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PLUTONIC NODULES IN LAMPROPHYRIC CARBONATITE DYKES NEAR FREDERIKSHÅB, SOUTH-WEST GREENLAND

BY

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WITH 1 FIGURE AND 7 TABLES IN THE TEXT,
AND 4 PLATES

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Abstract

A swarm of thin NW–SE lamprophyric carbonatite dykes of Mesozoic age occurs south of Frederikshåb associated with a contemporaneous, parallel swarm of thick dolerites.

Apart from local country rock material, inclusions in the lamprophyric carbonatites are mainly of the following types:

1) Single crystals of olivine which were probably mainly derived from the upper mantle. 2) Relatively unmodified garnet- and pyroxene-granulite nodules brought up from a lower crustal level. 3) Nodules, and single crystals, consisting mainly of hornblende and salite which are considered to have formed by metasomatic reaction between the carbonatite magma and mainly acid to intermediate lower crustal rocks, possibly at relatively low levels in the dykes. Hornblende shows various stages of growth from initially small, iron-rich crystals to larger, iron-poor crystals which have commonly replaced pyroxenes. The pyroxenes show a similar but less pronounced development. 4) Alkaline nodules which are again thought to have developed by metasomatic reaction between the magma and country rock inclusions, but possibly at higher levels in the dykes. 5) Phlogopite megacrysts which may be partly xenocrystal but which are thought to have mainly crystallised from the contaminated magma.

Complete chemical analyses of lamprophyric carbonatites and partial analyses of individual minerals are presented.

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THE REGIONAL SETTING

Lamprophyric carbonatite dykes were discovered in the Iluilârssuk area in 1964 during the Geological Survey of Greenland's regional mapping in the Frederikshåb district (Walton, 1966). The area lies immediately south-east of Frederikshåb in South-West Greenland and is composed predominantly of granodioritic gneisses, and amphibolites with ultrabasic lenses (fig. 1). A few thin discordant amphibolite dykes, which must once have been dolerites, also occur. All these rocks show metamorphism in the almandine-amphibolite facies and are of pre-Ketilidian age (approx. 2600 m.y.). They have been affected by four or five episodes of pre-Ketilidian folding. Intruded into the metamorphic rocks are swarms of dolerite dykes with individual members commonly attaining thicknesses of up to 80 m. The older dykes, which are also pre-Ketilidian, are referred to as MDs (standing for metadolerite, although in this particular area the dykes are not metamorphosed). The oldest generation, MD1s, are thin dykes and trend NNW but do not occur in the area of fig. 1. They are strongly sheared. Most of the NE trending dykes of fig. 1 are MD2s and these show some shearing along their margins, as do the WNW trending MD3s. A swarm of WNW trending dykes, younger than the MD3s and probably post-Ketilidian, is present a few kilometres NE of the area of fig. 1. These are distinctly porphyritic dykes with abundant phenocrysts of plagioclase. Younger than the porphyritic dykes are NE trending dolerites of Gardar age (approx. 1200 m.y.) which are not common in the Iluilârssuk area, although the two younger NE dykes immediately north of Iluilârssuk and the dyke at Ikerasârssuk belong to this group.

The youngest dolerites, which are of Mesozoic age, trend NW, parallel to the coast (Watt, 1969). In the Iluilârssuk area these increase in thickness and abundance towards the west of the region. The lamprophyric carbonatites occur as veins and dykes from a few mm to 1 m in thickness, and are restricted to the extreme west of the region. A similar dyke occurs in the Nigerdlikasik area about 30 km to the NE (Andrews, 1968) and this dyke is of particular interest since it is a true kimberlite. The lamprophyric carbonatite dykes trend NW and are present where the Mesozoic dolerites are thickest and most abundant.

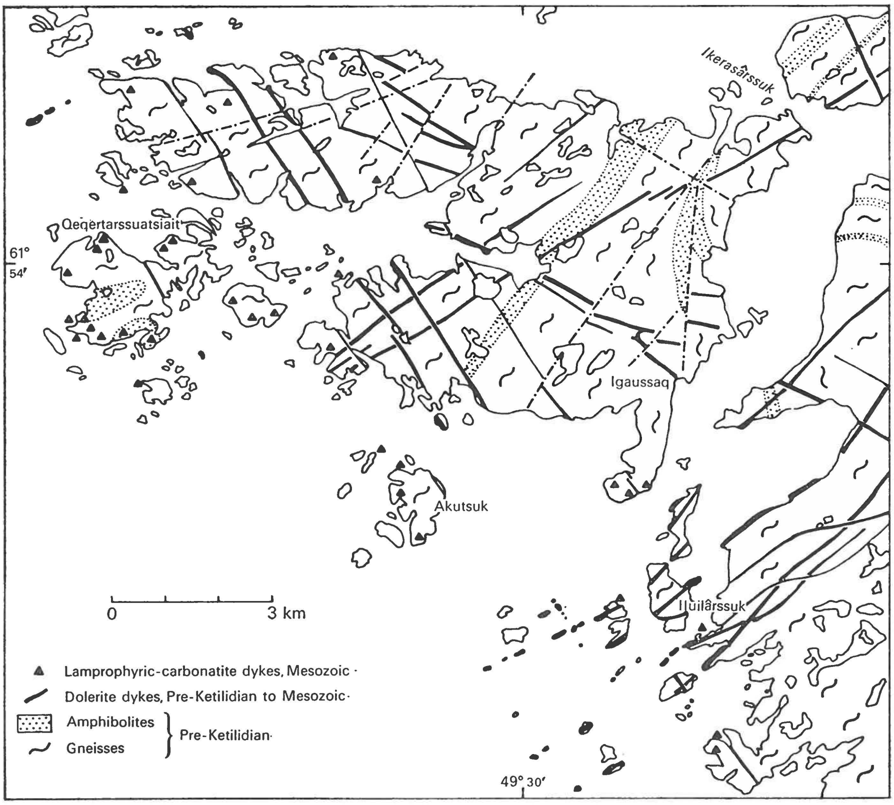

Fig. 1. Outline geology of the Iluilârssuk area.

At one locality a lamprophyric carbonatite is intruded into the margin of a thin Mesozoic dolerite. It is thought that the lamprophyric carbonatites may be genetically related to the Mesozoic dolerites and approximately contemporaneous with them, or slightly younger. The carbonatites may have developed together with the dolerites by a zone refining process similar to that described by Harris (1957). Watt (1969) notes that these dolerites vary from slightly quartz normative to slightly olivine and nepheline normative. They are therefore not particularly alkaline. Although carbonatites and kimberlites are commonly associated with alkali basalts in continental areas, Verschure (1966) has drawn attention to the fact that such activity is generally preceded by a much more important phase of tholeitic eruption and has suggested that this is necessary to allow concentration of alkalies and volatiles.

Biotite from a lamprophyric carbonatite has been dated at 162 ± 5 m.y. (Larsen & Møller, 1968) by the K/Ar method, and N. H. Gale, Oxford University, has dated a Mesozoic dolerite at 138 m.y. (Watt, 1969). The date from the lamprophyric carbonatite should probably be regarded as a maximum age, because, as LOVERING & RICHARDS (1964) have suggested, if the biotite is partly xenocrystal it may contain excess radiogenic Ar incorporated in the mineral lattice under conditions of locally high Ar partial pressures. These conditions could have existed either within the mantle or at depth in the host magma, and anomalous ages have been reported for similar rocks from several regions (DAVIDSON, 1964, ZARTMAN et al., 1967). Some relic Ar might conceivably have been present in the micas if they were partially derived from older rocks. However, since phlogopite and calcite form a stable assemblage in the lamprophyric carbonatites and orthoclase is absent, these dykes must have crystallised above 300° C (BAILEY, 1966) and so retention of relic Ar seems unlikely. The Nigerdlikasik kimberlite dyke has recently been dated by whole rock K/Ar at 609 ± 36 m.y. (J. ANDREWS, personal communication). The micas in this rock occur in the groundmass but the dyke has the same trend as the lamprophyric carbonatites and on geological evidence could be about the same age. The possibility therefore exists that an anomalously high age has been obtained.

About 50 km NE of Frederikshåb in the Oqúmiaq area, thin lamprophyric dykes occur trending WNW. These dykes are on the whole less carbonatitic and more alkaline than those in the Iluilârssuk area. Their age is unknown except that they can be seen to be younger than MD3s. Within this immediate region plutonic nodules occur only in the lamprophyric carbonatites, the lamprophyres of the Oqúmiaq area and in the Nigerdlikasik kimberlite. In the Iluilârssuk area these nodules have been found only in dykes on Qeqertarssuatsiait and the islands immediately to the east, including the large island on which Igaussaq mountain is situated. From a locality on the south west shore of Qeqertarssuatsiait a specimen measuring $30 \times 15 \times 10$ cm was collected (Plate 1 a). It was estimated that this contained at least 400 ultramafic nodules.

The Iluilârssuk area has been affected by important fault movements of greatly varying ages. Some of these movements were certainly pre-Ketilidian, but in addition the younger dykes have been affected by transcurrent and dip-slip faults. However, very little faulting seems to have occurred subsequent to the intrusion of the Mesozoic dolerites and no faults have been found cutting the lamprophyric carbonatites.

LAMPROPHYRIC CARBONATITES

These rocks consist of a fine- to medium-grained groundmass of primary carbonate, magnetite and small crystals of mafic minerals with megacrysts up to 1 cm in length of phlogopite, salite, hornblende and olivine (Plate 1b). The primary carbonate is mainly calcite but dolomite, and sometimes ankerite (Table 7, no. 1), also occurs. In a few cases relatively magnetite-free orbs of carbonate are developed around the megacrysts (Plate 1b). Nodules are only rarely found and these are most commonly leucocratic gneisses showing the same mineralogy as the country rocks. However, some leucocratic nodules can be shown to come from a lower crystal level, and this is believed to be generally the case with the rarer ultramafic nodules. The nodules tend to occur in very localised abundance rather than being distributed throughout the dykes.

The dyke rocks are clearly lamprophyric and from the analyses of Table 1 can be seen to be ultrabasic and carbonatitic in composition. They are richer in SiO_2 than some carbonatites because they carry a rather high proportion of xenocrystic material. In this respect they are similar to kimberlites. Other similarities to kimberlites include the presence of phlogopite and olivine crystals, the carbonate groundmass and the similar contents of certain trace elements (Table 2). The similarity between the "kimberlite" analysis given in Table 1 (WATSON, 1955) and the lamprophyric carbonatite analysis 73632 may be noted. The "kimberlite" is richer in MgO and somewhat poorer in CaO and CO_2 as would be expected. On the other hand these rocks differ from true kimberlites by not containing ultrabasic nodules characterised by such high-pressure minerals as pyrope and jadeitic pyroxene (DAWSON, 1967). The same applies to WATSON's "kimberlite" and this rock is probably more correctly described as a mica peridotite (DILLER, 1892). The lamprophyric carbonatites are also very similar to alnöites (VON ECKERMANN, 1963 and 1967) although melilite has not been identified in these rocks. In particular, they closely resemble alnöitic dyke rocks termed aillikites in Labrador (KRANCK, 1953) and the possibility exists that the two groups of dykes were one prior to the onset of continental drift. GRASTY, RUCKLIDGE & ELDERS (1969) give age determinations of dykes on the Labrador coast which are all older than 500 m.y. but these may include anomalously high values.

The common appearance of normative nepheline and olivine for the lamprophyric carbonatites (Table 3) indicates their silica undersaturation. Also their alkaline character is shown by the appearance of norma-

Table 1. *Chemical analyses of lamprophyric carbonatites from the Iluilârssuk area, compared with analyses of a "kimberlite" and average carbonatite*

	73632	73642	73701	73681	73703	60242	60240	a	b
SiO_2	22.91	22.91	29.03	20.35	14.99	30.47	nd	22.86	10.29
TiO_2	2.71	2.74	2.41	2.70	3.06	1.19	nd	2.98	0.73
Al_2O_3	5.02	4.18	4.54	4.95	5.67	9.72	nd	3.78	3.29
Fe_2O_3	5.05	4.62	3.01	7.85	9.21	1.54	nd	4.79	3.46
FeO	6.94	5.48	6.31	5.22	5.34	6.52	nd	5.32	3.60
MnO	0.30	0.25	0.10	0.26	0.25	0.30	nd	0.17	0.68
MgO	7.81	9.55	18.36	7.40	9.79	2.82	nd	14.58	5.79
CaO	24.82	26.19	11.74	25.44	29.53	19.36	nd	22.24	35.10
Na_2O	1.46	0.53	1.04	2.27	0.71	5.07	nd	0.33	0.42
K_2O	1.53	2.40	2.60	1.80	1.23	0.49	nd	1.52	1.36
P_2O_5	1.63	0.20	0.26	1.03	1.40	1.69	nd	1.32	2.09
CO_2	17.00	16.00	19.00	18.00	17.30	16.10	nd	14.84	28.52
S	nd	nd	nd	nd	nd	nd	nd	na	0.56
H_2O	3.17	2.54	1.51	1.88	1.89	4.12	nd	5.07	1.44
Ce	0.03	0.03	nd	0.06	0.03	0.03	tr	na	na
La	0.01	tr	tr	0.01	0.01	0.01	0.01	na	na
Ba	0.09	0.10	0.10	0.10	0.06	0.30	0.12	na	0.36
Zr	0.05	0.04	0.01	0.05	0.03	0.03	0.04	na	na
Nb	0.01	0.01	0.008	0.01	0.008	0.02	0.01	na	na
Sr	0.25	0.20	0.15	0.20	0.20	0.20	0.20	na	0.39
Y	0.005	nd	tr	0.005	tr	0.005	0.005	na	na
Rb	0.002	0.002	0.005	0.002	nd	nd	0.002	na	na
Cu	0.010	nd	nd	nd	nd	nd	nd	na	na
Ni	0.01	nd	nd	nd	nd	nd	nd	na	na
Co	0.005	nd	nd	nd	nd	nd	nd	na	na
V	0.03	nd	nd	nd	nd	nd	nd	na	na
Cr	0.02	nd	nd	nd	nd	nd	nd	nd	na
	100.877	97.972	100.183	99.587	100.678	99.655		99.80	100.00

a) "Kimberlite" analysis by Watson, 1955 p. 573, kimberlite analysis No. 3 from Bachelor Lake, Quebec.

b) Average carbonatite analysis given by Heinrich, 1966 p. 222.

Major elements for 60242–73703 and trace elements for 73632 were determined by Ib Sørensen. Trace elements for 60240, 60242 and 73642–73703 were determined by A. Livingstone.

nd = not determined na = not available tr = trace

Table 2. *Trace element contents of lamprophyric carbonatites compared with published analyses of "kimberlites", carbonatites and other ultrabasic rocks, in p.p.m.*

	lamprophyric carbonatites	"kimberlites" (1)	carbonatites (1)	other ultrabasic rocks (1)
La	0–100	370	284	0
Ba	600–3000	740–1000	2240	1
Zr	100–500	97–445	83	30
Nb	80–200	70–240	1690	1
Sr	1500–2000	200–1140	2450	1
Y	0–50	40–46	96	0
Rb	0–50	11–250	0	2
Cu	100	60–100	2.5	20
Ni	100	450–1200	8	2000
Co	50	40–70	17	200
V	300	120–170	–	40
Cr	200	1000–1500	–	2000

(1) from DAWSON, 1967, table 8.6.

Table 3. *One-cation molecular norms for the lamprophyric carbonatites, a "kimberlite" and an average carbonatite*

	73632	73642	73701	73681	73703	60242	a	b
corundum	–	–	0.02	–	1.43	0.92	–	1.18
zircon	0.06	0.04	0.02	0.06	0.04	0.04	–	–
orthoclase	9.00	–	14.35	10.50	–	2.85	9.05	7.75
albite	11.74	–	8.75	1.30	–	37.30	2.95	2.78
anorthite	2.60	1.95	–	–	4.96	–	4.40	–
leucite	–	11.48	–	–	–	–	–	–
nepheline	0.78	2.88	–	8.94	3.75	4.74	–	0.52
kaliophilite	–	–	–	–	4.29	–	–	–
acmite	–	–	–	3.12	–	–	–	–
diopside	2.68	16.68	–	5.12	–	–	2.80	–
hypersthene	–	–	17.08	–	–	–	3.72	–
olivine	17.97	15.14	8.25	13.23	19.95	6.99	27.30	8.13
larnite	–	1.74	–	–	7.00	–	–	–
magnetite	5.25	4.89	2.94	6.93	6.48	1.61	5.04	3.50
hematite	–	–	–	–	2.00	–	–	–
chromite	0.03	–	–	–	–	–	–	–
ilmenite	3.76	3.86	3.14	3.72	4.20	1.64	4.18	0.98
apatite	3.39	0.43	0.51	2.14	2.88	3.50	2.78	4.22
fluorite	–	–	–	–	–	–	–	0.88
pyrite	–	–	–	–	–	–	–	0.47
calcite	42.74	40.92	21.42	44.94	43.04	34.24	37.78	62.86
magnesite	–	–	23.52	–	–	6.16	–	6.74
	100.00	100.01	100.00	100.00	100.02	99.99	100.00	100.01

a) "Kimberlite" norm modified from WATSON, 1955, p. 573, Table 3.
b) Average carbonatite norm derived from HEINRICH, 1966, p. 222, Table 8–2, column 5.
Concerning one-cation molecular norms see ESKOLA, 1954.

Table 4. *Modal analyses of nodule-free lamprophyric carbonatites*

	73632	73642	73700	73681B	73702C
carbonate	54.0	45.0	45.0	63.5	54.3
magnetite	10.9	23.0	4.0	14.7	9.7
quartz	–	–	2.5	–	–
phlogopite	0.7	19.6	30.5	9.5	10.5
salite	15.4	6.4	–	5.7	23.8
hornblende	18.8	2.5	–	6.2	1.7
olivine (largely carbonated)	–	3.3	18.0	–	–
	99.8	99.8	100.0	99.6	100.0

tive acmite, leucite and kaliophyllite in addition to nepheline. 73642 and 73703 have such low silica/lime ratios that the C.I.P.W. scheme breaks down and it is necessary to substitute larnite (Ca_2SiO_4) for some of the normative anorthite. Modal analyses of some nodule-free lamprophyric carbonatites are given in Table 4.

PLUTONIC NODULES IN THE LAMPROPHYRIC CARBONATITES

These range in size up to 6 cm in diameter and can be divided into two main groups: 1) those of possible upper mantle origin, including olivine megacrysts and a hornblende olivine pyroxenite nodule, and 2) those of lower crustal origin. The second group may itself be divided into a) nodules of relatively little modified regional metamorphic origin including garnet- and hypersthene-granulite, and b) nodules of probable metasomatised crustal origin including pyroxene hornblendites, hornblende pyroxenites, biotite pyroxenites and glimmerites.

Many of the nodules are very similar to rocks described from intrusive alkaline and carbonatitic complexes (*e.g.* UPTON, 1967); however there are no intrusive complexes exposed in this part of West Greenland. It is conceivable that such complexes might be present at depth, but the detailed evidence of these nodules suggests that they probably represent mainly upper mantle material and modified lower crustal metamorphic rocks. WATERS (1955) has suggested that hornblende- and biotite-rich rocks may have become concentrated in the dwindling roots of tectogenes as more and more of the granitic fraction was sweated upward. Thus these ultramafic nodules might represent relic metamorphic rocks. However RINGWOOD & GREEN (1966) have argued that the lower crust consists mainly of quartzo-feldspathic granulites. VON ECKERMANN (1948) proposed that the mafic silicate rocks at Alnö developed by reaction between the carbonatite magma and sialic country rocks. The carbonate melt became enriched in silicon and alkalies at the same time as the wall rocks were depleted in these elements. HOLMES (1950, 1956) suggested a similar process for the development of the O.B.P. series (olivine-biotite-pyroxene nodules) in East Africa with the concomitant formation of alkali-rich katungite. Dunite inclusions were regarded as part of a continuous series including augite peridotites and pyroxenites by HOLMES & HARWOOD (1937), and other authors have found continuous variation from dunites to clinopyroxenites (BROTHERS, 1960; WHITE, 1966). However DE ROEVER (1963) found a hiatus between these two rock types and this seems to be the case in the Iluilârssuk area. Although the process described by HOLMES and VON ECKERMANN would probably account for most of the clinopyroxenite and hornblendite inclusions of this area, it is thought that the olivine nodules and xenocrysts are more likely to have come from the upper mantle. WILSHIRE & BINNS (1961) suggested that hornblendites, glimmerites, hornblende biotite pyroxenites and peridotites may all occur as local differentiates within the mantle. Unfortunately in the lamprophyric carbonatites very few rocks have been found which are transitional between peridotites and clinopyroxenites on the one hand, or between clinopyroxenites and granulites or gneisses on the other.

Nodules of possible upper mantle origin

Nodules of possible upper mantle origin are very rare in these dykes, but single crystals of olivine, which are thought to have come from the upper mantle, are common (Plates 2a and 3a). In the lamprophyric carbonatites the olivines are most often thoroughly carbonated, although some partly fresh crystals do occur. Even when completely altered the mineral is easily recognised by its relic shape, relief and characteristic pattern of internal veining. Relic olivines up to 1.5 cm across have been found but, because of the alteration, it is difficult to know whether these represent single crystals or aggregates.

The olivines might be 1) phenocrysts grown from the carbonatite magma, 2) xenocrysts derived from peridotites emplaced within the crust, or from dolerite dykes, or 3) xenocrysts derived from the upper mantle. They were clearly out of equilibrium with the carbonatite magma so if they were phenocrysts they must have formed at lower levels in the dykes. The crystals are generally sub-rounded and it has not been possible to distinguish a second generation of smaller, more euhedral olivines as was done by DAWSON (1962). The rounding may have occurred due to attrition as the crystals were carried upwards; olivine is not particularly common in the dolerite dykes of this area, and ultrabasic lenses carrying olivine make up only a small proportion of the country rocks. On the whole it seems likely that most of these crystals were derived from the upper mantle. One olivine has been analysed, using an electron microprobe, as Fo_{81} (Table 7, no. 4). This composition is somewhat more Fe-rich than that which is typical for olivines occurring in dunites or as inclusions in basalts (ROSS, FOSTER & MYERS, 1954). Possibly the olivines in the lamprophyric carbonatites were derived from the upper mantle as crystals poorer in Fe than the carbonatitic magma. If so they could have suffered metasomatic Fe enrichment. As shown subsequently the same process appears to have operated in the conversion of granulite inclusions to pyroxenite or hornblendite nodules.

Table 5. *Modal analyses of plutonic nodules in the lamprophyric carbonatites*

	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17
hornblende	14.4	–	–	–	8.3	3.3	80.9	88.3	96.5	80.7	100.0	39.2	42.6	0.8	–	–	13.7
salite/diopsidic augite.............	21.0	35.1	13.9	3.8	30.0	23.0	16.2	6.8	–	16.9	–	59.3	33.8	86.0	10.1	–	15.3
aegirine augite.......	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–	40.0	5.1
hypersthene	40.3	–	8.7	12.2	–	–	–	–	–	–	–	–	–	–	–	–	–
olivine	18.2	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–
biotite	–	–	0.9	–	–	–	–	–	–	–	–	–	20.8	11.4	80.7	54.0	–
garnet	–	20.0	46.0	6.4	34.8	–	–	–	–	–	–	–	–	–	–	–	–
plagioclase...........	–	33.9	21.1	73.0	20.6	71.8	–	–	–	–	–	–	–	–	–	–	–
albite	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–	2.3	3.4
quartz	–	–	9.4	–	–	–	–	–	–	–	–	–	–	–	–	–	1.1
carbonate	1.9	3.8	–	–	4.7	–	–	1.9	3.3	–	–	0.9	0.5	–	–	2.9	8.6
perovskite...........	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–	46.1
magnetite	4.1	7.2	–	4.5	1.6	1.9	0.2	2.9	1.2	2.3	–	0.5	2.3	1.6	9.2	0.7	6.6
apatite..............	–	–	–	–	–	–	2.7	–	–	–	–	–	–	–	–	–	–
Total	99.9	100.0	100.0	99.9	100.0	100.0	100.0	99.9	101.0	99.9	100.0	99.9	100.0	99.8	100.0	99.9	99.9

1) hornblende olivine pyroxenite 88480 6A/2 (the olivine and orthopyroxene are completely carbonated).
2) garnet granulite 73641 a.
3) garnet granulite 73641 b.
4) garnet granulite 73641 c/2.
5) garnet granulite 60237.
6) pyroxene granulite 73641 d/1.
7) pyroxene hornblendite 88480 6B/4.
8) pyroxene hornblendite 88480 3A/1.
9) hornblendite 88480 3A/3.
10) pyroxene hornblendite 88480 3B/1.
11) hornblendite 88480 3B/2.
12) hornblende pyroxenite 73641 d/2.
13) biotite pyroxene hornblendite 73641 c/1.
14) biotite pyroxenite 73641 c/3.
15) pyroxene glimmerite 73702 c.
16) pyroxene glimmerite 73637 –1.
17) hornblende pyroxene perovskite rock 60246 a.

One nodule in these dykes is a hornblende olivine pyroxenite (for the mode see Table 5, no. 1). This differs from the other pyroxenite nodules in that it contains carbonated olivine and another carbonated mafic mineral, which from its general form appears to have been originally orthopyroxene. Since olivine and orthopyroxene are of similar composition one might expect them to react similarly to carbonation whereas other minerals in the rock have been practically unaffected. The rock has a xenomorphic granular texture and an average grain size of about 0.7 mm with all minerals being of approximately equal size (Plate 2a). The hornblende is pleochroic in shades of light green and appears different from the amphibole in the other pyroxenite or hornblendite nodules. Partial analyses of the hornblende and diopsidic pyroxene from this nodule are given in Table 6, nos. 6 and 13. Although this rock is a pyroxenite, the most reasonable interpretation of its mineralogy and texture in this particular setting would seem to be that it is derived

Table 6. *Partial analyses of amphiboles and clinopyroxenes in the lamprophyric carbonatites*

	1	2	3	4	5	6	7	8	9	10	11	12	13
SiO_2 . . .	46.8	48.1	46.3	46.6	41.5	45.15	52.5	53.4	54.9	49.6	49.3	50.6	48.40
TiO_2 . . .	3.0	3.4	2.2	2.9	0.4	1.29	1.3	2.1	0.9	0.6	0.7	0.04	0.20
Al_2O_3 . .	12.6	12.6	11.3	8.0	14.6	12.74	6.0	4.7	4.2	5.2	2.7	1.4	2.29
FeO* . .	15.8	8.8	11.5	12.7	14.7	6.17	8.9	6.1	7.5	9.2	9.4	20.5	7.81
MgO . .	10.1	14.3	11.7	12.0	12.0	17.35	9.6	12.0	10.4	11.9	9.6	4.2	15.54
CaO . . .	10.9	11.9	11.4	14.8	9.5	13.60	19.9	21.2	21.1	21.1	22.2	10.3	22.15
Na_2O . .	0.8	0.8	2.5	3.1	4.8	1.94	1.8	0.4	0.9	1.1	0.6	7.7	0.39
Total§ .	100.0	99.9	96.9	100.1	97.5	98.24	100.0	99.9	99.9	98.7	94.5	94.74	96.8

1) Core of amphibole xenocryst 73632.
2) Margin of amphibole xenocryst 73632.
3) Amphibole in hornblende pyroxenite nodule 60247.
4) Brown amphibole in garnet granulite nodule 60237.
5) Blue amphibole in garnet granulite nodule 60237.
6) Amphibole in hornblende olivine pyroxenite nodule 88480 6A/2.
7) Core of clinopyroxene xenocryst 73632.
8) Margin of clinopyroxene xenocryst 73632.
9) Clinopyroxene in hornblende pyroxenite nodule 60247.
10) Clinopyroxene in garnet granulite nodule 60237.
11) Core of clinopyroxene in mela-syenite nodule 60246.
12) Margin of clinopyroxene in mela-syenite nodule 60246.
13) Clinopyroxene in hornblende olivine pyroxenite nodule 88480 6A/2.

*) Total iron determined as FeO.
§) Some of the totals obtained were originally in excess of 100 %.
These values have been reduced proportionately to total 100. The analyses were by electron microprobe and are considered accurate to ± 5 %.

from the upper mantle. In addition to WILSHIRE & BINNS (1961), KUNO (1967) has described pyroxenite nodules which he believes to be derived from the upper mantle. The primary mineral assemblage of the rock from the Iluilârssuk area was probably orthopyroxene, olivine, diopside, hornblende and magnetite, and this might represent a zone of parent mantle from which basaltic material has not been removed (OXBURGH, 1964; HARRIS, 1967). If this is true the only change this nodule has suffered during its ascent is the carbonation of its more susceptible, *i.e.* lime-free, silicate minerals.

Nodules of crustal metamorphic and metasomatic origin

Some nodules appear to be relatively little modified regional metamorphic rocks brought up from lower levels of the crust. One of these consists of plagioclase, potash feldspar, quartz, hypersthene, biotite and magnetite, an assemblage which is stable in the hornblende-orthopyroxene-plagioclase granulite subfacies (WINKLER, 1967). The hypersthene occurs as irregular crystals up to 4 mm in diameter and shows only slight alteration to carbonate marginally and along veins. The small amount of biotite is closely associated with the hypersthene, and the potash feldspar sometimes shows microcline twinning. Distinct gneissic banding is present.

Probably derived from a lower level is a nodule with the assemblage almandine, plagioclase, diopside, quartz, magnetite and apatite, which is characteristic of the higher grade clinopyroxene-almandine granulite subfacies. Other granulite nodules show unstable assemblages e.g. plagioclase, hypersthene, almandine, hornblende, diopside and magnetite. In one of these rocks the hypersthene, although clearly identifiable, is largely altered to carbonate and takes on an appearance very similar to that in the hornblende-olivine pyroxenite described previously. The hornblende is green-brown in thin section and occurs in only small quantities. The assemblage might be the result of polymetamorphism either progressive or retrogressive in the subfacies hornblende-orthopyroxene-plagioclase granulite and hornblende-clinopyroxene-almandine granulite or in the latter subfacies combined with the orthopyroxene-plagioclase granulite subfacies. Alternatively the development of hornblende and diopside in this rock may be due to metasomatism resulting from the immersion of this nodule in the carbonatite fluid. The process would be essentially one of removal of alkalies and silica from the nodule together with enrichment in calcium, iron and magnesium. There is evidence from other nodules that this has possibly happened at relatively low levels within the dykes.

Partial analyses of a brown hornblende and a clinopyroxene from this rock are given in Table 6 (analyses 4 and 10). The clinopyroxene is a salite with the structural formula $(Na,Ca)_{0.93}$ $(Mg,Fe,Al,Ti)_{1.08}$ $Si_{1.87}Al_{0.13}O_6$. On the margins of some of the brown hornblendes, minute spots of blue amphibole can be seen in thin section. Analysis shows that these spots have a sodic composition (Table 6, no. 5, pargasitic-richteritic hornblende with structural formula $(Na, Ca)_{2.89}$ $(Mg,Fe,Al,Ti)_{5.28}Si_{6.18}$ $Al_{1.82}$ O_{22} $(OH)_2$). The nodule must have suffered slight alkali enrichment at a late stage, and possibly at a fairly high level in the dyke. Where the garnet of this nodule is in contact with the carbonatite a dark green kelyphitic reaction rim rich in magnetite is developed. The garnet is almandine (Table 7, nos. 5,6) and was clearly unstable in this environment. Individual garnet crystals have not been found in these dykes.

Table 7. *Partial analyses by electron microprobe of a carbonate, micas, an olivine, and garnets from the lamprophyric carbonatites*

	1	2	3	4	5	6
SiO_2	–	39.47	nd	40.61	37.75	38.34
TiO_2	–	nd	nd	–	0.08	0.08
Al_2O_3	–	17.25	12.54	–	23.07	21.68
FeO*	4.35	5.09	5.88	13.82	24.32	24.51
MgO	16.75	19.00	21.32	46.82	9.51	8.95
CaO	32.38	nd	nd	–	5.86	5.91
Na_2O	–	nd	nd	–	0.32	0.32
K_2O	–	9.19	9.08	–	–	–
CO_2	46.50	–	–	–	–	–
	99.98	90.00	48.82	101.25	100.91	99.79

1) ankerite in lamprophyric carbonatite 60124.
2) phlogopite in lamprophyric carbonatite 60239.
3) phlogopite in lamprophyric carbonatite 60124.
4) olivine in lamprophyric carbonatite 60236.
5) almandine in garnet granulite nodule 60237.
6) almandine in garnet granulite nodule 60237.
*) Total iron determined as FeO.

The most abundant ultramafic nodules in these dykes are pyroxene hornblendites and hornblende pyroxenites consisting mainly of varying proportions of salitic pyroxene and hornblende. The texture of these nodules is variable but is predominantly xenomorphic granular. The hornblende shows particularly interesting variations in composition and texture which can perhaps be best understood when the behaviour of the hornblende xenocrysts is considered. Many of the individual hornblende crystals in thin section show a green core and a brown margin, with of course pleochroism in these shades (Plate 2b). The margins are sharply defined and the shapes of the cores tend to follow the crystal outlines which are more or less sub-rounded. In some cases aggregates of hornblende crystals show zoning which is confined to the margin of the aggregate, and cuts across from one crystal to another. This suggests that the zoning is due to metasomatic reaction with the carbonatite fluid. Partial analyses of these zoned hornblendes show that the margins, compared with the cores, are much poorer in iron and richer in magnesium and calcium. Dawson (1964) has shown that carbonatite magmas probably exist as multi-cation magmas containing both alkaline ions and basic ions, and that relative to the calcium and magnesium ions, the alkali-metal ions and (to a less extent) the iron ions appear to be much more ready to react with the wall rocks. The hornblende crystals have suffered marginal depletion in iron as a result of contact with the carbonatite magma (Table 6, nos. 1, 2). The margins have been correspondingly enriched in magnesium, calcium and titanium, but the content of other components is little changed. As already noted the conversion of a feldspathic granulite to a hornblendite or pyroxenite would involve alkali and silica depletion and it is suggested that this process may have operated at relatively low levels in the dykes, but at higher levels the process may have been reversed. Where the country rocks were amphibolites or pyribolites less alkali and silica depletion would be necessary to convert them to hornblendites or pyroxenites (Parsons, 1961; Dawson, 1964).

As far as the nodules in the lamprophyric carbonatites are concerned, there is evidence that suggests that hornblendes formed at an early stage of the metasomatism were relatively iron-rich minerals showing dominantly green pleochroism in thin section. As metasomatism continued the green hornblendes became marginally zoned to iron-poor brown amphibole, and as the process reached completion the mineral became a uniform brown colour. The analysis of a nodule hornblende given in Table 6, no. 3 is for a dominantly brown, edenitic-pargasitic hornblende which has been calculated to have a structural formula of $(Na,Ca)_{2.48}\,(Mg,Fe,Al,Ti)_{4.90}\,Si_{6.76}\,Al_{1.24}\,O_{22}\,(OH)_2$. The amphiboles in this rock show slight indications of greenish cores, and as would be expected the mineral has an intermediate content of iron (compare with Table 6, nos. 1 and 2). The various stages in the process are well represented in the nodules and together with the change in composition there is a corresponding change in the texture of the amphibole. In the early stages the hornblende forms discrete crystals with equidimensional form and a typical average grain size of somewhat less than 1 mm. They characteristically show 120° points of contact with pyroxene crystals and, although this texture might be produced in a cumulate as a result of crystals sintering together under the influence of interfacial tension (VOLL, 1960), in these rocks it is more likely to be due to simultaneous metasomatic crystallisation of hornblende and salite under equilibrium conditions. With continued metasomatism the hornblendes have grown to large irregular crystals commonly enclosing smaller pyroxenes (Plates 3a and 3b). If the hornblendes have grown at the expense of pyroxenes hydration of the nodule must have occurred.

The pyroxenes in the nodules and the xenocrysts show generally less pronounced changes than do the hornblendes. However quite marked zoning is apparent in many of the pyroxene xenocrysts such that neutral to pale pink margins are developed around light green cores as seen in thin section. Table 6, nos. 7 and 8 show analyses of a zoned pyroxene which indicate that the margin is again relatively depleted in iron and enriched in magnesium and calcium, but to a lesser extent than in the hornblende. However the increase in titanium from the core to the margin is relatively greater in the pyroxene and this may be the main cause of the colour zoning. In the nodules pyroxenes rarely show zoning and the crystals are essentially pale green and non-pleochroic, but occasionally aggregates of pyroxenes show a neutral or pale pink margin against the carbonatite matrix with the margin traversing from one crystal to another. The texture of the pyroxenes only varies in so far as with increasing metasomatism the crystals become enclosed within the hornblendes. The partial analyses of pyroxene xenocrysts, and, of pyroxenes from a garnet granulite and from the common type of hornblende pyroxenite nodule, show these minerals to be salites. They resemble pyroxenes in similar rocks from New South Wales as described by WILSHIRE & BINNS (1961).

The iron depletion of the nodules and xenocrysts appears to have followed an earlier stage of iron enrichment in which the pyroxenites and hornblendites were formed, and the olivines became enriched in iron. This change in direction of the iron metasomatism could be due to the development of a fluidised system which would allow migration of the more mobile ions (in this case iron) into the high fugacity medium.

One of the features which indicates that these hornblendite and pyroxenite nodules may well be metasomatised crustal rocks is the presence of distinct traces of a presumably relic foliation, which tends to be more common in the less modified rocks. Minor constituents in these nodules include magnetite, apatite and calcite (Table 5, nos. 7–12).

Other ultramafic nodules show signs of alkali enrichment. The marginal sodic spots of hornblendes in a granulite nodule have already been noted, and more distinct small, blue amphibole crystals occur in some of the hornblendites. In one pyroxene hornblendite nodule minute, euhedral, pleochroic lilac-coloured amphibole crystals have grown against brown hornblende in an interstitial area together with carbonate. Some perovskite occurs in this rock, and this mineral is particularly abundant in another nodule (Table 5, no. 17) where it is very closely associated with minute crystals of sodic amphibole. This nodule shows other features of alkali enrichment in that a small amount of interstitial albite is present, the diopsidic pyroxenes have sodic margins (Table 6, nos. 11, 12), and small euhedral crystals of aegirine-augite are developed (Plate 4a). Analysis of the zoned pyroxene shows that the margin is strongly enriched in iron as well as sodium, with a corresponding reduction in the content of magnesium and calcium. The pyroxene core is a salite with structural formula $(Na_{0.05}, Ca_{0.94})(Mg,Fe,Al,Ti)_{0.97}Si_{1.98}Al_{0.04}O_6$, and the rim is aegirine-augite with formula $(Na_{0.58},Ca_{0.43})(Mg,Fe,Al)_{0.96}Si_{1.98}Al_{0.02}O_6$. Iron enrichment occurred together with alkali metasomatism of the nodule, possibly at a relatively high level in the dyke, and is the reverse of the alkali and iron depletion which may have occurred at a lower level. SAETHER (1948) has demonstrated that diffusion processes in a magma with great vertical extension lead to the concentration of volatiles at the top of the chamber, with the volatiles, through their acid character, attracting basic ions (Na^+, K^+, Ca^{++}, and Ba^{++}). The mineralogy of this nodule also shows that sodium metasomatism has occurred, in this case, without potassium metasomatism.

Another nodule consists predominantly of aegirine-augite with minor amounts of apatite and magnetite. Some of the pyroxenes are zoned with greener margins, but the most interesting thing about this rock is its texture. This consists of an irregular arrangement of small blebs of pyroxene which could well have resulted from gas streaming (Plate 2a).

Potassium enrichment of nodules is presumably indicated by the development of biotite or phlogopite. One hornblende pyroxenite contains a biotite crystal, nearly 1 cm across, poikiloblastically including crystals of salite partly rimmed by brown hornblende. This nodule also contains a small amount of albite associated interstitially with carbonate. A hornblendite nodule containing green hornblende shows a fairly good biotite orientation which is probably following an earlier metamorphic texture. Biotite orientation of a different kind is seen in a diopside glimmerite nodule where zones of granulation are developed parallel to the foliation trace and many of the mica crystals are bent. Another glimmerite nodule consists of large, disoriented crystals of biotite poikiloblastically enclosing disoriented, small, euhedral crystals of aegirine-augite. A little albite and carbonate occurs interstitially. In this case the nodule appears to have been enriched in both potassium and sodium.

One of the most interesting and significant nodules is a biotite salite hornblendite with a pure hornblendite rim (Plate 4b). The core of the nodule shows a distinct foliation trace defined by biotite and hornblende crystals, which is presumably inherited from the original metamorphic rock. This texture appears to have been disturbed in some parts by gas streaming. The core also contains salite, apatite and interstitial albite. The outer rim consists of virtually pure, green hornblende, and between this and the core is an inner rim consisting mainly of albite and biotite. It would seem that this xenolith originally suffered alkali and silica removal possibly at a low level in the dyke to convert it to a salite hornblendite before being carried to a high level where it became re-enriched in potassium and sodium, producing biotite and albite. After this the nodule may have fallen to a level where alkali depletion produced the hornblende rim. The biotite-albite rich inner rim may have developed while the inclusion was suffering alkali enrichment, or it may have formed by metasomatic differentiation with the growth of the hornblende rim. Finally the nodule was brought up to the present level. This kind of behaviour may be expected in the turbulent conditions which would have existed in the magma if the dyke was emplaced by a fluidisation process (Reynolds, 1954). As the dykes do not exceed 1 m in thickness they were probably too narrow for turbulence to be very effective. At the same time the average diameter of the nodules is only about 2 cm.

A few trachytic inclusions have been found in the lamprophyric carbonatite dykes. These consist of small, disoriented crystals of albite set in a matrix which contains some minute augite crystals but consists mainly of magnetite (Plate 3b). Therefore it seems likely that these inclusions have suffered some iron enrichment presumably at high levels in the dykes. No trachyte dykes or plugs have been found but trachyte flows may have formed part of the cover at the time of intrusion of the lamprophyric carbonatites.

Phlogopite forms an important constituent of many of the dykes occurring as elongated, subhedral crystals, the orientation of which indicates the direction of flow of the carbonatite magma. On the other hand phlogopite or biotite is not very common in the nodules. Whereas individual olivine, salite and hornblende crystals seem to represent mainly modified xenocrystal material, phlogopite may have developed mainly as phenocrysts i.e. by crystallisation from the lamprophyric carbonatite magma. It seems likely that phlogopite or biotite would crystallise from an ultrabasic magma of this type which was reacting strongly with sialic country rock inclusions. Some of the clinopyroxene and hornblende probably crystallised from the magma, partly as small crystals in the groundmass, and partly in continuity with the metasomatised rims of the xenocrysts. Magnetite granules are very abundant in the dykes and these might represent some of the iron removed from included material.

CONCLUSIONS

At the time of intrusion of the Mesozoic dolerite dykes the West Greenland region must have been a stable continental area subject only to broad crustal warping and deep fracturing probably associated with the onset of continental drift (*cf.* Bailey, 1964). Basaltic magma rose quietly through the crust without becoming significantly contaminated or, at this erosion level, suffering much differentiation. Volcanic fluids, particularly in this region carbon dioxide and water, may have become concentrated near the base of the crust by a zone refining process, and were then able to rise and bring with them fragments of upper mantle and lower crust. Possible upper mantle fragments are represented mainly by carbonated olivine xenocrysts. It is thought that lower crustal xenoliths, mainly of acidic to intermediate composition and of granulite type, may have suffered alkali and silica depletion, and enrichment in calcium, iron and magnesium in lower levels of the dykes so that they became converted to hornblendites and pyroxenites. During fluidisation iron depletion and hydration of the mafic minerals occurred, and modified hornblende grew at the expense of pyroxene. Some nodules became enriched in alkalies and iron with the result that aegirine-augite, alkali amphibole, biotite, albite and perovskite developed, possibly at higher levels in the dykes. One nodule shows a central zone of alkali enrichment and a marginal zone of alkali depletion. This may have been carried up and down within the dyke so that it suffered different processes at different levels. Changes in direction of the alkali and iron metasomatism may have occurred as a result of change in the chemical potential of these elements as the pressure, temperature and composition of the carbonatite magma changed during its ascent. The composition of the magma relative to the inclusions would change with the gradual incorporation of sialic material. Salite and hornblende xenocrysts in the dykes show the same behaviour as these minerals in the nodules. The contaminated carbonatite magma precipitated mainly calcite, dolomite, phlogopite and magnetite.

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PLATES

Plate 1

Fig. 1 a.

Lamprophyric carbonatite from Qeqertarssuatsiait. The specimen was estimated to contain about 400 ultramafic nodules.

Fig. 1 b.

Zoned hornblende and salite xenocrysts in a lamprophyric carbonatite. Relatively magnetite-free orbs of carbonate are developed around the xenocrysts. Plane polarised light.

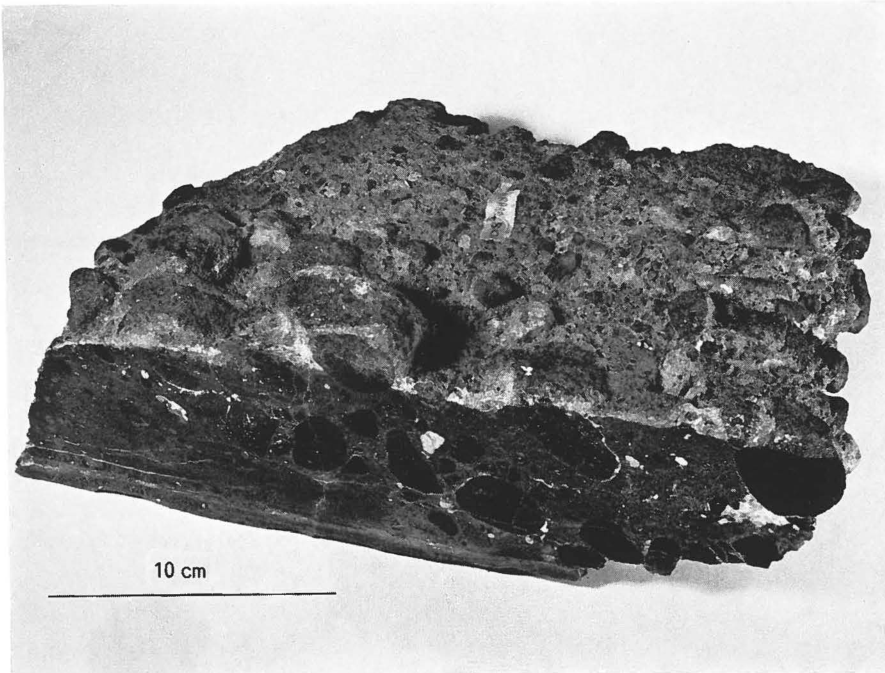

Fig. 1 a.

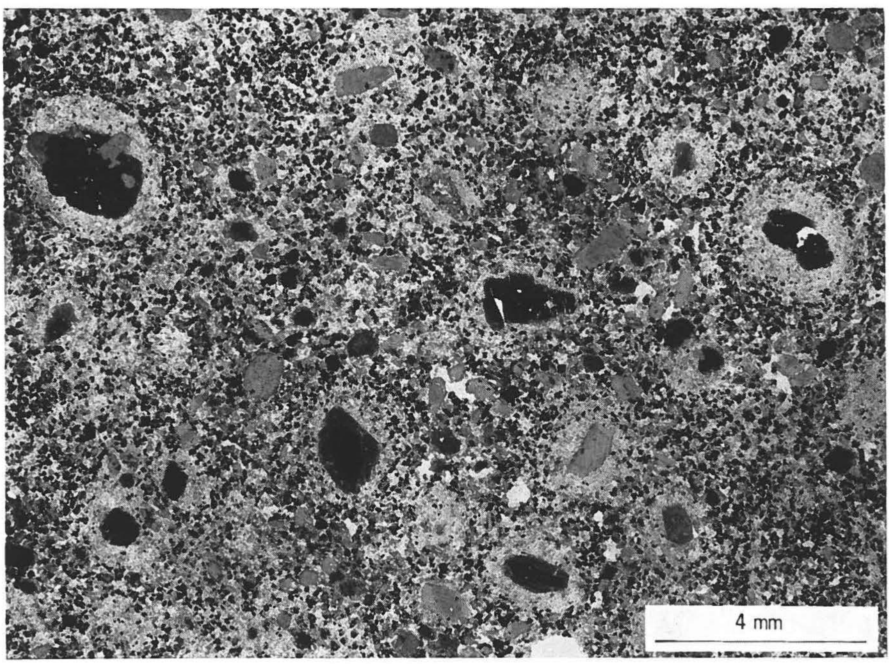

Fig. 1 b.

Plate 2

Fig. 2a.

Aegirine augite nodule with a possible gas-streamed texture (a), a carbonated olivine xenocryst and a nodule consisting of carbonated orthopyroxene, olivine, diopside, hornblende and magnetite (b). Plane polarised light.

Fig. 2b.

Hornblende xenocryst with Fe-poor rim. The groundmass shows fluxion structure. Plane polarised light.

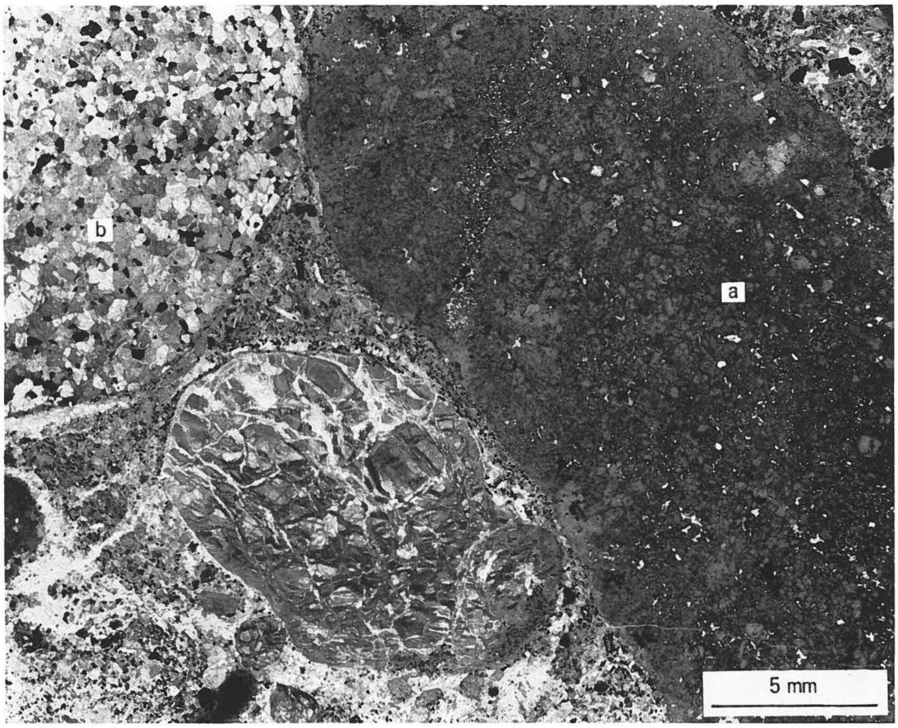

Fig. 2 a.

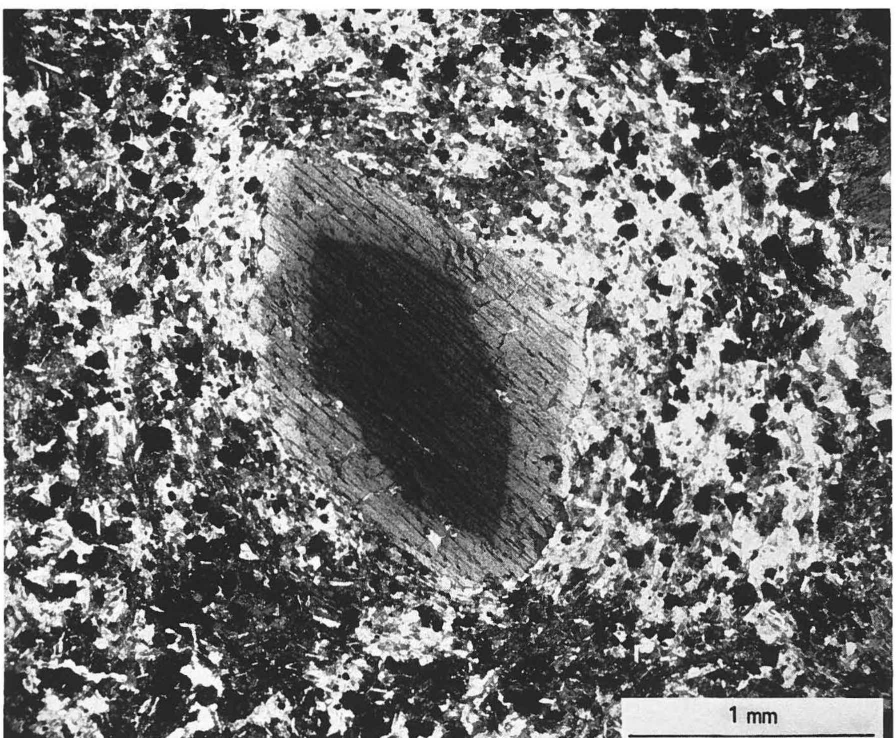

Fig. 2 b.

Plate 3

Fig. 3 a.

Salite hornblendite nodule (a). Megacrysts of brown hornblende (b) and fresh olivine (c). Plane polarised light.

Fig. 3 b.

Hornblende clinopyroxenite nodule with relatively late brown hornblende (h). Hypersthene granulite (g) and trachyte inclusion (t). Plane polarised light.

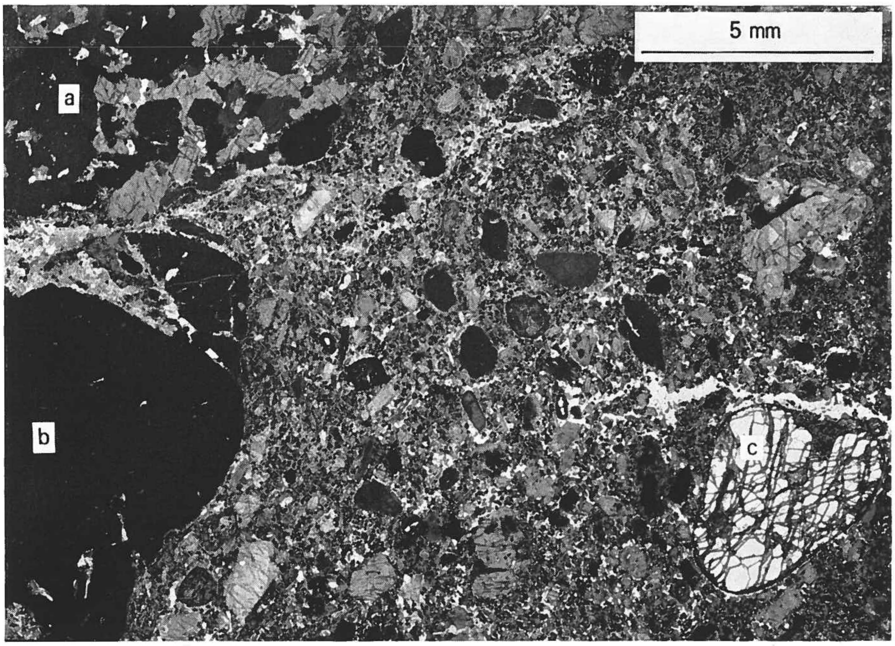

Fig. 3 a.

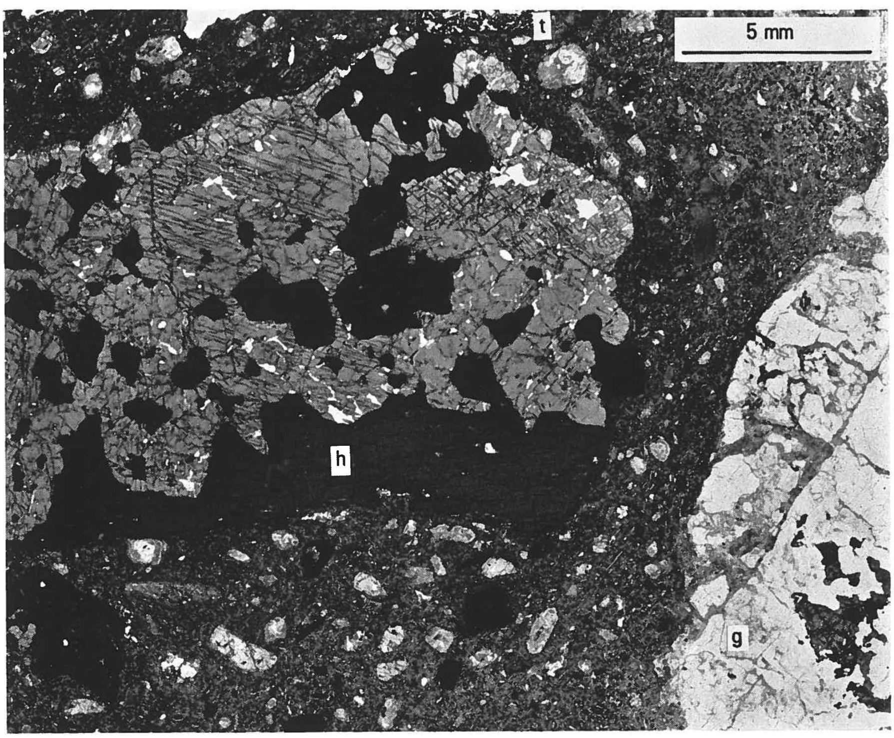

Fig. 3 b.

Plate 4

Fig. 4a.

Mela-syenite nodule showing salite with Na, Fe rich rims, small crystals of aegirine-augite developed within albite (a), and perovskite (p). Plane polarised light.

Fig. 4b.

Biotite salite hornblendite nodule with a pure hornblendite rim (h).
Plane polarised light.

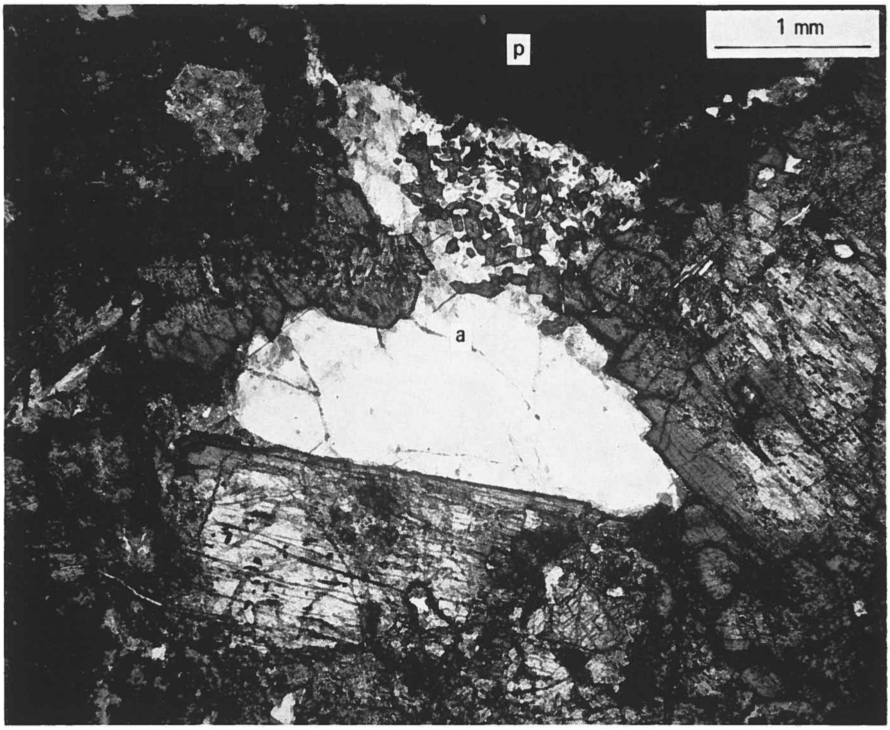

Fig. 4a.

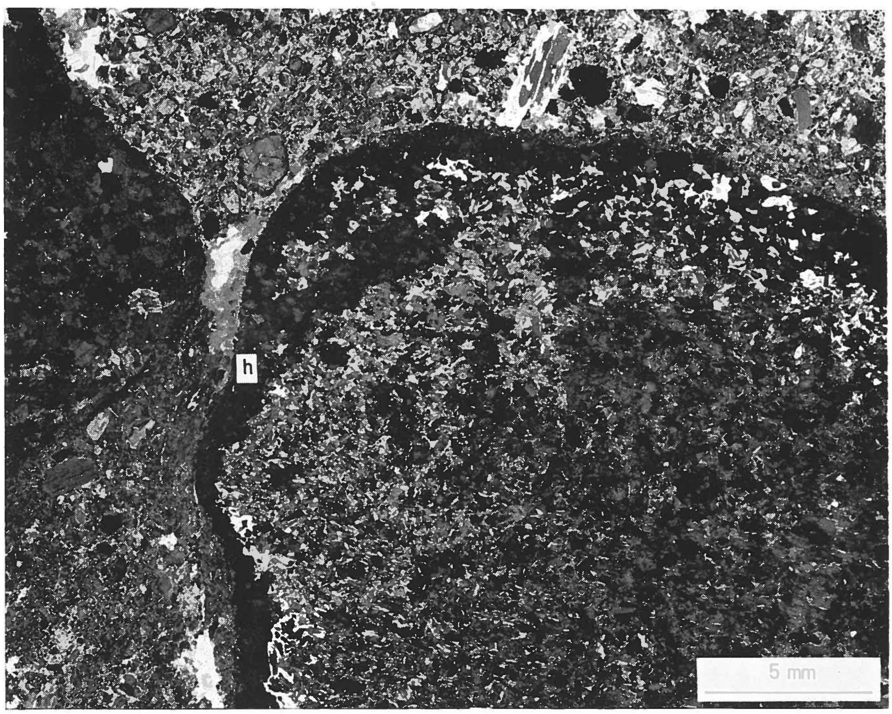

Fig. 4b.