Th		22	12	25		18					19			24	46
U		6.5	3.4	5.8		5.8					5.4			5.0	4.9
La		126	46.6	167		80.6					105		43.2	105	240
Ce		232	94	340		154					205		84.5	195	400
Nd		102	42	147		62					88		46.4	100	207
Sm		20.9	7.40	25.2		11.9					16		10.2	22	24
Eu		5.18	1.90	5.74		2.90					3.9		2.97	4.9	4.8
Tb		2.2	.7	1.6		1.0					1.4			1.8	1.4
Dy		11.5	4.4	8.0		5.7					7.4		5.65	5.7	6.3
Yb		2.81	1.16	2.24		1.55					1.9		2.13	1.9	1.7
Lu		.41	.20	.33		.26					0.3		.38	.37	.23
Li	9	16	10	13	26	20	18	14	15	19	15			76	31
Au		.015	.002	.008		.013					.01				

* All samples contain ≤ 50 ppm Cl, < 1 ppm Se, < 0.6 ppm Br, < 8 ppm Mo, < 1.5 ppm Ag, < 0.2 ppm Cd. ** Averages represent volatile-free concentrations (except H_2O, CO_2). FeO: total Fe expressed as FeO. Values from Bergman (1987) and Rock (1984). *** Volatile-free basis, except H_2O, CO_2. # Total Fe as FeO. & Kajanite (from the Kajan River of western East Kalimantan: Brouwer 1909); analyses A from Lacroix (1926), B from Wagner (1986). Major elements in weight %, trace elements in ppm; Mg# = MgO*100/(MgO + FeO^T) (atomic proportions).

(Fig. 5). The Fe^{3+} contents are uniform and in the range 0.4–0.6 wt.% Fe_2O_3. In contrast, Sr is extremely variable and is most enriched in the margin (≈0.6 wt.% SrO) compared with the core (≈0.05 wt.%). Andesine ($An_{52}Or_3Ab_{45}$) is very rarely observed in the more evolved rocks. Where present in appreciable concentrations, Ba is most enriched (0.2 wt.% BaO) in the margins. In general, however, Ba is present in concentrations below the detection limit (<0.01 wt.% BaO).

The LM sanidine (≈0.5 wt.% $Fe_2O_3{}^T$, 0.5–1.1 wt.% CaO, 2–5 mole % An) is distinct from sanidine from the Shaw's Cove minette (0.2–1.6 wt.% $Fe_2O_3{}^T$, <0.3 wt.% CaO, <1 mole % An; Or_{34-100}: S. Bachinski, written comm. 1985), as well as that from potassic mafic rocks of the Colima Graben (0.7–0.9 wt.% Fe_2O_3, 1–15 mole % An: Allan & Carmichael 1984), but are similar to sanidines from the Permian minettes of Devonshire (0.5–0.9 wt.% Fe_2O_3, 0.5–1.1 wt.% CaO, 3–6 mole % An: Jones

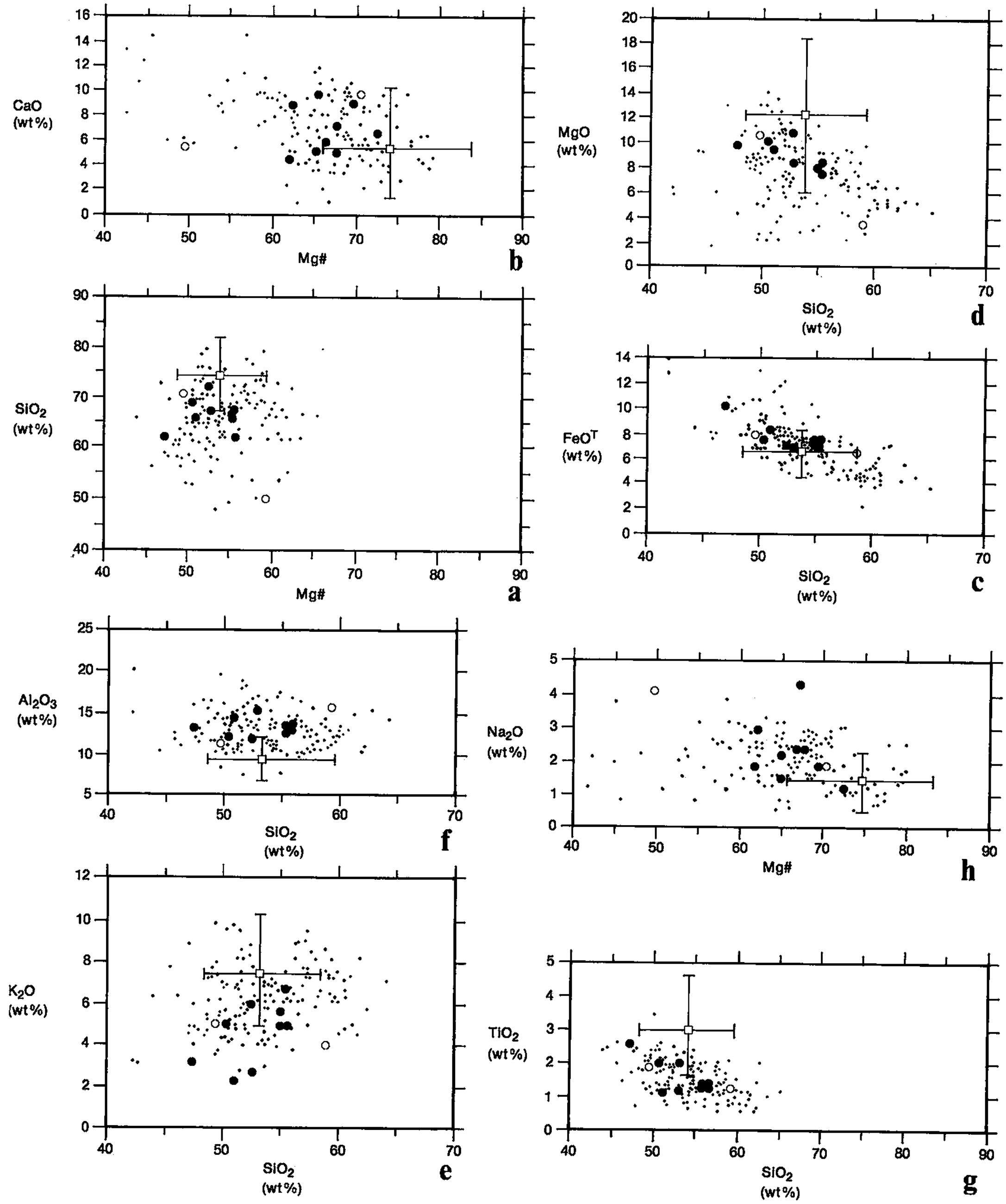

FIG. 8. Major-element variation diagrams (SiO_2 and Mg# as the absissa) showing the compositions of Linhaisai minette (large filled and open circles) compared with minettes from the literature (small dots; Rock 1984), and the average and range (2σ) in lamproite compositions (Bergman 1987). Open circles represent the two samples for which mineral chemical data exist. All analyses have been recast on a volatile-free basis.

& Smith 1986) as well as minettes from Arizona (Jones & Smith 1983). Ba and Sr contents of the LM sanidine are slightly lower than those of sanidine in the Devonshire minette.

Fe–Ti oxides

Titaniferous magnetite, the principal Fe–Ti oxide observed, typically occurs as 10–25-μm grains in the groundmass. Ilmenite rarely occurs. The magnetite grains are relatively homogeneous with 7–9 wt.% TiO_2, 1–3 wt.% MgO, 1–2 wt.% Al_2O_3, 0.05 wt.% Cr_2O_3, 1 wt.% MnO and 77–81 wt.% FeO^T. Ilmenites contain relatively low Mg ($\approx$1 wt.% MgO) and moderate Mn (1–2 wt.% MnO) contents.

WHOLE-ROCK GEOCHEMISTRY

Major and selected trace-element determinations (26 elements) were performed on 10 texturally distinct samples representing the entire range of relatively fresh and unaltered rock types observed in the Linhaisai River exposures. More extensive trace-element determinations (an additional 28 elements) were made on four samples spanning the entire compositional range. Results of analytical data are presented in Table 3; analytical methods are summarized in the Appendix.

Major elements

Nearly all major elements display ranges in concentrations: 44–58 wt.% SiO_2, 10–15 wt.% Al_2O_3, 1.0–2.5 wt.% TiO_2, 7–11 wt.% MgO, 3–10 wt.% FeO, 4-9 wt.% CaO, 1-4 wt.% Na_2O, 2–6 wt.% K_2O and 0.6–1.3 wt.% P_2O_5. Although the LM varies from mafic to intermediate, the average Linhaisai composition is nearly identical to that of the average minette (Table 3). The LM ranges from relatively primitive compositions (K1, K5, S6-3; 62–70 Mg#, 44–47 wt.% SiO_2) to extremely evolved compositions (K4; 50 Mg#, 58 wt.% SiO_2). Compatible and incompatible trace-element variations largely follow these major-element features (see below).

Variation diagrams illustrating the compositional relationships between major elements in the LM suite and typical minettes from the literature are illustrated in Figure 8. The variation displayed by the ten LM samples is similar to that displayed by the entire range in published minette compositions. In addition, the LM possesses an extremely wide range in various elemental ratios, such as K/Al (0.17–0.58) and K/Na (0.4–3.1, molar), emphasizing the probability of some subtle alteration.

Elements that decrease with increasing evolution as indicated by decreasing Mg or Mg# and increasing Si include P, Fe, Ti, and Na, whereas Al and K are variable but generally behave in an opposite manner. Some elemental ratios vary systematically with the degree of differentiation within the suite. Most dramatic is the change of Al/Ca from 1.2 to 3 (molar) with increasing differentiation.

As observed in Table 3, the average LM composition is nearly identical to the average minette, with the exception of a slight enrichment in K in the former. The average LM shows a marked compositional contrast to the "average lamproite" with respect to Ti, Al, Mg, Ca, Na and K. The kajanite reported by Brouwer (1909) and Wagner (1986) resembles the more primitive members of the LM suite. The primitive LM samples display elevated Si and Al and depressed Mg, Fe and P contents, and similar K, Na, Ca and Ti contents relative to those of the kajanite reported by Wagner (1986).

TABLE 4. CIPW NORMATIVE COMPOSITION OF THE LINHAISAI MINETTE AND COMPARABLE ROCKS

	K-1	K-2	K-4	K-5	K-5-2	K-5-4	K-SR	S-6-1	S-6-2	S-6-3	Average Linhaisai Minette (n=10)	Average Minette (n=50)*	Kajanite** M146
q	5.3	3.9	8.9	0	0	2.8	0	1.1	2.1	0	2.4	2.7 ± 5	-
or	13.9	39.8	21.7	29.0	16.0	32.0	35.8	28.6	29.0	17.9	26.4	34.5 ± 8	28.4
ab	12.4	15.5	34.2	15.3	35.4	19.1	10.9	19.7	20.3	19.6	20.2	13.6 ± 7.4	3.2
an	26.2	6.9	13.6	8.1	12.1	7.4	7.2	11.1	10.8	12.8	11.6	5.6 ± 4.6	3.4
ne	0	0	0	0	1.2	0	0	0	0	2.8	0.4	1.7 ± 3.3	6.1
di	10.1	3.9	2.4	17.4	13.6	5.4	11.2	5.7	4.4	16.7	9.1	14.9 ± 10.5	23.1
hy	18.2	16.0	8.1	1.6	0	20.5	17.6	23.8	22.7	0	12.8	6.5 ± 6.8	-
ol	0	0	0	12.9	11.6	0	6.2	0	0	15.4	4.6	6.2 ± 6.3	15.0
mt	6.1	2.9	5.0	5.9	5.1	4.9	2.7	2.2	3.7	5.7	4.4	3.6 ± 2.5	8.8
il	2.1	3.2	2.6	3.6	2.3	3.2	3.7	2.9	2.9	5.0	3.1	3.3 ± 2	3.5
hm	2.4	3.9	0	0	0	0	0	0	0	0	0.6	1.3 ± 2.5	-
ap	2.4	3.0	1.4	3.6	1.5	2.5	3.2	1.9	1.9	3.0	2.5	2.7 ± 2.1	4.5
cc	0.8	0.2	1.4	1.7	0.7	1.4	0.5	1.9	1.2	0.7	1.1	2.1 ± 3.9	-
∑ foid	0	0	0	0	1.2	0	0	0	0	2.8	0.4	2.0	6.1
∑ salic	57.9	66.3	78.5	52.6	64.8	61.5	54.1	60.7	62.4	53.2	61.2	59.0	41.1

Data in wt.%. Norms were calculated using the reported FeO and Fe_2O_3 contents.

*from Bergman (1987)

**from Wagner (1986)

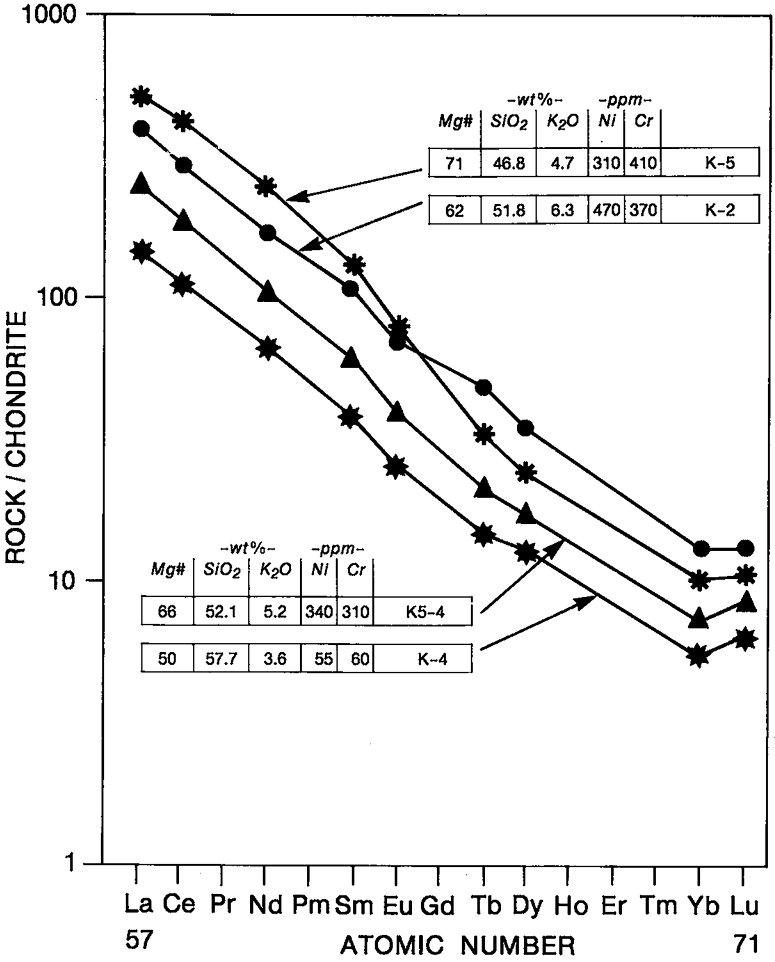

FIG. 9. Chondrite-normalized (chondrite from Hanson 1979) rare-earth-element patterns of four Linhaisai minette samples spanning the full range in major-element composition. Selected whole-rock compositional features are listed for ease of comparison.

Although CIPW norms are of questionable relevance to exotic, mica-rich rocks like minettes, they are useful in defining broad compositional features (Table 4). The LM ranges in color index (wt.% femic phases) from 21.5 to 47.4, and compositions are slightly quartz-normative (up to 9 wt.%) to slightly nepheline-normative (up to 3 wt.%); eight of the ten samples analyzed are silica-saturated or -oversaturated. This range in normative composition is similar to that of typical minettes. As borne out by the major-element compositions (Table 3), the LM is similar (within the mean ± standard deviation) to other minettes with respect to normative components although slight differences exist (especially in ab, or, an, and hy). None of the LM rocks analyzed exhibits normative acmite, a component of most peralkaline rocks and some minettes.

Despite their common affiliation with calc-alkaline lamprophyres (*e.g.*, Streckeisen 1978, Rock 1977) minettes fall in the alkali-calcic group of Peacock (1931) and are more alkaline than the typical calc-alkaline andesites associated with subduction-related magmatic arcs (*e.g.*, compilation of Brown 1982). The LM rocks likewise fall into Peacock's alkali-calcic group, yet display Fe/Mg values typical of calc-alkaline andesites (≈1–2).

Trace-element geochemistry

The trace-element concentrations of the LM and comparable lithologies are listed in Table 3. Some trace-element variations are disproportionately excessive compared to the subtle range in major-element compositions. Most apparent are systematic variations in several trace-element contents with various differentiation indices (*e.g.*, Mg#, Si, Σ salic, *etc.*). For example, in going from the most "primitive" to the most "evolved" compositions in terms of major elements (*e.g.*, K-5, K-SR, K-5-2, K-1, S-6-1, S-6-2, S-6-3 primitive to K-2, K-5-4, K-4 evolved), Cr and Ni decrease by an order of magnitude, La (and other *LREE*) decrease by a factor of 3, and Yb (and *HREE*), Co, Cu, Hf, U and Th decrease by a factor of 2.

Elements such as Li, Rb, Sr, Ba, Zr, Y, Ta, and Nb show no well-developed systematic variation related to Mg, Mg# or Si. Ba and Sr are generally most enriched in the most evolved rocks. *REE* patterns (Fig. 9) are subparallel (except K-5, which displays a Ce/Yb value twice that of the other three samples analyzed) and systematically decrease in their degree of *LREE* enrichment with increasing Si. No negative Eu anomaly is observed, even for the more evolved samples. This feature is characteristic of all minettes and results from the restriction of feldspar to the groundmass, indicating a lack of feldspar fractionation.

In terms of trace elements, the average LM is very similar to the average minette except for Cu, Hf, Ni, Cs, Zr and Nb (Table 3). As with major elements, the LM differs greatly from the average lamproite, with the exception of similar Sc, Cs, Y and U contents. LM Nb contents are uniformly low (10–20 ppm) and similar to the levels observed in calc-alkaline andesitic rocks from subduction zones and some lamprophyres (*e.g.*, Caledonian dykes from Scotland: Rock *et al.* 1986). The average minette, however, is enriched in Nb by a factor of four relative to the LM.

The primitive LM specimens possess much greater *LREE* (5×) and similar *HREE* (1–2×) contents compared with the cocite of Wagner & Velde (1986) and the kajanite of Wagner (1986). The kajanite and cocite also possess much lower Ce/Yb ratios ($\approx$40) than any of the LM rocks analyzed (80–150).

Many incompatible elements are positively correlated with each other in the LM. Particularly well developed is a La–Th correlation that follows the kimberlite trend, in contrast to that of minettes and other lamprophyres, as reviewed by Rogers *et al.* (1982). In addition to La–Th, the LM falls outside of other compositional fields for minettes defined by Rogers *et al.* (1982).

PETROGENESIS

The major- and trace-element variations noted above can be interpreted in a number of ways. First and foremost, the Mg#, Mg, Ni and Cr contents of the more primitive LM melts, (Mg# 68–73, 8–10 wt.% MgO, Ni 280–470 ppm, Cr 280–510 ppm) demonstrate that they represent primary melts of a mantle peridotite with olivine of Fo_{88-92} (Frey *et al.* 1978). The general mantle-derivation of minettes is supported by the populations of included xenoliths (*e.g.*, Ehrenburg 1982) as well as experimental studies (*e.g.*, Ruddock & Hamilton 1978). The compatible element compositions of some LM specimens indicate that olivine has accumulated; however, some of the more "evolved" melts (MgO 6–9 wt.%, Mg# 62–67) contain equally high Ni and Cr contents ($\approx$280 ppm Ni, 300 ppm Cr). These LM rocks may, as an alternative, represent melts of a slightly different bulk-composition. Relative to the primitive rocks, the most "evolved" LM are an order of magnitude lower in Ni and Cr contents and a factor of 2–3 lower in Mg (MgO 3.4 wt.%, Mg# 50, Ni 55 ppm, Cr 64 ppm). The evolved magma clearly results from crystal fractionation.

Several elements do not follow well-defined differentiation trends on Harker diagrams (*e.g.*, Ca, Al: Fig. 8), although others (*e.g.*, Fe, Mg, Si, K) display expected crystal fractionation-controlled variation trends. Apatite fractionation is suggested by the progressive decrease in P with increasing Si (and decreasing Mg or Mg#) and the presence of apatite in all members of the suite. Other phenocryst or microphenocryst phases include olivine, clinopyroxene, phlogopite, amphibole and magnetite. Least-squares mixing routines (Bryan *et al.* 1969) involving various combinations of olivine, clinopyroxene, phlogopite, amphibole, magnetite, or apatite, lead to the best solutions (sum of squares of residuals <0.2) for the following model:

K-5 (parent) → K-4 (daughter):
51% loss of 7.2% magnetite, 39.7% phlogopite, 46.3% clinopyroxene, 6.8% apatite

The major-element contrasts between K-5 and K-4 are consistent with loss of clinopyroxene, phlogopite and apatite (all observed phenocryst phases). Trace-element variations follow these major-element fractionation models, although some elements show disparate relationships. The *REE* trends support the view that both partial melting and fractional crystallization have controlled elemental variation. Samples K-2, K-5-4 and K-4 seem to be related by fractionation as indicated by similar Ce/Yb values (80–100), whereas Ni decreases from 470 to 55 ppm; however, K-5 possesses a much higher Ce/Yb ratio (151) at high Ni (310 ppm), indicating that K-5 can represent a smaller degree of partial melt (1/10) rather than a less-fractionated version of K-2 using a reasonable source estimate (10% garnet, 10% clinopyroxene, 25% orthopyroxene, 55% olivine). Therefore, crystal fractionation involving phases which preferentially partition *REE* (relative to a melt) such as apatite led to the progressive decrease in *REE* contents in the LM as fractionation proceeded. This feature is a general characteristic of minettes and other lamprophyres. Consideration of other trace-element relationships (*e.g.*, Sr, Rb, *etc.*) indicates a multiplicity of petrogenetic processes because fractionation of the above-noted phases alone cannot produce the full spectrum of trace-element features. The least-squares mixing routines, therefore, are gross simplifications that provide only broad constraints on the extent to which fractionation could have played a role in controlling the major-element variation.

Considering the possible partial melting of a lherzolitic source, in major-element chemistry the LM falls well below the MORB–alkali basalt trend on Ca/Ti and Al/Ti *versus* Ti plots (after Sun *et al.* 1979). This indicates that the LM magmas may have been derived from a source depleted in Ca and Al (or clinopyroxene and garnet) relative to MORB/alkali basalt sources (for comparable Ti contents). The range in Ca/Ti and Al/Ti values displayed by the more primitive LM rocks suggests a range in degree of partial melting of 5–10% of a slightly anomalous peridotite composition. These compositional relations suggest that the LM source must have been considerably enriched in large-ion lithophile (incompatible) elements relative to the sources for alkali basaltic melts suggested by Frey *et al.* (1978). We interpret the silica-saturated to -oversaturated character of the LM to reflect a low $CO_2/(H_2O+CO_2)$ source, in keeping with the experimental findings of Mysen & Boettcher (1975a,b).

The zoning profiles of clinopyroxene and phlogopite largely reflect crystallization from magmas that were changing significantly in composition as a result of differentiation. There are, however, sutble anomalies (*e.g.*, Cr, F and Na in phlogopite) that reflect more complex histories of crystallization. Widely varying Cr contents in phlogopite of comparable Mg# hint at a complex evolution, the origin of which was initially addressed by Métais *et al.* (1962). The amphibole phenocrysts in the most evolved rock exhibit firm evidence for the inheritance of crystals from a slightly evolved magma that precipitated amphiboles with "normal" zoning profiles. Such amphibole crystals were added to a more primitive batch of minette magma, itself evolving in a normal manner. However, the more primitive LM rocks studied do not show evidence of this contamination (the variable Cr values in phlogopite are possible exceptions). The globular or ocellar textures observed in several samples suggest that late-stage liquid immiscibility may have occurred during the crystallization history of the LM, based on the studies summarized in Rock (1984).

Whereas some of the major- and trace-element features indicate that the LM is similar to minettes, others indicate differences (*e.g.*, the similarity in Nb contents in the LM relative to arc andesites, yet different relative Rb, Ba, Sr, V, *REE*, Hf and Zr contents). This Nb similarity is anomalous, especially when considering the contrasts in ratios of various incompatible elements such as K/La (0.25 ×), K/Cs (0.2 ×), K/Rb (0.4 ×), Cs/La (0.2 ×), Sr/La (0.25 ×) and Ba/La (0.5 ×) between the LM and arc andesites (based on the compilation of Morris & Hart 1983). In contrast, the LM rocks display values of these ratios closer to those of the alkali basalts of ocean islands as compiled by Morris & Hart (1983).

The most primitive LM rocks are highly enriched in incompatible elements, and require as source a metasomatized mantle peridotite (*e.g.*, Lloyd & Bailey 1975, Wilshire & Nielson-Pike 1975, Menzies & Murthy 1980, Boettcher & O'Neil 1980, Bailey 1982, Dawson 1984). The nature of this sub-Borneo mantle metasomatism, as reflected in the trace-element geochemistry of the LM, both contrasts with and is similar to that defined for the asthenospheric wedge overlying subduction zones (*e.g.*, Gill 1981, Morris & Hart 1983) and possesses many affinities with those of typical source-regions for alkali basaltic magma.

The various petrogenetic models thus far proposed for minettes have been addressed by Rock (1984). However, modern petrological, geochemical and isotopic data substantiate a broad genetic model for minettes involving the production of melts by small but variable degrees of partial melting of a metasomatized mantle peridotite (± zone refining) with subsequent differentiation through crystal fractionation, liquid immiscibility, and volatile-phase transfer with variable but small amounts of contamination by crustal material. We interpret the petrological features enumerated above in the Linhaisai rocks to support this general petrogenetic scenario. The extensive major-element variation displayed by the LM intermediate-to-mafic sequence is interpreted to be largely the result of crystal fractionation and magma mixing. Despite Rock's (1984) preference for extensive crustal modification of mantle-derived minette melts in explaining minette petrogenesis, we prefer to view minettes as relatively unmodified mantle melts whose crustal interaction is the exception rather than the rule. The compositional modification of primary mantle-derived melts is related to "primary" magmatic processes such as crystal fractionation and volatile phase transfer. We concur with Bachinski & Scott (1979, 1980) in viewing the *LILE*-enriched and exotic composition of minettes as reflecting the nature of the mantle source and complexity in subsequent petrological processes (*e.g.*, crystal fractionation, volatile phase transfer), which, we contend, occurred largely in the upper mantle.

As far as the tectonic and petrogenetic affiliations of the LM are concerned, we consider that the age, major-, and trace-element data argue against a direct association with subduction-related magmatism. Clearly, no known subduction zone was interacting with central and northern Borneo at 8 Ma. The timing of the LM intrusion occurs in a magmatic lull in Borneo, between two major regional magmatic pulses in the Oligocene–Eocene (16–30 Ma) and Quaternary (< 1 Ma). The only other known magmatic rocks in the area of this age (7–10 Ma) occur in the Sintang area to the west of the LM (P. Williams, written comm. 1987). Despite a slight "shoshonitic" association (*e.g.*, Joplin 1968, Morrison 1980) displayed by the LM on the basis of major elements, we prefer to group the LM with intraplate alkaline magmatism resulting from postorogenic melting of the subcontinental mantle lithosphere, a process that may, in turn, have resulted from crustal thickening associated with the regional compressive deformation in Late Cretaceous to mid-Tertiary times.

Rock (1984, 1987) and Bergman (1987) summarized the petrologic similarities and differences between minettes and lamproites. Of all mafic to intermediate rock-types, minettes are the closest to mafic to intermediate lamproites and it is most probable that these two groups form a compositional continuum. Nevertheless, certain features of the LM are in marked contrast to those of lamproites, for example (*cf.* Bergman 1987):

a) The LM are more evolved (lower in Mg and Mg#) and possess higher Na and Al and lower K, Ba, Sr, Rb, Nb, Th, *LREE,* Ta, Zr and Hf contents than lamproites.

b) The LM mafic phases possess anomalously high Al contents compared with those of lamproites (amphibole: 10–14 wt.% *versus* 0.8 ± 1.7 wt.% Al_2O_3; clinopyroxene: 1.2–4.4 wt.% *versus* 0.5 ± 0.7 wt.% Al_2O_3; phlogopite: 12.5–14.6 wt.% *versus* 10 ± 3 wt.% Al_2O_3, respectively).

c) The LM sanidine possesses a high Na content (4 wt.% *versus* 0.5 wt.% Na_2O) and Ca (1 wt.% *versus* 0.2 wt.% CaO) content and lower Fe (0.5 wt.% *versus* 2.3 wt.% Fe_2O_3) and Ba (0.1 wt.% *versus* 1 wt.% BaO) contents than those of lamproites.

In contrast to the kajanite characterized by Wagner (1986), which possesses many phases characteristic of lamproites (olivine + phlogopite + leucite + diopside + richterite), the mineralogical assemblage in the LM (olivine + phlogopite + sanidine + augite + hornblende + titaniferous magnetite) is much more allied to typical minettes (*e.g.*, Velde 1971, Rock 1977, 1984). The major- and trace-element geochemical contrasts between the LM on the one hand, and kajanite or cocite on the other (data of Wagner & Velde 1986, Wagner 1986) are sufficient to justify a fine-scale petrogenetic distinction between these groups. Furthermore, the LM mineral compositions are likewise slightly distinct from those of kajanite and cocite. So, whereas kajanite is interpreted by Wagner (1986) as a member of the kamafugitic clan, an interpretation with which we agree, the LM is strictly interpreted as a member of the minette clan. We contend that these two clans possess many similarities and may be one and the same. In other words, whereas the LM and nearby kajanite may not be part of the same comagmatic suite (in time and space), they may represent a compositionally similar rock group in a more global sense.

We do not infer a lamproitic parent melt *sensu stricto* for the LM or related kajanite melts because of the mineral chemistry and whole-rock geochemistry contrasts. Based on available data, we group the kajanite of Brouwer (1909) and Wagner (1986) in a very broad mafic-to-intermediate potassic (possibly kamafugitic) petrogenetic series, and include the LM. We infer that the LM and petrologically similar melts, such as kajanites, do not form the source rocks that hosted the ubiquitous alluvial diamonds present in Cretaceous and post-Cretaceous deposits of Central and West Kalimantan (*e.g.*, Koolhoven 1935, Wing Easton 1894, van Bemmelen 1949, Ubaghs 1941, Witkamp 1932). These diamonds, therefore, probably owe their existence to an unidentified lamproite or kimberlite in Borneo or adjacent Southeast Asia. The widespread post-Cretaceous sedimentary deposits in the area could easily conceal a diamondiferous lamproite or kimberlite diatreme. Alternatively, a local pre-Cretaceous kimberlitic or lamproitic source may have been largely removed by erosion or a more distant igneous source may be suggested. The general tectonic setting of Kalimantan suggests a lamproite, rather than kimberlite, as a source for the local diamonds (Bergman 1987), in that Kalimantan is situated on the margin of the Sundaland craton. The closest known lamproites or rocks closely allied to lamproites occur in the Fitzroy Basin, northwestern Australia (*e.g.*, Wade & Prider 1940), at Coc Pia, North Vietnam (Lacroix 1933, Wagner & Velde 1986), and at Chelima, India (Bergman & Baker 1984), but all of these are over 1000–2000 km from the LM, even in Cretaceous–Tertiary reconstructions of southeast Asia (*e.g.*, Owen 1983). In keeping with the suggestions of R. Halligan (pers. comm., 1983), we prefer to consider a yet-undiscovered, more local source for the Central Kalimantan diamonds. However, the lack of any classic kimberlite or lamproite "indicator minerals" in diamondiferous sediments and alluvium throughout Kalimantan suggests a complicated multicycle sedimentary evolution for Indonesian diamonds. The identification of the primary source of Kalimantan diamonds clearly requires much more study.

ACKNOWLEDGEMENTS

This study was supported by the Anaconda Minerals Company, a former division of the Atlantic Richfield Corporation, as part of its diamond exploration effort in Borneo and the U.S.A. Field assistance was provided by Z. Masalayuk and other members of the P.T. Aneka Tambang staff of Jakarta and Banjar Baru. Discussions with L. Noddles and M. Bergman proved worthwhile. Ken Foland kindly provided the K–Ar age determinations. Discussions with R. Halligan, W. Turner, and P. Williams improved our understanding of the regional geology. We thank P. Williams for preprints and unpublished data. We thank Danielle Velde, R. Halligan, W. Laughlin, Nicholas Rock and Robert Martin for very thoughtful reviews of an early version of this manuscript. We appreciate the assistance of J. Toney and D.J. Henry in the electron-microprobe work. We thank ARCO Oil and Gas Co. for assistance and permission to publish this paper.

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Received December 23, 1986; revised manuscript accepted April 22, 1987.

APPENDIX

Analytical Methods

Major and trace-element data were determined in the laboratories of X-Ray Assay Laboratories, Ltd. of Toronto, using standard X-ray-fluorescence techniques for major elements and S, Cl, Cr, Rb, Sr, Y, Zr and Ba; instrumental neutron-activation techniques for *REE*, Hf, Ta, W, Cs, Mo, Sb, Sc, Th, U, Au, Cr, As, Se, and Be; atomic absorption for Li; and direct current plasma spectroscopy for Pb, Bi, Ag, Cd, V, Co, Cu, Zn and Ni, and wet methods for H_2O, F, CO_2 and FeO. Based on the analyses of standard rock powders, combined precision and accuracy of elements present in quantities > 10 wt.% is 2–4 relative %; that of trace elements ranges from < 10 relative % for some INA and XRF analyses to < 50% for XRF and INA analyses close to detection limits.

High-precision electron-microprobe analyses (wavelength dispersion) were performed in the ARCO Plano Research Laboratories using a fully-automated 1983 Cameca MICROBEAM instrument and standard operating techniques (15 kV potential, 20 nA beam current, a rastered beam at 8,000–20,000 ×). A ZAF correction procedure was used to convert X-ray intensities to concentrations of elements. Estimated errors for major elements (10–70 wt.% oxide) are $<$ 1–2 relative %, for minor elements (< 10 wt.% oxide) $<$ 5–10 relative % and for extremely minor elements (< 1 wt.% oxide), 10–50 relative %.